III. SOIL MOISTURE CHARACTERISTIC CURVE

In order to represent the h-θ relation in one graph the moisture characteristic curve is often a semi-log presentation, being the log (- h cm WH) versus the moisture content. The value of log (- h cm WH) is called the pF-value and the graphical presentation the pF-curve (figure 17 and figure 18).
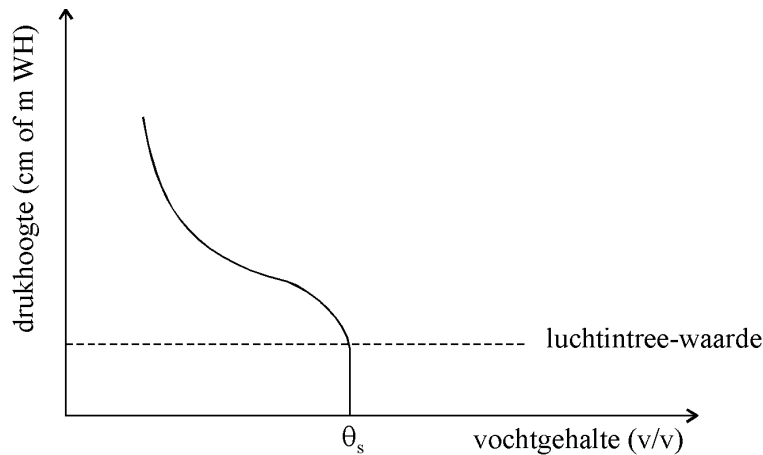

Figure 17. The soil moisture characteristic curve

When to a saturated soil sample an increasing over-pressure of under-pressure (decreasing matric potential or soil water pressure head) is applied, no outflow may occur until, a certain critical value is exceeded at which the largest pores of entry begins to empty. This critical matric potential or soil pressure head is called the *air-entry value*. Its value is generally small (in absolute value) in coarse-textured and well aggregated soils. However, since in coarse-textured soils the pores are often more of nearly uniform in size, these soils may exhibit critical air-entry phenomena more distinctly and sharply than do fine-textured soils.

As the pressure is further increased, more water is drawn out of the soil and more of the relatively large pores, which cannot retain water against the pressure applied, will empty out. A gradual increase in pressure will result in the emptying of progressively smaller pores, until at low matric potential or soil water pressure head values, only the very narrow pores retain water. Similarly, a decrease in soil water pressure head is associated with a decreasing thickness of the hydration envelopes covering the soil particle surfaces. The latter is due to adsorption forces.

When the soil is composed of a large range of particle sizes than a steady decrease in moisture content occurs with decreasing matric potential or soil water pressure head. The h-θ curve of a coarse sandy soil, which is composed of soil particles of more or less

equal size, however shows a plateau because it consists of a large amount of large pores of equivalent diameter which are emptied at a specific soil water pressure head.

Increasing the applied pressure is thus associated with decreasing soil wetness. The amount of water remaining in the soil at equilibrium is a function of the sizes and volumes of the water-filled pores and hence it is function of the soil water pressure head. This function is usually measured experimentally and it is represented graphically by a curve known as the soil-moisture retention curve, or the soil moisture characteristic.

Recalling the capillary equation the equivalent diameter (d) of the pores which are emptied at a specific soil water pressure head e.g. - 0,3 m equals:

$$d = 2\,r = \frac{2 \,.\, 2 \,.\, \gamma}{\rho_w \; h \; g} \;=\; \frac{4 \,.\, 0{,}075}{10^3 \,.\, 0{,}3 \,.\, 10} \;\approx\; 1 \,.\, 10^{-4} \; m \;=\; 100 \; \mu m$$

This relation is valid for non-swelling soils and moreover gives only a rough idea since the pores are in reality not always spherical cavities interconnected by straight capillaries.
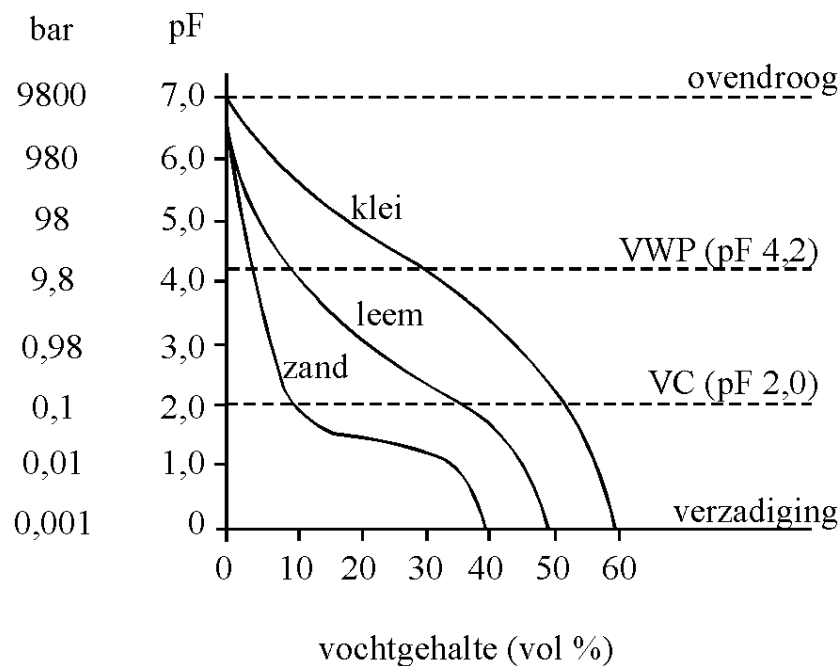

Figure 18. Soil moisture characteristic curve of a sand, loam and clay soil.

In figure 18 are given the pF-curves of different soils. It is obvious that in sandy soils the capillary forces are more important than in clayey soils where adsorption forces are dominant.

Apart from the particle size distribution (texture), also the soil structure (pore geometry) affects the shape of the soil moisture characteristic curve, particularly in the wet range (high soil water pressure head range). The amount of water retained at relatively high matric potential (say between 0 and − 1 bar) depends primarily upon the capillary effect and the pore size distribution, and hence is strongly affected by the structure of the soil. On the other hand, water retention in the low matric potential range is due increasingly to adsorption and is thus influenced less by the structure and more by the texture and specific surface of the soil material. The water content at a matric potential of − 15 bar (often taken to be the lower limit of soil moisture availability to plants) is fairly well correlated with the surface area of a soil and would represent, roughly about 10 molecular layers of water if it ware distributed uniformly over the particle surfaces.

It should be obvious from the foregoing that the soil moisture characteristic curve is strongly affected by soil texture. The greater the clay content, in general, the greater the water retention at any particular matric potential, and the more gradual the slope of the curve. In a sandy soil, moist of the pores are relatively large, and once the large pores are emptied at a given matric potential, only a small amount of water remains. In a clayey soil, the pore size distribution is more uniform, and more of the water is adsorbed, so that decreasing the matric potential causes a more gradual decrease in water content (Figure 18).

Soil structure also affects the shape of the soil moisture characteristic curve, particularly in the high matric potential range. The effect of compaction upon a soil is to decrease the total porosity, and, especially, to decrease the volume of the large inter-aggregate pores.

This means that the saturation water content and the initial decrease of water content with the application of low pressure are reduced. On the other hand, the volume of the intermediate-size pores is likely to be somewhat greater in a compact soil (as some of the originally large pores have been squeezed into intermediate size by compaction), while the intra-aggregate micropores remain unaffected and thus the curves for the compacted and uncompacted soil may be nearly identical in the low soil water pressure head range (Figure 19). In the very low soil water pressure head range, the predominant mechanism of water retention is adsorptive rather than capillary, and hence the retention capacity becomes more of a textural than a soil-structural attribute.
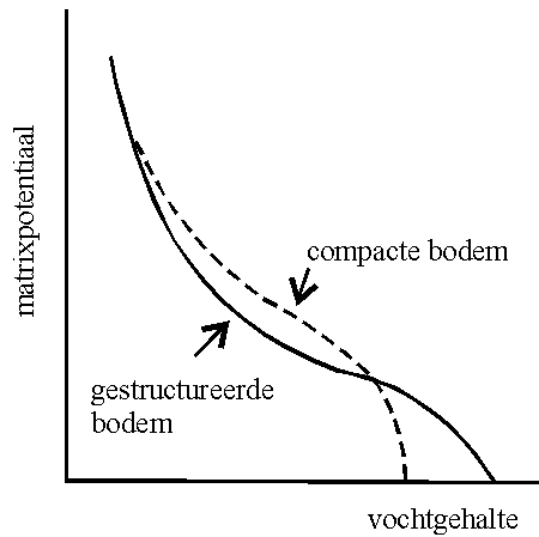
Figure 19. The effect of soil structure on soil water retention.

