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SMR/104-17

COLLEGE ON SOIL PHYSICS

19 September - 7 October 1983

NUCLEAR TECHNIQUES IN SOIL PHYSICS STUDIES

Klaus REICHARDT

Head, Soil Fertility, Irrigation
and Crop Production Section
I.A.E.A.
Wagramerstrasse 5
A-1400 Vienna
Austria

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by Klaus Reichardt

1. Definitions

Given a soil sample of total mass M and total volume V, these can be split into the three main fractions: solid, liquid and gaseous:

$$M = m_s + m_l + m_g \quad (1)$$

$$V = v_v + v_l + v_g \quad (2)$$

From these fractions we can define:

Soil water content

(1) Gravimetric $u = m_l/m_s$ $g.g^{-1}$ (3)

(11) Volumetric $\theta = v_l/V$ $cm^3.cm^{-3}$ (4)

Soil bulk density

(i) dry $BD_d = m_s/V$ $g.cm^{-3}$ (5)

(ii) wet $BD_w = (m_s + m_l)/V$ $g.cm^{-3}$ (6)

Soil particle density

$$PD = m_s/v_s \quad g.cm^{-3} \quad (7)$$

Soil total porosity

$$\alpha = (v_l + v_g)/V = (V - v_s)/V \quad cm^3.cm^{-3} \quad (8)$$

Some important relations

$$\theta = BD_d \cdot u \quad (9)$$

$$\alpha = (1 - BD_d/PD) \quad (10)$$

Energy status of water will be measured as an equivalent height of water (engergy per unit weight) called hydraulic head (cm H_2O):

$$H = h + z \quad (11)$$

where:

H: total hydraulic head (cm H_2O)

h: pressure head (cm H_2O)

z: gravitational head (cm H_2O)

Equation 11 neglects osmotic and other types of heads. The pressure head h can be positive (positive hydraulic pressures, mostly in saturated and water logged soils) and can be negative (hydraulic tensions in unsaturated soil due to air/water interfaces).

Darcy's steady-state water flux equation defines the soil hydraulic conductivity K (cm/sec):

$$q = -K(\theta) \frac{\partial H}{\partial x} \qquad (12)$$

In this equation q is the flux of water defined as a volume (cm^3) of water crossing a section (cm^2) per unit of time (sec), and x is a position coordinate. $\partial H/\partial x$ is the total hydraulic gradient.

For a given rigid porous material the hydraulic conductivity is strongly dependent on soil water content θ. The K(θ) relation is for most soils esponential, K decreasing rapidly with decreasing θ. The K(θ) relation is for most soils exponential, K decreasing rapidly with decreasing θ. The energy status of water h in a rigid porous material is also related to θ, the relation being however affected by hysteresis. h(θ) is called soil water characteristics.

Darcy's equation for horizontal flow (no gravitational head) can also be written in the form:

$$q = -D(\theta) \frac{\partial \theta}{\partial x} \qquad (13)$$

where D is the soil water diffusivity, defined as:

$$D(\theta) = K(\theta)\, \partial h/\partial \theta \ cm^2 . sec^{-1} \qquad (14)$$

2. Soil water determinations: classical methods:

2.1. Direct sampling

The measurements of u, θ, BD_d, BD_w, PD and involve measurments of masses and volumes. Given a soil sample, the determination of the mass of its fractions is much easier than the determination of their volumes. The size of the sample is important for the representativity in the soil profile and for the accuracy and precision of the determinations. Most common sample sizes range from 50 to 500 g, with volumes of 30 to 300 cm^3.

The mesurement of u does not envolve measurements of volumes and therefore destructive auger samples can be taken. The information about the volumne of the sample is, however, lost, when this method is used. In practice equation (3) is rewritten in the form:

$$u = \frac{W_{ws} - W_{ds}}{W_{ds}} \qquad (15)$$

were W_{ws} is the weight of the wet sample, W_{ds} is the weight of the same sample after drying in oven at 105° C (up to constant weight which in most cases is achieved after 24 hrs). The value of u is also often presented in percent.

Example 1:

$W_{ws} = 358$ g

$W_{ds} = 296$ g

$$u = \frac{358 - 296}{296} = 0.209 \text{ g/g or } 20.9\%$$

To measure θ and bulk densities, the volume of the sample must be known.

One way is to use cylinders of known volumne which are driven into the soil in order to get a soil sample of the size of the cylinder. The operation has to be performed carefully in order to avoid compaction of the sample.

