SOIL WATER DISTRIBUTION

A STATE OF THE ART REPORT

National Committees for the International Hydrological Decade in Denmark, Finland, Iceland, Norway, and Sweden
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Editor: E. Danfors

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National Committees for the International Hydrological Decade in Denmark, Finland, Iceland, Norway and Sweden

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PREFACE

The soil water zone is a key zone for the circulation of water in nature. The processes of this zone determine to what degree incoming precipitation will return to the atmosphere as transpiration, move horizontally as surface runoff, or percolate down through the soil profile to recharge the groundwater reservoir. Much interest has therefore been devoted to the unsaturated zone during the International Hydrological Decade 1965-74 (IHD). Such has also been the case within the Nordic region both on the regional and national level.

A Nordic working group was established by the IHD committees of the five Nordic countries Denmark, Finland, Iceland, Norway and Sweden already in 1966. The task of this working group was to analyze the most convenient techniques for soil water measurements as well as the possibilities to organize a Nordic soil water network. In many of the thirteen representative basins in these countries, soil water has been measured on a monthly basis. The book should be seen as the main result of the Nordic IHD cooperation within the field of soil water. It intends to give guidance as to both measurement of soil water in the point and to the extrapolation of point values to areal values of soil water storage.

The idea that a state of the art report on soil water should be elaborated was born in this working group already early during the decade. The present book has been produced within this group, and most of the authors are members of the group. Editor has been Erik Danfors, chairman of the working group. Assistance in the editorial work has been given by Mr Lennart Åberg of the Swedish IHD secretariat.

Stockholm, March 1975

Malin Falkenmark
Executive Secretary
Swedish National IHD Committee
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1. INTRODUCTION

1.1. Soil water in the water cycle

The soil water zone, located in the uppermost layer of the earth's crust, constitutes in fact a key zone in which incoming precipitation water is directed and partitioned in various ways. Accordingly, the nature of this zone will determine the extent of storage, runoff, downward percolation to the ground water and the degree of upward return to the atmosphere through evapotranspiration. Storage within the soil water zone is important for plant growth, percolation is important for groundwater replenishment and the downward transport of dissolved and suspended materials in the soil, whilst runoff produces useful surface water.

The effective input of rainwater into the soil water zone consists of the precipitation water subtracted by the amount that leads to immediate surface runoff. This net amount has been referred to as soil moistening by Lvovich and is the water available for evaporation and ground water recharge. Lvovich 1973, in his attempts to map the large scale variations of the main water balance elements, presents a map of the global distribution of the soil moistening, see figure 1:1. According to these estimations the annual soil moistening in Fennoscandia is of the order of 350 mm, whereas it exceeds 500 mm in Middle Europe. In most parts of India it exceeds 800 mm except in the NW deeper area. In the semi-arid Sahelian zone of Africa the average annual soil moistening exceeds about 400 mm.

The output processes of the soil system are still rather unclear. As to the runoff some scientists differentiate between a lateral and a vertical output where the former leads to groundwater flow. Different terms such as throughflow and interflow are often used to characterize the lateral component of the runoff process. The significance of the soil water zone on runoff is most clearly seen from the poor correlation which generally exists between the rainfall intensity and the output responses in the form of stream discharge and ground water table movement. This can apply even for very closely related hydrological
Figure 1: Total soil moistening (annual renewal) according to Lvovich 1973.
systems. The great variability in the characteristics of the soil cover is the main cause for this poor interrelationship between rainfall and runoff, see figure 1:2.

As to output from the soil water zone in the form of evapotranspiration, meteorological factors constitute the driving force whilst soil and vegetation characteristics modify the actual rate. The dependence of actual evapotranspiration on soil water deficit is shown in figure 1:3.

1.2. Soil water in the geochemical cycle
The close connections between hydrological processes and hydrochemical characteristics of the passing water, makes the study of the soil water zone important also from water quality aspects. In this context efforts are being made to discriminate the runoff according to its different origins or more correctly according to its main pathways through the soil medium. Different types of runoff are thereby being partitioned and related to different hydrochemical characteristics. Obviously such endeavours constitute important steps on the way to the development of water quality models that may forecast chemical changes in water due to human activities such as drainage, cultivation, fertilization, etc.

The bound character of soil water introduces some conceptual difficulties due to the gradational mobility that this property inflicts on the water body. Thus a large part of the soil water is so tensely bound to the soil matrix that it becomes almost immovable, whereas other fractions are rather free and can pass through the soil strata fairly rapidly. The transit times for various fractions are therefore widely varying according to how tightly held they are. These conditions have a major influence on the hydrochemical composition of the soil water, since the degree to which uptake of new materials from the surrounding soil occurs is dependent on the time the water spends in this surrounding. Water of different transit times therefore has different dissolving conditions.
Figure 1:2 Rainfall and runoff data from Verkaän, Sweden 1967. Note the poor correspondence between runoff and rainfall changes.

Figure 1:3 Reduction in evapotranspiration due to increased soil water deficit. From Aalyng 1966.
1.3 Soil water and human society

Action in order to counteract negative effects of soil water deficit within agriculture was one of the earliest human experiences, irrigation being the basis for a number of early civilizations. The ability to regulate soil water is well developed today. Excess of water is thus removed by drainage manipulations, deficit of water is met with by irrigation. In fact the extent of human modifications on soil water is today so large that it is not often possible to study a really natural water cycle. Instead, most hydrologically studied areas represent hybrids between natural and modified conditions.

Water balance can be affected and regulated not only by changing the capacity of surface reservoirs, but also through modification of the soil itself and the land use conditions. Thus ploughing, for example, has significant effects on the surface depression storage, infiltration capacity and permeability of the soil. The effects of this are especially evident in Norden in the spring time during thawing, when runoff is significantly reduced through increased uptake of melt-water in the surface depressions and in the increased pore volume of the ploughed soil.

Trafficability across various land types is also closely dependent on soil moisture conditions. The relationship between the shear strength of a mineral soil and its water content is shown in figure 1:4. Artificial drainage through tiling, ditch systems, etc facilitate runoff and reduce periods of poor load bearing capacity.

The soil water zone is also involved in the pollution problem. Here such questions as to what degree fertilizers introduced into the soil water zone will be transmitted to the groundwater zone and later emitted to the river system that dewater the area are becoming of prominent concern. This problem area is relevant to both agriculture and forestry. The soil is furthermore becoming a disposal system for a variety of wastes, where the ability of the upper layer to render the waste harmless, is utilized so that contamination of valuable water resources is avoided. The crucial status of the soil body within the continental water circulation cycle implies that the soil water zone is
involved also in other pollution problems: industrial acidification of precipitation and the effects which this might have on the productivity of the soil and on the quality of the water leaving a land area as run-off; the effect of salting roads during winter to reduce iciness and thereby slipperiness; the effects of fallout from chimneys, etc. Also in the direct effects of urbanization on infiltration, soil properties become important. The ground is used as a foundation for buildings. Therefore much damage can result if the bearing capacity of the ground is reduced due to changes in the water distribution. Such changes frequently result as a consequence of pavements etc being laid on the surface and to excessive loss of soil water due to facilitated drainage conditions.
The properties of the soil and the characteristics of the vegetation thus affect the hydrological cycle.

The influence of the soil type is manifested by lack of plant available water in some periods of the growing season. This causes a reduction of cessation of the plant's transpiration ability, and may be due to low water holding capacity of the soil and/or shallow root depth. The influence of the vegetation is due to differences in the life cycles of the different species. For permanently vegetated areas (forests, etc) the evaporative demand may be less for deciduous than for non-deciduous plants (Wind, 1958). For arable soils the type of crops cultivated, and the way they are harvested influences the evapotranspiration.

![Figure 2:3. Monthly precipitation (P), potential evapotranspiration (EP) according to Penman and actual evapotranspiration measured from barley (B), long grass (G1) short grass (G2) and sugar beets (S) at Højbakkegaard, Denmark. Averaged for the years 1969-1972.](image-url)
of the Arctic Circle there is then 24 hours of daylight. Norden may be divided into five plant geographical regions, namely from north to south: Arctic-subarctic, Alpine-subalpine, Boreal coniferous, North European mixed forest and North European deciduous forest regions, see Appendix 2.

Soil water hydrology has in Norden, as in other parts of the world, represented a rather neglected area of hydrological research. It has been studied primarily within agricultural and forestry research with special reference to soil biological aspects, agricultural techniques and water-regulative activities. Most studies have been of rather local character.

5.5 Some soil water activities during the IHD

During the lapse of the IHD the interest focused on soil water hydrology has successively increased. Knowledge of the water availability in the soil water zone necessitates measurements of the distribution of the water content and water retention properties in time and space. To reach this knowledge there is a need for techniques for evaluating both the flow of water passing a given point and the water content at this same point. Within IHD, most of the interest has been paid to the problems of measuring water content. The nuclear measuring technique for in situ measurements based on the interaction between fast neutrons and hydrogen atoms has been developed to the degree of routine standardization. Within the International Atomic Energy Agency a manual has been prepared.

Besides the problem of point measurements of soil water there is the problem of choosing representative sites for the measurements, and the problem of determining areal mean values from the point data. Kutilek 1970 has proposed a method of extrapolation of point values based on an ecological classification, taking into consideration the duration of different water contents in relation to field capacity and wilting point.

In the Nordic region, the IHD study of soil water has been approached both in a number of national projects within the different countries and
by the work of a joint Nordic working group. In Denmark soil water has been studied as a factor in plant production with special reference to assessing deficits and irrigation requirements while in Finland surplus effects and drainage needs have been a major field of interest. Systematic information has in Sweden been gathered on the water retention properties of agricultural soils and on the dependence of retention on clay content, see figure 6:5 from a recent study by Andersson & Wiklert, 1972. The Nordic working group has concentrated its efforts on measurement techniques and on means of estimating storage changes on a monthly basis as an element of water balance of representative basins. The experience of large statistical variations of point water content values, has lately led to efforts to develop a systematic biogeological zoning procedure, based on the expected coupling between soil water and landscape factors.

