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SOIL WATER CONTENT and POTENTIAL: MEASURING TECHNIQUES

SOIL WATER CONTENT

Soil water is important in itself as a feature of the physical environment but especially prominent in its relationships with climatology and with the surface and the subsurface hydrologic regimes as a component in the terrestriel water balance. To understand the behaviour of soil water one must measure it. This have long proved a difficult task both from the instrumental aspect as because of the complexity of the soil body surpasses that of the vegetated layer and the atmosphere above.

The soil water content (wetness) can be expressed in terms of either mass or volume ratios or fractions.

- *Mass wetness or dry mass fraction of water (w).*

$$w = \frac{M_w}{M_s} \tag{1}$$

This is the mass of water relative to the mass of dry (105°C) soil particles where w is the mass wetness, M_w water mass and M_s dry (105 °C) soil mass.

- *Volume wetness or volume fraction of water (θ).*

$$\theta = \frac{V_w}{V_s + V_w + V_a} \tag{2}$$

It is the dimensionless ratio of the water volume (V_w) relative to total bulk soil volume V_t. The latter is the sum of the volume of solids (V_s), water (V_w) and air (V_a) (Fig.1).

The two expressions can be related to each other as follows:

$$\frac{\theta}{w} = \frac{V_w \; M_s}{V_t \; M_w} \tag{3}$$

where: ρ_b = M_s/V_t (mass of dry soil per unit bulk volume and usually lies in the range of 1.1 and 1.7 g cm^{-3})

ρ_w = M_w/V_w (mass of water per unit volume of water and is approximately equal to 1 g cm^{-3})

Equation (3) becomes

$$\theta = \frac{w \, \rho_b}{\rho_w} \tag{4}$$

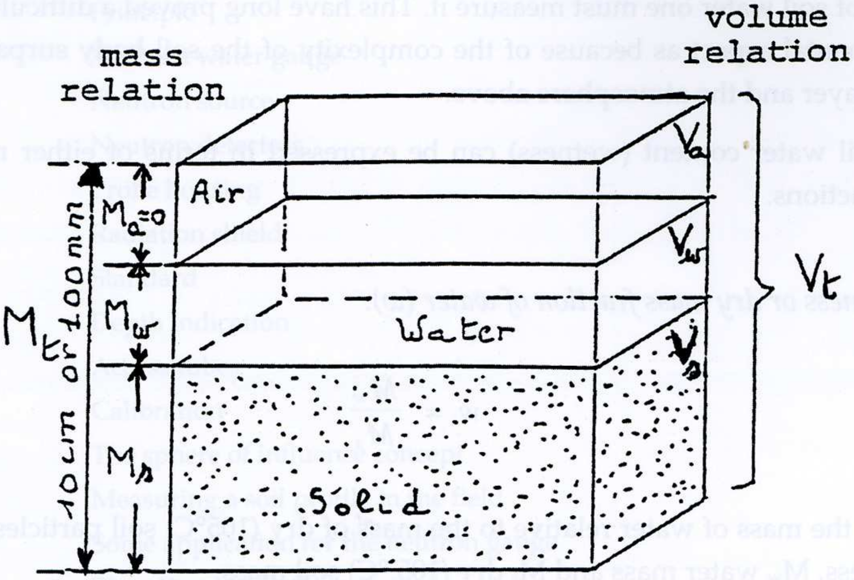

Fig. 1 - Schematic diagram of the soil as a three-phase system.

So, in order to obtain the soil water content on a volume basis the mass wetness is multiplied with the dry bulk density. Both w and θ are usually multiplied by 100 and reported as percentages by mass or volume.

The usefulness of the soil water content by volume lies in the fact that it can be converted easily into head units of water therefore it is compatible with quantities of rainfall or irrigation water applied. All calculations involving the water balance of the soil, including calculation of irrigation deficits, water application efficiency and recharge of soil water reservoir by rainfall, involve the use of the volumetric soil water percentage (vol% of water = mm of water per 10 cm cm soil depth).

If the soil water content is measured at different depths of the profile than the depth-interval Δz_i for which θ_i is valid, is taken as the vertical distance between the measuring point and the points located half-way respectively the above and underlying measuring point (Fig. 2).

$$S = \sum_{i=1}^{i=n} \theta_i \, \Delta z \tag{5}$$

The soil water storage is then equal to the equivalent waterhead (mm or cm) present in the profile until a depth z and is obtained through the summation of the water contents over each depth interval: with the same unit as that of z which is compatible with the amount of precipitation or evaporation.

Two water content profiles, measured at time t_1 and t_2 respectively, allows us to calculate the change in water storage $\Delta S(z_1, z_2)$ during the period Δt.

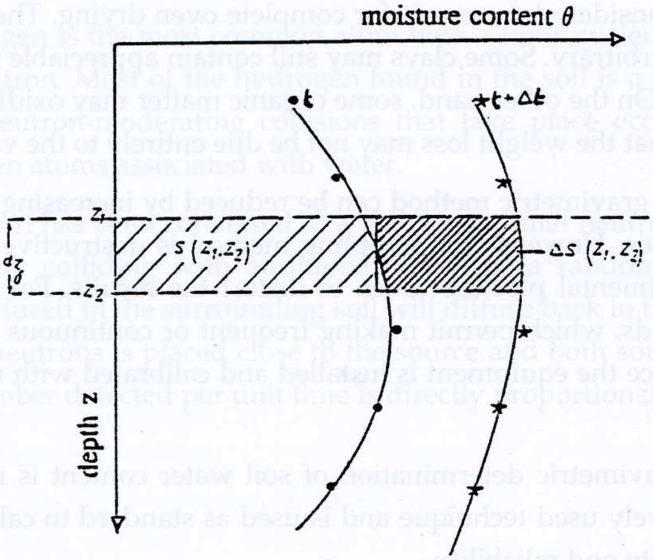

Fig. 2 - Water content profiles.

Measuring Soil Water Content

There are numerous methods available for measuring soil water content. Since each method has advantages and limitations over the other, one should consider both the purpose for which determinations are to be made and the features of each possible method including necessary equipment, operation and maintenance costs.

Gravimetric method

The gravimetric method of measuring soil water content consists of collecting a sample by augering into the soil and then determining its wet and dry weights. The wet weight is determined by weighing the sample at the time of sampling and the dry weight is obtained after drying the sample to a constant weight in an oven. The standard method of drying is to place the sample in an oven at 105°C for 24 hours.

The mass wetness is the ratio of the weight loss in drying to the dry weight of the sample:

$$w = \frac{\text{wet weight - dry weight}}{\text{dry weight}} = \frac{\text{weight loss by drying}}{\text{weight of dried sample}}$$

The method, depending on sampling, transporting and repeated weighings, entails practically inevitable errors. It is also laborious and time consuming, since a period of at least 24 hours is usually considered necessary for complete oven drying. The standard method of oven drying is itself arbitrary. Some clays may still contain appreciable amounts of adsorbed water even at 105°C. On the other hand, some organic matter may oxidize and decompose at this temperature so that the weight loss may not be due entirely to the vaporization of water.

The errors of the gravimetric method can be reduced by increasing the sample size and the number of samples. However the sampling method is destructive and may disturb an observation or experimental plot too much to distort the results. For these reasons, many prefer indirect methods, which permit making frequent or continuous measurements at the same location and once the equipment is installed and calibrated with much less time, labor and soil disturbance.

Although the gravimetric determination of soil water content is rather laborious, it is still the most extensively used technique and is used as standard to calibrate other methods because of its simplicity and reliability.

Neutron scattering technique

The neutron scattering technique is one of the two nuclear methods available for the measurement of the soil water content. It is based on the interaction between high energy neutrons (fast neutrons) and the nuclei of hydrogen atoms in the soil. The other method is based on the attenuation or backscattering of gamma rays as they pass through the soil. Both methods use portable equipment for taking measurements at permanent observation sites. These methods are therefore non-destructive and have the additional advantage of yielding data from the same location at each observation. Both methods require careful calibration with the soil in which the equipment is to be used.

However the nuclear methods have common drawbacks. The equipment is expensive and requires considerable maintenance. It is necessary to train the operator not only to produce satisfatory results but also to observe necessary safety precautions. Indeed improper use in handling the equipment might involve radiation hazards.

Principle. If a radioactive source, emitting high energy neutrons (fast neutrons), is placed in the soil (Fig. 3), the fast neutrons are emitted radially into the soil from the source at high speed, colliding on their way with nuclei of various elements found in the soil. When a fast neutron collides with a heavy nucleus, its direction will be changed but its kinetic energy will be relatively unaffected. However, when the fast neutron collides with a light nucleus (especially H), an appreciable part of its energy gradually lose their energy by this interaction until they have been moderated to "slow" or "thermal" neutrons (through elastic, like billard-ball collisions).

In the soil, hydrogen is the most common atom with a light nucleus and which has the same mass as the neutron. Most of the hydrogen found in the soil is a component of water. So the majority of neutron-moderating collisions that take place occur between the fast neutrons and hydrogen atoms associated with water.

After a fast neutron has been converted to a slow or thermal neutron it will continue to travel through the soil colliding with additional atoms at a random path. Some of the thermal neutrons produced in the surrounding soil will diffuse back to the source so that, if a detector for thermal neutrons is placed close to the source and both source and detector are small enough, the number detected per unit time is directly proportional to the water content of the medium.

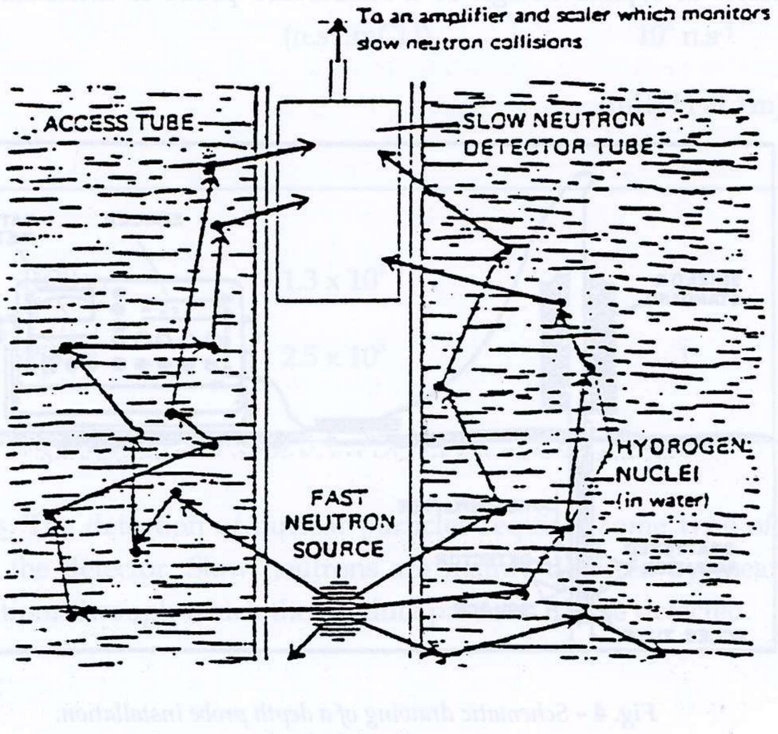

Fig. 3 - Operational principle of a neutron water gauge.

Excluding the presence in the soil of hydrogen atoms not due to water and of the other atoms with light nuclei which may also participate in fast neutron moderation, it can be concluded that the higher the volumetric moisture content of the soil the greater the likelihood that any single fast neutron will collide with the hydrogen atom and become moderated within a given distance from the source.

Conversely the lower the soil water content the greater the likelihood that a fast neutron will traverse a certain distance without colliding with a hydrogen atom. At a given uniform soil water content the number of unmoderated or fast neutrons per unit volume of soil decreases with distance from the source.