Therefore it is important to determine the soil moisture characteristic curve on undisturbed samples. The inverse of the slope of the curve, which is the change of water content per unit change of matric potential, is generally termed the differential (or specific) water capacity "C (θ)":

$$C (\theta) = \frac{d\theta}{dh}$$

This is an important property in relation to soil moisture storage and availability to plants as water transport in general. The actual value of C (θ) depends upon the wetness range, the texture and the hysteresis effect.

The estimation of the relation between the matric potential versus the moisture content based on soil properties is not always successful (see chapter IV). The adsorption and pore geometry effects are often too complex to be described by a simple model.

3.1.   Determination of the soil moisture characteristic curve.

The determination of the soil moisture characteristic curves as presented in Figure 18, is carried out in the laboratory through a systematic desorption of water starting from a saturated soil sample. As mentioned earlier the determination should be carried out on undisturbed soil samples, certainly at high soil water pressure heads (0 till -1/3 atm.). At lower soil water pressure heads the influence of soil structure on the further shape of the curve (ψ_m(h) - θ relation) is negligible.

The methods consists in the determination of the water content of the soil sample at the moment that hydraulic equilibrium is reached for a certain applied matric potential. It takes mostly some days before equilibrium is reached depending on the particle size distribution, size of the sample and the applied pressure.

3.1.1. Hanging water column (figure 20a and 20b).

A hanging water column consists of a water-saturated, highly permeable porous ceramic plate connected on its underside to a water column terminating in a reservoir open to the atmosphere. Water-saturated soil samples held in rings are placed in contact with the flat porous plate when the height of the water reservoir is even with the top of the plate. Then the reservoir is lowered to a new height a distance z below the top of the porous plate (figure 20). By the equilibrium principle, water will flow from the soil sample through the ceramic plate to the outflow until the total water potential of the system is constant.

At hydraulic equilibrium follows:

$$\psi_{t_A} \; (soil\,water) \; = \; \psi_{t_B} \; (outflow\,water) \; = \; \text{constant}$$

or from what has been discussed in chapter I:

$$\psi_{m_A} + \psi_{h_A} + \psi_{e.p_A} + \psi_{g_A} + \psi_{0_A} = \psi_{m_B} + \psi_{h_B} + \psi_{e.p_B} + \psi_{g_B} + \psi_{0_A} \qquad (9)$$

- The system is under atmospheric pressure and thus:

$$\psi_{e.p_A} \; = \; \psi_{e.p_B} \; = \; 0$$

- The salt concentration, if it occurs, is by diffusion everywhere the same:

$$\psi_{0_A} \; = \; \psi_{0_B}$$

- The hydraulic pressure potential does not exist:

$$\psi_{h_A} \; = \; \psi_{h_B} \; = \; 0$$

- The reference level is taken at the soil sample height and so:

$$\psi_{g_A} \; = \; 0$$

- At the outflow : $\psi_{m_B} = 0$, so that eq. (9) becomes:

$$\psi_{m_A} = \psi_{g_B} \quad \text{of} \quad \frac{1}{g} \, \psi_{m_A} = \frac{1}{g} \, \psi_{g_B} = -h$$

At equilibrium the moisture content (θ) is determined and one obtains consequently one point of the soil moisture characteristic curve (h_1 - θ_1).

As the tube is adjustable, different under-pressures or matric potentials can be applied, whereby the air entry value of the porous ceramic plate may not be exceeded. The lowest matric potential that can be applied by this method is similar to that for a tensiometer being $\approx$ - 8 m WH.

A set up whereby more than one sample can be subjected to a hanging water column is given in figure 20b. The so-called "sandbox". Here a layer of calibrated sand replaces the porous ceramic plate. A matric potential from 0 till - 100 cm WH can be applied. Sometimes the sand layer is replaced by a mixture of sand and kaolinite which resists to a hanging water column of 500 cm WH (matric potential $\approx$- 0,5 bar), which can be applied through a vacuum pump.
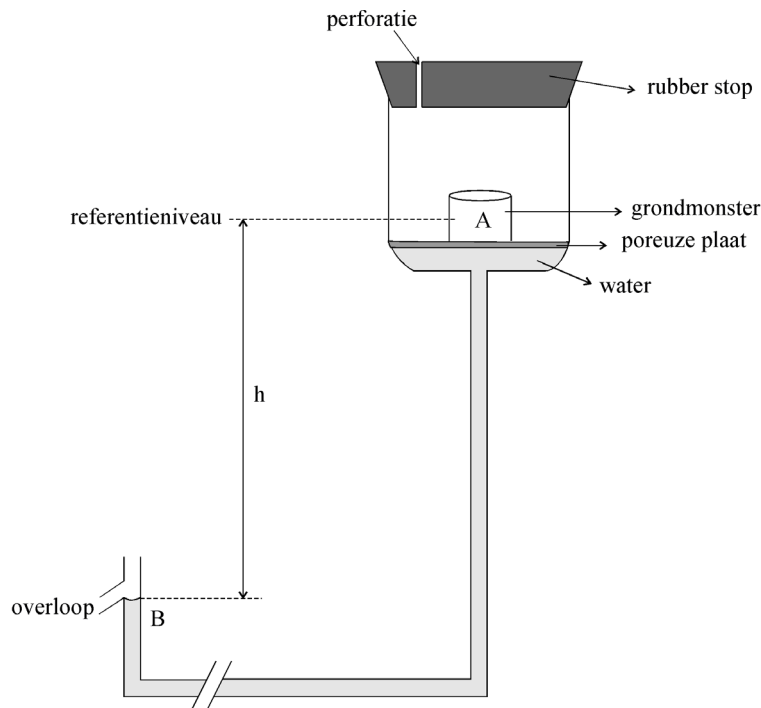
(a)
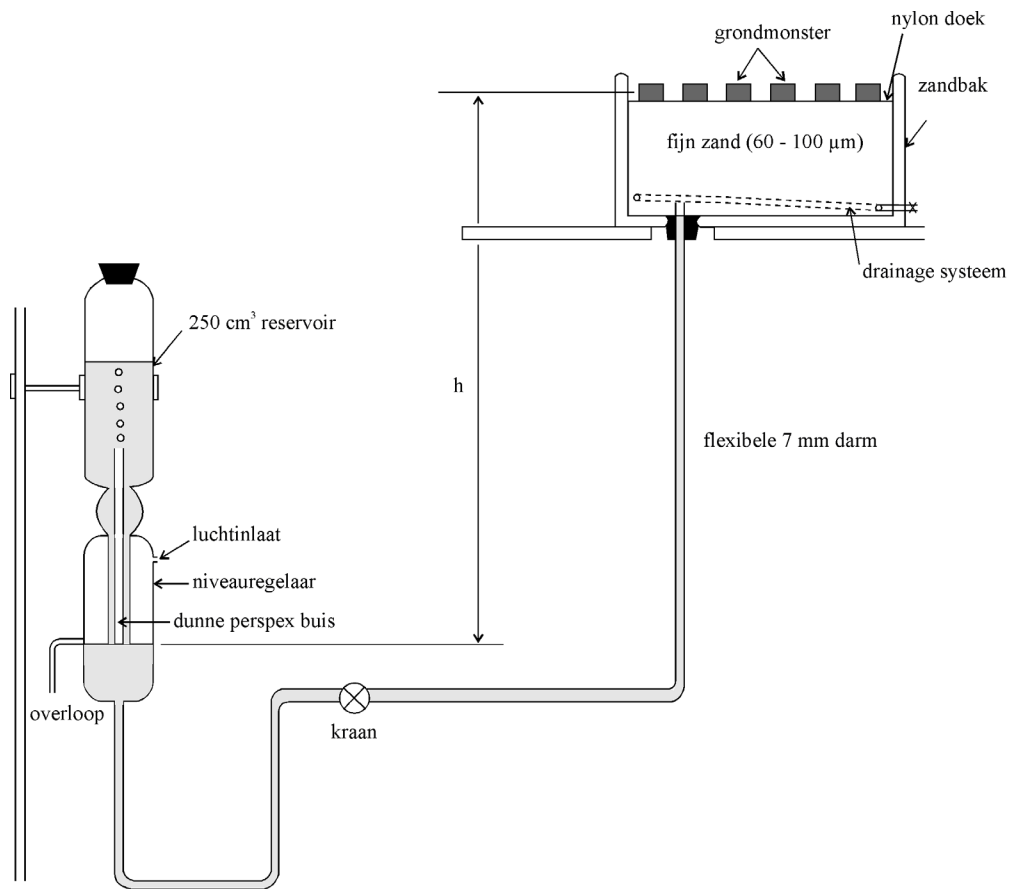
(b)

Figure 20.   Desaturation of saturated soil samples to a desired energy state with a hanging water column.