The present book presents the results of the Nordic studies on measurement problems. It discusses the reasons for measuring soil water within hydrology and how such measurements might be performed. It starts with a description of the soil water zone, how it can be described and the terminology used. The publication should be seen as the main result of the Nordic IHDCooperation within the soil water field, the scope of this Nordic cooperation being primarily to develop guidance as to the assessment of soil water in catchment studies.
2. CHARACTERIZATION OF WATER IN THE SOIL

2.1. Defining the nature and extent of soil water

The term "soil" may have different meanings, depending on the connection in which it is used. For agricultural and hydrological purposes, it is generally defined as being the loose material that covers the bedrock (Vemb 1971). The soil may be of mineral or organic origin, and these materials may be present in mixtures and/or in different layers in the soil profile.

The soil is a very complex system. A given volume of soil is made up of solid material of varying size and shape, with intervening pores (the soil matrix). The pores may be more or less occupied by water. The part of the pores not occupied by water contains the soil air. At and below the ground water level, all the soil pores are water filled.

The water in the soil is not pure water, but contains different kinds of suspended and dissolved materials in varying concentrations, depending on application of fertilizer and weathering of the primary soil particles. In the upper soil layers also living organisms may be present.

Soil water is the part of the hydrosphere where water is held in the soil, either by adhesive forces existing between water and soil material or by capillary forces caused by the soil pores and the surface tension of the water. Soil water is generally in a state of tension, i.e., the hydrostatic pressure in a body of soil water is less than the atmospheric pressure.

In the ground water, on the contrary, the hydrostatic pressure is greater than the atmospheric pressure. The level where the pressure in the water embodied in the soil equals the atmospheric pressure is then by definition the ground water level (or ground water table).

According to the definition, all the water held in the soil between the soil surface and the ground water level is soil water. This water is
also referred to as vadose water (Todd 1966). Since the distance between the soil surface and the ground water table may be considerable in many locations, and thus not all the soil water is influenced by the plant root activity, a subdivision or zoning of the soil water regime may be desirable.

2.2. Zoning of soil water regime

The soil water regime may be usefully differentiated into 5 main zones as shown in Figure 2. Starting from the soil surface, and moving vertically toward the ground water level, the zones may be characterized as follows:

The *evaporation zone* is the layer of soil material between the soil surface and what may be termed the thermocline level. Most soils will exhibit a thermocline level (discontinuity or Sprungschicht layer) which is characterized by a dramatic change in the gradient of temperature versus soil depth (Björ 1972). The evaporation zone which lies above the thermocline is a transition or barrier zone between the atmosphere and the soil proper. It is generally rather thin, and may at times be difficult to define, for example under a dense vegetation and/or under very humid conditions. The zone is subject to occasional drying in excess of the ability of the plants, i.e., by direct evaporation to the atmosphere. The evaporation zone may consist of more or less loosened mineral or organic soil, plant residues, different kinds of mulching material, or mixtures of these materials.

The *transpiration zone* is the layer of soil between the thermocline level and the maximum root depth level. The *maximum root depth* is the maximum depth from which plants extract water. The drying out in this zone is mainly caused by the process of plant transpiration.

Together the evaporation zone and the transpiration zone constitute the *root zone*.

The *intermediate zone* is the layer of soil between the maximum root depth level and the capillary level. The *capillary level* is the maximum height above the water table to which water can rise by capillary
Figure 2:1 Profile zoning of soil water regime
action. The water content in this zone only varies due to water vapour diffusion to the root zone in dry periods or due to infiltration in periods with surplus precipitation.

The capillary zone is the layer of soil between the capillary level and the saturation level. The saturation level is the height above the ground water level to which all pores are filled by capillary action.

The four zones mentioned together make up the saturation zone, which is characterized by having a fraction of air-filled pores.

The capillary fringe zone is the layer of soil between the saturation level and the ground water level (Childs 1957). The ground water level is the level in the soil in which the hydrostatic pressure equals the atmospheric pressure. This level can be measured with tensiometers or piezometers or by measuring the water surface level in a suitable well. The ground water level here defined thus corresponds to the phreatic ground water surface.

The capillary fringe zone together with the four previously mentioned zones makes up the soil water zone. In this zone the water is normally subject to a hydrostatic pressure less than the atmospheric pressure (in a state of tension). It is only during periods with infiltration that the hydrostatic pressure may temporarily exceed atmospheric pressure, thereby giving rise to ephemeral or ponded ground water.

In many textbooks soil water (or soil moisture) is used to imply water in the root zone only. We find this limitation less practical and suggest soil water to be used for all water above the ground water level.

The capillary zone and the capillary fringe zone will extend to a distance above the ground water level determined by the pore size and pore size distribution in the soil. In coarse soils with relatively large pores, the capillary fringe zone may be very limited in extent or even vanish as may also the capillary zone. In fine-textured soils the two zones may reach to a considerable height above the ground water level. The two zones will fluctuate in height according to the variation of the ground water level. When the ground water level comes within
one meter or less of the soil surface, the intermediate zone will disappear and the capillary zone will merge directly into the transpiration zone. Under conditions of varying ground water level, both the water content and extent of capillary rise will depend on the vertical variation of the pore geometry of the soil profile.

The water content through an imaginary soil profile is shown schematically in figure 2:2. The water occupies all the pores up to the capillary fringe level. Through the capillary zone the water content declines to a value corresponding to field capacity (FC), at which it theoretically remains through the intermediate zone. The water content curve through the root zone depicts a situation that might occur after a long dry period. At the thermocline level the soil water content passes below the wilting point (WP).

![Figure 2:2: Schematic representation of the distribution of soil water content with depth through the unsaturated zone. The curve applies to a homogeneous soil profile that has been dried from a more or less saturated state without complications due to infiltration or an undulating groundwater level.](image-url)
In conclusion it should be pointed out that the profile zones defined above are generally difficult to delineate precisely under field conditions. The reasons for this are partly due to the effects of continuously changing force fields that seldom allow soil water distribution to reach states of equilibrium and partly due to the heterogeneous structure of the soil medium that cause non-reversible hysteresis effects.

The zoning that has been presented is important, however, as a model for interpreting and understanding the spatial variability of soil water.

2.3. Soil Water as an Element in the Hydrological Cycle

A part of the world's immense stock of water is at any time participating in a circulation known as the hydrological cycle. Fed by the sun's energy, water is transferred from the liquid or solid phase, to the vapour phase, and brought into the atmosphere. In the atmosphere it condenses and returns to the earth's surface as precipitation (rain, snow, hail, fog, dew).

The precipitation that reaches the earth may be intercepted by vegetation and via the vegetation reach the soil surface, or it will reach the soil surface directly, where no vegetation is present. Arriving at the soil surface, the liquid precipitation may stream away on the surface or infiltrate into the soil or both.

The infiltrated part of the water restores the soil water reservoir in the root zone, if this zone is below the field capacity. When the root zone water is replenished, the infiltrated water will percolate to the ground water, causing this to rise, and eventually promote run-off. If the percolation rate is lower than the infiltration rate, a ponded ground water may be established near the soil surface, and cause a sub-surface or increased surface run-off.

The hydrological cycle is influenced by the soil by virtue of the buffering effect of the soil. This buffering effect is rather small for non-vegetated soil (essentially limited to the evaporation zone), while it can be considerable for vegetated soil, where water will be taken from the transpiration zone, and via the plants carried to the atmosphere.
Figure 2:3 shows the monthly values of actual and potential evapotranspiration, together with the precipitation. The crops used are the most common agricultural crops. The potential evapotranspiration is calculated according to Penman's equation (Penman 1948; Aslyng 1965). The actual evapotranspiration is calculated as precipitation plus decrease in the water content in the root zone. This water content was measured by the neutron scattering method. The measurements were made at Højbakkegaard, 20 km west of Copenhagen.

The actual plant water use during the growing season (April-September incl.) is given in Table 2:1. The evaporation in the winter period (October-March, incl.) amounts to 70-90 mm at Højbakkegaard.

Table 2:1. Potential and actual evapotranspiration measured from different crops during the growing season, (April-Oct, incl.). Højbakkegaard, Denmark.

<table>
<thead>
<tr>
<th>Year</th>
<th>Potential (Penman)</th>
<th>Barley</th>
<th>Sugar beets</th>
<th>Grass (hay)</th>
<th>Grass (grazing)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1969</td>
<td>491</td>
<td>332</td>
<td>362</td>
<td>345</td>
<td>306</td>
</tr>
<tr>
<td>1970</td>
<td>471</td>
<td>353</td>
<td>362</td>
<td>370</td>
<td>351</td>
</tr>
<tr>
<td>1971</td>
<td>519</td>
<td>339</td>
<td>353</td>
<td>323</td>
<td>-</td>
</tr>
<tr>
<td>1972</td>
<td>477</td>
<td>358</td>
<td>369</td>
<td>399</td>
<td>356</td>
</tr>
<tr>
<td>Average</td>
<td>489</td>
<td>345</td>
<td>362</td>
<td>360</td>
<td>338</td>
</tr>
</tbody>
</table>

The results are for crops grown on a soil with a relatively high amount of water available for the plants (about 20 per cent by volume) and great root depth (about 150 cm). Therefore the actual evapotranspiration is mainly determined by the need of the crops (active leaf area index) and only to a minor extent by lack of available water.

For crops grown on soils with insufficient available water in the root zone (low water holding capacity and/or shallow root depth) the actual evapotranspiration would tend to be less and the differences between the crops tend to be greater than reported above, depending on the length and time of dry periods in the growing season.
The properties of the soil and the characteristics of the vegetation thus affect the hydrological cycle.

The influence of the soil type is manifested by lack of plant available water in some periods of the growing season. This causes a reduction of cessation of the plant's transpiration ability, and may be due to low water holding capacity of the soil and/or shallow root depth. The influence of the vegetation is due to differences in the life cycles of the different species. For permanently vegetated areas (forests, etc) the evaporative demand may be less for deciduous than for non-deciduous plants (Wind, 1958). For arable soils the type of crops cultivated, and the way they are harvested influences the evapotranspiration.