In this case:

$$\theta = \frac{W_{ws} - W_{ds}}{V_{cylinder}} \qquad (16)$$

in equation 16 the mass of water, equal to $W_{ws} - W_{ds}$ is taken equal to its volume.

$$BD_d = W_{ds} / V_{cylinder} \qquad (17)$$

$$BD_w = W_{ws} / V_{cylinder} \qquad (18)$$

Example 2: imagine that the soil sample of example 1 was collected with a cylinder of 250 cm^3. Then:

$$\theta = \frac{358 - 296}{250} = 0.248 \text{ cm}^3/\text{cm}^3 \text{ or } 24.8\%$$

$$BD_d = \frac{296}{250} = 1.184 \text{g/cm}^3$$

$$BD_d = \frac{358}{250} = 1.432 \text{g/cm}^3$$

Note that for the same soil sample u is different from θ, that $BD_w - BD_d = \theta$ and that $\theta = 1.184 \times 0.209$ (equation 9).

Another common way to measure the volume of the soil sample, if its structure permits, is to cover the sample with a thin film of parafin to make it impermeable to water and then measure its volumne through the difference of its weight when hanging in air and hanging in water (Archimede's law).

To measure soil particle density, the volumne of the solid fraction v_s has to be measured. If a dry soil sub-sample (of exampled 2) of, lets say 50 grams is immersed completely in 50 cm^3 of a liquid like ethilic alcool (to avoid trapped air in porous spaces) and its volumne increases to 68.9 cm^3, it is clear that $v_s = 68.9 - 50 = 18.9$ cm^3. And so:

$$PD = 50/18.9 = 2.645 \text{g/cm}^3$$

Using equation (10) we can calculate :

$$\alpha = (1 - 1.184/2.645) = 0.552 \text{ or } 55.2\ \%$$

2.2 Sensors

The measurement of soil water content has also extensively been made by different types of sensors. A few of them are briefly described here.

Porous resistance blocks

They constitute of two electrodes which are embodied in a porous block (gypsum, fiberglass, etc.) and connected to an electric circuit. The resistance between the electrodes is a function of the water content of the porous block which in turn should be in equilibrium with soil water content. The instruments have to be calibrated empirically, normally their resistance is expressed as a function of θ or h.

They are more suitable for the dry range of soil water contents, i.e. between water pressure heads of -1,000 to - 15,000 cm H_2O. Their calibration is problematic, it changes with time and with changes in soil solution quality.

Porous plates under suction. (tension)

A water saturated porous plate (ceramic or fritted glass) can hold water under suctions ranging from 0 to about -800 cm of H_2O, depending on its pore size. A typical arrangement is shown in figure 1.

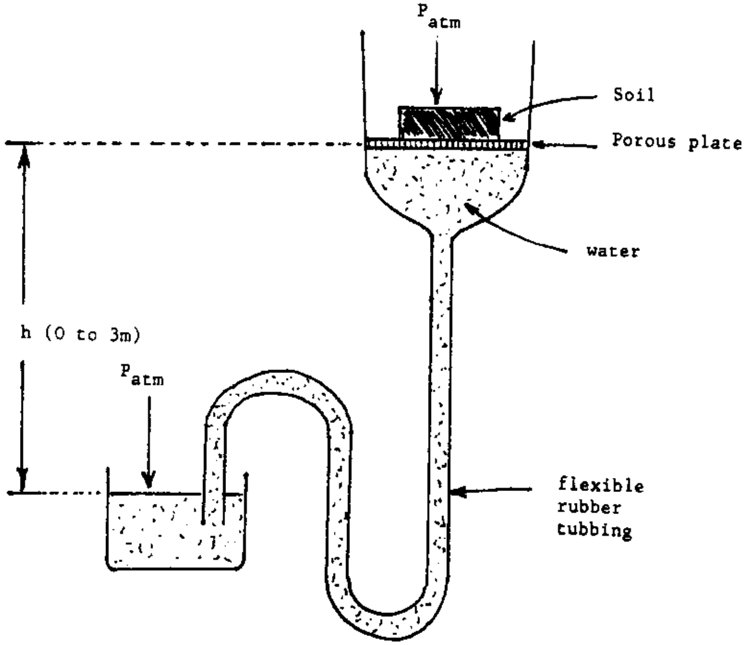

FIGURE 1

If a soil sample is set in good contact with the porous plate, and equilibrium is attained, the energy status of the water in the soil sample is $-h$ (the thickness of the plate and of the soil sample should be negligible). Equilibrium is attained when h does not change with time and then the soil water content θ of the sample should also be constant with time.

Starting with small values of h and obtaining respective values of θ at equilibrium, a relation between h and θ can be established up to values of about -800 cm H_2O. The relation $h(\theta)$ is called soil water characteristics. Typical relations are shown in Figure 2 for two extreme texture soils.