3.1.2. Pressure method (Pressure plate method).

For the determination of the h-θ relation from saturation (h = 0) till a matric potential of ± - 20 bar the pressure plate method is commonly used (figure 21).

The pressure plate consists of an airtight chamber enclosing a water-saturated, porous ceramic plate connected on its underside to a tube that extends through the chamber to the open air. Saturated soil samples are enclosed in rings and placed in contact with the ceramic plate on the topside. The chamber is then pressurized, which squeezes water out of the soil pores, through the ceramic, and out the tube (figure 21). At equilibrium, flow through the tube will cease.
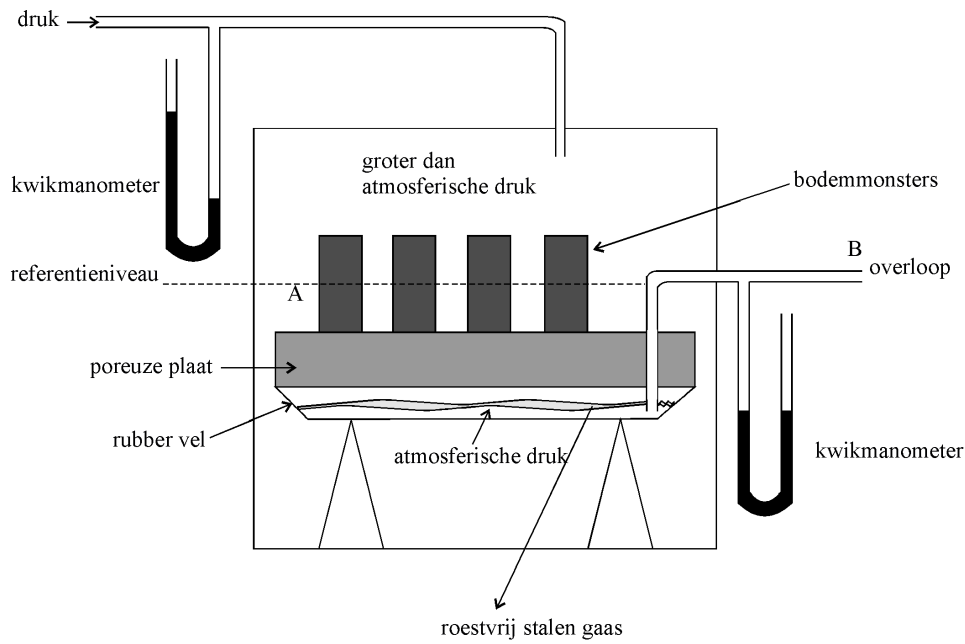
Figure 21.   Desaturation of saturated soil samples to a desired energy state with a pressure plate.

At hydraulic equilibrium:

$$\psi_{t_A} = \psi_{t_B} = \text{constant}$$

- Salt, if present, diffusion occurs through the porous plate and:

$$\psi_{0_A} = \psi_{0_B}$$

- No hydrostatic pressure is present or: $\psi_{h_A} = \psi_{h_B} = 0$

- In B we have free water ($\psi_{m_B} = 0$) under atmospheric pressure ($\psi_{e.p_B} = 0$).

- The differences in gravitational potential between A en B may be neglected so that:

$$\psi_{g_A} = \psi_{g_B}$$

(eventually = 0 if the reference level is taken at the level of the sample).

So the equation (9) becomes:

$$\psi_{m_A} + \psi_{e.p._A} = 0 \quad or \quad \psi_{m_A} = - \psi_{e.p._A}$$

40

When equilibrium is reached, the chamber may be depressurized and the water content of the samples measured and one point of the h-θ relation is obtained. An assumption is made in this method whereby the matric potential of the sample does not change as the air pressure is lowered to atmospheric.

This method may be used up to air gauge pressures of about 15 bars (pF = 4,2, being permanent wilting point, see 3.4) if special fine-pore ceramic plates are used. Since these devices have a very high flow resistance, it may require a substantial amount of time to remove the last small amount of water from the soil. Thus, the time of equilibrium is difficult to estimate.

3.1.3. Vapor pressure equilibrium.

For matric potentials lower than – 15-bar vapor pressure equilibration of salt solutions are used. By adding pre-calibrated amounts of certain salts, the energy level of a reservoir of pure water may be lowered to any specified level. If this reservoir is brought into contact with a moist soil sample, water will flow through the vapor phase from the sample to the reservoir. If the sample and the reservoir are placed adjacent to each other in a closed chamber at constant temperature, water will be exchanged through the vapor phase by evaporation from the soil sample and condensation in the reservoir until equilibrium is reached. Since the reservoir is a pool of salt solution, at equilibrium the total potential will be $\psi_t = \psi_{so}$ of the solution. In the soil $\psi_t = \psi_m + \psi_o$ since the air-water interface acts as a solute membrane. Thus $\psi_m = \psi_{so} - \psi_o$, the difference between the solute potentials of the reservoir and the soil. In practice, the soil will usually not be saline enough for its solute potential to be significant compared to ψ_o in the range where these measurements are made. Knowing that the relative humidity is related to the pool of salt solution one obtains that:

$$\psi_m + \psi_0 = \frac{R\,T}{M}\,\ln\frac{P}{P_0} = \psi_w \ (water\ potential) \tag{10}$$

at t = 20°C, and with:
   R = 8,3 J mol^{-1} K^{-1}
   T = 273 + 20 = 293 K;
   M = 18 10^{-3} kg mol^{-1}
   P = vapor pressure of soil water
   P$_o$ = saturated vapor pressure (vapor pressure of pure free water)

Equation 10 becomes:

$$\psi_m + \psi_0 \approx \frac{8,3 \cdot 293}{18 \cdot 10^{-3}}\,\ln\frac{P}{P_0} \approx 1,3 \cdot 10^5 \cdot 2,303\ \log\frac{P}{P_0}$$

$$\psi_m + \psi_0 \approx 3.10^5 \log \frac{P}{P_0} \quad (\text{J kg}^{-1})$$

$$\rho_w \, (\psi_m + \psi_0) \approx 3.10^8 \log \frac{P}{P_0} \quad (\text{Pa})$$

$$\frac{1}{g} \, (\psi_m + \psi_0) \approx 3.10^4 \log \frac{P}{P_0} \quad (\text{m})$$

or

$$3.10^6 \log \frac{P}{P_0} \; (cm) \;=\; 3.10^6 \log \frac{RH}{100} \quad (cm) \;=\; h$$

RH, being the relative humidity and equals 100 (P/P₀). The relative vapor pressure P/P₀ seems to be a measure of the sum of the matric and osmotic potential of the soil water. It is now possible to describe the shape of the relation between the relative vapor pressure or relative humidity in function of pF-values (figure 22). From the equation

$$\psi_m + \psi_o \;=\; \frac{RT}{M} \ln \frac{P}{P_o}$$

follows for a non-saline soil ($\psi_o = 0$)

$$\psi_m \;=\; \frac{RT}{M} \ln \frac{P}{P_o}$$

or

$$\frac{1}{g}\psi_m \;=\; h \;=\; 3.10^6 \log \frac{P}{P_o} \quad (cm)$$

So finally one obtains:

$$pF \;=\; \log(-h) \;=\; \log\left(3.10^6 \log \frac{P}{P_o}\right)$$

$$pF \;=\; 6{,}5 \;+\; \log \left(\log \frac{100}{RH}\right)$$

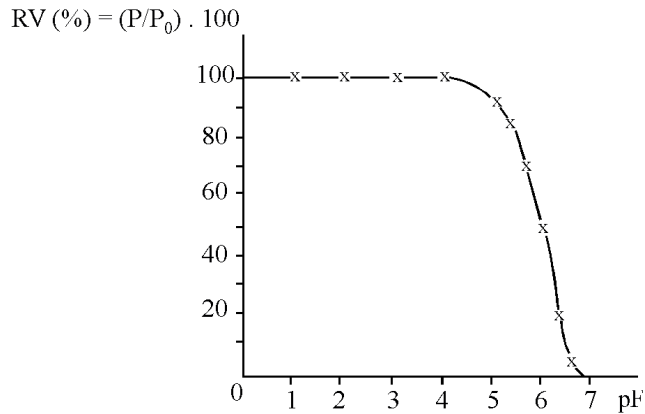$$pF = 6{,}5 \;+\; \log \left(2 - \log RH\right) \tag{11}$$

Figure 22. Relation between relative humidity and pF-values.