Figure 2:3. Monthly precipitation (P), potential evapotranspiration (E_p) according to Penman and actual evapotranspiration measured from barley (B), long grass (G_1) short grass (G_o) and sugar beets (S) at Højbakkegaard, Denmark. Averaged for the years 1969-1972.
3. FACTORS AFFECTING SOIL WATERY

3.1. Physical properties of the soil

The amount of water which can be retained in a given volume of soil depends on the size distribution of the primary particles of the soil (the soil texture), and the packing together of these particles (the soil structure). The soil texture is a primary characteristic of the soil. It determines to which degree the soil can develop a structure, and the degree to which this soil structure can be created or changed by external actions of physical or chemical nature. Soil texture and soil structure are decisive for most of the other soil characteristics.

3.1.1. Soil texture

The soil texture is an expression of the proportions of the various size-groups of mineral particles (the particle-size distribution) and of organic matter (humus) that constitutes the soil.

The analysis of particle size distribution is normally based on a combination of sedimentation and sieving techniques. The soil must be pretreated physically and chemically in order to ensure a single-grain condition. The finer fractions are determined by sedimentation according to Stokes' law, which predicts the relation between particle size and sedimentation velocity.

The sedimentation analysis may be carried out by extracting a known amount of the suspension at a specified depth at specified time intervals. The aliquots are dried out and the dry weight less the calculated weight of dispersing agent gives the mass of particles in the samples taken. Measurements can also be made by determining the density of the suspension at specified time intervals. The most commonly used method is the one introduced by Bouyoucos (1951). This method uses a float with a long narrow neck. The neck extends above the surface of the suspension, and is graduated to read either the density of the suspension or the mass of the suspended material in grams per litre. Special care must be taken in interpreting the readings, because the
suspension density, and thereby the depth to which the float sinks, depends on the amount of fine particles in the specimen analysed and the sedimentation time. Methods, precautions and treatments in the use of the method are described more in detail in the literature e.g. Keen 1931, Aslyng 1952, Hansen 1961, Day 1965, De Boodt 1967.

The coarser fractions are separated by passing the suspension through sieves with specified openings, followed by drying and weighing of the material retained on each sieve.

The organic fraction (Humus) can be determined by combustion with determination of the loss in weight, or by measuring the amount of carbon dioxide produced. Loss in weight may be caused by evaporation of crystal water (clay soils) and by carbon dioxide liberation from possible lime in the soil. The latter will also cause faults in the carbon dioxide liberation method mentioned. Lime must be removed from the soil, e.g. by treatment with acids. The weight-loss method can only be used on soils low in clay content. Methods for determination of organic matter in the soil are described by e.g. Aslyng 1952, and Aslyng 1956.

The soil texture is characterized in the Nordic region by means of various systems of particle size distribution classifications based on the content of the particle size fractions that are defined by the International Soil Science Society, see Table 3:1. The textural composition is expressed as the percentage by weight of each of the fractions. A soil containing 3 % humus, 15 % clay, 20 % silt, 30 % fine sand and 22 % coarse sand may be characterized as follows:

$$3Hu, 15Cl, 20Si, 30Sa, 22cSa$$

For more details regarding actual soil textural classifications in use in the Nordic Region see Appendix I. In this chapter the general principles involved will be presented.
Table 3:1. International soil particle classification.

<table>
<thead>
<tr>
<th>Name of separate</th>
<th>Abbreviation</th>
<th>Diameter (range)</th>
<th>Unit</th>
</tr>
</thead>
<tbody>
<tr>
<td>Humus</td>
<td>Hu</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Clay</td>
<td>Cl</td>
<td>0 - 2</td>
<td>μm</td>
</tr>
<tr>
<td>Silt</td>
<td>Si</td>
<td>2 - 20</td>
<td>-</td>
</tr>
<tr>
<td>Fine sand</td>
<td>fSa</td>
<td>20 - 200</td>
<td>-</td>
</tr>
<tr>
<td>Coarse sand</td>
<td>cSa</td>
<td>200 - 2000</td>
<td>-</td>
</tr>
<tr>
<td>Gravel</td>
<td>Gr</td>
<td>2 - 20</td>
<td>mm</td>
</tr>
<tr>
<td>Stone</td>
<td>St</td>
<td>20 - 200</td>
<td>-</td>
</tr>
<tr>
<td>Boulders</td>
<td>Bo</td>
<td>above 200</td>
<td>-</td>
</tr>
</tbody>
</table>

Table 3:2. Textural classification of soils of known particle size distribution.

<table>
<thead>
<tr>
<th>Textural type</th>
<th>Per cent by weight of clay</th>
<th>Nordic usage</th>
<th>English literature</th>
</tr>
</thead>
<tbody>
<tr>
<td>Clay free soils</td>
<td>less than 2</td>
<td>Gravels</td>
<td>Gravels</td>
</tr>
<tr>
<td>Slightly clayey soils</td>
<td>2 - 5</td>
<td>Slightly clayey sands</td>
<td>Loamy sands and silts</td>
</tr>
<tr>
<td>Clayey soils</td>
<td>5 - 15</td>
<td>Clayey sand</td>
<td>Sandy loams</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Clayey silt</td>
<td>Loams</td>
</tr>
<tr>
<td>Light clays</td>
<td>15 - 25</td>
<td>Sandy light clay</td>
<td>Silt loams</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Silty light clay</td>
<td></td>
</tr>
<tr>
<td>Medium clays</td>
<td>25 - 40</td>
<td>Medium clay</td>
<td>Clay loams</td>
</tr>
<tr>
<td>Heavy clays</td>
<td>40 - 60</td>
<td>Heavy clay</td>
<td>Clays</td>
</tr>
<tr>
<td>Extremely heavy clays</td>
<td>more than 60</td>
<td>Extremely heavy clay</td>
<td>Heavy clays</td>
</tr>
</tbody>
</table>
The textural soil types mentioned in Table 3:2 can roughly be estimated on the basis of consistence in the dry state (powderiness) and in the moist state (plasticity).

Soils with a clay content less than about 5% are characterized by the dominating sand fraction. A subdivision according to the dominating sand fraction therefore is desirable by adding to the name fine or coarse.

Clay free soils: Wet material will not form a thread when rolled. It does not stick to the fingers when worked. Separation between coarser and finer types of clay free soils is done on the basis of the appearance of the single grains.

Slightly clayey soils: Wet material will only with difficulty form a thread when rolled. Sticks slightly to the fingers when worked. Division between coarser and finer types of slightly clayey soil is done on the basis of the appearance of the single grains.

For soils with more than 5% clay, the method described by Vemb 1971 is suggested. Water is added to the soil until the soil is just plastic. A sample is gently rolled with the fingers, to form a thread. When the thread starts to show breaks, the thickness of the thread is measured. The classification then is made according to Table 3:3.

Table 3:3. Thickness of thread made from different soil types by rolling wet soil.

<table>
<thead>
<tr>
<th>Soil type</th>
<th>Approximate thread diameter, mm</th>
</tr>
</thead>
<tbody>
<tr>
<td>Clayey soils</td>
<td>3</td>
</tr>
<tr>
<td>Light clays</td>
<td>2</td>
</tr>
<tr>
<td>Medium clays</td>
<td>1 - 1.5</td>
</tr>
<tr>
<td>Heavy clays</td>
<td>1</td>
</tr>
<tr>
<td>Extremely heavy clays</td>
<td>less than 1</td>
</tr>
</tbody>
</table>

The method is delicate and requires some skill. Its advantage is its quickness and that it can be used directly in the field.
If the soil contains organic matter, a phrase characterising the estimated humus content should be added to the soil name, e.g. poor or rich in humus, if the estimated content is below or above 4 per cent by weight.

Soils having organic matter contents above 30 per cent by weight are characterized as organic soils. Depending on the degree of decomposition the names peat, mucky peat, peaty muck and muck are used.

If the content of material coarser than 2 mm is very great (above 20 per cent) the soil is characterized as gravelly, stony or blocky, depending on the size-group that dominates.

In order to characterize a given soil, samples may be required from several depths, depending on the layering and homogeneity of the soil.

3.1.2. Soil structure

The soil structure refers to the aggregation of primary soil particles which are separated from adjoining aggregates by surfaces of weakness (U. S. D. A. 1954). An aggregate is a group of two or more primary particles which cohere to each other more strongly than to surrounding particles (Kemper & Chepil 1965).

Soil structure and soil texture are together decisive for physical properties of the soil such as density, porosity, capillarity, hydraulic conductivity and water retaining capacity.

The soil structure may be described by the form, character, size and size-distribution of the aggregates. The aggregates in the upper part of the soil will change in time due to climatic and possible soil tillage impacts. The structure below will remain more constant, although it may be subject to change by chemical and biological effects (fertilization, root activity, activity of soil animals) and by physical effects (deep tillage, frost action, particle migration).

In discussing soil structure both the size-distribution of the aggregates and the mechanical stability of the aggregates are important. The size distribution can be found for dry soil by gently sieving the natural soil.
through a system of different sieve-sizes, which are operated in a fixed way, or by wet treatment either by sieving under water, sedimentation or elutriation (Kemper & Chepil 1965). The stability of the aggregates may in principle be found by varying the intensity of the treatments used in the size-distribution analysis (Kemper 1965). Methods for aggregate distribution and aggregate stability analyses are given by Kemper & Chepil 1965, Kemper 1965, De Boodt 1966 and Davies 1969. The methods are delicate since the results are influenced by the history of the aggregates and by the exactness by which the different treatments can be repeated.

For the more stable type of soil structure, i.e. the structure which occurs below the top layer of regular tillage treatment, the structure may be described according to appearance. Such a description should be based on details which are easy to observe. Classification into four main types, as suggested by Nikiforoff 1941, is recommended.

I. **Platy structure**: Aggregates with horizontal axes longer than the vertical.

II. **Prismatic structure**: Aggregates with horizontal axes shorter than the vertical.

III. **Blocky structure**: Aggregates with horizontal and vertical axes more or less equal.

IV. **Granular structure**: Aggregates that are more or less rounded.

Each of the types of soil structure should be classified according to size. The classification suggested by USDA Soil Survey Manual is suggested, see Figure 3:1.