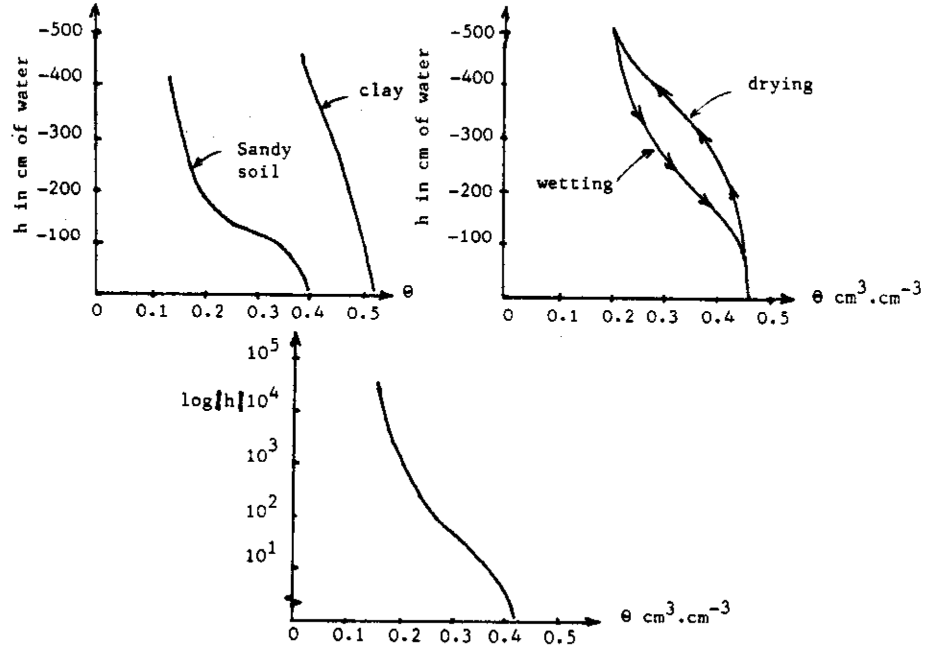

FIGURE 2.

The soil water characteristic depends on type of soil, its structure and also responds to hysteresis, that is, different curves are obtained by wetting and by drying procedures.

Soil water characteristic curves are use to convert h in θ and vice-versa. In the definition of D (equation 14) $\partial h/\partial\theta$ is the slope of the soil water characteristic curve at a given value of θ.

Porous plates under pressure

A water saturated porous plate under pressure can also hold water. Pressures can range from 0 to 20,000 cm H_2O (the agronomically useful range of soil water pressure heads is 0 to 15,000 cm H_2O) depending on its pore size. Different apparatus are designed for different pressure ranges:

(i) Porous Plate funnels 0 - 300 cm H_2O
(ii) "Pressure cooker" 100 - 3,000 cm H_2O
(iii) Richards Pressure Extractor 1,000 - 15,000 cm H_2O

Figure 3 gives a schematic view of these apparatus. All these pressure systems can similarly be used to obtain soil water characteristics. Since h varies several orders of magnitude while θ varies its full range, many times the relation of h versus θ is plotted in semi-log paper, using $|h|$.

Porous cups with water under suction (tension)

The mecanism is the same as in the porous plates. A typical arrangement is the tensiometer, shown in Figure 4.