From this it is seen that the radius of the volume of soil which effectively participates in neutron moderation is itself a function of soil moisture content.

Neutron water gauge. The basic parts of a neutron water gauge are the neutron source, the neutron detector and the electric power supply and a scaler (readout) unit. Source and detector are contained in a probe housing. For safety reasons the source is shielded by a neutron and gamma radiation shield. To check instrument stability a standard is used. In sub-surface measurements there must be some means of depth indication. In most cases sub-surface probes are lowered into the ground through an access tube. In some cases surface and subsurface probes are combined neutron water and gamma (backscatter or transmission) density probes. These probes also contain a gamma source and a gamma radiation detector. A typical design of a subsurface probe is illustrated schematically in Fig. 4.

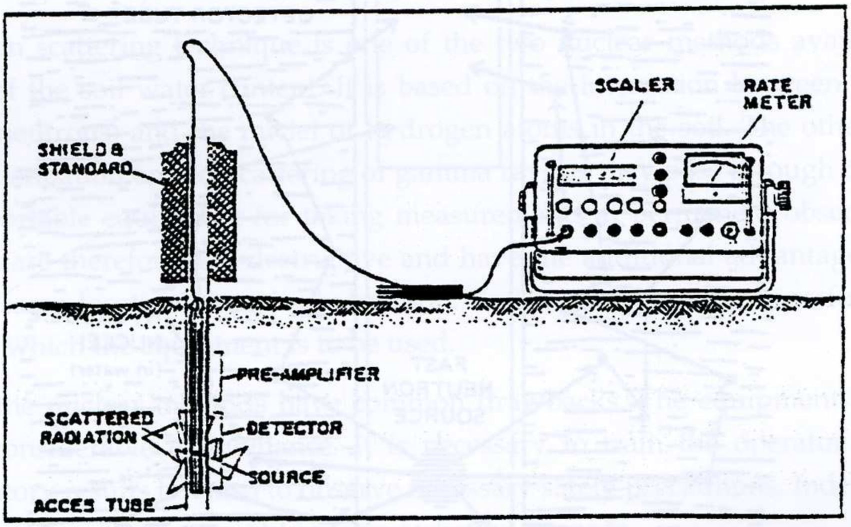

Fig. 4 - Schematic drawing of a depth probe installation.

Neutron source. The most common radio-isotope neutron sources used in water gauges are $^{226}Ra^{9}$-Be and $^{241}Am^{9}$-Be viz., an a-source mixed with beryllium. In these sources, alpha particles emitted by radium or americium are adsorbed by the nucleus of ^{9}Be with the subsequent emission of a neutron, the final product nucleus being ^{12}C. The process may be described symbolically as follows:

$$Be_9^4 + \alpha_2^4 \rightarrow C_{12}^6 + 5.74 MeV$$

The 5.74 MeV is surplus energy of the nuclear reaction and is known as the reaction energy. Most neutron sources emit also gamma radiation.

Radium-beryllium sources have a much higher output of gamma-rays than americium-beryllium sources. Most of the gamma radiation from americium-beryllium is of low energy and thus needs very little shielding. Sources are usually encapsulated in cylindrical double-walled stainless steel containers with welded or silverbrazed seals. Table 1 gives some useful data for these sources.

***Table 1.** Basic data for some neutron sources*

Source	Half-life	Emission ($n.s^{-1}.mCi^{-1}$)	Gamma dose rate per 10^6 $n.s^{-1}$ (mR/H at 1m)
Ra-Be	1620 yr	1.3×10^4	60
Am-Be	458 yr	2.5×10^3	1

Neutrone detectors. The detection of nuclear particles requires some type of interaction of the radiation with the detector. Slow neutrons are mainly detected by means of neutron-induced transmutations through which the product particles can be detected.

Reactions used in neutron detectors for water gauges are :

$$^{10}B\ (n, a)^{7}\ Li$$

$$^{3}He\ (\ n, p)^{3}\ H$$

$$^{6}Li\ (n, a)^{3}\ H$$

$$^{113}Cd\ (n, g)^{114}Cd$$

The first two reactions are used in gas-filled proportional counters. The third reaction is used in scintillation detectors and the last one in cadmium-covered Geiger-Müller tubes.

At presence the most widely used detector is the $^{10}BF_3$-filled proportional counter. The very low sensitivity to gamma radiation makes it possible to have very compact source-detector geometry without shielding between source and detector. On the other hand, this fact makes it difficult to use the BF_3-tube as the detector in a gauge for simultaneous water and density measurement. The BF_3-tube is not sensitive to epithermal or fast neutrons. The main disadvantage of the BF_3-tube is its comparatively low slow-neutron detection efficiency which can, to some extent , be overcome by the use of large tubes.

^{3}He-filled proportional counters have come into use comparatively recently. Their efficiency for slow neutron detection is higher than the BF_3-tubes. They have some efficiency for epithermal and fast neutron detection and a low gamma-detection efficiency. In some countries Geiger-Müller counters covered with cadmium foil are used. They have the advantage that the associated electronic equipment is simple, however with the disadvantage of a low efficiency for slow neutron detection.

Scintillation detectors for neutron water probes make use of europium-activated lithium iodide crystals or lithium-containing glass scintillators. In both cases the lithium used is highly enriched in ^{6}Li. The main advantage of the lithium-containing scintillators compared to gas-filled detectors is their very high neutron detection efficiency. As they are solids, a large amount of the element necessary for the detection of neutrons can be present in a small volume. They also have a high efficiency for gamma-radiation detection and, therefore, a few centimeters of lead shielding between source and detector are necessary. Their sensitivity to gamma radiation makes scintillation detectors useful in gauges for combined water and density determination. In this case both neutrons and gamma rays are detected by the same detector and the pulses separated by means of electronic pulse height discrimination in the counting component of the unit.

Probe housing. A housing is placed around the source and detector to make a robust, watertight measuring probe. For sub-surface probes, cylindrical tubing of aluminium or stainless steel and of minimum diameter is used. The wall thickness particularly for steel is small to minimize absorption of thermal neutrons. Usually the probe housing also contains an electronic preamplifier or impedance converter to permit the use of a long cable without distortion of the pulse signal from the detector.

Radiation shield. Users of neutron water gauges have to be shielded against the neutron and gamma radiation emitted by the source. As neutron shields, hydrogen-containing material such as paraffin or plastics are used. When Am-Be source is used the neutron shield will also shield against the low-energy gamma radiation emitted by the source. When Ra-Be sources are applied it is necessary to use lead to shield against the high-energy gamma radiation emitted.

With sub-surface probes the shield is a separate unit, usually a cylinder into which the probe fits when not in use. Generally, the shield has a hole running along its axis through which the probe can be lowered into the ground. The shield acts as a transport and storage container for the probe. It should be fitted with a lock so that the probe cannot be removed from the shield by unauthorized persons. It should have a proper handle for convenient carrying and placing the probe on top of the access tubes.

Standard. The stability of the gauge has to be checked at regular intervals by carrying out measurements in a standard medium made from a non-hygroscopic, mechanically-stable, hydrogen-containing material. For sub-surface probes the shield container is used as standard. To avoid errors due to instrumental instabilities it is often suggested that the probe is calibrated in terms of N/N_s against volumetric soil water content θ, where N is the count rate in the soil and N_s the count rate in the standard either being the shield container or a water standard.

Depth indication. For sub-surface probes used to determine water profiles it is necessary to know the depth at which the measurement is being made and marks on the cable are commonly used for this purpose. Alternatively, mechanical registers driven by a wheel held against the cable can be used. Sub-surface probes should be supplied with a cable clamp to hold the probe at the required depth.

Access tubes. Sub-surface probes for soil water measurements are mostly introduced into the ground through access tubes which are left in position as long as measurements are required. Access tubes are usually made from aluminum or steel (ordinary or stainless). Aluminum tubes have the advantage that aluminum is not a strong slow neutron absorber. In stony ground or for deep holes aluminum may not be strong enough and steel tubes may have to be used. If the soil has corrosive properties, stainless steel tubes will have to be used.

Access tubes should be closed at the bottom to avoid the entry of water into the tube. When not in use, they should be covered with a stopper to prevent the entry of rain and dirt.

moisture content (vol %)

Installation of the access tube must ensure as tight a fit as possible between the tube and the soil without causing unnecessary soil disturbance.

Calibration. The calibration of a neutron water probe is usually presented as a curve relating the count ratio CR (N/N_s) to the volumetric water content of the soil dried at 105°C. Although manufacturer of probes supply a generalized calibration with each unit, it is generally considered necessary to recalibrate the probe for each soil type since different soil parameters influence the neutron water gauge readings.

Those soil parameters are:

- dry bulk density: as the neutron flux depends on the dry bulk density of the soil, variations in density cause a change in water content. More exactly, a small variation of dry density will involve a displacement of the calibration curve, but when this curve is a straight line, the slope will not be changed very much. Fig. 5 illustrates the influence of soil dry bulk density on water gauge readings. The curves are practically parallel above 25% water content. They converge slightly towards lower water contents.
- soil composition: not all water is expelled at 105°C, some remains as water of crystallization or hydration of various minerals. Hydrogen present does not always stem from water. Hydrogen in organic compounds is not expelled at 105°C but affects the count rate exactly as if it is water.

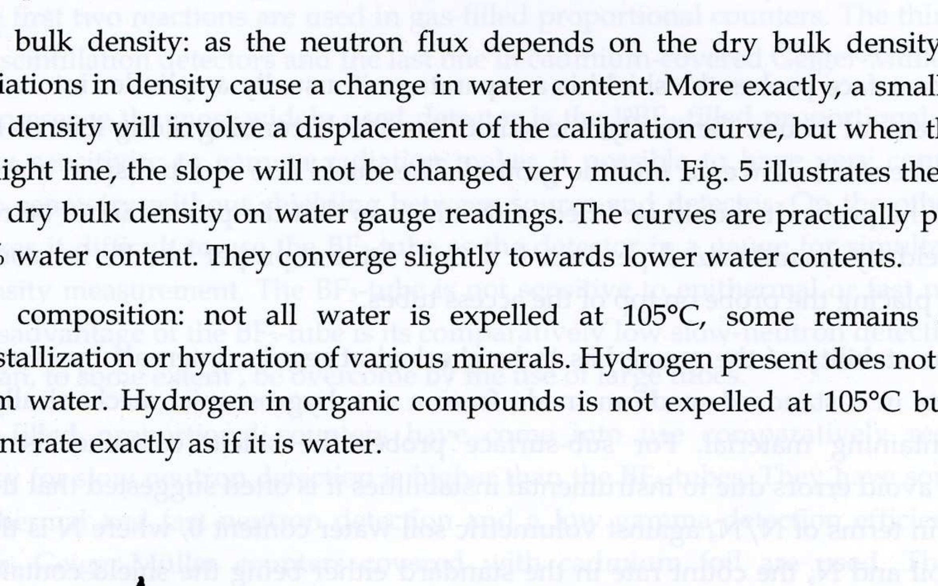

Fig. 5 - Calibration curves of a soil at different dry densities

The presence of neutron absorbing elements in the soil leads to lower thermal neutron flux and gauge readings. Particularly important elements are B, Cl, Mn, Fe, K which have high absorption probabilities.