Using this relation, part of the moisture characteristic can be established. The method consists of putting in a closed environment, containing a salt saturated solution, a soil sample (figure 23). Water transport will take place through the vapor phase until equilibrium will be reached between the ψ_t of the soil water and the ψ_t of the water in the vapor phase. Equilibrium is reached when a constant weight of the soil sample is reached. At that moment the moisture content is determined.

From figure 22 it is clear that within the range of 10% to 75% R.H. the relation is almost linear and very steep which shows that a change in R.H. has only a relative small effect on the pF-value (range of pF = 5,5 till pF = 6,5). On the other hand in the low pF-range (important for plant growth), a small change in R.H. results in a relative large change in pF-values. Since changes in temperature strongly influence the shape of the curve in that range this technique is only applicable to establish the h-θ relation in the range > pF ± 5. To obtain a certain vapor pressure a specific salt saturated solution is used.
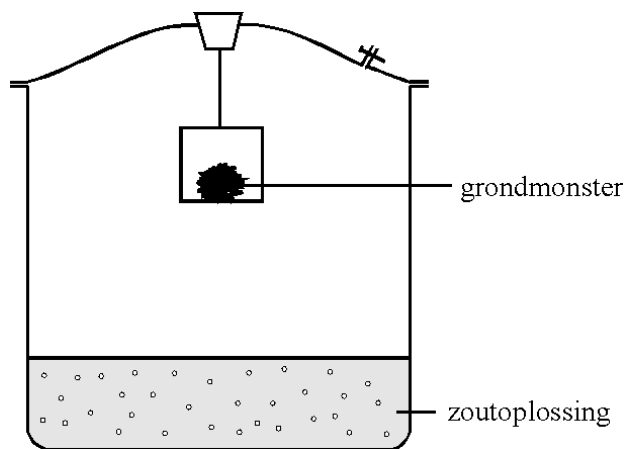

Figure 23.   Set up to obtain a certain pF-value by means of vapour pressure equilibrium.

43

If in the salt saturated solution there is still salt present in solid form then, when water uptake or water release occurs, the concentration by dissolving of salts or the precipitation of salts will automatically be adapted. Since the solubility of most salts depends on temperature, one has to use those which are less temperature dependent (table 2).

Table 2.   Relation between the relative humidity and the pF-values of some salt saturated solutions at 20°C.

Salt	Relative humidity (%) at equilibrium	pF
Ammonium oxalate	98,8	4,2
Potassium sulfate	97,1	4,6
Potassium chromate	88,0	5,2
Sodium chloride	75,8	5,6
Potassium carbonate	44,0	6,0

If salts are present in the soil sample a correction for the osmotic potential has to be introduced otherwise the obtained results cannot be combined with those obtained by the hanging water column and the pressure plate method.

3.2.   In situ determination.

Apart from laboratory determination also the possibility exists to estimate the soil water characteristic curve in the field. A special situation is when equilibrium occurs in situ e.g. with shallow ground water table. In such a case the matric potential or soil water pressure head at each location above the ground water table equals the negative value of the distance from the point under consideration to the ground water table. The moisture content in relation to that distance covers a part of the soil moisture characteristic curve.

Normally the matric potential is measured by means of a tensiometer in the soil water pressure range > - 800 cm WH (wet range) or with an electrical resistance unit in the range < - 1000 cm WH (dry range), while simultaneously the moisture content is measured.

For the range where the tensiometer is not useful (< - 1 bar) a method has been developed (psychrometer technique) allowing the in situ measurement of ψ_m or $\psi_m + \psi_0$ (water potential) based on vapor pressure measurements in the soil.

At equilibrium, the potential of soil moisture is equal to the potential of the water vapor in the ambient air. If thermal equilibrium is assured and the gravitational effect is neglected, the vapor potential can be taken to be equal to the sum of the matric and osmotic potentials, since air acts as an ideal semi-permeable membrane in allowing only water molecules to pass.

The vapor pressure P will normally decrease in the presence of the soil matric and the amount of salt in solution versus the vapor pressure P_0 of free and pure water at the same temperature. The system consists of a small porous cup that is installed in the soil (figure 24). Since diffusion of the vapor phase takes place through the porous cup the vapor pressure within the porous cup will be in equilibrium with that of the ambient air.

A psychrometer is an instrument designed to indicate the relative humidity of the atmosphere in which it is placed, generally by measuring the difference between the temperatures registered by a wet bulb and dry bulb thermometer. The dry bulb thermometer indicates the temperature of a non-evaporating surface in thermal equilibrium with the ambient air. The wet bulb thermometer indicates the generally lower temperature of an evaporating surface, where latent heat is adsorbed in proportion to the rate of evaporation. If the atmosphere has a low relative humidity (i.e. is relatively dry) its evaporative demand and hence the evaporation rate will be higher, resulting in a greater depression of wet bulb temperature relative to dry bulb temperature, and vice versa. The relative humidity of a body of air in equilibrium with a moist porous body will depend upon the temperature, as well as upon the state of water in the porous system – i.e., on the constraining effects of adsorption, capillarity, and solutes, all of which act to reduce the evaporability of water relative to that of pure, free water at the same temperature. Hence the relative humidity of a unsaturated soil's atmosphere will generally be under 100% and the deficit to saturation will depend upon the soil water potential, due to the combined effects of the matric and osmotic potentials. However, throughout most of the range of variation of soil moisture, this deficit to saturation is found to be very small, generally less than 2%. If we are to attempt to measure soil water potential in terms of its effect upon the equilibrium relative humidity of the ambient air, we must have a very accurate psychrometer.

The development of highly precise miniaturized thermocouple psychrometers, makes indeed possible the in situ measurement of soil water potential. A thermocouple is a double junction of two dissimilar metals. If the two junctions are subjected to different temperature, they will generate a voltage difference. If, on the other hand, an electromotive force (emf) is applied between the junctions, a difference in temperature will result. Depending on which way a direct current is applied, one junction of which is equilibrated with the soil atmosphere by placing it inside a hollow porous cup embedded in the soil, while the other junction is kept in an insulated medium to provide a temperature lag. During operation, an emf is applied so that the junction exposed to the

soil atmosphere is cooled to a temperature below the dew point of that atmosphere, at which point a droplet of water condenses on the junction, allowing it to become, in effect, a wet bulb thermometer. This is a consequence of the so-called *Peltier effect*. The cooling is the stopped, and as the water from the droplet reevaporates the junction attains a wet bulb temperature which remains nearly constant until the junction dries out, after which it returns to the ambient soil temperature. While evaporation takes place, the difference in temperature between the wet bulb and the insulated junction serving as dry bulb generates an emf, which is indicative of the soil water potential.

The relative humidity (i.e. the vapor pressure depression relative to that of pure, free water) is related to the soil water potential according to

$$\psi_m + \psi_0 = \frac{R\,T}{M}\,\ln\frac{P}{P_0} = \psi_w \quad \text{(water potentiaal)} \tag{12}$$

where P is the vapor pressure of soil water, P_0 the vapor pressure of pure free water at the same temperature and air pressure, and R the gas constant for water vapor.

The measurement of vapor pressure or relative humidity is obviously highly sensitive to temperature changes. Hence the need for very accurate temperature control and monitoring. Under field conditions, the accuracy claimed is of the order of about 0.5 bar of soil water potential. The instrument is thus not practical at high soil water potential, but can be quite useful considerably beyond the suction range of the tensiometer. It is used mostly in research and manufactured commercially.

To measure the relative humidity a thermocouple of chromel/constantan is used whereby the cold junction is suspended in the porous cup while the other one is fixed in teflon to be used as reference and kept at constant temperature.