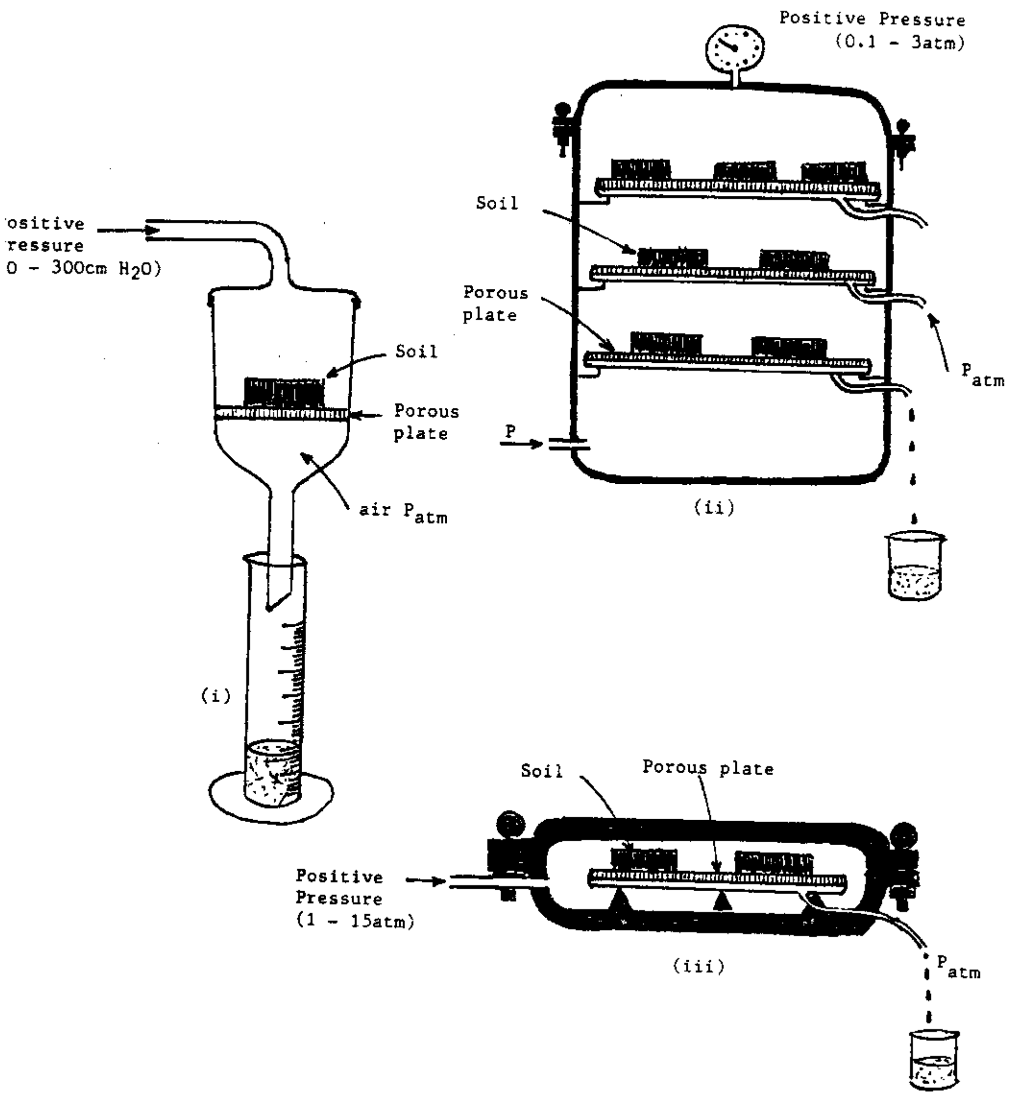

FIGURE 3.

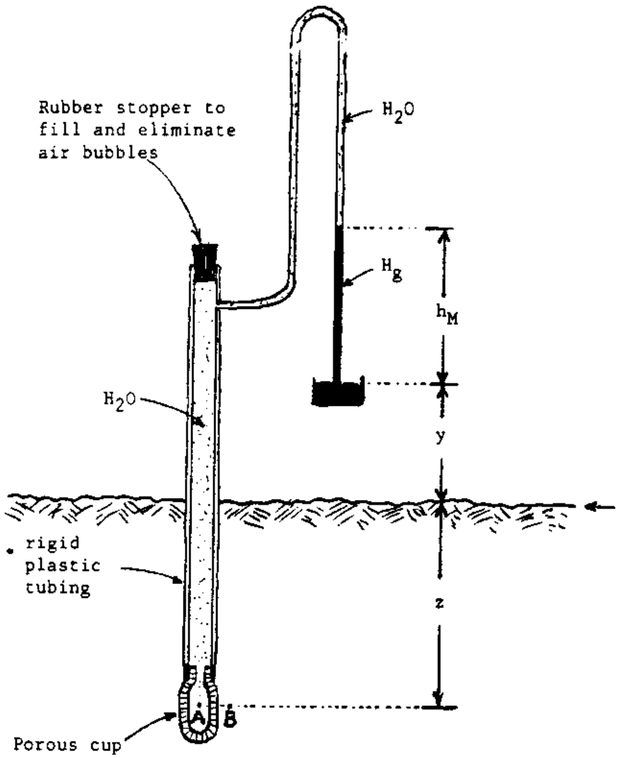

FIGURE 4

At equilibrium the energy status of soil water around the cup (Point B) should be the same as that of the water inside the cup (point A). Then, the hight of mercury in the manometer h_M is a measure of soil water energy status h at depth z. It can be shown that:

$$h = -12.6\, h_M + y + z \qquad (19)$$

where h = soil water pressure head at depth z given in cm H_2O

h_M = high of mercury in manometer given in cm

z = depth of tensiometer cup measured from soil suface, given in cm.

y = hight of mercury/atmospher interface measured from soil surface, given in cm.

Tensiometers are very useful, reversible systems. As soil dries out h_M increases and as soil becomes wetter h_M decreases. Special care has to be taken in their maintance and their useful range is 0 to - 800 cm H_2O. The cup is about 5 cm long and 2 cm in diameter.