The degree of development may be differentiated as follows:

*Structureless*. No aggregation can be observed. When particles are coherent the structure is designated as massive and when non-coherent as single grain.
### Types and Classes of Soil Structure

<table>
<thead>
<tr>
<th>Class</th>
<th>Thickness of plate</th>
<th>Horizontal axis</th>
<th>Axes</th>
<th>Diameter</th>
</tr>
</thead>
<tbody>
<tr>
<td>Very fine</td>
<td>Less than 1 mm</td>
<td>Less than 10 mm</td>
<td>Less than 5 mm</td>
<td>Less than 1 mm</td>
</tr>
<tr>
<td>Fine</td>
<td>1 to 2 mm</td>
<td>10 to 20 mm</td>
<td>5 to 10 mm</td>
<td>1 to 2 mm</td>
</tr>
<tr>
<td>Medium</td>
<td>2 to 5 mm</td>
<td>20 to 50 mm</td>
<td>10 to 20 mm</td>
<td>2 to 5 mm</td>
</tr>
<tr>
<td>Coarse</td>
<td>5 to 10 mm</td>
<td>50 to 100 mm</td>
<td>20 to 50 mm</td>
<td>5 to 10 mm</td>
</tr>
<tr>
<td>Very coarse</td>
<td>More than 10 mm</td>
<td>More than 100 mm</td>
<td>More than 50 mm</td>
<td>More than 10 mm</td>
</tr>
</tbody>
</table>

Figure 3:1 Types and classes of soil structure
Weak. Aggregates are poorly formed and only barely observable. When disturbed, the soil breaks easily into a mixture of unaggregated material and a small amount of entire aggregates.

Moderate. Aggregates are well-formed and distinct. They are moderately durable when soil is disturbed.

Strong. Distinct aggregation. When soil is disturbed it breaks down mainly into entire aggregates and forms very little or no unaggregated material.

Soil bulk density ($\rho_b$) is ratio of the dry mass to the volume of soil particles plus pores. The volume is determined either by use of a sampler with known volume (core method) or by excavating a sample and determining the volume of the excavation. Especially core sampling requires a rather high water content in the soil, and it is difficult to practice in stony soils. The soil bulk density can also be obtained indirectly by radiometric methods (transmission or scattering of gamma radiation). This method requires, however, simultaneous measurement of the soil water content. The methods are described by Blake 1965 and Aslyng 1968.

The total porosity ($P$) of the soil can be calculated from the soil bulk density and the particle density ($\rho_p$). The latter refers to the density of the soil particles collectively, and expresses the ratio of the total mass of the solid particles to their net volume, excluding pore spaces between particles. The particle density is measured by means of a pycnometer method as described by e.g. Blake 1965 and Aslyng 1968. The porosity is given by:

$$P = \frac{\rho_p - \rho_b}{\rho_p}$$

$P$ is the volume fraction of pores in the natural soil. The value of $\rho_p$ is for normal mineral soil 2.6 to 2.7 (g cm$^{-3}$), and for peat soils 1.4 to 1.6 (g cm$^{-3}$).

More information about the soil structure can be achieved by the determination of the pore-size distribution.
Current use of pore-size distribution as a soil characteristic depends on acceptance of the capillary model as representing soil pore space (Vomicil 1965). The pore radius can be calculated from the equation of capillary rise:

\[ r = \frac{2a \cos \theta}{pgh} \]

- \( r \) = pore radius
- \( h \) = height of capillary rise, or equivalent suction required to empty a water-filled pore of radius \( r \)
- \( a \) = surface tension of the liquid
- \( g \) = acceleration due to gravity
- \( \theta \) = contact angle of liquid to pore wall
- \( p \) = density of the liquid (water)

The units should be kept consistent. In the cgs-system and for ordinary water at 20°C, the following constants apply:

- \( a = 73 \text{ dyn cm}^{-1} \)
- \( p = 1.0 \text{ g cm}^{-3} \)
- \( g = 982 \text{ cm sec}^{-2} \)
- \( \cos \theta = 1.0 \)

Introducing these constants, the above equation reduces to:

\[ r = \frac{0.15}{h} \]

If an initially water-filled soil is subject to drainage by stepwise increase of \( h \), either by suction or pressure in a suitable apparatus (see 3.2.2) the volume of water liberated by each step is identical to the volume of pores having effective radii between the ones that can be calculated according to the capillary equation. Methods are described more in detail, by e.g. Richards 1941, Richards 1949, Vomicil 1965, Aslyng 1968.

The soil structure can be characterized by the volume of pores in different size groups as suggested in Table 3:4. The pore radius is given in \( \mu \text{m} \) and \( h \) in cm of water column.

Normally the characterization is not carried out to more than \( 10 \log h = 4.2 \), which generally is considered to be the maximum degree to which plants can dry the soil. Characterization in excess of this value requires vapour-pressure equilibrium methods and is hardly meaningful in terms of pore dimensions. As \( 10 \log h \) approaches and exceeds 4.2 surface adsorption effects determine water retention.
Table 3:4. Characterization of the soil pores.

<table>
<thead>
<tr>
<th>( r, \mu m )</th>
<th>( h, \ cm )</th>
<th>( 10^{\log h, \ cm} )</th>
<th>Pore class</th>
</tr>
</thead>
<tbody>
<tr>
<td>Below 0.01</td>
<td>Above 160000</td>
<td>Above 5.2</td>
<td>Micro-pores</td>
</tr>
<tr>
<td>0.1 - 1.0</td>
<td>160000-16000</td>
<td>5.2 - 4.2</td>
<td>Very fine pores</td>
</tr>
<tr>
<td>1.0 - 10</td>
<td>1600 - 160</td>
<td>3.2 - 2.2</td>
<td>Fine pores</td>
</tr>
<tr>
<td>10 - 100</td>
<td>16 - 16</td>
<td>2.2 - 1.2</td>
<td>Medium pores</td>
</tr>
<tr>
<td>Above 100</td>
<td>Below 16</td>
<td>Below 1.2</td>
<td>Coarse pores</td>
</tr>
</tbody>
</table>

For peat soils a special classification system based on the relative occurrence of various peat constituents in peat samples has been developed (cf Heikurainen and Hulkari 1952, Heikurainen 1964). The peats are usually divided into three groups based on the dominant peat constituent as follows: Sphagnum peats (S peats), sedge peats (C peats) and woody peats (L peats).

In addition to the peat type, the stage of decomposition of the peat affects its structure. The classification of peats according to their stage of decomposition is usually carried out using the scale presented by von Post (1922). According to this method, the degree of humification of peat is determined by means of ocular study of peat samples. von Post’s scale includes ten classes of humification: H 1 refers to fresh undecomposed plant material and H 10 to fully decomposed peat.

As peat decomposes it also becomes denser; its bulk density increases. The bulk density of peat is determined from the oven-dry weight (105°C) of a core of undisturbed and saturated peat.
According to the literature concerning peat, the humification percentage (Pjavitshenko 1958) and the fiber content (Farnham and Finney 1965) of peat have also been frequently used to indicate the stage of decomposition.

3.2. Physical properties of soil water

3.2.1 Soil water content

The water content (or moisture content) of the soil can be expressed on the basis of mass ($W_m$):

$$W_m = \frac{\text{Weight of wet soil} - \text{Weight of dry soil}}{\text{Weight of dry soil}}$$

By dry soil is understood soil dried in an oven at 105°C, until its weight remains constant.

Knowing the bulk density of the soil ($\rho_b$) and of the soil water ($\rho_w$), the soil water content can be expressed on a volume basis ($W_v$):

$$W_v = W_m \frac{\rho_b}{\rho_w}$$

The value of $\rho_w$ can safely be assumed to be unity.

The water content ($W_m$ or $W_v$) may be given as fractions of 1 as above or (more commonly) in percent by multiplying by 100.

Finally the soil water content can be expressed in units of length (as for precipitation). Obviously the $W_v$ value expressed on a fraction basis is mm water per mm soil layer, and in percent as mm water per 10 cm soil layer.
The soil water is confined in the pores between the solid soil particles. The maximum water content of the soil therefore is $W_v = P$. The maximum water content normally only occurs below and immediately above the ground water level (the zone of saturation). Only during periods with rainfall or snow-melting in excess of the percolation ability of the soil, will the maximum water content occur in the other soil zones.

Total dryness will hardly occur under natural conditions, but may be approached in the evaporation zone of coarse textured soils during periods with intensive evaporation. Between these two extremes, the soil water is found to be more or less intensively bound to the soil matrix.

3.2.2. Soil water potential

The soil water is in nature subject to forces originating from the presence of the solid soil material, the dissolved matter in the soil water and the earth’s gravitational field. Forces are measured in the unit dyn (mass times acceleration). When speaking of soil water status, the term "potential" is often used. The soil water potential is the specific energy of unit mass or unit volume of water, i.e. potential = force times length per unit water = dyn cm g$^{-1}$ = erg g$^{-1}$, when unit mass is used.

Unit mass is convenient since the potential can be made to assume the dimension of length by division by the gravitational constant. It is then usually expressed as cm water column.

As the absolute potential cannot be defined, the potential has to be related to a specified reference niveau. When dealing with soil water, the ground water level is often a convenient level to use. (Soil surface or other defined level can also be chosen.

The total potential of the soil water ($\psi$) is made up of potentials caused by the soil matrix (matric potential, $\psi_m$), the solutes in the soil water (osmotic potential, $\psi_s$), the earth’s gravity (gravitational potential, $\psi_g$) and a possible potential caused by differences in pneumatic pressure (pneumatic potential, $\psi_p$), i.e.:

$$
\psi = \psi_m + \psi_s + \psi_g + \psi_p
$$
The soil water potentials are defined by I. S. S. S. (1963) as:

\( \psi = \) the amount of work that must be done per unit quantity of pure water in order to transport reversibly and isothermally an infinitesimal quantity of water from a pool of pure water at the reference level at atmospheric pressure to the soil water at the point under consideration.

\( \psi_m = \) The amount of work that must be done per unit quantity of pure water in order to transport reversibly and isothermally an infinitesimal quantity of water from a pool containing a solution identical in composition to the soil water at the elevation and the external gas pressure of the point under consideration to the soil water.

\( \psi_a = \) The amount of work that must be done per unit quantity of pure 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 to a pool containing a solution identical in composition with the soil water at the point under consideration, but in all other respects identical to the reference pool.