The system is based on the Peltier effect whereby the junction of the thermocouple cools of (figure 25, section AB) when during a short period a small current has been sent through. The temperature of the cold junction reaches the dewpoint (T_D) and condensation of a small water droplet occurs. When cooling continues ($T < T_D$) still more condensation of the vapor takes place (figure 25, section BC).
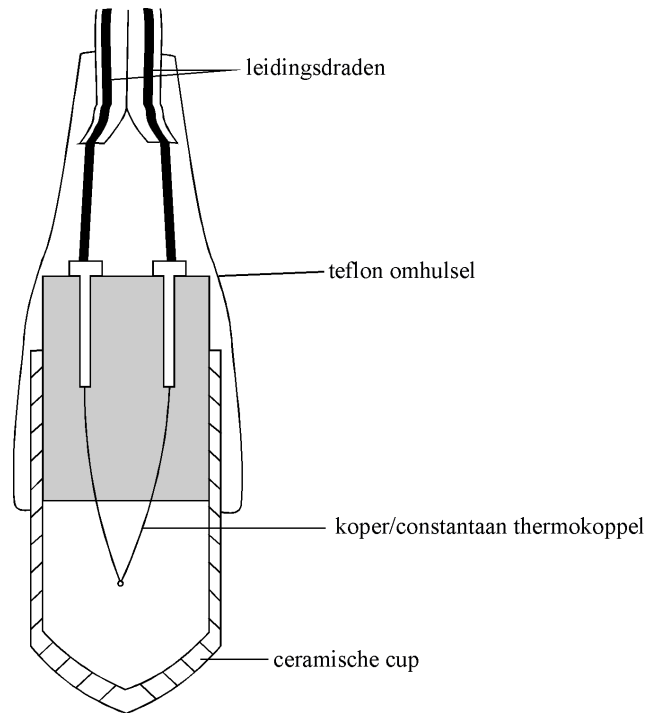
Figure 24.   Cross section of a thermocouple psychrometer contained in an air-filled ceramic and used to measure the $\psi_m + \psi_0$ of the soil water in situ.
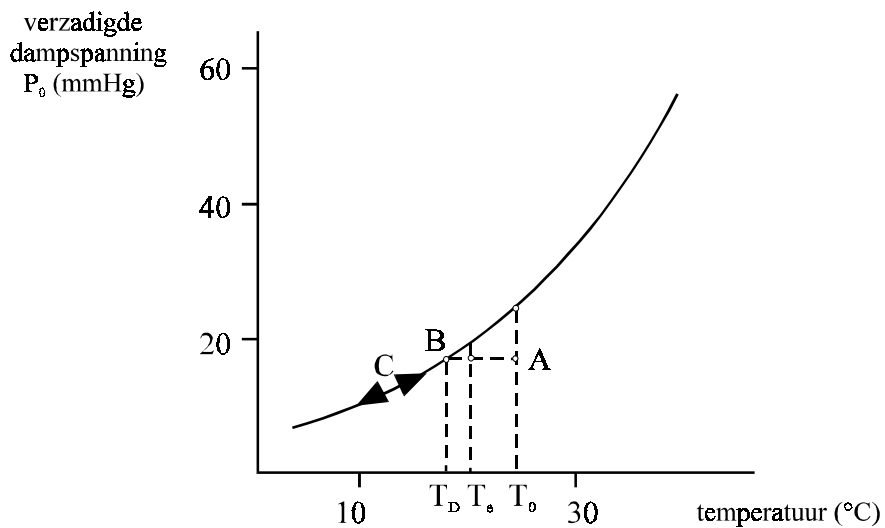

Figure 25. Saturated vapor pressure versus temperature.

After the current is stopped the temperature of the cold junction will increase a little bit until an equilibrium temperature T_e and the water condensed evaporates. The velocity of evaporation is influenced by the relative humidity of the ambient air. The resulting tension (emf) difference as a consequence of temperature difference between the wet and dry junction is measured and is a measure of the relative humidity and consequently of

the soil water potential (ψ_m + ψ_0). Once the whole water droplet disappeared through evaporation the cold junction will reach a temperature equals that of the ambient air.

The technique consists in converting the measured tension difference as a consequence of temperature difference into a vapor pressure and finally to know the ψ_m or ψ_m + ψ_0 of the soil water. Therefore the thermocouple should be calibrated using different solutes (figures 26 and 27).
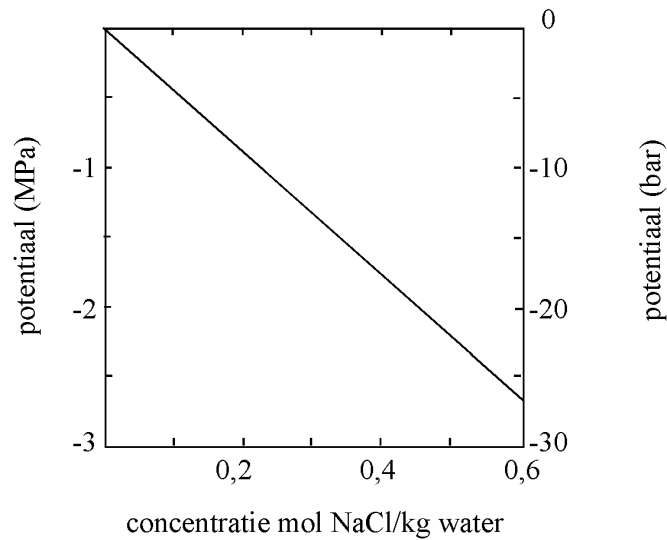
Figure 26.   Relation between different solute concentrations and the osmotic potential used for the calibration of the thermocouple-psychrometer.
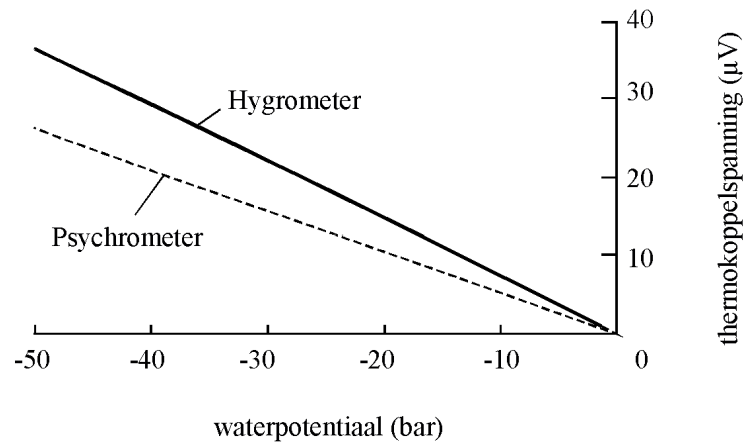
Figure 27. Calibration curve of soil (thermocouple-) psychrometer.

3.3. Hysteresis effect.

The relation between matric potential and soil moisture content can be obtained in two ways:

- in desorption, by taking an initially saturated sample and applying increasing pressure to gradually dry the soil while taking successive measurements of moisture content versus soil water pressure head or matric potential
- in sorption, by gradually wetting up an initially dry soil sample while reducing the pressure.

Each of these two methods yields a continuous curve, but the two curves will in general not be identical. The equilibrium moisture content at a given soil water pressure head is greater in desorption (drying) than in sorption (wetting). This dependence of the equilibrium content and the state of soil water upon the direction of the process leading up to it is called hysteresis. Figure 28. shows a typical soil water characteristic curve and illustrates the hysteresis effect in the soil water equilibrium relationship.

The existence of hysteresis makes that the ψ_m-θ relation is rather complex. In practice the desorption curve is used.
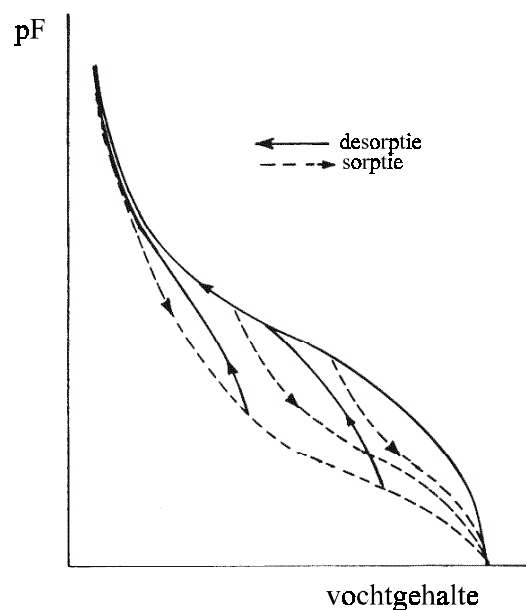

Figure 28. Soil water characteristic curves in sorption and desorption. The intermediate loops are scanning curves, indicating transitions between the main branches.