Porous cups with air under suction

The relative humidity of the air inside a wet porous cup which is in equilibrium with soil water, is a function of the energy status of soil water. These systems are called psychrometers and are much smaller than tensiometer cups, they are of about 1 cm length and 0.5 cm diameter. The measurement of the relative humidity of the air within the porous cup is done electrically, using the dew-point principle. These instruments are relatively new and still have their problems in practical applications.

3. Soil water determinations: nuclear methods.

3.1. Use of tracers

Tracers can be used to label water and to measure its dynamics, that is, water fluxes and, indirectly, soil hydraulic conductivities and soil water diffusivities. Water can be labelled with Tritium (^{3}H) which is low energy beta emiter with a half life of 12,346 years, which is suitably detected in liquid scientillation counters. The use of natural stable isotopes like ^{2}H (Denterium) and ^{18}O to label water is also a common technique. The natural abundance ratios of $^{2}H/^{1}H$ and $^{18}H/^{16}O$ are artificially changed in order to label the water. The detection of these non-radioactive isotopes is done with mass-spectrometers. Many times natural changes in these ratios are used to trace water, mainly in hydrological and environmental studies. Water can also be labelled through its solutes. This, however, brings problems when the solutes interact with the soil. The use of ^{36}Cl is very common, but a great number of radioisotopes are used in these studies.

3.2 Gamma-radiation Attenuation

Gamma-rays (also called photons) are electro-magnetic waves of very high frequency and very short wavelengths which are emitted by some unstable nuclei. Their energy is usually measured in MeV. Some of the commonly used gamma emitters are:

^{60}Co: emitts two gammas, one of 1.17 MeV and one of 1.33 MeV; and has a half life of 5.3 years.

^{137}Cs: Emitts one gamma of 0.661 MeV and has a half life of 30 years.

^{241}Am: emitts one gamma of 0.08 MeV and has a half life of 458 years.

When a gamma source is enclosed in a lead shield (for exposure protection) of proper size and with a whole (or slid) a gamma-ray beam is obtained. Such a beam can be used to study physical properties of soils, e.g. bulk densities, soil water contents.

When a gamma-ray beam interacts with matter (in our case, soil) three processes might occur, the intensity of each mostly depending on the energy of the gammas. These processes are i: photoelectric effect, ii: Compton effect, and iii: Pair production. Processes i and iii are absorbtions, i.e. the photon disapears during the interaction; and process ii is a scatering i.e. photons loose energy and change path-way.

Having a gamma-ray beam of an intensity I_o (No.of photons per unit area and per unit time) passing through a soil sample, its intensity is decreased to I, due to the three processes cited above. The decrease of the intensity depends on the energy of the gamma radiation and the properties of the soil sample. Figure 5 shows a diagram of a gamma-beam system used to measure soil water content and bulk density.

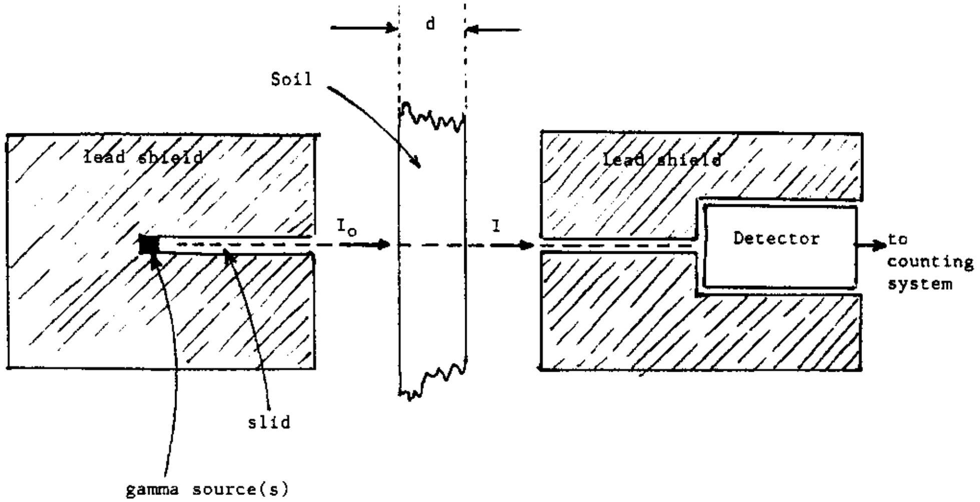

FIGURE 5

It can be demonstrated that

$$I = I_o \exp[-d(\mu_s \cdot BD_d + \mu_w \cdot \theta)] \quad (20)$$

Where d is the thickness of soil material crossed by the beam, μ_s the soil absorbtion coefficient and μ_w the water absorbtion coefficient. Both μ_s and μ_w depend on the energy of the gamma radiation, and u_s of soil type.