One may note that all materials in the soil will, to some extent, thermalize and absorb neutrons. Hence the final flux of slow neutrons at the detector depends on the scattering and absorbing properties of all the atomic nuclei in the soil. Besides soil parameters the calibration curve is also influenced by measurements parameters such as:

- diameter and thickness of the access tube: as the calibration curve is also a function of the diameter of the access tube, variations in the diameter will affect the gauge readings. Similarly, variations in the thickness of the access tube change not only the amount of absorption of the thermal neutrons (particularly if an iron tube is used), but the average distance from the probe to the measured medium. To obtain high efficiency and good calibration curve, the access tube should have small diameters in the order 4-6 centimeters.
- the distortion caused by drilling and insertion of the access tube.
- the fall in neutron flux for near surface measurements: when the probe is near the surface, the loss of neutrons leads to a decrease in count rate and the apparent water content measured with neutron water gauge is too low (see sphere of influence).

Consequently it might be necessary to calibrate a neutron probe at various depths within a soil profile. It should further be remembered that "until the effects of non-water hydrogen, bulk density, soil constituents etc. are all resolved, it is necessary to calibrate the neutron gauge separately for different soils to be studied".

There are three basic ways for calibrating soil water against count rate "N" or the count ratio "CR = N/Ns".

(a) Drum calibration in the laboratory.

A laboratory calibration is performed on a large sample of the soil which is dried, mixed and packed into a large drum to its original bulk density. However, only soils which are homogeneous in chemistry and texture, which can be repacked uniformly in the laboratory to something like their field conditions and which do not change their dry bulk density with water content are suitable. According to these criteria only gravel, sandy and silty soils are suitable for drum calibration. The drum must be at least 1 m deep and 1.5 m in diameter to give a valid result (if a smaller drum is used, the results may be erroneous to the escape of neutrons, particularly at the dry end of the water content range). An access tube is installed in the center of the drum and a count rate profile through the central part of the drum. The probe is set in the center of this plateau and a large number of counts are taken and averaged. The mean count ratio is plotted against the measured water content of the drum, giving the first calibration point. The volumetric water content is obtained through undisturbed soil samples taken at the measuring depths.

The soil should now be removed and wetted to a certain water content and leave for equilibrium. Afterwards the drum should be refilled to the same bulk density and the measuring procedure repeated. This procedure should continue over a large water content range. Finally a calibration curve can be obtained using the count ratio values and the corresponding water contents (Fig. 6a).

(b) Field calibration.

This is the common and the standard way for calibration but due to soil heterogeneity and various sampling errors there is often a fairly wide scatter in the calibration points. Many points are required from each site, therefore to perform a linear regression it usually takes a year to finalise the calibration. Since the count ratio is usually interpreted as volumetric soil water content, a knowledge of the soil bulk density is required for calibration. This is probably the largest single source of error in the calibration procedure. Precise count ratio's are obtained from the appropriate depth or depths and at the end of the calibration year at least six known volume soil cores are taken to estimate the bulk density. Fig. 6b gives an example of a field calibration curve.

The sphere of influence concept. Some knowledge of the volume measured by neutron water gauges is of importance for several reasons. First, if laboratory calibration or any other laboratory experiment is planned, samples of proper dimensions must be used. Secondly, in field measurements the volume actually measured should be known to estimate the optimum measurement intervals in sub-surface measurements.

The problem is difficult to deal with and several different simplified approaches have been used. One of these is the concept of the "sphere of influence". It was defined as the sphere around the neutron source which contains 95% of all thermal neutrons. The concept of the sphere of influence has a number of shortcomings. One important shortcoming is that the neutrons in the outer part of the sphere will, for hydrogenous media, have little chance to get to the counter in case the latter is situated near or at the source. A different concept viz. the "sphere of importance" was introduced and it was defined as the sphere around the source, situated in a moderating medium, which if all soil and water outside the sphere was removed, will yield a neutron flux at the source which is 95 % of the flux obtained if the medium is infinite. If the medium is smaller than the sphere of importance, the neutron leakage becomes significant and the count rate will therefore decrease.

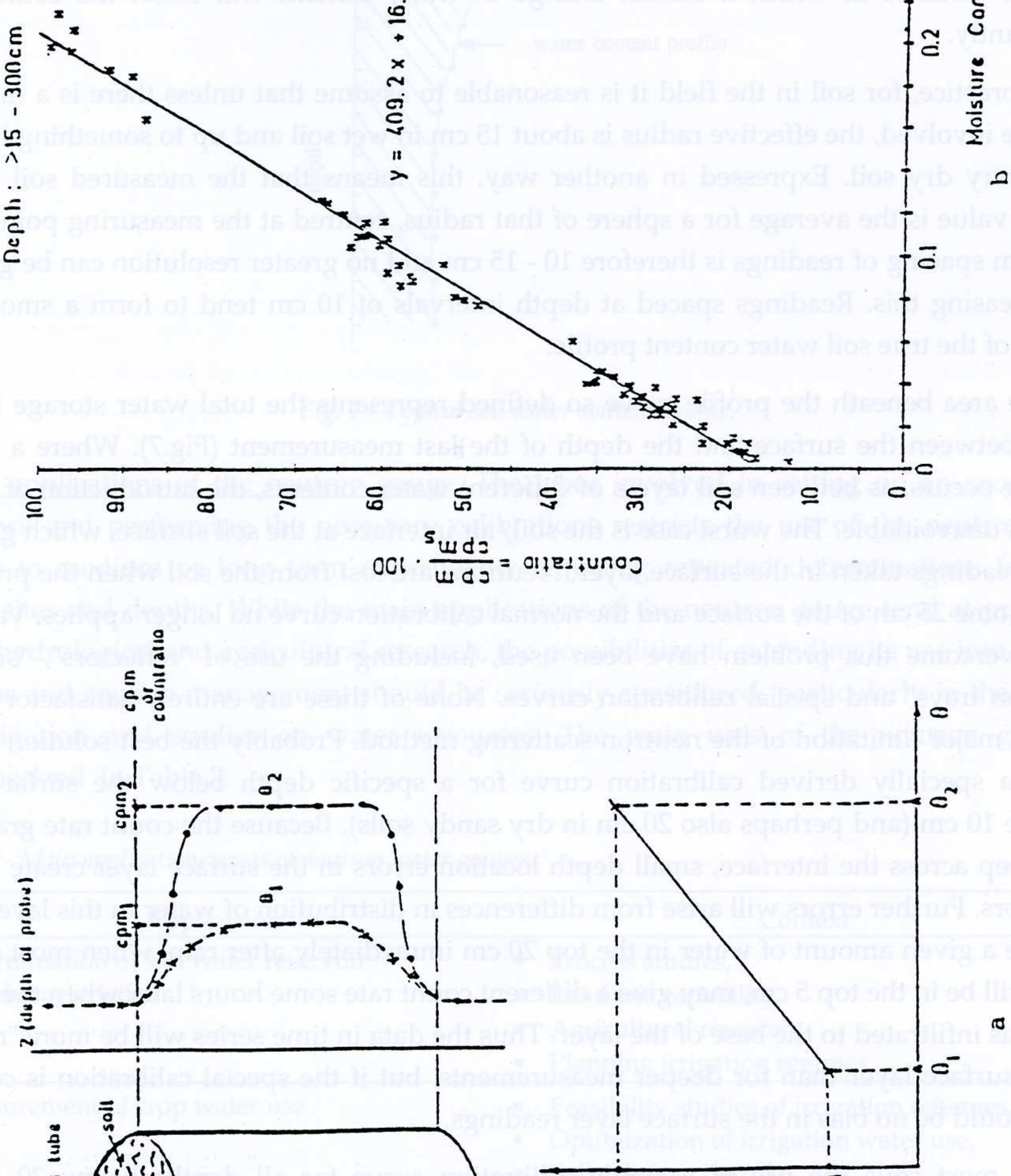

Fig. 6 - Laboratory (a) and field (b) calibration of neutron water gauge.

Measuring a soil water profile in the field. The count rate of slow neutrons depends upon the probability of any fast neutron emitted eventually being scattered back to the detector. Neutrons which become thermalised farthest away from the detector have very little chance of getting back to it since thermal neutrons are easily absorbed. It is thus very difficult to define a distance at which a certain change in water content will affect the count rate significantly.

In practice, for soil in the field it is reasonable to assume that unless there is a marked interface involved, the effective radius is about 15 cm in wet soil and up to something like 30 cm in very dry soil. Expressed in another way, this means that the measured soil water content value is the average for a sphere of that radius, centred at the measuring point. The optimum spacing of readings is therefore 10 - 15 cm and no greater resolution can be gained by decreasing this. Readings spaced at depth intervals of 10 cm tend to form a smoothed version of the true soil water content profile.

The area beneath the profile curve so defined represents the total water storage in the profile between the surface and the depth of the last measurement (Fig.7). Where a sharp interface occurs, as between soil layers of different water contents, the introduction of small errors is unavoidable. The worst case is the soil/air interface at the soil surface, which greatly affects readings taken in the surface layer. Neutrons are lost from the soil when the probe is within some 25 cm of the surface and the normal calibration curve no longer applies. Various ways overcome this problem have been used, including the use of "reflectors", "surface extension trays" and special calibration curves. None of these are entirely satisfactory and this is a major limitation of the neutron scattering method. Probably the best solution is the use of a specially derived calibration curve for a specific depth below the surface, for example 10 cm (and perhaps also 20 cm in dry sandy soils). Because the count rate gradient is so steep across the interface, small depth location errors in the surface layer create count rate errors. Further errors will arise from differences in distribution of water in this layer. For example a given amount of water in the top 20 cm immediately after rain, when most of the water will be in the top 5 cm, may give a different count rate some hours later when the same water has infiltrated to the base of the layer. Thus the data in time series will be more "noisy" for the surface layer than for deeper measurements, but if the special calibration is correct there should be no bias in the surface layer readings.

For most soils the use of a single calibration curve for all depths below 20 cm is justifiable but in strongly layered soils as such, for example, a podzol or a clay overlying a gravel, the use of different calibration curves for each principal soil layer may be necessary.

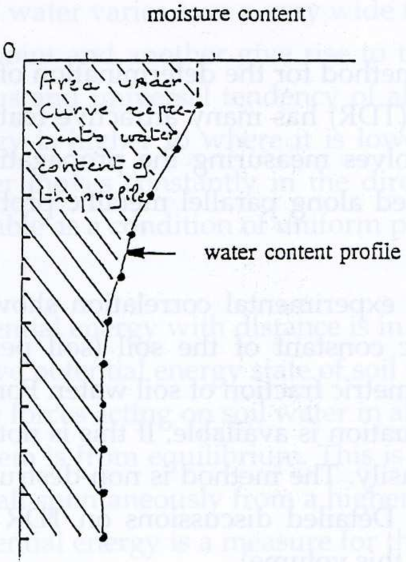

Fig. 7 - Typical soil water content profile.

Some applications of the neutron gauge. The labor involved in setting up an access tube network and performing the necessary calibrations restricts the use of the neutron water gauge to medium or long term experiments requiring repeated determinations from the same sites and depths. While the main applications of the neutron gauge tend at present to be in hydrological and agricultural research, the possibilities of extending its use into applied studies and routine management should be seriously considered, particularly in the context of irrigation and studies on water resources. The main uses of the neutron gauge is summarized in Table 2.

***Tab.** 2 - Main applications areas of neutron water gauges.*

Application	Context
Determination of soil water reservoir characteristics.	• Process studies, • Land use capability, • Agricultural research, • Planning irrigation regimes.
Measurement of crop water use	• Feasibility studies of irrigation schemes, • Optimization of irrigation water use, • Soil salinity control, • Ground water recharge measurements, • Infiltration and drainage studies, • Studies of chemical pollution in unsaturated zone.
Measurement of soil water storage for water balance studies	• Catchment and plot studies for various hydrological studies

Time Domain Reflectometry

The recently developed method for the determination of soil water content by means of Time Domain Reflectometry (TDR) has many attractive features and is already being used worldwide. The method involves measuring the propagation velocity (travel time) of an electromagnetic pulse launched along parallel metallic probes (a sensor) embedded in the soil.