The hysteresis effect may be attributed to several causes:

1. the geometric nonuniformity of the individual pores, which are generally irregularly shaped voids interconnected by smaller passages, resulting in the *ink-bottle effect*, illustrated in figure 29.

desorptie   sorptie

(a)   (b)
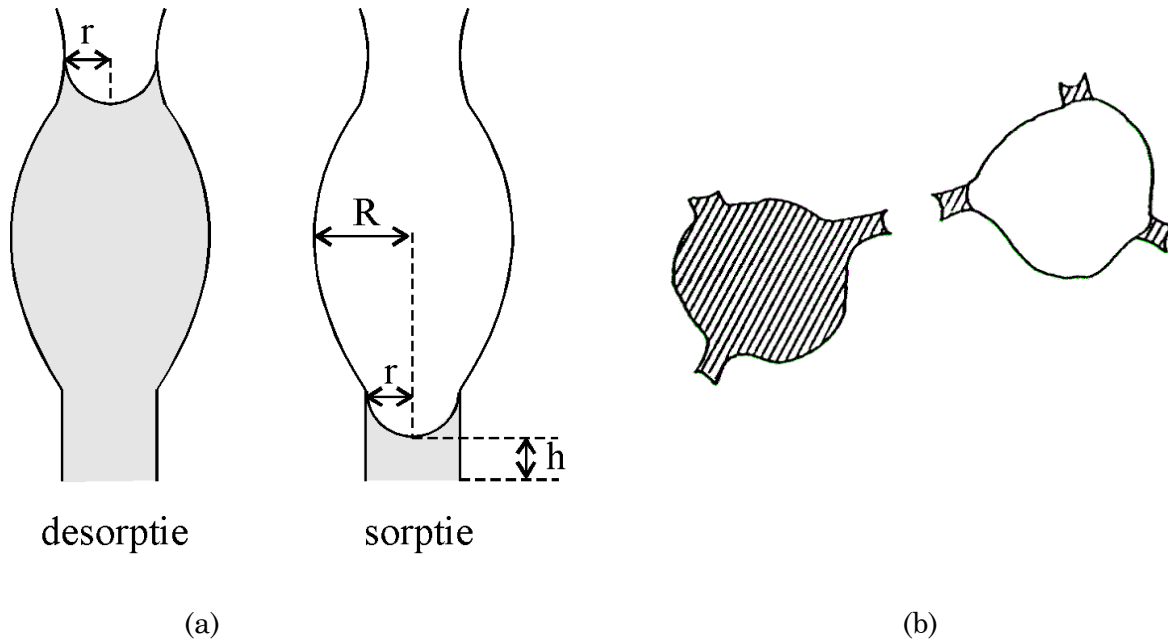
Figure 29.   Ink-bottle effect determines equilibrium height of water in a variable-width pore: in a capillary drainage (desorption) and in capillary rise (sorption).

Consider the hypothetical pore shown in figure 29. This pore consists of a relatively wide void of radius R, bounded by narrow channels of radius r. If initially saturated, this pore will drain abruptly the moment the applied suction exceeds the retention capacity or the soil water pressure should became lower than

$$\Psi_{m(r)} = - \frac{2\,\gamma}{r\,\rho_w}$$

For this pore to rewet, however, the suction must decrease or the soil water pressure head should became greater than

$$\Psi_{m(R)} = - \frac{2\,\gamma}{R\,\rho_w}$$
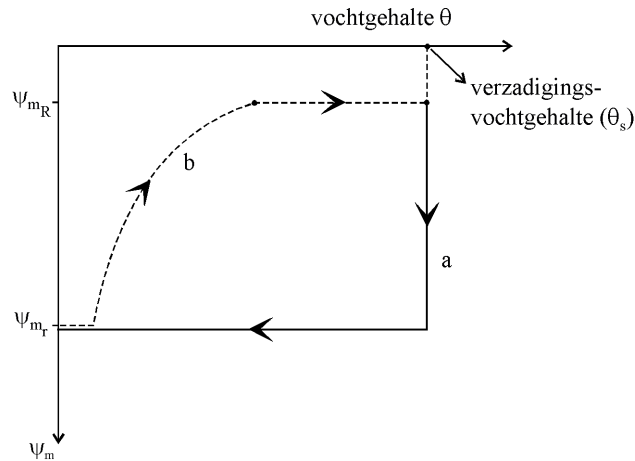
whereupon the pore abruptly fills.(figure 30).

Figure 30.   Desorption(a) and sorption(b) of a pore as presented in figure 29.

Since R > r, it follows that $\psi_{m(R)} > \psi_{m(r)}$, which means that the moisture content in the pore in one case corresponds to a potential $\psi_{m(r)}$ and in the case of sorption to a potential $\psi_{m(R)}$. Desorption depends on the narrow radii of the connecting channels, whereas sorption depends on the maximum diameter of the large pores.

Figure 29 clearly shows that the curvatures at the gas-water interface in both situations are in equilibrium (potential equilibrium), although the moisture content strongly differs. Hysteresis effect is in general more pronounced in coarse-textured soils in the high soil water pressure range, where pores may empty at an appreciable lower matric potential than that at which they fill.

2. the contact angle effect by which the contact angle (α) is greater and hence the radius of curvature is greater, in an advancing meniscus than in the case of receding one. According to following equation:

$$\psi_m = -\frac{2\ \gamma\ \cos\alpha}{r\ \rho_w}$$

one obtains for <u>desorption</u>:

$$\psi_{m(d)} = -\frac{2\ \gamma\ \cos\alpha_d}{r\ \rho_w}$$

and for <u>sorption</u>:

$$\psi_{m(s)} = -\frac{2\ \gamma\ \cos\alpha_s}{r\ \rho_w}$$

51

The contact angle is always greater at sorption($\alpha > 0°$, $\cos \alpha_s < 1$) than at desorption ($\alpha = 0°$, $\cos \alpha_d = 1$), due to impurities at the surface of soil particles or the roughness of the surface so it follows that:

$$\psi_{m(d)} \, \langle \, \psi_{m(s)}$$

The matric potential $\psi_{m(d)}$, whereby the pore with radius r is emptied, is always smaller than $\psi_{m(s)}$, whereby the same pore is filled with water.

3. entrapped air, which further decreases the water content of newly wetted soil. Failure to attain true equilibrium can accentuate the hysteresis effect.

4. swelling and shrinkage phenomena, which result in differential changes of soil structure, depending on the wetting and drying history of the sample. The gradual solution of air, or the release of dissolved air from soil water, can also have a differential effect upon the soil water pressure head – moisture content relationship (pF- curve) in wetting and drying systems.

In the past, hysteresis was generally disregarded in the theory and practice of soil physics. This may be justifiable in the treatment of processes entailing monotonic wetting (e.g. infiltration) or drying (e.g. evaporation). But the hysteresis effect may be important in cases of composite processes in which wetting and drying occur simultaneously or sequentially in various parts of the soil profile (e.g. redistribution). It is possible to have two soil layers of identical texture and structure at equilibrium with each other (i.e. at identical energy states) and yet they may differ in wetness or water content if their wetting histories have been different. Furthermore, hysteresis can affect the dynamic, as well as the static, properties of the soil (i.e. hydraulic conductivity and flow phenomena).

The two complete characteristic curves, from saturation to dryness and vice versa, are called the main branches of the hysteretic soil moisture characteristic. When a partially wetted soil commences to drain, or when a partially desorbed soil is rewetted, the relation of soil water pressure head to moisture content follow some intermediate curve as it moves from one branch to the other. Such intermediate spurs are called scanning curves. Cyclic changes often entail wetting and drying scanning curves, which may form loops between the main branches (figure 28). The h-θ relationship can thus become very complicated. Because of its complexity, the hysteresis phenomenon is too often ignored, and the soil water characteristic curve, which is generally reported, is the desorption curve, also known as the soil water release curve. The sorption curve, which is equally important but more difficult to determine, is seldom even attempted.