\( \psi_g = \) The amount of work that must be done per unit quantity of pure water in order to transport reversibly and isothermally an infinitesimal quantity of water from a pool containing a solution identical in composition to the soil water at a specified elevation at atmospheric pressure, to a similar pool at the elevation under consideration.

\( \psi_p = \) The amount of work that must be done per unit quantity of pure water in order to transport reversibly and isothermally an infinitesimal quantity of water from a pool containing a solution identical in composition to the soil water at atmospheric pressure at the point under consideration to a similar solution at a pneumatic pressure, identical to the one of the soil air.

\( \psi_p \) was not defined by I. S. S. S. (loc. cit.) The definition given here is taken from Jensen (1969). \( \psi_p \) is of no significance under natural (field) conditions, but in laboratory determinations of soil water retention known pneumatic potentials are utilized in the pressure membrane apparatus.
The matric potential is composed of potentials caused by adsorption of water molecules to the surface of the soil material, curved miniscus at the air-water interface and the interaction of ions and soil particle surfaces (Gardner 1968). These potentials cannot easily be measured individually and are therefore collectively designated as the matric potential.

Often it is practical to use special combinations of soil water potentials. The sum of matric and gravitational potentials is called hydraulic potential, and is often used in description of water transport processes in the soil. The osmotic potential generally is of minor importance in this process. The sum of matric and osmotic potentials is referred to as the water potential, and is of special interest in connection with the availability of water to plants.

The soil water potentials may be positive or negative, depending on the nature of the acting forces and the reference level chosen. Using the ground water level as reference, \( \phi_g \) is positive above and negative below, and \( \phi_m \) is negative above and positive below the groundwater table, except in situations with ponded water in the soil above the groundwater. \( \phi_g \), if present, is always negative. A sketch of the potentials is given in Figure 3:2.

![Figure 3:2. Schematic course of soil water potentials through the different soil zones (groundwater level as ref.). The matric potential (\( \phi_m \)) is in equilibrium with the gravitational one (\( \phi_g \)) (note the signs) through the groundwater and capillary zones. \( \phi_m \) remains constant (simplified through the intermediate zone and declines in the root zone, depending on the dryness of the soil). The total potential \( \phi \) with (broken line) and without accompanying osmotic potential (\( \phi_{mg} \)) are also shown.](image-url)
The water potential is often referred to as the binding force of the soil water \( F = h_m + h_s \) and is generally used on the \( 10 \log \) - basis \( pF \):

\[
pF = 10 \log (h_m + h_s)
\]

\( h_m \) and \( h_s \) is the height in cm of the water column \( \rho_w = 1.0 \) in equilibrium with the potentials \( \gamma_m \) and \( \gamma_s \) respectively (Schofield 1935). 

\( pF \) is often (incorrectly) used for \( h_m \) alone. For many soils and situations this can be justified, however, due to the small value of \( h_s \).

\( F \) (and \( pF \)) can be approximated by measuring freezing-point depression, the water vapour depression, or by the equilibration method (Richards 1965a, Rose 1966, Aslyng 1968, Jensen 1969). The relation of \( F \) to the water vapour depression is given by:

\[
F = \frac{RT}{Mg} \left( \ln \frac{e_o}{e} - \ln e \right)
\]

\( R \) = Gas constant \( (8.314 \times 10^7 \text{ erg} \ \text{deg}^{-1} \text{ mol}^{-1}) \)

\( T \) = Absolute temperature \( (^\circ K = 273 + ^\circ \text{C}) \)

\( M \) = Molecular weight of water \( (18 \text{ g mol}^{-1}) \)

\( g \) = Acceleration due to gravity \( (= 982 \text{ cm sec}^{-2}) \)

\( e_o \) = Vapour pressure of pure water \( (\text{dyn cm}^{-2}) \)

\( e \) = Vapour pressure of soil water \( (\text{dyn cm}^{-2}) \)

\( h_m \) can be measured separately by tensiometers (Richards & Gardner 1936, Richards 1965b) if its numerical value is below the atmospheric pressure.

The relation between \( h_m \) and soil water content can be established by means of suction or pressure plate apparatus (Richards 1949, Richards 1965a), see Figure 3:3. The apparatus can in principle be used for suction (0-1 atmosphere) and pneumatic pressure up to the limit of the strength of the apparatus, in general not more than about 16 atmospheres. The porous plate (or membrane) must have pores small enough to remain water-filled at the maximum pneumatic pressure imposed on it.

A curve showing the relationship between \( 10 \log h_m \) and soil water...
content is generally referred to as the soil water retention curve (or soil moisture retention curve or retention curve). Soil water retention curves for two different soils (sandy loam and clay) are shown in Figure 3:4. For water retention in peat soils, see Boelter 1969, Paivanen 1973.

Figure 3:3 Pressure membrane (plate) apparatus. Pressures below 1 atm. can be established by lowering the water reservoir (W). Pneumatic pressures by increasing the gas pressure in the chamber above atmospheric pressure. h: inlet for gas, C: porous ceramic plate, R: rubber diaphragm. Between the diaphragm and the plate is a sheet of fine wire netting to facilitate water movement to the outlet (O).

Figure 3:4 Soil water retention curves for a clay loam (CL) and a loam sand (LS). The right-side scale shows the corresponding pore radii. The soil water class limits as defined in Table 3:5 are given.
The soil water may be classified according to energetic conditions. The recommended classification is given in Table 3.5 using the terms suggested by Kutilek (1971) in the left side, and those generally used in agricultural science to the right. The range-limits are in terms of $h_m$ which is also referred to as tension or suction.

The range limits (constants) are as follows:

**Full saturation (FS)** This implies that all pores are filled with water. The water content is equivalent to porosity, $h_m$ is zero.

**Saturation (S)** The point when $\log h_m$ is 1.2.

**Retention capacity (RC)** A laboratory substitute for field capacity (FC), which is rather difficult to characterize satisfactorily since it depends on field variables. RC is the water content at $\log h_m$ equal to 2.2.

**Point of decreased availability (PDA)** represents the moisture content at which plant roots experience limitations in the supply of water with consequent reduction in growth rate. The suction value may vary for different plants, but is here set at a value of $\log h_m$ equal to 3.2.

**Wilting Point (WP)** is the point beyond which plant roots are unable to extract water from the soil, $\log h_m$ is equal to 4.2.

**Hygroscopic coefficient (HC)** corresponds to a soil in equilibrium with an atmosphere having a relative humidity of about 89% (atmosphere above a 22% sulfuric acid solution). Kutilek (1971) uses 95% relative humidity, but for systematic reasons we suggest 89%, corresponding to $\log h_m$ of 5.2. Cf definition given in Appendix III.

**Dry state (D)** implies that the soil has been dried to constant weight at a temperature of 105°C, corresponding to $\log h_m$ of 7.0.
<table>
<thead>
<tr>
<th>Soil water</th>
<th>Definitions</th>
<th>Soil water</th>
</tr>
</thead>
<tbody>
<tr>
<td>Range</td>
<td>Classification</td>
<td>$10 \log h_m/cm$</td>
</tr>
<tr>
<td>Full saturation (FS) to Saturation (S)</td>
<td>Wet</td>
<td>below 1.2</td>
</tr>
<tr>
<td>Saturation (S) to Retention capacity (RC)</td>
<td>Humid</td>
<td>1.2-2.2</td>
</tr>
<tr>
<td>Retention capacity (RC) to Point of decreased availability (PDA)</td>
<td>Semihumid</td>
<td>2.2-3.2</td>
</tr>
<tr>
<td>Point of decreased availability to Wilting point (PDA) (WP)</td>
<td>Semiarid</td>
<td>3.2-4.2</td>
</tr>
<tr>
<td>Wilting point (WP) to Hygroscopic coefficient (HC)</td>
<td>Arid</td>
<td>4.2-5.2</td>
</tr>
<tr>
<td>Hygroscopic coefficient (HC) to Dry (D)</td>
<td>Dry</td>
<td>above 5.2</td>
</tr>
</tbody>
</table>
3.3 Movement of soil water

3.3.1 Hydraulic conductivity

Water in the soil is subject to forces of physical and/or chemical nature. The soil water will tend to move in response to gradients in these forces, i.e., due to differences in gravity, matric and osmotic potentials. Under natural conditions, water transport, such as downward movement of precipitation or irrigation water and unidirectional movement of water to plant roots, drain tiles, etc., the two first mentioned potentials are of far greater importance. They are, therefore, often referred to as the hydraulic gradients. In terms of cm water column the hydraulic head is \( H = h_m + h_g \).

The amount of water that can be transported through a given soil, can be formally expressed by Darcy's equation:

\[
Q = K \frac{H_2 - H_1}{l} \text{ At}
\]

where

- \( Q \) = water quantity \( (\text{cm}^3) \)
- \( K \) = hydraulic conductivity \( (\text{cm sec}^{-1}) \)
- \( H_2 - H_1 \) = hydraulic head difference \( (\text{cm}) \)
- \( A \) = cross section area \( (\text{cm}^2) \)
- \( l \) = distance over which the hydraulic head \( (H_2 \text{ to } H_1) \) is acting \( (\text{cm}) \)
- \( t \) = time \( (\text{sec}) \)

The soil is characterized by the hydraulic conductivity, which assumes the dimension of velocity.

Other dimensions may be practical depending on the problems involved. For further information see e.g., Childs 1957, Rose 1966, Slatyer 1967, Jensen 1970, Hillel 1971.

The hydraulic conductivity \( (K) \) depends strongly on the pore geometry of the soil and on the soil water content. As the pore geometry of a given soil (subsoil) may remain rather stable, the drainage properties of this soil can be ascertained by measuring \( K \) at specified water con-
Generally this determination is limited to the case where the soil is at full saturation (FS), for which situation the hydraulic conductivity is termed $K_s$. In contrast to $K_u$, the unsaturated conductivity, $K_s$ depends, however, also on the physical conditions of the water (fluid) i.e. primarily on the temperature, as both the specific density ($\rho_w$) and the viscosity ($\eta$) of the water vary with temperature ($K = \frac{\rho_w}{\eta}$).