If d, μ_s and μ_w are known, measuring I and I_o, it is possible to calculate BD_d and θ. This can be done in several ways:

i: Using a gamma beam of one energy.

Since we have two unknows BD_d and θ, we first measure BD_d in an oven dry sample. In this case $\theta=0$ and

$$I=I_o \exp(-d.\mu_s BD_d) \qquad (20a)$$

Example: For a system as shown in Figure 5 having a $^{137}C_s$ source, the following values were obtained:

$I_o=50{,}468$ cpm
$I=20{,}501$ cpm
$D=8.2$ cm
$\mu_s=0.075\ cm^2.g^{-1}$ (measured separately)

$$BD_d = \frac{\ln(I_o/I)}{d.\mu_s} = 1.465 g.cm^{-3} \qquad (20b)$$

If the sample is now wetted we can measure θ, assuming that BD_d does not change with the wetting (which is true only for rigid porous materials, e.g. non-swelling soils with good structure). For the above example, after wetting, I decreased to 16.018 cpm. Then:

$$\theta = \frac{\ln(I_o/I)-d.\mu_s.BD_d}{d.\mu_w} = 0.367 cm^3.cm^{-1} \qquad (20c)$$

$\mu_w=0.082 cm^2.g^{-1}$ (measured seperately).

The coefficients μ_s and μ_w can be obtained for a given gamma energy using the same equations for samples of known BD_d and θ. For u_s we use oven dry soil samples of known BD_d, e.g. those obtained with cylinders described on page 5. For u_w we use pure water, for which $\theta = 1 cm^3.cm^{-1}$

If for an oven dry soil column we measure BD_d at fixed points, say every cm, we can then measure θ during any wetting process at the same points. Many infiltration and redistribution studies have been made with this technique. The main advantage is that the technique is non-destructive and changes of θ with time can be measured in the same sample, which is not possible with the classis methods.

ii. Using a gamma beam of two energies.

Since equation (20) has two unknowns BD_d and θ, a beam with two energy gammas could be used in order to obtain a set of two equations with the same unknowns. The gamma rays have to be of different energies in order to detect them separately and to have different attenuation coefficients μ_s and μ_w for soil and water.

The system is the same as in figure 5, having only two gamma sources in the lead shied. For this method there is no need to oven dry sample and BD_d and θ are measured simultaneously. The porous system does not have to be rigid, swelling and shrinking soils can also be tested.

Equation (20) can be written for two different gamma energies a and b as follows:

$$I_a=I_{oa}\exp[-d(\mu_{sa}.BD_d+\mu_{wa}.\theta)] \qquad (20d)$$

$$I_b=I_{ob}\exp[-d(\mu_{sb}.BD_d+\mu_{wb}.\theta)] \qquad (20e)$$

and in matrix form:

$$\begin{bmatrix} \ln(I_{oa}/I_a) \\ \\ \ln(I_{ob}/I_b) \end{bmatrix} = d \begin{bmatrix} u_{sa} & u_{wa} \\ \\ u_{sb} & u_{wb} \end{bmatrix} \begin{bmatrix} BD_d \\ \\ \theta \end{bmatrix} \qquad (20f)$$

iii. Using three energies.

As it can be seen from the above, the thickness of the soil sample d which is crossed by the beam has to be known. Small errors in d can lead to big errors in BD_d and θ. Most recently attempts are being made to use 3 gamma energies in order to have also d as an unknown in equation (20)

Gamma-ray attenuation to measure BD_d and θ has several advantages over classical methods, but its application has to be done properly. FERRAZ and MANSEL (1979) describe in detail the method, its advantages and difficulties, and mainly experimental errors associated with the technique. Updated information can also be found in FERRAZ (1983).

For field work, there are available gamma probes, based on the attenuation principle. Two parallel rods are pushed into the soil down to a desired depth, and one having the gamma source and the other the detector. Figure 6 shows the geometry of such equipment.

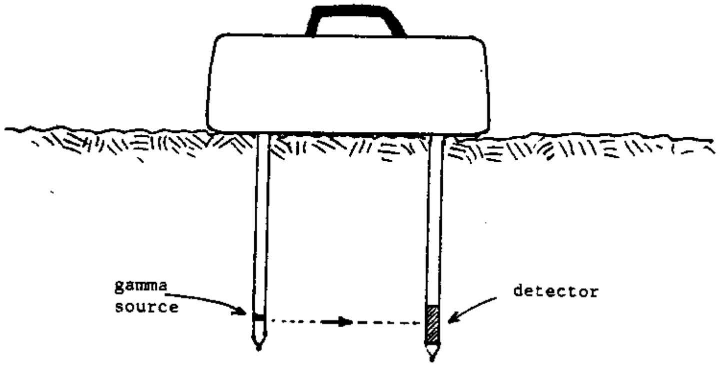

FIGURE 6

In general they are calibrated empirically and equation 20 is not directly used.