Theoretical analysis and experimental correlation show that the pulse travel time is proportional to the dielectric constant of the soil (soil permittivity), which in turn is a sensitive measure of the volumetric fraction of soil water. For the latter relationship a widely used empirical calibration equation is available. If this is not accurate enough a soil specific calibration can be obtained easily. The method is non-destructive, yield results immediately and can be fully automated. Detailed discussions on TDR can be found in a subsequent chapters (Santini and D'Urso, this volume).

SOIL WATER POTENTIAL

Soil water content is not sufficient to specify the entire status of water in soil. For example, if soils with a same water content but with different particle size distribution are placed in contact with each other, water will flow from a coarse textured soil to a fine textured soil.

One needs to define a property that will help to explain this observation.

Perhaps the following analogy will help. Heat content (analogous to soil water content) is a property of a material that is useful for many purposes. It will not, however, tell us directly whether heat will flow. Therefore a heat intensity term, temperature, has been defined which permits to determine the direction of heat flow. The soil water term that is analogous to temperature (i.e. the intensity with which the water is in the soil) is called the soil water potential. Water potential is a much more complicated property than temperature.

Energy State of Soil Water

Soil water, like other bodies in nature, contains different quantities and forms of energy. Classical physics recognizes two principal forms of energy, kinetic and potential. Since the movement of water in the soil is quite slow, its kinetic energy, which is proportional to the velocity squared, is generally considered to be negligible. On the other hand, the potential energy, which is due to position or internal condition, is of primary importance in determining the state and movement of water in the soil.

The potential energy of soil water varies over a very wide range. Differences in potential energy of water between one point and another give rise to the tendency of water to flow within the soil. The spontaneous and universal tendency of all matter in nature is to move from where the potential energy is higher to where it is lower and to equilibrate with its surroundings. In the soil, water moves constantly in the direction of decreasing potential energy until equilibrium, definable as a condition of uniform potential energy throughout, is reached.

The rate of decrease of potential energy with distance is in fact the moving force causing flow. A knowledge of the relative potential energy state of soil water at each point within the soil can allow us to evaluate the forces acting on soil water in all directions, and to determine how far the water in a soil system is from equilibrium. This is analogous to the well-known fact that an object will tend to fall spontaneously from a higher to a lower elevation, but that lifting requires work. Since potential energy is a measure for the amount of work a body can perform by virtue of its stored energy. Knowing potential energy state of water in the soil and in the growing plant can help us to estimate how much work the plant must expend to extract a unit amount of water.

Clearly, it is not the absolute amount of potential energy "contained" in water which is important in itself, but rather its relative level in different regions within the soil. The concept of soil-water potential is a criterion, for this energy. It expresses the specific potential energy (= per unit mass) of soil water relative to that of water in a standard reference state. The standard state generally used is that of a hypothetical reservoir of pure free water (i.e. water not influenced by the solid phase), at atmospheric pressure, at the same temperature as that of soil water (or at any other specified temperature) and at a given and constant elevation.

Assuming a zero value potential to free and pure liquid water is a accepted convention. Since the elevation of this hypothetical reservoir can be set at will, it follows that the potential which is determined by comparison with the standard is not absolute, but by employing even an arbitrary criterion we can determine the relative magnitude of the specific potential energy of water at different locations or times within the soil.

The concept of soil water potential is of great fundamental importance. This concept replaces the arbitrary categorizations which prevailed in the early stages of the development of soil physics and which purported to recognize and classify different forms of soil water such as gravitational water, capillary water, hygroscopic water.

New definition by soil physics terminology committee of the International Soil Science Society provided more clarity in what used to be a rather complicated theoretical set of criteria. The total potential of soil water was defined as follows : " the amount of work that must be done per unit quantity (mass, volume or weight) of pure free water in order to transport reversibly and isothermally an infinitesimal quantity of water from a pool of pure water at a specified elevation at atmospheric pressure (standard reference state) to the soil water at the point under consideration in the soil-plant-atmosphere-system (Fig. 8).

If work is required the potential is positive, but if water in the reference state can accomplish work in moving into the soil the potential is negative.

Soil water is subjected to a number of force fieldswhich cause its potential to differ from that of pure free water. Such forces result from the attraction of the solid matrix for water, as well as from the presence of dissolved salts and the action of the local pressure in the soil gas phase and the action of the gravitational field. Accordingly the total potential (Φ) of soil water relative to a chosen standard rate can be thought of as the sum of the separate contributions of the various components as follows :

$$\Phi = \psi_g + \psi_o + \psi_m + \psi_{e.p} + \dots$$

where: Φ = total soil water potential

ψ_g = gravitational potential

ψ_o = osmotic potential

ψ_m = matric potential

$\psi_{e.p}$ = external gas pressure potential

… = additional terms are theoretically possible

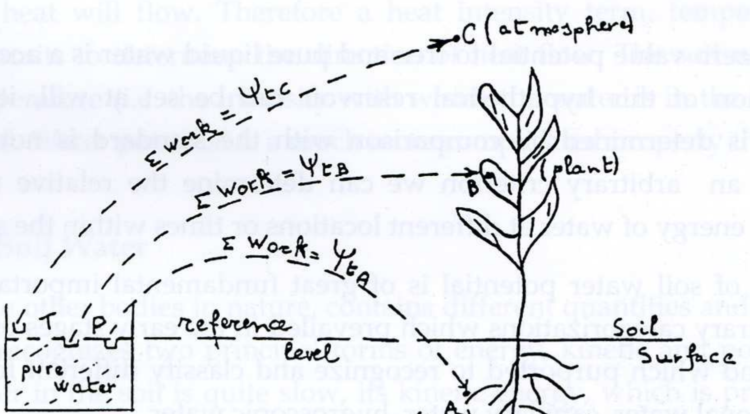

Fig. 8 - Potential of soil water, water in plant cell and water in the atmosphere

The main advantage of the total potential concept is that it provides a unified measure by which the state of water can be evaluated at any time and location within the soil-plant-atmosphere system.

Quantitative Expression of Soil Water Potential

The dimensions of the soil water potential are those of energy per unit quantity of water and the units depend on the way the quantity is specified. Common alternatives used are :

a) Energy per unit mass of water (J/kg)

This method of expression is not widely used.

b) Energy per unit volume of water (pressure) (J/m^3 or N/m^2)

This is the most common method of expressing potential and can be written with units of either Pascal, bar or atmosphere.

c) Energy per unit weight of water (head) (J/N = Nm/N = m)

This method of expressing potential is also common and has units of length.

Table 3 can be used for conversion from one unit to another

Gravitational Potential

Every body on the earth's surface is attracted towards the center of the earth by a gravitational force equal to the weight of the body, that weight being the product of the body's mass multiplied by the gravitational acceleration. To rise a body against this attraction, work must be expended and this work is stored by the rised body in the form of gravitational potential energy. The amount of this energy depends on the body's position in the gravitational force field.

The gravitational potential of soil water at any point is determined by the elevation of the point relative to some arbitrary reference level. If the point in question is above the reference, ψ_g is positive. If the point in question is below the reference, ψ_g is negative. Thus the gravitational potential is independent of soil properties. It depends only on the vertical distance between the reference and the point in question.

At a height z below a reference level (e.g. the soil surface) the gravitational potential of a mass M of water, occupying a volume V is:

$$-Mgz = -\rho_w V g z$$

where: ρ_w = density of water

g = acceleration of gravity

Gravitational potential can be expressed:

- per unit mass: $\psi_g = -g\,z$ (J/kg)
- per unit volume: $\psi_{gv} = \psi_g\,\rho_w = -\rho_w\,g\,z$ (Pa)
- per unit weight: $\psi_{gw} = \dfrac{\psi_g}{g} = -z$ (m)

Osmotic Potential

The osmotic potential is attributable to the presence of solutes in the soil water. The solutes lower the potential energy of the soil water. Indeed, the fact that water molecules move through a semi-permeable membrane from the pure free water into a solution (osmosis) indicates that the presence of solutes reduces the potential energy of the water on the solution side (Fig. 9). At equilibrium sufficient water has passed through the membrane to bring about significant difference in the heights of liquid. The difference (z) in the levels represents the osmotic potential.

Since the osmotic potential of pure free water is zero the osmotic potential of a solution at the same temperature of free water is negative (water flow occurs from point of high potential to one with lower potential).

Differences in osmotic potential only play a role in causing movement of water when there is an effective barrier for salt movement between the two locations at which the difference in ψ_o was observed. Otherwise, the concentration of salts will become the same throughout the profiles by the process of diffusion and the difference in ψ_o will no longer exist. Therefore osmotic potential does not act as a driving force in water flux. This potential is of importance in water movement into and through plant roots, in which there are layers of cells which exhibit different permeabilities to solvent and solute.

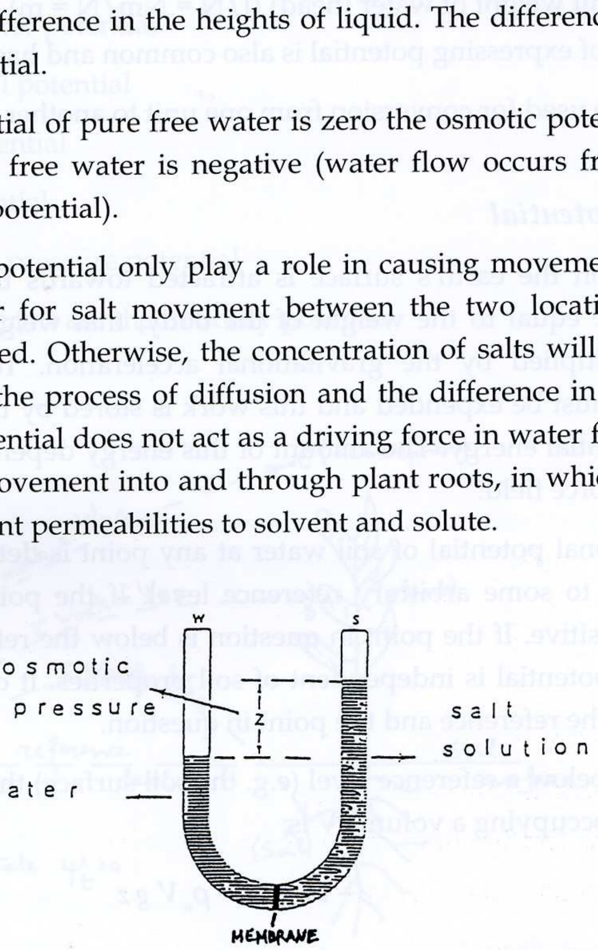

Fig. 9 - Schematic presentation of osmosis.

Matric Potential

Matric potential results from forces associated with the colloidal matric and includes forces associated with adsorption and capillarity. These forces attract and bind water in the soil and lower its potential energy below that of bulk water. The capillarity results from the surface tension of water and its contact angle with the solid particles. In an unsaturated (three-phase) soil system, curvature of meniscus obeys the equation of capillarity

$$P_i - P_a = \Delta P = \gamma\left(\frac{1}{R_1} + \frac{1}{R_2}\right)$$

where: P_a = atmospheric pressure, conventionally taken as zero

P_i = pressure of soil water, can be smaller than atmospheric

ΔP = pressure deficit

γ = surface tension of water

R_1, R_2 = principal radii of curvature of a point on the meniscus, taken as negative when the meniscus is concave

As we assume the soil pores to have a cylindrical shape (Fig. 10) the meniscus has the same curvature in all directions and equation above becomes :

$$P_i = \Delta P = \frac{-2\gamma}{R}$$

since :

$$R = \frac{r}{\cos\alpha}$$

($\alpha = 0$: contact angle between water and the soil particle surface)

$$P_i = \Delta P = \frac{-2\gamma}{r} \qquad \text{with } \Delta P \text{ equals } - h\,\rho_w\,g$$

where: h = height of capillary rise

ρ_w = density of water (10^3kg/m^3)

g = acceleration of gravity (9.81 m/s^2 ≈ 10 m/s^2 ≈ 10 N/kg)

r = radius of the capillary tube

If soil was like a simple bundle of capillary tubes, the capillarity equation might be sufficient to describe the relation of negative pressure potential or matric potential to the radii of water filled soil pores. However, in addition to the capillarity phenomenon, the soil also exhibits adsorption, which forms hydration envelopes over soil particle surfaces. These two mechanisms of soil water interaction are illustrated in Fig. 11.