3.4. Utility of the soil water characteristic curve.

From the above discussion it is obvious that the lower the matric potential the less the soil water is available for the plant (figure 31). Experiments have shown that soil water at a matric-potential of approximately -15 bar = -1500 J kg-1 = -1500 kPa (pF 4,2) cannot be taken up by plants and permanent wilting occurs. This point is called *"permanent wilting point" (P.W.P.)*. This is the lower limit of the amount of available water for plant growth. Also an upper limit exists and is based on following considerations:

- in the field short after a heavy rainfall or irrigation the soil is not any more saturated since that amount which cannot be retained will drain through gravity.
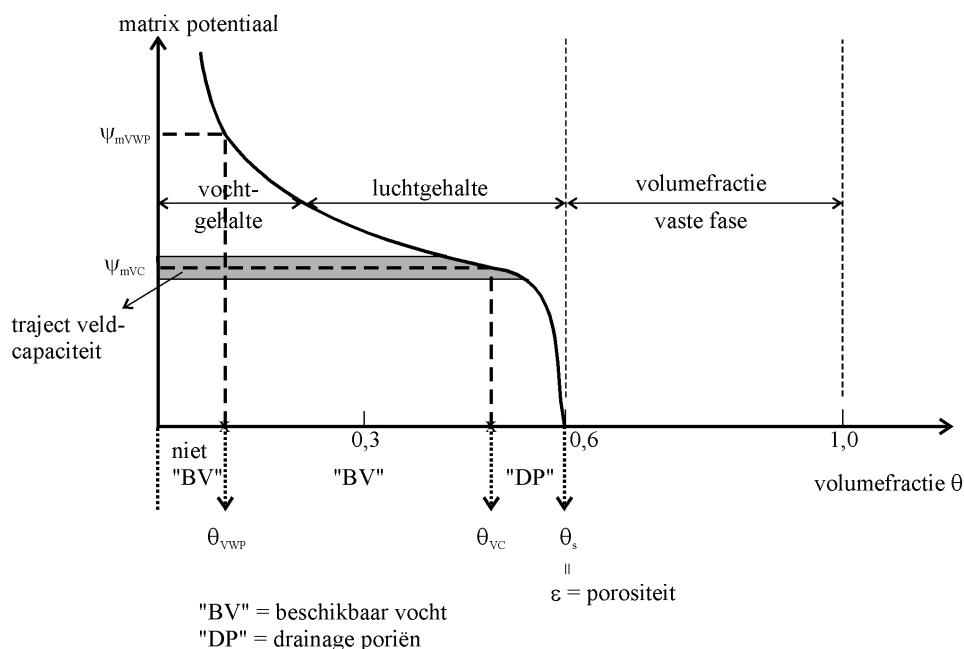- secondly in a saturated soil plant growth is impossible due to lack of oxygen

Figure 31. Soil water regime in function of the soil water pressure head.

On this basis the upper limit of the plant available water content *"field capacity (FC)"* was introduced. This value is rather difficult to define precisely since the amount of water retained in the soil against gravity depends on the boundary conditions of the soil system (e.g. depth of ground water table).

For example in the Netherlands where the ground water table is mostly located at shallow depth (1 to 2 m) a pF = 2 (h = -100 cm water column) is a useful definition for field capacity.

For loamy soils in Belgium, where the ground water table is located at great depths, it was found that the moisture content at -1/3 bar (pF ≈ 2,54) is a useful definition for field capacity, while in sandy soils the moisture content at -0,1 bar (≈ pF 2) reflects the reality.

When a shallow ground water table is present the matric potential at equilibrium following desorption and corresponding to the definition of field capacity is equal to the negative value of of the distance from the point under consideration to the depth of the ground water table.

This discussion shows that field capacity is only a useful definition for a specific situation in the field. It is generally defined as the amount of water, which remained in the soil 2 or 3 days after a heavy rain and whereby evaporation was avoided. The presumed water content at which internal drainage allegedly ceases, termed the field capacity, had for a long time been accepted almost universally as an actual physical property, characteristic of a constant for each soil. However the term field capacity concept may have done more harm than good since when and how can one determine that the redistribution has virtually ceased, or that its rate has become negligible or practically zero.

The redistribution process is in fact continuous and exhibits no abrupt "breaks" or static levels. Although its rate decreases constantly, in the absence of a water table the process continues and equilibrium is approached, if at all, only after very long time. Such example is given in figure 32, where it is shown that the internal drainage of a saturated sandy soil was not ceased after three weeks of internal drainage.

However the field capacity concept has some practical aspects when calculating the amount of irrigation water to be applied at any particular time on the basis of the deficit to the field capacity of the soil depth zone to be wetted without to much water loss through deep percolation.
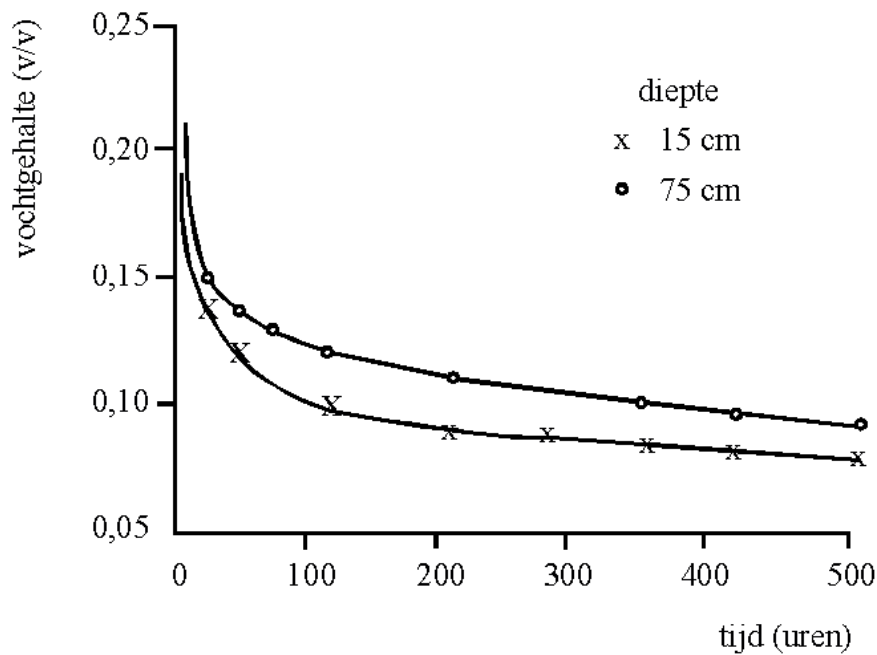
Figure 32.   Change of moisture content in function of time during an internal drainage in a sandy soil and whereby evaporation was avoided.

The moisture content/matric potential of a soil has to be estimated in the field. Therefore a plot, where tensiometers are installed at different depths, is saturated, followed by an internal drainage, whereby evaporation has to be avoided. During the internal drainage the soil water pressure head is measured at regular times. After some days the rate of decrease is strongly reduced and a "quasi" equilibrium is reached and the matric potential at that time is taken as the matric potential at field capacity. Simultaneous soil sampling will allow the estimation of the corresponding moisture content.

Where the moisture content at P.W.P. is determined by the composition of the soil (texture and organic matter) the moisture content at F.C. is function of its composition, soil structure and field situation as well.

The concept of soil water availability has for many years excited controversy among adherents of different schools of thought. The first school claimed that soil water is equally available throughout a definable range of soil wetness, from an upper limit (field capacity) to a lower limit (the permanent wilting point), both of which are characteristic and constant for a given soil. They postulated that plant functions remain unaffected by any decrease in soil wetness until the permanent wilting point is reached, at which plant activity is curtailed abruptly. This schematized model, though based upon arbitrary limits, enjoyed widespread acceptance for many years, particularly among workers in the field of irrigation management.

A second school however produced evidence indicating that soil water availability to plants actually decreases with decreasing soil wetness, and that a plant may suffer water stress and reduction of growth considerably before the wilting point is reached.

A third school seeking to compromise between the opposing views, attempted to divide the so-called available ranges, and searched for a critical point somewhere between field capacity and wilting point as an additional criterion of soil water availability.