3.3. Gamma-Radiation scattering.

If a gamma-ray detector is shielded from direct radiation of a gamma-source, it can detect scattered radiation. If the gamma energy is less than 1 MeV, the probabilities of Compton scattering and phoelectric absorbtion are very great. As the number of electrons per unit volume of soil (which is proportional to soil bulk density) is increased, the Compton scattering power of the medium increases proportionately. With each scattering process, however, the gamma ray loses some of its energy. Thus, although an increase in electron density of the medium increases the probability of multiple scattering of the gamma ray, the probability that the gamma ray will suffer photoelectric absorbtion, before it can reach the detector, is also increased. The combined effect of these two probabilities is that a smaller number of gamma rays will reach the detector as the surrounding material becomes more dense. Figure 7 shows the geometry and principle of gamma-density depth and surface probes.

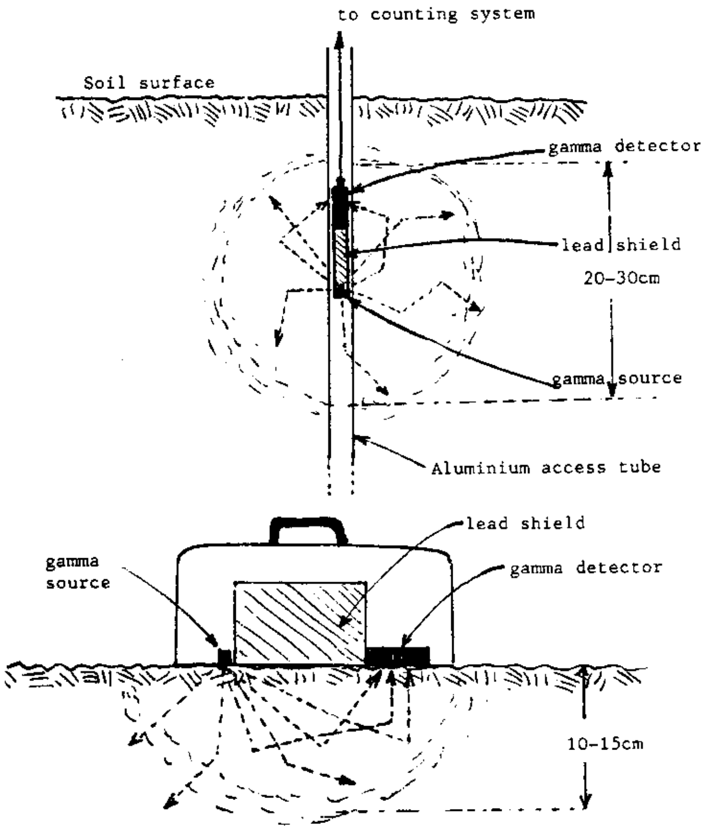

FIGURE 7

For the practical use of these instruments they are empirically calibrated in the laboratory or in the field.

3.4: Neutron moderation.

Soil water content can also be measured through the use of Neutron Moisture Meters, which operate using neutron moderation phenomena. If a source of fast neutrons (2 to 10 MeV) are introduced into the soil, neutrons which are radially emmitted will collide elastically with nuclei of the surroundings. The loss of energy per collision is maximal when the target nucleus is of about the same mass. These nuclei are mainly those of the Hydrogen in water. The number of collisions needed to slow down a fast neutron (2MeV) to become a thermal neutron (0.025MeV) can be seen below for a few target nuclei:

Isotope	No of collisions
^{1}H	18
^{2}H (Deuterium)	25
^{4}He	43
^{12}C	115
^{16}O	152
^{238}U	2172

The slowing down process by which fast neutrons become slow (or thermal is called neutron moderation or thermalization.

Simultaneously to the moderation phenomena, neutrons can be captured by nuclei, depending on neutron/nucleus affinity and on the energy of the neutron. This capture is a nuclear reaction and the product is a new stable or radioactive nucleus. The probability of capture is measured through "cross-sections" which are a function of target nucleus and neutron energy.

During the slowing down process, if a neutron is not captured, it will disintegrate with a half-life of about 12 minutes. Free neutrons are not stable particles.

Therefore, when a fast neutron source is introduced into the soil, due to the three phenomena described above, the number of slow neutrons surrounding the source will attain to an equilibrium after a few seconds. It was observed that this number of slow neutrons is directly proportional to the concentration of H (the most effective nucleus for moderation) of the surrounding media. These slow neutrons diffuse around the source building up a "neutron cloud" of spherical shape and of a diameter of 10-30cm, depending on H concentration.