Especially the variation in $\eta$ can be important, e.g. the viscosity for water is 0.0101 centipoise (= g cm$^{-1}$ sec$^{-1}$) at 20°C and 0.0179 at 0°C. $K_s$ can be measured in the laboratory or in situ. Care should be taken not to disrupt the original soil structure when making the measurements, as the value of $K_s$ depends on the structural conditions of the soil. Methods for measuring saturated hydraulic conductivity ($K_s$) are described by several authors, e.g. by Reeve et al. 1957, Klute 1965a, Boersma 1965a and 1965b, Aalayng 1968. Methods for measuring unsaturated hydraulic conductivity ($K_u$) are described by Klute 1965b and Hillel 1971. The saturated hydraulic conductivity ($K_s$) may be classified according to the system proposed by Smith & Browning 1946, see Table 3:6.

Table 3:6. Classification of saturated hydraulic conductivity, $K_s$, in mineral soils

<table>
<thead>
<tr>
<th>Class</th>
<th>$K_s$, cm per hour</th>
</tr>
</thead>
<tbody>
<tr>
<td>Extremely slow</td>
<td>Below 0.0025</td>
</tr>
<tr>
<td>Very slow</td>
<td>0.0025 – 0.025</td>
</tr>
<tr>
<td>Slow</td>
<td>0.025 – 0.25</td>
</tr>
<tr>
<td>Moderate</td>
<td>0.25 – 2.5</td>
</tr>
<tr>
<td>Rapid</td>
<td>2.5 – 25</td>
</tr>
<tr>
<td>Very rapid</td>
<td>Above 25</td>
</tr>
</tbody>
</table>

*The hydraulic conductivity of peat soils under saturated conditions is strongly affected by the peat type, the degree of humification, the bulk density, and the sampling depth (Boelter 1965, Paivinen 1973).*
The unsaturated hydraulic conductivity \( (K_u) \) depends on the water content. As the water content also depends on the matric potential, \( K_u \) is a function of \( h_m \) \( (K_u = f(h_m)) \). Gardner 1950 found the following empirical relation:

\[
K_u = \frac{a}{(h_m)^n + b}
\]

where \( a, b \) and \( n \) are constants. The ratio \( a/b \) is identical to \( K_s \) and \( n \) depends primarily on the texture of the soil. It is about 2 for fine-textured soils and about 4 for coarse-textured soils (Gardner, loc. cit.).

In Figure 3:5 \( h_m \) and \( K_u \) are plotted against volumetric water content for a clay loam soil (Philip 1957). \( K_u \) drops quickly as water is removed from the soil (note the logarithmic scale) due to reduction in the flow area as air enters the pores (the greatest and most effectively conducting pores empty first). The great reduction in \( K_u \) as a result of moderate decline in the water content of the soil is the primary reason why many soils retain water at field capacity (FC) in excess of the water content corresponding to the actual ground water level.

![Figure 3:5 Unsaturated hydraulic conductivity (\( K_u \)) and matric potential in cm water column (\( h_m \)) as a function of soil water content (After Philip 1957)](image-url)
In unsaturated soils, water movement also takes place in the vapour phase, due to differences in vapour concentrations. Gradients may be caused by differences in matric potential, osmotic potential and temperature. Generally, the water vapour transport can be described by the following expression (Statyer 1967):

\[
Q_v = -\lambda f \frac{D_v (c_2 - c_1)}{l} t
\]

where

- \(Q_v\) = quantity of vapour (g)
- \(\lambda\) = tortuosity factor (cm cm\(^{-1}\))
- \(f\) = air-filled pore area (cm\(^2\))
- \(D_v\) = diffusion coefficient (cm\(^2\) sec\(^{-1}\))
- \(c_2 - c_1\) = difference in vapour concentration (g cm\(^{-3}\))
- \(l\) = distance (cm)
- \(t\) = time (sec)

The negative sign indicates that water vapour moves from high to low concentrations.

Under isothermal conditions, the transport in the vapour phase is quite small in the transpiration zone of the soil, since the maximum difference in vapour concentration \(c\) (at 20°C) is about 0.2 x 10\(^{-6}\) g cm\(^{-3}\), corresponding to differences in relative humidity from full saturation (100%) to wilting point (20%, 8%).

Differences in temperature can, however, create gradients of importance for water transport. For a temperature difference of 1°C (19-20°C), a vapour concentration difference of about 1 x 10\(^{-6}\) g cm\(^{-3}\) can be established.

3.3.2. Infiltration and distribution of soil water

Infiltration is the entry of water through the soil-atmosphere interface. As water continues to move through a wet soil to the groundwater, the movement is termed percolation. The maximum amount of water, that under specified conditions (soil surface just covered with
water) can infiltrate the soil is termed infiltrability (or infiltration capacity). The amount infiltrated per time increment is called infiltration rate (or water intake rate). The infiltrability $I$ is defined as the maximum amount of water ($Q$, cm$^3$) that in a given time ($t$, sec) can infiltrate through a given area ($A$, cm$^2$). Thus $I$, analogous with $K$, assumes the dimension of velocity.

The maximum amount of water that at any time can enter into the soil depends on $I$ and the hydraulic gradient, and can analogous with the Darcy equation be written:

$$Q = I \frac{H_2 - H_1}{z} At$$

If the soil surface is chosen as reference level, the hydraulic head difference is $H_2 - H_1 = h_m(z) + h_g = h_m(z) + z$, where $h_m(z)$ is the matric hydraulic head at the depth $z$ to which water has infiltrated. The above equation then reduces to:

$$Q = I \left( \frac{h_m(z)}{z} + 1 \right) At$$

When the soil is saturated down to the ground water ($h_m(z) = 0$) only the force of gravity is operative and the gradient approaches unity. This is the minimum infiltrability for a soil already saturated to field capacity.

The infiltrability is measured in the field, either by artificial rain or by use of special infiltrometers, which are devices enclosing a defined area (e.g. ring infiltrometers) over which a constant water level can be maintained, and the water entry can be measured. Methods for determination of infiltrability are described by Bertrand 1965, De BooHt 1967 and Aslyng 1968.

The path of downward movement of water in soils following application of water at the soil surface is described in detail by Bodman & Colman 1944 and Colman & Bodman 1945. Water enters the soil and fills up nearly all the air pore space layer by layer from above, as the water supply proceeds. Situations of water entry into two initially dry soils of different waterholding capacities are shown in Figure 3:6 (after Bodman & Colman 1944). During the process of infiltration sur-
plus water (relative to field capacity) exists in the previous wetted part of the soil (the transmission zone). A sharp division between the wetted and non-wetted part of the profile appears. Figure 3·6 also illustrates the difference in infiltrability of the two soils.

![Figure 3:6 Soil water contents and infiltration times (in minutes indicated by numbers at the curves) for coarse (A) and fine (B) textured soil. (After Bodman & Coleman 1944)](image)

The infiltrability decreases as the process of infiltration proceeds, primarily due to the continuous increase in length of the transmission zone. When the whole soil profile is saturated, the infiltrability reaches its minimum value.

The change of infiltrability with time is shown in Figure 3·7 for three different soils (Aslyng 1968). The curves may be described by an empirical relation, as demonstrated by Philip (1964):

\[ I = \frac{s}{2\sqrt{t}} + a \]

where \( s \) and \( a \) are constants related to the sorption capacity of the soil and the gravity field respectively. As \( t \) (sec.) increases, the first term on the right hand side approaches zero, and \( I \) consequently approaches \( a \).
As the infiltrability depends on initial water content, water distribution and pass of time (Figure 3:7), it is only practical to characterize the infiltrability of a soil by its minimum value \( (I_{\text{min}}) \). Such a classification is suggested in Table 3:7.

**Table 3:7. Classes of minimum infiltrability \( (I_{\text{min}}) \) of soils.**

<table>
<thead>
<tr>
<th>Class</th>
<th>( I_{\text{m}, \text{cm per hour}} )</th>
</tr>
</thead>
<tbody>
<tr>
<td>Extremely slow</td>
<td>Below 0.0025</td>
</tr>
<tr>
<td>Very slow</td>
<td>0.0025 - 0.025</td>
</tr>
<tr>
<td>Slow</td>
<td>0.025 - 0.25</td>
</tr>
<tr>
<td>Moderate</td>
<td>0.25 - 2.5</td>
</tr>
<tr>
<td>Rapid</td>
<td>2.5 - 25</td>
</tr>
<tr>
<td>Very rapid</td>
<td>Above 25</td>
</tr>
</tbody>
</table>

The classes and criteria are identical to the ones recommended for saturated hydraulic conductivity (Table 3:6).
3.4 Effects of vegetation

Besides soil factors, vegetation also has very significant effects on soil water. Thus, having dealt with the main abiotic factors affecting soil water in the preceding section, it will be the purpose of the present section to take up important biotic aspects.

The characteristics of a particular soil water regime can be regarded as an expression of the combined effects of soil, vegetation and climate. The interaction between these three components (of the ecosystem) is very close. Plants absorb energy from the sun and transform this in such a way that materials from both the atmosphere and soil are accumulated as organic matter both living and dead on the earth surface. Certain parts of the vegetation extend above the surface of the soil to form a cover which protects it from the atmosphere and from the effects of climate. Other parts penetrate the soil as a kind of network, the root system, which acts to bind together the soil particles, thus affecting both the structure and the permeability of the soil. The hydrological significance of vegetation thus becomes obvious, but also difficult to characterize in simple terms. Some of the more important consequences of vegetation on the soil water regime will be presented in this section. Furthermore an attempt will be made to show how soil vegetation complexes can be classified and used to delineate various types of soil water regimes.

3.4.1 Effects of vegetation on precipitation

Plants depend on the soil for their water supply. This water is intermittently renewed to the soil by precipitation, but continuously depleted from it during the growing season by transpiration. The renewal is interfered with through interception of the rainfall by the vegetation cover itself. Since the extent of the cover greatly varies between various vegetation types, a separation will be made in the following discussion between forests and arable land.