The presence of water in films as well as under fully filled pores are is important in clay soils. At high suctions or low potential, energy state of soil water is influenced by double layer and exchangeable cations present in soils. In sandy soils adsorption is relatively important and the capillary effect predominates.

In general, however, the matric potential results from the combined effect of the two mechanisms, which cannot easily be separated since the capillary "wedges" are at a state of internal equilibrium with the adsorption "films" and one mechanism cannot be changed without affecting the others. Hence matric potential denotes the total effect resulting from the affinity of water for the whole soil matric, including its pores and particle surfaces together.

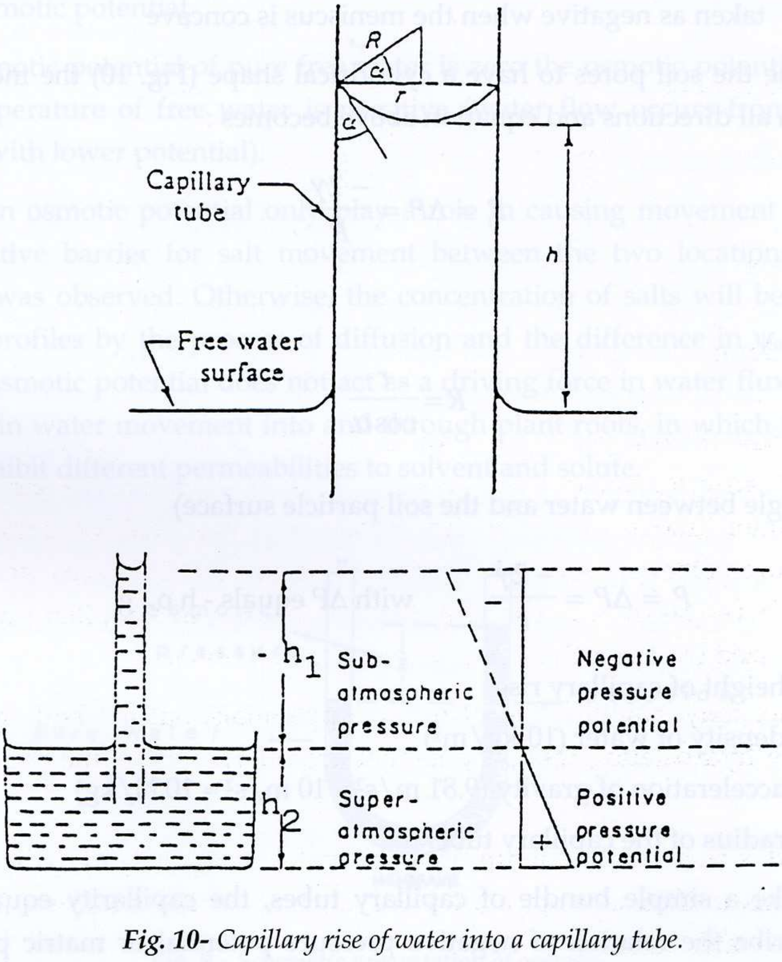

Fig. 10- Capillary rise of water into a capillary tube.

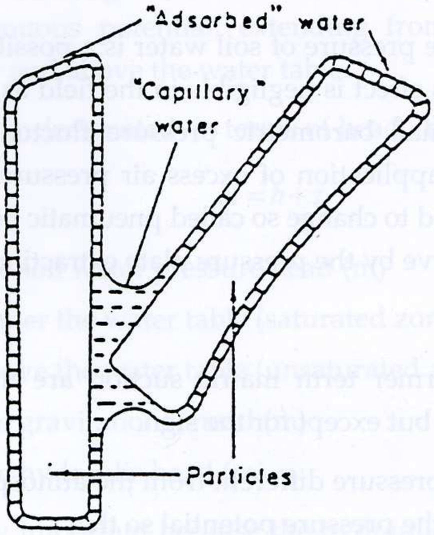

Fig. 11 - Water in an unsaturated soil is subject to capillary and adsorption, which combine to produce a matric potential.

The matric potential can be expressed :

- per unit mass: $$\psi_m = -g\,h = \frac{-2\gamma}{\rho_w\, r} \qquad \text{(J/kg)}$$

- per unit volume: $$\psi_m\,\rho_w = -\rho_w\, g\,h = \frac{-2\gamma}{r} \qquad \text{(Pa)}$$

- per unit weight: $$\psi_m \frac{1}{g} = -h = \frac{-2\gamma}{\rho_w\, g\, r} \qquad \text{(m)}$$

In saturated soil (below the ground water level) the liquid phase is at hydrostatic pressure greater than atmospheric and thus its pressure potential is considered positive (Fig. 10). Thus water under a free water surface is at a positive pressure potential (hydrostatic pressure potential ψ_h), while water at such a surface is at zero pressure potential (assuming atmospheric pressure in the soil) and water rising in a capillary tube above that surface is characterized by a negative pressure or matric potential.

Since soil water may exhibit either of the two potentials, but not both simultaneously, the matric and the hydrostatic pressure potential are referred to as the pressure potential (ψ_p).

Nevertheless it is an advantage in unifying the matric potential and hydrostatic pressure potential in that this unified concept allows one to consider the entire profile in the field below or above the water table.

External Gas Pressure Potential

A factor which may affect the pressure of soil water is a possible change in the pressure of the ambient air. In general this effect is negligible in the field as the atmospheric pressure remains nearly constant and small barometric pressure fluctuations are not significant. However, in the laboratory the application of excess air pressure to change the soil water pressure is a common practice used to change so called pneumatic potential (see determination of the soil water characteristic curve by the pressure plate extraction apparatus)

Final remarks:

1) Matric potential and the former term matric suction are numerically equal - when expressed in the same units - but except for the sign.
2) The effect of an external gas pressure different from the atmospheric (reference) pressure is generally also included in the pressure potential so that :

$$\psi_p = \psi_m + \psi_h + \psi_{e.p}$$

Thus the total potential being in the form

$$\psi_t = \psi_g + \psi_o + \psi_p$$

can fully characterize energy state of soil water and the gradients of these three parameters are the basis for water transport theory.

Hydraulic Head

The total potential is obtained by combining the relevant component potentials :

$$\psi_t = \psi_g + \psi_o + \psi_p$$

Equilibrium, which is defined as the situation where mass transfer of water in the liquid phase is absent, is obtained when the value of the total potential at different points in the system is constant. Usually, single sufficient condition for equilibrium is that the sum of the component potentials, ψ_o being ignored, is constant. The equilibrium condition states then that :

$$\psi_g + \psi_p = constant = \psi_H \qquad (6)$$

called hydraulic potential.

As already stated, the external gas pressure or pneumatic potential in the field is assumed to be zero. Also the soil water within a profile may exhibit either matric or hydrostatic pressure potential (Fig. 10) but not simultaneously. Therefore it is an advantage to define a single continuous potential, extending from the saturated region into the unsaturated region below and above the water table.

The potential is usually designated in terms of head, thus equation (6) becomes :

$$H = h + z$$

where: h = the soil water pressure head (m)
> 0 under the water table (saturated zone)
< 0 above the water table (unsaturated zone)
z = the gravitational head (m)
H = the hydraulic head (m)

The definition is very important because the hydraulic gradient between the two given points in a soil is the driving force for water movement.

In Fig. 12 the condition is applied to a vertical soil column in equilibrium with water table. No water movement occurs in the column. The water table is taken as the reference level for the gravitational potential.

Under the water table, matric potential equals zero, but in this region pressure potential called hydrostatic pressure must be considered, and its value equals to the value of h.

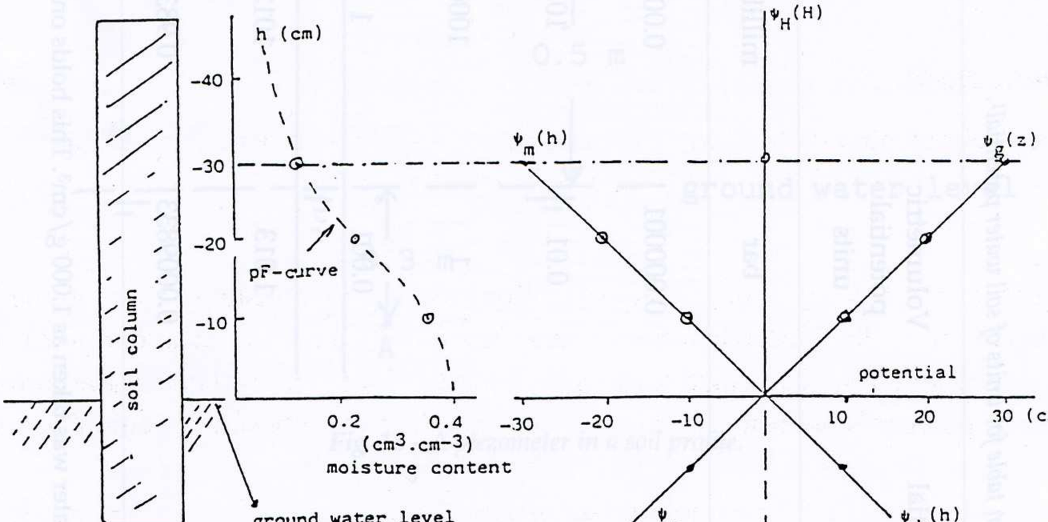

Fig. 12 - Equilibrium condition in a soil column.

***Tab. 3** - Conversion table for units of soil water potential*[1].

Specific potential units	Volumetric potential units				Weight potential units
Joule/kg	bar	millibar	Pa	atmosphere	m
0.0001	0.000001	0.001	0.1	0.000000987	0.001017×10^{-2}
1	0.01	10	10^{3}	0.00987	10.17×10^{-2}
100	1	1000	10^{5}	0.987	10.17
0.1	0.001	1	10^{2}	0.000987	1.017×10^{-2}
101.3	1.013	1013	1.013×10^{5}	1	10.30
0.09833	0.0009833	0.9833	98.33	0.0009703	10^{-2}

[1] The density of water was taken as 1.000 g/cm^3. This holds only at 4°C but is approximately correct at other temperatures.

Measuring Soil Water Potential

Piezometers

As discussed earlier the hydrostatic (positive) pressure potential ψ_h under field conditions applies to saturated soils and is measured with a piezometer (Fig. 13). A piezometer is a tube of a few cm inner diameter, open at both ends, which is installed in the soil profile. If the lower end is below the groundwater table, a piezometer is partially filled with water. By determining the height of the water level in a piezometer it is possible to calculate the (positive) hydrostatic pressure potential of the soil water at the lower end of the tube. The diameter of piezometers is chosen large enough that capillary rise and resistance to water flow are negligible. As a result, any variation in hydraulic potential that may arise inside the piezometer, is instantaneously equalized. Thus, even if the hydrostatic pressure potential at the lower end is changing rapidly, the water inside a piezometer goes through a series of static equilibrium and at any moment it can be assumed that the hydraulic head is uniform and equal to the hydraulic head of the soil water at the open lower end. The static hydraulic head in piezometers can be determined by measuring the depth of the water level, since at the flat air-water interface the pressure potential is zero.