Since the H concentration of a soil is the combination of the hydrogen from soil matrix (which is constant) and the hydrogen from soil water (variable) it whould be possible to measure soil water contents θ by measuring the "intensity" of the slow neutron cloud which surrounds the source.

There are detectors which are specific for slow neutrons, even in the presence of fast neutrons. These are Brown triflouride, Helium-3 and scintillation detections. Figure 8 is a schematic view of a neutron probe and detection systems. To introduce the probe into the soil an aluminium access tube is used, which is transparent to neutrons.

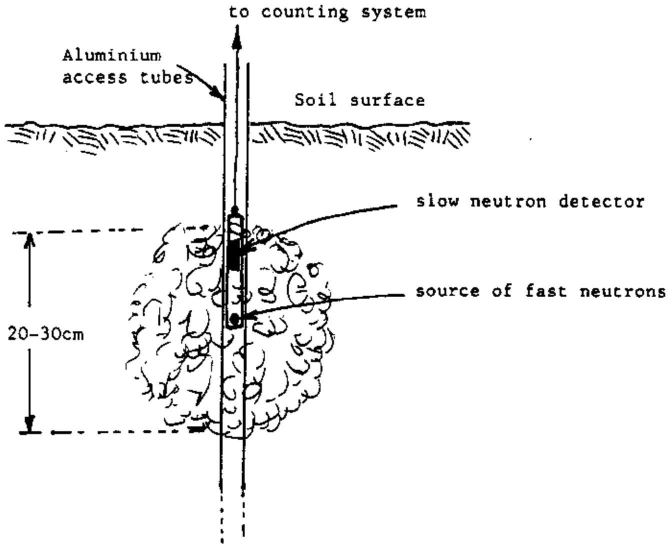

FIGURE 8

These instruments are empirically calibrated by finding a relation between the number of slow neutrons detected per unit of time (c.p.m.) and soil water content θ. Calibration curves are in general straight lines, but are a function of soil type and its bulk density. An example is given in Figure 9.

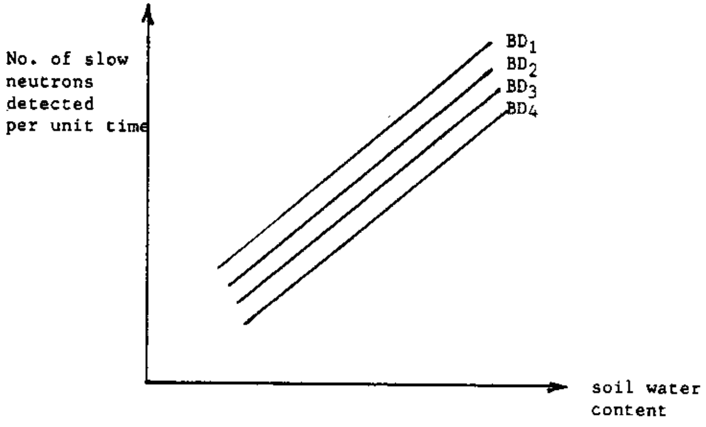

FIGURE 9

The measurement of θ through a neutron probe, is an average value for a volume of soil corresponding to a sphere of diameter 20 to 30 cm. The wetter the soil the smaller the diameter. There are practical procedures to measure this diameter. Due to this problem, special care has to be taken when measuring close to the soil surface. The cloud of slow neutrons should always be inside the soil. There are also practical procedures to minimize these errors.

There are also surface neutron moisture meters, which operate according to the same principles, and do not need access tubes. They measure soil water contents of a semi-sphere f radius 10-20cm. Figure 10 shows the diagram of such equipment.

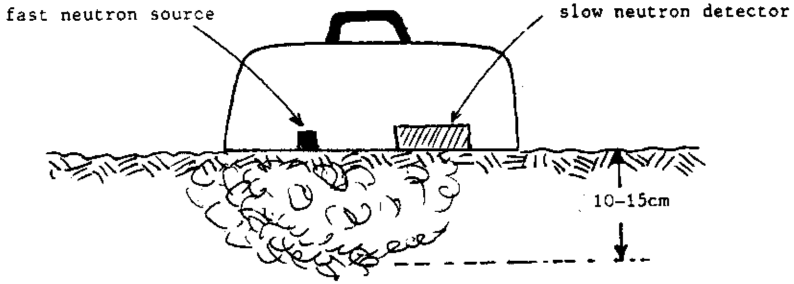

FIGURE 10

More details about the calibration, installation of access tubes, measurement errors and use of neutron moisture meters are found in IAEA (1976), Couchat(1983) and Vauclin et al (1983).

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