Forests

In forests a considerable part of the precipitation is intercepted by the tree crowns. The amounts are dependent on the tree species,
Figure 3:8 Throughfall during variable rains under an individual tree. The points are averages of four sites, expressed as per cent of the gross precipitation. Curve 1 shows average precipitation at 125 cm and curve 2 at 50 cm from the trunk.
(MITSCHERLICH, 1971)

Figure 3:9 Interception, throughfall and stemflow. The dashed lines on the top of the diagrams indicate the standard error.

In Figure 3:9 the results from interception investigations in Middle- and Western Europe are gathered. In coniferous forest between 25 and 35 per cent of the precipitation is retained by the crowns. The
values for the broadleaved species presented vary correspondingly between 10 and 20 per cent, calculated as rough annual values. The bottom of the diagrams indicates the rate of stemflow with the lowest values for conifers.

DELF (1967) found the interception losses of conifers and hardwoods in Germany so different that hardwoods were recommended as watershed cover. ZINKE (1967) suggest that sizeable amounts of intercepted water may be absorbed into living plant tissue to join the transpiration stream, thus further complicating the role of interception. In forest stands, even the bottom vegetation and litter, retain considerable amounts of water from the crown throughfall, see Table 3:8.

Table 3:8

<table>
<thead>
<tr>
<th>Moisture intercepted and absorbed, in mm</th>
</tr>
</thead>
<tbody>
<tr>
<td>Thickness, cm</td>
</tr>
<tr>
<td>----------------</td>
</tr>
<tr>
<td>6.0</td>
</tr>
<tr>
<td>2.0</td>
</tr>
<tr>
<td>0.25</td>
</tr>
<tr>
<td>0.10</td>
</tr>
</tbody>
</table>

Arable land

According to PENMAN (1963) information concerning interception by agricultural crops is scanty. The amount of rainfall intercepted varies with plant species, spacing, age and size of plant, and with the amount of rainfall. Presented results reveal an enormous variability under canopy. PENMAN (1963) gives, as an example, the results for different crops from an investigation made by WOLLNY in 1890 (Table 3:9). Similar variations are found in investigations from later years, and it is difficult to find results of more general value.
Table 3:9

Interception (%) by agricultural crops - detail
(Woilny)

<table>
<thead>
<tr>
<th>Period</th>
<th>Rain (mm)</th>
<th>1-5</th>
<th>21-25</th>
<th>11-15</th>
<th>6-10</th>
<th>16-20</th>
<th>16-20</th>
<th>26-31</th>
<th>6-10</th>
</tr>
</thead>
<tbody>
<tr>
<td>Maize 9m⁻²</td>
<td>86</td>
<td>53</td>
<td>40</td>
<td>30</td>
<td>15</td>
<td>9</td>
<td>2-7</td>
<td>0-6</td>
<td></td>
</tr>
<tr>
<td>Oats 16m⁻²</td>
<td>4</td>
<td>32</td>
<td>47</td>
<td>10</td>
<td>15</td>
<td>38</td>
<td>-21</td>
<td>14</td>
<td></td>
</tr>
<tr>
<td>Beans 25m⁻²</td>
<td>27</td>
<td>22</td>
<td>19</td>
<td>32</td>
<td>6</td>
<td>12</td>
<td>4</td>
<td>57</td>
<td></td>
</tr>
<tr>
<td>Peas 25m⁻²</td>
<td>1</td>
<td>-</td>
<td>10</td>
<td>18</td>
<td>11</td>
<td>41</td>
<td>-</td>
<td>-</td>
<td></td>
</tr>
<tr>
<td>Clover (broadcast)</td>
<td>44</td>
<td>-</td>
<td>26</td>
<td>44</td>
<td>46</td>
<td>56</td>
<td>-</td>
<td>-</td>
<td></td>
</tr>
</tbody>
</table>

3.4.2 Effects of vegetation on surface runoff

Surface runoff occurs when rain intensity exceeds the infiltration rate. A plant cover significantly affects the infiltration rate of the soil surface and thus reduces the surface runoff. Of particular interest is the runoff in spring when accumulated snow is melting. MOLCHANOV (1963) gives a broad survey of research in this field, and Tables 3:10 and 3:11 are examples from this publication.

A plant cover forms a rougher surface and the root systems loosen the soil. Decayed roots even leave porous channels into the ground. These are all factors which lower the surface runoff.

3.4.3 Effects of vegetation on evaporation

The term evaporation is here used for vapor transfer to the atmosphere from the soil surface and the plant cover. In the literature, the terms: Evaporation, evapotranspiration, and vaporization can be found to be used synonymously.

In the first place evaporation governed by the fact that energy is needed in the order of 600 cal. per g. water. Furthermore the process depends on the transport of water vapor from the evaporation surface to the surrounding air. Several equations have been suggested to calculate the potential evaporation from meteorological parameters.
Table 3:10

Possible surface runoff on 5° slope, in various types of forest and at different slope exposures (in %)

<table>
<thead>
<tr>
<th>Type of forest (and exposure)</th>
<th>Length of slope, in m</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>2 100      200 300 400 500 600 700 800 900 1000 1200 1400 1600 1800</td>
</tr>
<tr>
<td>Spruce stand on north-eastern slope, with disturbed cover and very trampled litter; soil overgrown with grass.</td>
<td>38 18 16 16 16 15 15 15 14 14 14 13 13 13</td>
</tr>
<tr>
<td>Id., on southern slope</td>
<td>47 23 21 20 20 19 19 18 18 18 18 18 17 17</td>
</tr>
<tr>
<td>Id., on northern slope; litter undisturbed.</td>
<td>3 1 0 0.9 0.9 0.9 0.8 0.8 0.8 0.8 0.8 0.7 0.6 0.8</td>
</tr>
<tr>
<td>Beech stand</td>
<td>1 0 0.5 0.5 0.4 0.4 0.4 0.4 0.4 0.4 0.4 0.4 0.4 0.4</td>
</tr>
<tr>
<td>Pasture land</td>
<td>86 51 46 45 44 43 43 42 42 41 41 40 39 38 37</td>
</tr>
</tbody>
</table>

Table 3:11

Runoff on elementary plots, in mm

<table>
<thead>
<tr>
<th>Year</th>
<th>Autumn</th>
<th>Winter</th>
<th>Spring</th>
<th>Summer</th>
<th>Runoff during the year</th>
<th>Precipitation</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>forest</td>
<td>meadow</td>
<td>forest</td>
<td>meadow</td>
<td>forest</td>
<td>meadow</td>
</tr>
<tr>
<td>1934/35</td>
<td>1</td>
<td>39</td>
<td>-</td>
<td>1</td>
<td>26</td>
<td>98</td>
</tr>
<tr>
<td>1935/36</td>
<td>21</td>
<td>161</td>
<td>1</td>
<td>13</td>
<td>59</td>
<td>76</td>
</tr>
<tr>
<td>1936/37</td>
<td>0</td>
<td>4</td>
<td>0</td>
<td>33</td>
<td>55</td>
<td>113</td>
</tr>
<tr>
<td>1937/38</td>
<td>0</td>
<td>12</td>
<td>0</td>
<td>11</td>
<td>38</td>
<td>116</td>
</tr>
<tr>
<td>Average</td>
<td>6</td>
<td>54</td>
<td>1</td>
<td>14</td>
<td>44</td>
<td>101</td>
</tr>
</tbody>
</table>
and using a standard vegetation cover (PENMAN 1963).

Figure 3:10 shows a potential evaporation map for Scandinavia. The values will be found to decrease from 600 mm in the south to less than 200 mm in the north.

Figure 3:10. (WERNER JOHANNESEN, 1970)
The mean annual potential evaporation from grass surface (mm).
Actual evaporation varies with the type of vegetation, the stand density and the vegetative height. Water supply limitations usually reduce the potential values of evaporation. Figure 3:11 gives average values for precipitation and actual and potential evaporation in Denmark. It shows a precipitation deficit during the summer which is reflected in the partition between the curves for potential and actual evaporation.

Soil water deficits relative to field capacity will arise if the root system is extracting more water than can be redistributed either by rain or by capillary rise from the groundwater. No infiltration to the groundwater will take place until this deficit in the root zone is recharged.

Figure 3:11. (ASLYNG, 1966)

Monthly average precipitation (N), potential (Ep) and actual (E_a) evaporation, at K. V. Sta. 1955-64.
BAUMGARTNER (1970) has in Figure 3:12 grouped together the evaporation rates in relation to annual precipitation from different lysimeter investigations. It shows that there is a significant variation both within and between different types of evaporation surfaces.

Lawn (water table near surface)

Open Water

Forests

Grassland

Vegetables

Cereals

Bare Soil

of annual Precipitation

Figure 3:12

Lysimeter values for specific earth surfaces and vegetation covers in the middle latitudes. Evaporation rates are related to the annual precipitation.

3.4.4. Effects of vegetation on groundwater Level and Streamflow.
The interception and evaporation effects of vegetation are difficult to measure and are therefore seldom included in soil water investigations. Values of these parameters can therefore only be obtained from the literature as very rough estimates.

The total effects of vegetative cover on soil water is reflected, however, on the groundwater level variations. Figure 3:13 shows the groundwater level changes during the year before and after clear-cutting of a beech stand in Denmark. On small cutting areas the effect
is very much influenced by soil texture. When the groundwater level lies near the ground surface as in peat lands drained for forestry both thinning, clear cutting and fertilisation have an effect on the water regime of the soil (Heikurainen & Päivänen 1970, Päivänen 1972).

MOLCHANOV (1963) draws the following conclusions: "On sands, forest has only a very slight lowering effect on the water table. When the level of groundwater lies deeper than 3.3 m, there is no observable difference in water table between forest and field. Some variation in the water table level starts when the groundwater lies at a depth of 2 m. As the water table becomes shallower, the influence of vegetation increases. When the water table is between 1.5 and 1 m, one observes the greatest lowering of the groundwater level during the growing season. In forest, when the water table is not deeper than 1.5 m, its level is above that observable on field. The strongest lowering of the water table is caused by rye; buckwheat is less effective. Under bare fallow the water table is shallower than under sown field."