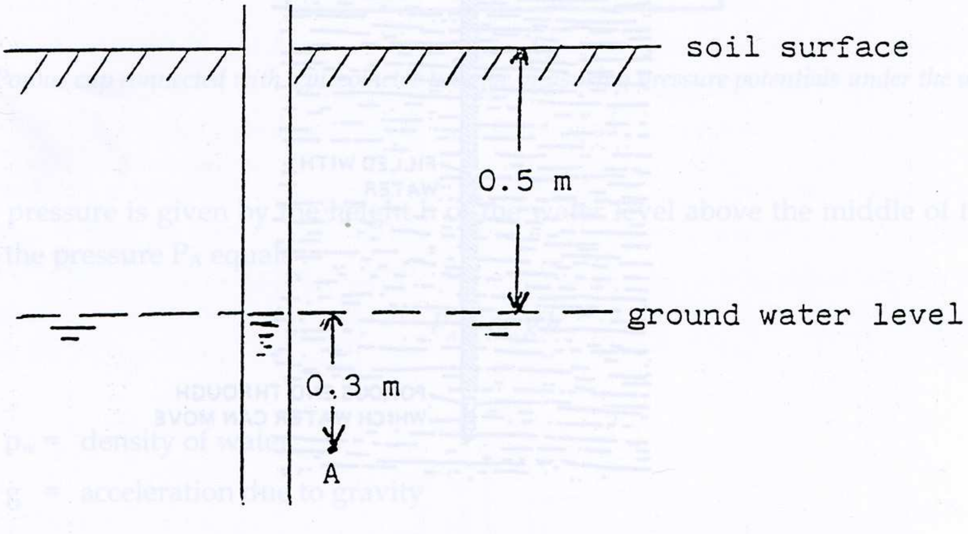

Fig. 13 - A piezometer in a soil profile.

Figure 13 shows a piezometer in a soil profile in which the water is at static equilibrium. The reference point z = 0 is taken at the soil surface. In the piezometer at the water level H = h + z = 0 – 0.5 m = - 0.5 m. Thus at point A, H must also be – 0.5 m (H = h + z = 0.30 m – 0.80 m = - 0.50 m).

The hydrostatic pressure potential expressed per unit weight of water at any point in the soil under water table is the distance between the point and the water level in the piezometer tube.

The water level in a piezometer tube is at the same level as the groundwater table under static equilibrium, independent of the depth of the lower end.

Tensiometers

Piezometers cannot be used to measure negative pressure potentials because in unsaturated conditions, water flows out of the tube into the soil leaving the tube dry. The negative pressure or matric potential can be measured with so-called tensiometers.

The tensiometer consists of a liquid filled porous cup, mostly of ceramic material and connected to a pressure measuring device such as a mercury manometer or vacuum gauge via a liquid-filled tube (Fig. 14).

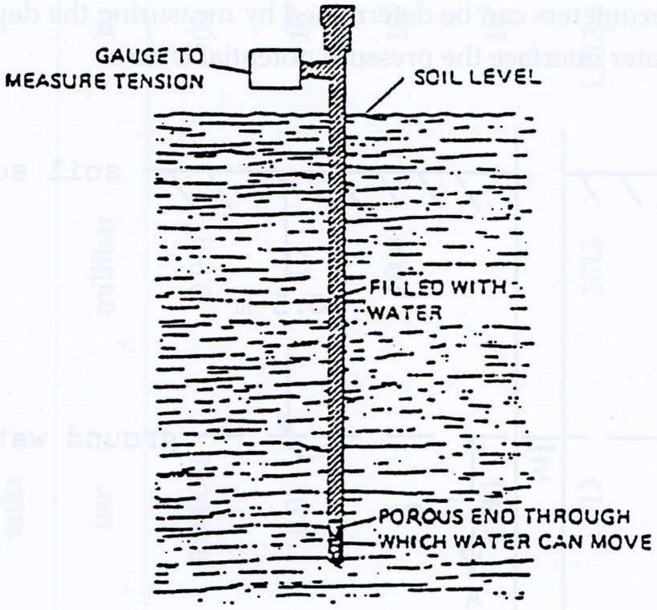

Fig. 14 - Tensiometers with a vacuum gauge.

If the ceramic cup is embedded in soil, the soil solution can flow into or out of the tensiometer through the very small pores in the ceramic cup. Analogous to the situation discussed for piezometer, this flow continues until the (negative) pressure potential of the liquid in the cup has become equal to the (negative) pressure potential of the soil water around the cup. Thus the (negative) pressure potential called matric potential ψ_m of soil water can be measured with a tensiometer, and is therefore also often called tensiometer pressure potential.

Principle. When the cup is placed in a water reservoir (Fig. 15), the water inside the cup comes into hydraulic contact with the water in the reservoir through the water-filled small pores in the ceramic walls. The water level in the tube will indicate the level of the water in the reservoir.

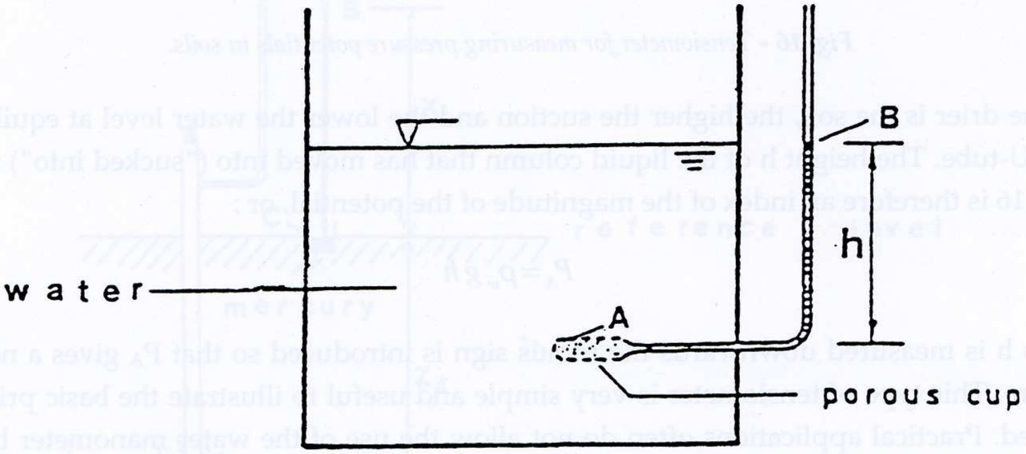

Fig. 15 - Porous cup connected with a piezometer tube for measuring pressure potentials under the water table.

The pressure is given by the height h of the water level above the middle of the porous cup and the pressure P_A equals :

$$P_A = \rho_w g h$$

where: ρ_w = density of water

g = acceleration due to gravity

If we place now the porous cup, connected with an U-shape water filled tube in a soil then the bulk water inside the cup will come in hydraulic contact with the liquid phase in the soil. When initially placed in the soil, the water in the tensiometer is at atmospheric pressure. Soil water in unsaturated soil has a negative pressure and therefore exercises a suction which drawn out a certain amount of water from the rigid and air-tight tensiometer, causing a drop in the water level at the open end of the U-tube (Fig. 16).

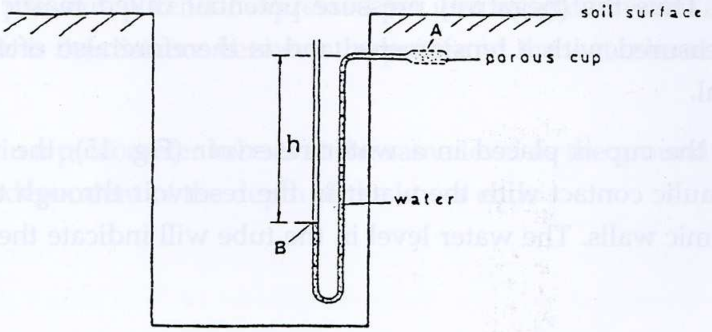

Fig. 16 - Tensiometer for measuring pressure potentials in soils.

The drier is the soil, the higher the suction and the lower the water level at equilibrium in the U-tube. The height h of the liquid column that has moved into ("sucked into") the soil in Fig. 16 is therefore an index of the magnitude of the potential, or :

$$P_A = \rho_w \, g \, h$$

As h is measured downwards the minus sign is introduced so that P_A gives a negative pressure. This type of tensiometer is very simple and useful to illustrate the basic principles involved. Practical applications often do not allow the use of the water manometer because the U-tube extends below the level of the tensiometer cup and measurements thus requires inconvenient, deep pits. Therefore open manometers, filled with immiscible liquids of different densities such as mercury are used so that these problems do not arise (Fig. 17).

Using mercury implies that a relatively short height can indicate a relatively large pressure difference in the manometer (1 cm of mercury corresponds to 13.55 cm of water). Besides the simple water or mercury manometer, a vacuum gauge or an electrical transducer can also be used.

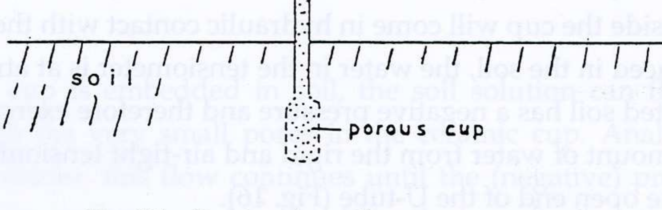

Fig. 17 - Tensiometer with mercury manometer.

<u>Calculating pressure and hydraulic head.</u> Let x be the height of the mercury in the manometer (cm) and z the vertical axis. At the interface, water-mercury in the manometer, the pressure is the same in water and in mercury (being P_B). Thus hydrostatic pressure at point A equals (water column from A to B) to pressure of the mercury column, from point C to B (Fig. 18).

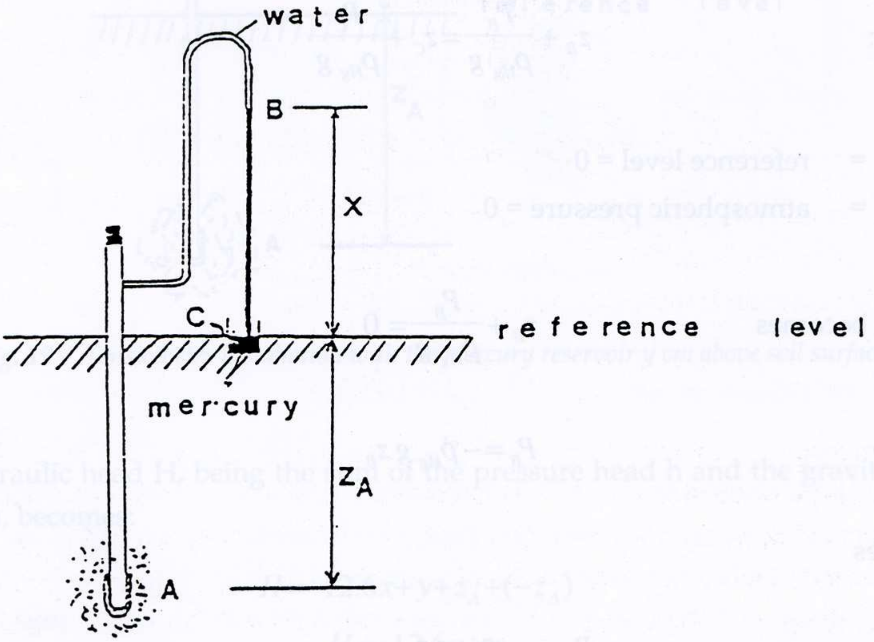

Fig. 18 - Tensiometer installation with the mercury level in the reservoir at the soil surface being the reference level.