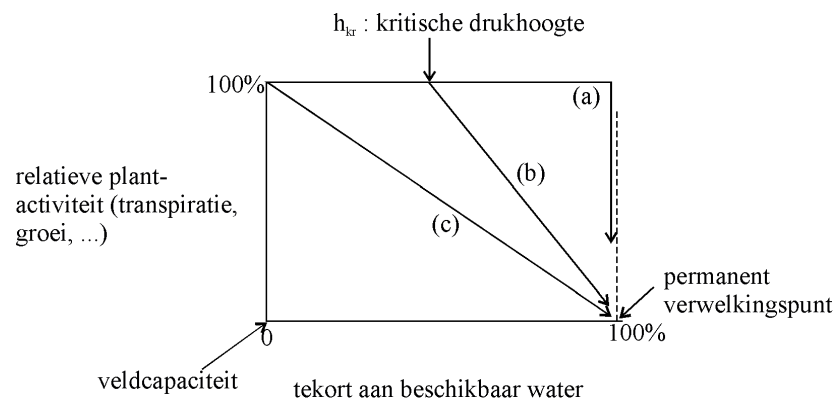These different hypotheses are represented graphically in figure 33.

Figure 33.   Three classical hypotheses regarding the availability of soil water to plants
a. equal availability from field capacity to wilting point
b. equal availability from field capacity to a critical moisture content beyond which availability decreases
c. availability decreases gradually as soil moisture content decreases

None of these school was able up to now to base its hypotheses upon a comprehensive framework that could take into account the array of factors likely to influence the water regime of the soil-water-atmosphere system as a whole.

Since it was recognized that soil wetness per se is not a satisfactory criterion for availability, attempts were made to correlate the water status of plants with the energy state of soil water, i.e. with the soil water potential. The soil water constants were therefore defined in terms of potential values, which could be applied universally, rather than in terms of soil wetness.

The availability of soil water is not only determined by the matric potential but also influenced by the osmotic potential of the soil water. Assume a soil with a $\psi_m = -300$ cm WH and a $\psi_0 = -14500$ cm WH – due to the presence of salts – than the sum of both potentials, also called the water potential ψ_w , is so low that wilting will occur. Due to the presence of solutes the "ψ_w-θ"-curve (pψ_w-curve, in analogy with the pF-curve) is

shifted and the available water is decreased as the moisture content, where wilting occurs, increases as the salt content in the soil water increases (figure 34).
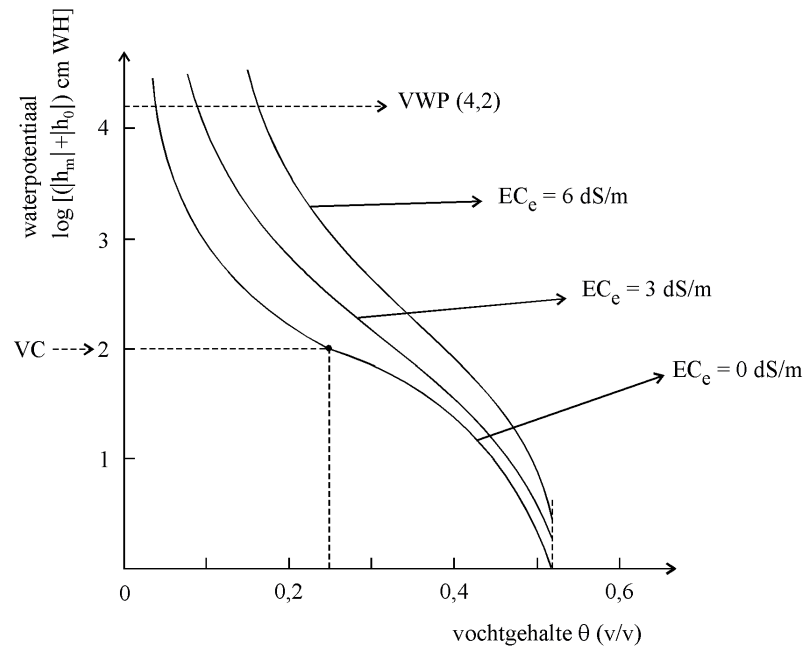

Figure 34.　　The change of the water potential $[\log\,(|h_m| + |h_0|)$ cm WH] in function of moisture content and this for different salinity contents.

So it becomes clear that one should be very careful when using the terms F.C., P.W.P. and available water.

In table 3 are presented some physical values for different textures. These are only indicative values since these can largely be influenced by local factors such as structure, the presence of organic matter etc… (figure 35). Apart from the above-mentioned utility of the soil moisture characteristic curve the h-θ relation is often indispensable when using some models for e.g.:
- the use of soils for landfill
- the transport and behavior of pesticides in the soil
- in relation with leaching of nitrates
- the effect of lowering of the ground water table on plant growth and plant production
- and finally to estimate the unsaturated hydraulic conductivity [K(θ)], a soil physical characteristic needed for solving simulation models of water and solute transport processes in soils
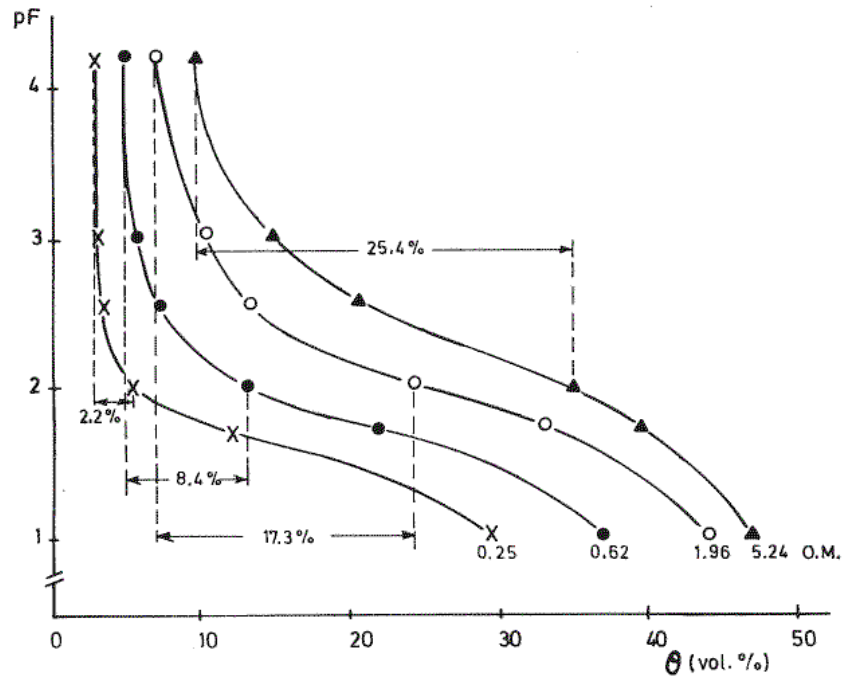
Figure 35.   Influence of organic matter (O.M.) on the amount of available water in sandy soils.

Table 3. Soil physical parameters of different soil textures (Israelson and Hanson, 1962)

Soil texture	Permeability (cm h^{-1})	T.P.V. (%)	ρ_b (g cm^{-3})	F.C. (mass %)	P.W.P. (mass %)	A.W. (mass %)	A.W. (volume %)	mm A.W. till 1 m depth
(1)	(a)	(2)	(3)	(4)	(5)	(6)	(b)	
Loamy sand	5	38	1,65	9	4	5	8	80
	(2,5 - 25)	(32 - 42)	(1,55 - 1,80)	(6 - 12)	(2 - 6)	(4 - 6)	(6 - 10)	
Sandy loam	2,5	43	1,50	14	6	8	12	120
	(1,3 - 7,6)	(40 - 47)	(1,40 - 1,60)	(10 - 18)	(4 - 8)	(6 - 10)	(9 - 15)	
Loam	1,3	47	1,40	22	10	12	17	170
	(0,8 - 2,0)	(43 - 49)	(1,35 - 1,50)	(18 - 26)	(8 - 12)	(10 - 14)	(14 - 20)	
Clay loam	0,8	49	1,35	27	13	14	19	190
	(0,25 - 1,5)	(47 - 51)	(1,30 - 1,40)	(23 - 31)	(11 - 15)	(12 - 16)	(16 - 22)	
Silty clay	0,25	51	1,30	31	15	16	21	210
	(0,03 - 0,5)	(49 - 53)	(1,30 - 1,40)	(27 - 35)	(13 - 17)	(14 - 18)	(18 - 23)	
Clay	0,5	53	1,25	35	17	18	23	230
	(0,01 - 0,1)	(51 - 55)	(1,20 - 1,30)	(31 - 39)	(15 - 19)	(16 - 20)	(20 - 25)	

1. American classification
2. T.P.V. = total pore volume
3. ρ_b = bulk density
4. F.C. = field capacity
5. P.W.P. = permanent wilting point
6. A.W. = available water

a. Strongly dependent on soil structure and structure stability
b. Easily available water is more or less 75% of the available water