Using the hydrostatic law for liquids inequilibrium one obtains per unit weight of liquid the following hydraulic head equation :

$$z + \frac{P}{\rho g} = constant$$

where: z = gravitational head

P/ρ g = pressure head

From Fig. 18 one obtains :

- in water: $$z_A + \frac{P_A}{\rho_w g} = z_B + \frac{P_B}{\rho_w g}$$

Because soil surface is taken as reference level for the gravitational potential, and point A is located below that level, the gravitational head is negative ($-z_A$).

$$P_A - z_A \rho_w g = P_B + z_B \rho_w g$$

$$P_A = P_B + z_B \rho_w g + z_A \rho_w g = P_B + \rho_w g (z_A + z_B) \qquad (7)$$

- in mercury:

$$z_B + \frac{P_B}{\rho_{Hg} g} = z_C + \frac{P_C}{\rho_{Hg} g}$$

because z_C = reference level = 0

P_C = atmospheric pressure = 0

the equation becomes

$$z_B + \frac{P_B}{\rho_{Hg} g} = 0$$

or:

$$P_B = -\rho_{Hg} g z_B \qquad (8)$$

(8) in (7) gives

$$\frac{P_A}{\rho_w g} = \frac{-\rho_{Hg} g z_B}{\rho_w g} + z_A + z_B \qquad (9)$$

since ρ_{Hg} = 13600 kg/m^3

ρ_w = 1000 kg/m^3

(9) becomes

$$h_A = -13.6 z_B + z_A + z_B$$

or

$$h_A = -12.6 z_B + z_A$$

$$h_A = -12.6x + z_A \qquad (10)$$

Normally the three surface of the mercury in the reservoir (point C) is located y cm above the soil surface (reference level) (Fig. 19).

Equation (10) becomes:

$$h_A = -12.6x + y + z_A \qquad (11)$$

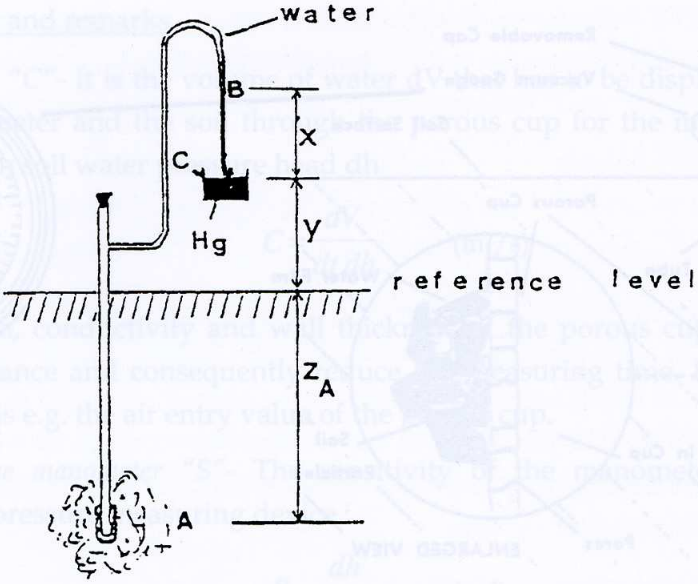

Fig. 19 - Tensiometer installation with the mercury reservoir y cm above soil surface.

The hydraulic head H, being the sum of the pressure head h and the gravitational head z, (H = h + z), becomes:

$$H=-12.6x+y+z_A+(-z_A)$$

$$H=-12.6x+y \qquad (12)$$

If the soil water pressure head around the porous cup changes, the height of the mercury level will change consequently. For a situation where e.g. two tensiometers are connected to the same mercury reservoir, the value y is the same and from equation (12) it follows that :

$$H_1-H_2=-12.6(x_1-x_2)$$

This means that tensiometers connected to the same mercury reservoir yield hydraulic head differences irrespective of the depths. So $H_1 - H_2 = 0$ where $x_1 = x_2$. Such installation provides at once the direction of the water flow: from the tensiometers with low mercury height to those with high mercury height.

The mercury manometer can be replaced by a Bourdon or dial vacuum gauge (Fig. 20) or by a pressure transducer (Fig. 21). For such manometers equation (11) and equation (12) becomes

$$h=h_{man}+z_o=h_{man}+z_1+z_2$$

$$H=h_{man}+z_1+z_2-z_2=h_{man}+z_1$$

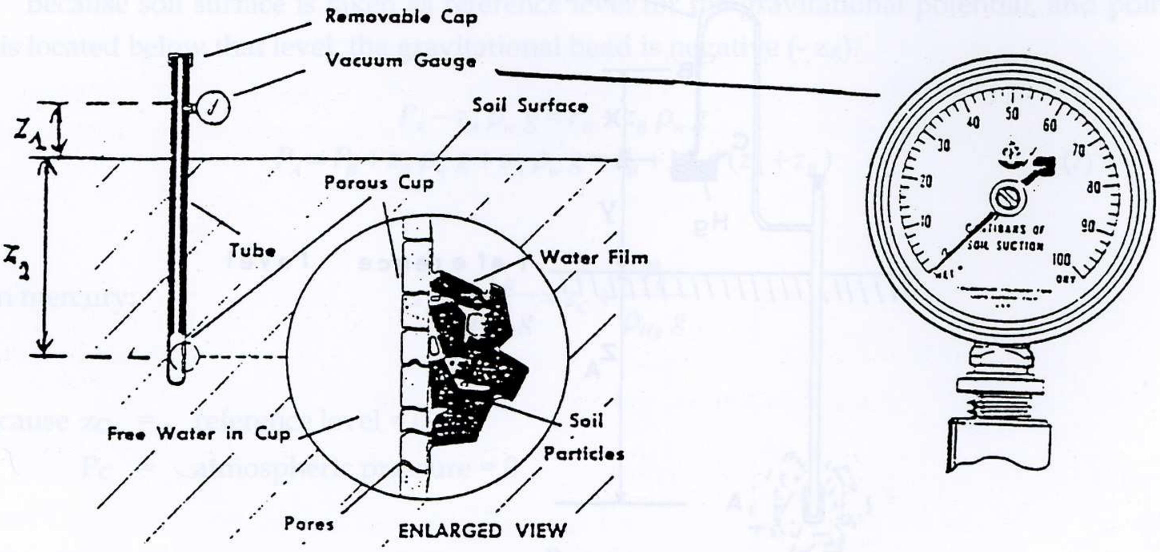

Fig. 20 - Principle of tensiometer with vacuum gauge.

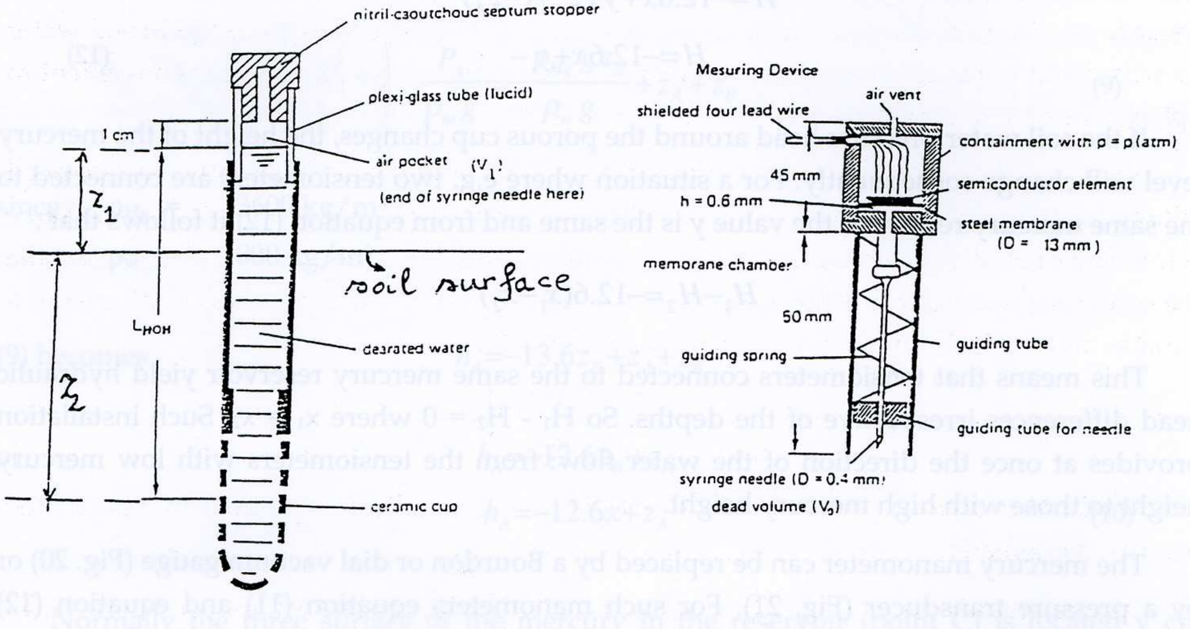

Fig. 21 - Diagram of tensiometer with septum stopper and pressure transducer with attached syringe needle.

Some basic concepts and remarks.

Cup conductance "C"- It is the volume of water dV that has to be displaced per unit time between the tensiometer and the soil through the porous cup for the measuring device to register the change in soil water pressure head dh

$$C = \frac{dV}{dt\,dh} \qquad (\mathrm{m^2/s})$$

The surface area, conductivity and wall thickness of the porous cup can be varied to increase its conductance and consequently reduce the measuring time. However there are practical limits to this e.g. the air entry value of the porous cup.

Sensitivity of the manometer "S"- The sensitivity of the manometer determines the performance of the pressure measuring device :

$$S = \frac{dh}{dV} \qquad (\mathrm{m^{-2}})$$

where dV is the volume of water displaced needed by the measuring device to register a pressure head change dh.

Water/mercury manometer - Water manometer has the lowest sensitivity as it requires a large volume of water to be displaced per unit soil water pressure head change. If the diameter of the manometer tube is large enough to eliminate capillary effects, soil water pressure head can be measured with a precision of ± 1 mm. Since large change in soil water pressure head requires an exchange of a large volume of water with soil, it is convenient to replace water by mercury for measurements in soils.

The mercury manometer reduces the height of the liquid column and thus the displaced volume of water by a factor 12.6 (equation 11) and so increases the sensitivity by the same factor. The overall precision of mercury manometers is often not better than 5 cm. Also the size of the manometer tube influences the sensitivity. Indeed if the cross-sectional area of a manometer tube is A cm^2 the water displacement head ΔV is A Δh so that :

$$S = \frac{\Delta h}{A\,\Delta h} = \frac{1}{A} \qquad (\mathrm{m^{-2}})$$

Bourdon/dial vacuum gauges - The sensitivity are generally comparable to that of the mercury manometer.

Pressure transducers - They have very high sensitivity since they require very small displacements of their sensing element to register a full scale pressure range. However the sensitivity of the tensiometer system is different from that of the pressure sensor. The volume of water that needs to be exchanged with the soil is generally larger because the tubing and air present in the system expand and compress with changes in pressure.

Response time "T_r" - It is the time needed to make an accurate measurement with a particular tensiometer system and is inversely proportional to the conductance of the porous cup of the tensiometer and the sensitivity of the pressure measuring device and is characterized as follows :

$$T_r = \frac{1}{C\,S}$$

This is for a system where the tensiometer is the limiting factor (in water or wet soil). In a soil-limited system (dry soil) the response is much slower under decreasing soil water pressure heads due to small water flow rates to the tensiometer cup as a result of low soil hydraulic conductivities. The flexibility of the tensiometer tubing and the compressibility of air bubbles inside the tubing decrease the effective sensitivity and thus increase the effective response time of a tensiometer system.

Practices and limitations. The purpose of the measurements with tensiometers is to characterize the existing pressure potential of the soil water. Water within the tensiometer should be continuous throughout the system to allow a correct transfer of pressure from the soil to the mercury. Occurrence of gas bubbles disrupts this continuity and makes the system inoperative. The fine porous cup has the function of not allowing entry of air from the unsaturated soil into the water-filled tensiometer tube, even though water can and should move through it. The fine pores inside the wall of the ceramic cup have a high air-entry value which is the pressure needed to remove the water from the pores in the cup replacing it by air. Even with a high air entry value breakdown of the system occurs due to entrapped air within the tensiometer tube or to air coming out of solution at reduced pressure.

Due to the fact that the manometer measures a partial vacuum relative to the external atmospheric pressure, measurements by tensiometers are generally limited to about - 850 cm of water. Use of tensiometers in the field is therefore only possible when pressures do not fall below this value. Moreover the pressure head in the tensiometer must stay above the air-entry value of the porous material which is given as follows :

$$h = \frac{-4\sigma \cos\theta}{\rho_w\, g\, d_{max}}$$

where h = pressure head at which air enters the cup through the largest pore
σ = surface tension of an air-water interface (72.8 dyne/cm at 20°C)
d_{max} = largest pore diameter
θ = contact angle

So to cover the practical tensiometer range to h = - 850 cm for a completely wetting material ($\theta = 0$) d_{max} must be smaller than 3.6 μm. If the air-entry value is exceeded air can enter the tensiometer and the soil will drain the tensiometer.

However, the limited range of pressure measurable by the tensiometers is not as serious as it may be seen at first sight. In many agricultural soils the tensiometer range accounts for more than 50% of the amount of soil water taken up by the plants. To what extend the available water range expressed e.g. as a percentage of the water between pF 2 and pF 4,2 is covered by the tensiometer depends on the shape of the water characteristic curve (pF-curve) as shown for three soil types in Fig. 22.

Thus where soil management (particularly in irrigation) is aimed at maintaining high pressure potential conditions which are mostly favorable for plant growth, tensiometers are definitely useful.

Air diffusion through the porous cup into the system requires frequent purging with deaired water. Tensiometers are also sensitive to temperature gradients between their various parts. Hence the above-ground parts should preferably be shielded from direct exposure to the sun. Therefore it is also suggested to make readings always at the same time of the day (e.g. at 08.00 a.m.).

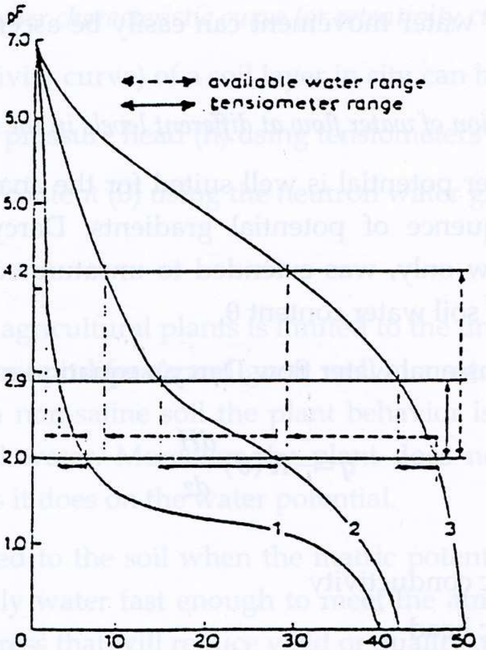

Fig. 22 - Part of the available water content range covered by tensiometers, depending on soil type 1. Sand 50% of available water; 2. Loam 75% of available water; 3. Clay 25% of available water.

When installing a tensiometer it is important for proper functioning that good contact be made between the porous cup and the surrounding soil. Generally the porous cup is pushed into a hole with a slightly smaller diameter to ensure good contact. If the soil is initially rather dry and hard, prewetting of the hole may be necessary. In a stony soil a small excavation should be made and filled with very fine sand into which the tensiometer can be placed.

With mercury manometers, even when small diameter nylon tubing (+/- 2 mm) is used, often a considerable volume of water must be adsorbed by the soil (during water uptake or drying process) or by the porous cup (replenishing by rainfall or irrigation) before the potential that really exists can be read off correctly. A very convenient modern device, the electronic transducer can be used which reacts to very small changes in pressure and converts these changes in a small electrical current which can be registered and amplified by a voltmeter. This system is very accurate but also very sensitive to the occurrence of small air bubbles in the tensiometer system. Moreover it is rather expensive.

Since the porous cup walls of the tensiometer are permeable to both water and solutes, the water inside the tensiometer tends to assume the same solute composition and concentration as soil water, and the instrument does not indicate the osmotic potential of soil water.

<u>Applications of measurements.</u> By the use of several tensiometers at different depths the direction and magnitude of water movement can easily be ascertained. Some examples are given here.

(a) Determination of the direction of water flow at different levels in the soil profile

The concept of the water potential is well suited for the analysis of water flow in soils, since all flow is a consequence of potential gradients. Darcy's law, though originally conceived for saturated flow only, was extended to unsaturated flow, with provision that conductivity is a function of soil water content θ.

For a vertical one dimensional water flow Darcy's equation can be written as follows :

$$q = -K(\theta)\frac{dH}{dz} \tag{13}$$

where q = flux

K(θ)= hydraulic conductivity

H = hydraulic head

= h + z with

h = soil water pressure head

z = gravitational head

The minus sign in the equation indicates that the flow is in the direction of decreasing potential. This means also that if we have two tensiometers located at depths z_1 and z_2 ($z_1 < z_2$) :

- q will be negative (upward flow - evaporation) if $H_2 > H_1$; the rise of mercury in manometer No. 2 is lower than in manometer No. 1
- q will be positive (downward flow - percolation) if $H_2 < H_1$; the opposite situation is observed
- q will be zero (zero flux plane) at a certain depth z when the curve H(z) will show a maximum or the rise of the mercury a minimum because $dH/dz = 0$. A graphical example is presented in Fig. 23

(b) Flux control at a certain depth

From agricultural point of view it would be of interest to know if there is a recharge of the water table or capillary rise. Therefore only 2 tensiometers are needed with a depth distance of say 25 cm in the control zone. A simple reading of the rise of mercury in the manometer will indicate the flow direction.

Knowing the water content θ at the depth between z_1 and z_2, the $K(\theta)$ relation of that soil and the hydraulic head gradient dH/dz, one can calculate the instantaneous water flux q (see equation 13).

(c) Determination of the soil water characteristic curve (or retentivity curve)

The h-θ relation (retentivity curve) of a soil layer in situ can be established :

- knowing the soil water pressure head (h) using tensiometers (see equation 11)
- knowing the soil water content (θ) using the neutron water gauge.

(d) Irrigation scheduling

The root zone for most agricultural plants is limited to the unsaturated part of the profile because the plant roots do not proliferate and grow well in a saturated soil where aeration is limiting. Consequently in a non-saline soil the plant behavior is largely determined by the matric potential of the soil water. Moreover the plant does not depend as much on the quantity of water present as it does on the water potential.

Water should be applied to the soil when the matric potential is still high enough that the soil can and does supply water fast enough to meet the atmospheric demands without placing the plant under a stress that will reduce yield or quality of the harvested crop.

Although the tensiometers function over only a limited part of the available water range (0 to - 800 cm water) it is usually in this range that plants should be irrigated.

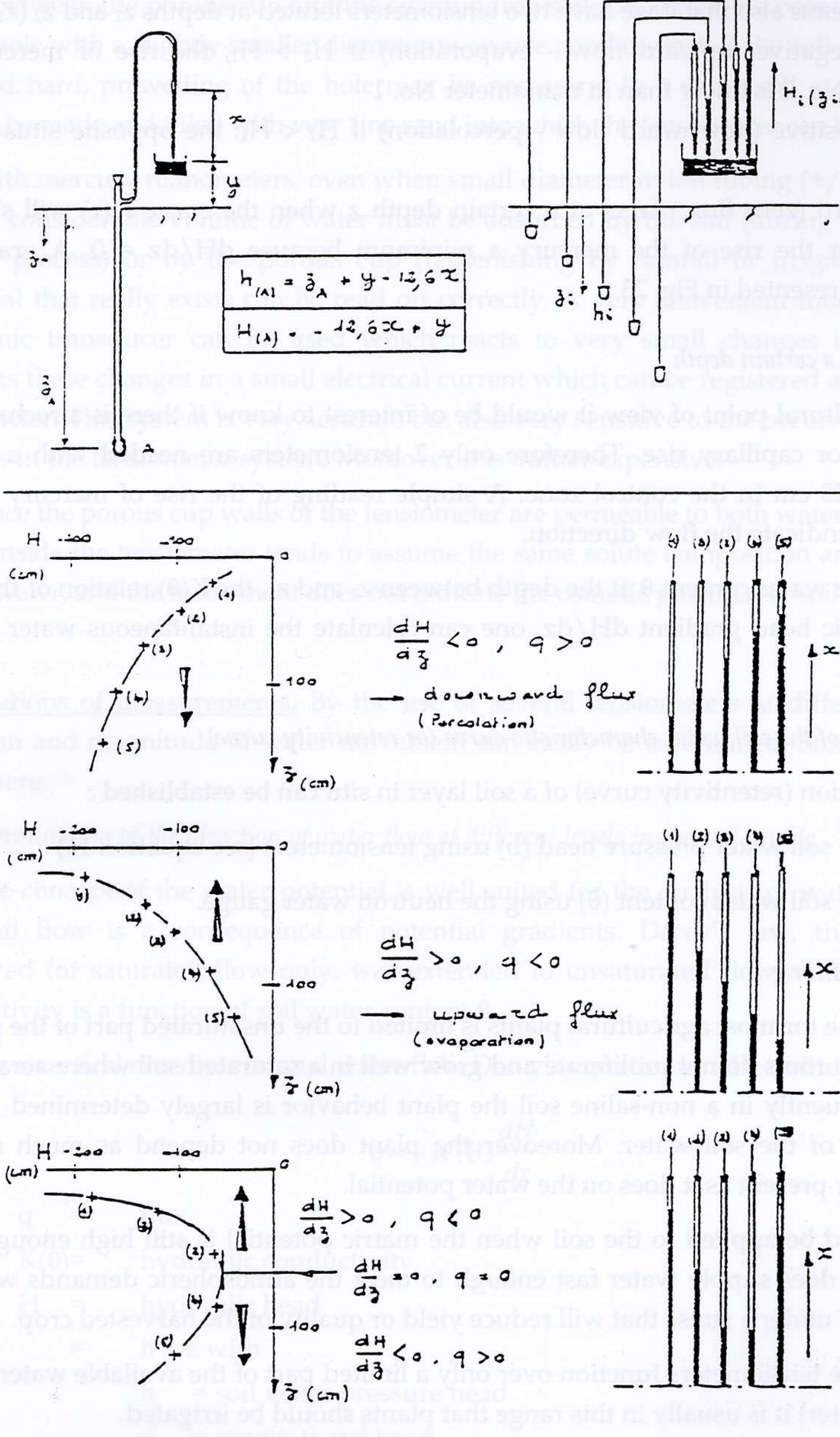

Fig. 23 - Hydraulic head profiles. The manometers from left to right increase with depth.

From practical point of view tensiometers are installed at minimum 2 locations. One unit should be placed in the zone of maximum root activity and another near the bottom of the active rootzone.

The time to irrigate is determined by following the matric potential readings in the zone of the greatest root activity. The exact value of the matric potential at which water should be applied is not the same for every crop. A good approximation of that matric potential is available for many common crops. For most crops it is time to irrigate when the top tensiometer reads - 300 to - 500 cm water and the bottom tensiometer begins to indicate drying.

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