Concerning the effect of vegetation on streamflow, the conclusions found in the literature are often contradictory. RAKHMANOV (1966) says that: "Forests are conducive to higher precipitation, lower evaporation, and to a greater volume of water within the drainage area. Deforestation results in the shoaling of rivers and a general desiccation of the locality." PENMAN (1963) gives the following comment to this: "It is safer to treat Rakhmanov's statement as a hypothesis rather than as a conclusion, and then look at the relevant evidence."

![Figure 3:13](HOLSTENER-JØRGENSEN, 1967)

Ground water fluctuations before and after clear cutting.
Dotted curve: 1956/7 before clear cutting
Solid curve: 1957/8 after clear cutting.
Figure 3:14 shows an example of the effect of forest on streamflow from a famous watershed experiment in Georgia. The dependence of streamflow on the forest stand development from the first to the second clearcutting is significant.

Figure 3:14. (HIBBERT, 1967)
Deviations from regression of annual streamflow for Coweeta Watershed 13 on annual streamflow for control watershed calibration (1936-9) and treatment (1939-64) periods.

A more varied presentation of data from various types of vegetation cover is presented in Figure 3:15. The regression line for forest shows the lowest streamflow in relation to precipitation.
Streamflow from areas with various surface cover.

3.4.5 Vegetation and ecosystem classification

Various systems have been developed for the classification of plant communities and vegetation stands. Since vegetation is closely inter-related with soil water conditions, it should be possible to develop a natural system of classification that would allow important predictions regarding the soil water status.

For the analysis of plant communities it is important to look at the structure (physiognomy) of the vegetation stand. For this purpose it is convenient to recognize four different strata of vegetation above the soil surface, see Figure 3:16:
1. **Tree layer** comprised of high wooded plants exceeding ca 3 meters in height.

2. **Shrub layer** also composed of wooded plants but less than 3 meters in height.

3. **Field layer** composed of brush, herbs and grasses.

4. **Bottom layer** made up of very low plants such as mosses and lichens growing close to the ground surface.

Some plant communities are composed of all four strata, e.g. deciduous forest, while others only have one or two strata, e.g. peat bogs. Molchanow 1960 investigated evaporation from the field layer of a number of pine, spruce, birch and oak forests by using weighable "evaporators" and found that the field layer accounted for 20-25% of the total evaporation from the forest.

**Figure 3:16.** Stratification of a vegetation stand into 4 layers (Påhlsson 1972).
Using the water and nutrient status of the soil as decisive parameters in differentiating habitat types and combining this with prevailing vegetation structure and composition (floristics), the following taxonomy has been developed to classify terrestrial ecosystems:

A. Marsh or bog
B. Meadow
C. Steppe
D. Heath

The system in its present form affords a simple type of ecosystem hierarchy, which can be applied to fairly large parts of the Nordic region. Furthermore it helps give some hydrological bearing on the significance of vegetation. It is to be hoped that efforts will be made by hydrobiologists to develop the classification system through intensified vegetation reconnaissance of watersheds. Systematic development of vegetation surveys will bring to light many important correlations and problem complexes that might otherwise pass unnoticed. The International Biological Programme that has been set up in the Nordic region is also presently making efforts to develop a more universal type of vegetation classification for mapping purposes in Norden (IBP i Norden No 11, 1973).

Table 3:12 and Figure 3:17 give an outline of the characteristics of the four terrestrial ecosystems mentioned. See also Appendix 2.

<table>
<thead>
<tr>
<th>Ecosystem</th>
<th>Dominant vegetation layers</th>
<th>Soil characteristics</th>
</tr>
</thead>
<tbody>
<tr>
<td>Heath</td>
<td>Field layer: heather, cranberry, grasses, some few herbs. Bottom layer: mosses and lichens</td>
<td>Podzol mor type of humus</td>
</tr>
<tr>
<td>Steppe</td>
<td>Field: some brushes, grasses, numerous herbs. Bottom: specific types of mosses and lichens</td>
<td>Brown soils - pararendzina. Often on calcareous soils</td>
</tr>
<tr>
<td>Meadow</td>
<td>Field: numerous grasses and herbs. No bushes. Bottom: only mosses</td>
<td>Brown forest soils and gray brown podzolic soils. Mull-humus</td>
</tr>
<tr>
<td>Bog</td>
<td>Tree and shrub layers</td>
<td>Organic soils</td>
</tr>
<tr>
<td></td>
<td>Field: sedge, dwarf shrubs. Extremely wet bushes</td>
<td>Bottom: mosses</td>
</tr>
</tbody>
</table>

See also Appendix 2.
Figure 3:17 Schematic arrangement of terrestrial ecosystems according to gradients of moisture and nutrients (Sjörs 1967).
4. FIELD MEASURING TECHNIQUES

4.1 Classification of methods

In order to facilitate the following presentation of methods for measuring soil water, some words must be said concerning the underlying aspects of approach for these methods. First of all it must be made clear that soil water can be characterized both as regards concentration and as regards energy status. Concentration is generally measured as percent by weight of the oven dry soil or as percent by volume of the soil in its natural state. The energy status of soil water is expressed as a potential, that is as a pressure difference of soil water in relation to some reference, generally atmospheric pressure. The units applied may be those of energy per unit mass or unit volume of soil water. For a given soil media there exists a relationship between the concentration and potential aspect of the soil water. If this functional relationship is known, then it becomes possible to derive the magnitude of one of these aspects from knowledge of the other.

Methods of determining soil water may also be differentiated according to the principle of detection of the water. When the water is determined by direct extraction from the soil medium, the methods are classified as direct methods. Such methods are generally based on sampling techniques and extraction of water from these by physical or chemical means. Methods based on properties of the soil medium or on properties of an object put in equilibrium with the soil that are dependent on the soil water conditions are classified as indirect methods, since it is not the water but some other property affected by the water that is measured. The direct methods involve such techniques as the gravimetric method with oven drying and, less commonly in use today, the pyknometer, densimetric and carbide methods. The indirect methods include the neutron, gamma ray, electrical resistance and capacitance, tensiometric and thermal conductivity methods. It is apparent that for stationary point measurements of soil moisture the direct methods are less convenient because they are elaborate and need sampling for each
moisture measurement. As the samples cannot be taken from exactly the same place, the observations are influenced by sampling errors, the magnitude of which depends on the non-homogeneous nature of the soil.

The indirect methods offer several advantages, especially for stationary point measurements. The measuring devices are permanently built into the soil, or they are inserted into access tubes. The physical quantities sensitive to moisture changes are measured, and the soil moisture is then read from calibration curves. Long-distance registration is possible with most methods. Some devices age during long-term measurement and as a consequence their calibration changes. It is therefore necessary to recalibrate them regularly. As the moisture is not measured directly, it is evident that the measured data will reflect, apart from the changes in moisture, changes in those soil properties which influence the physical quantities measured by the individual methods. The precision of the method then depends on the possibility of either eliminating the influence of these outside factors or determining how constant this influence is. Many physical properties of soils exhibit a hysteresis curve with respect to soil moisture similar to the suction moisture relation, Figure 4:1. If hysteresis takes place, the history of the soil moisture under measurement should be taken into account, which makes the method less useful as a standard type of measurement.

Figure 4:1 Soil water retention curve for state of sorption and desorption with intermediate hysteresis loops. (taken from Hillel 1971)
Apart from the above division, methods may also be divided according to the volume of soil for which the measurement is valid. If the method determines the moisture status in the immediate neighbourhood of the measuring device, the method is designated as a point method. Such are the gravimetric method in the usual system of measurement, the electrical resistance method, most tensiometer methods, and the thermal conductivity and electrical capacitance methods. Some methods determine the average moisture value of a larger volume of soil with a spatial extent of tens or even hundreds of centimetres. Such methods may be termed bulk methods and include the radiometric and lysimetric methods.

Regardless of the type of method applied, it is essential in every instance to determine the moisture in the soil at many test points in the area under examination, because of the high lateral and depth variability of the soil medium. A statistical analysis of the variability of the soil moisture values observed makes it possible to calculate the error of the mean and to find the limits within which the true mean result lies, which can be obtained only by an infinite number of determinations. This analysis of variance makes it possible to determine the number of test points required for solving specific problems.

### 4.2. Sampling techniques

Most methods involve making a hole either for the removal of a sample or for the insertion of an instrument. In addition, even indirect methods require samples, as they are calibrated with direct methods. Tools for these purposes range from trowels to power augers. A good sampler will allow uncontaminated samples to be taken as quickly as possible from the required depth with minimal damage to the site. Holes should be filled, but damage to plant roots and the introduction of variable drainage and infiltration characteristics may be unavoidable. Since much replication is needed to characterize a typical experimental site adequately, disturbance is often considerable. These limitations do not apply to in situ methods, since instruments may be installed in a few holes before measurements commence. For some projects samples may be taken more quickly with power tools, perhaps at the expense of more damage to the site.
The main purpose of sampling is to obtain a representative soil sample from the intended soil layer in as undisturbed a state as possible. However, this involves numerous difficulties, while the success of the sampling naturally has a considerable effect on the reliability of the results. The heterogeneity of the soil layer, as well as the moisture status and the sampling method, impose such great restrictions that ideal conditions can hardly ever be reached in practice. Thus, even under the best circumstances, the sample only approximately represents the actual moisture status of the soil and is more or less disturbed. Sampling is especially difficult when the soil is very dry or very wet and when it contains stones, rocks, and other objects which preclude easy cutting using sampling equipment.

The technique and equipment used for sample collection should be such that the samples do not lose or gain moisture, or otherwise become altered or contaminated during sampling and transportation. In sampling through a wet layer into a dry layer, care must be taken to keep the sampling equipment as dry as possible and to prevent water from running down the hole into the drier material. If there is free water in the soil, the measured moisture content will probably be less than the correct value because some water will run off as the sample is being removed from the ground, or some may be squeezed out by compaction during sampling.

When fine-textured sediments are in a dry, hard state, it is difficult to drive the core barrels or to rotate the augers. When dry, coarse-textured sediments are sampled, the sample may slide out of the core barrel or auger as it is withdrawn. Moraine soils are very difficult to sample, especially volumetrically, owing to the danger of hitting stones with the cutting edges of the equipment, and because the representative samples must be large. Soils that contain a considerable amount of roots and organic matter also present difficulties. In soil moisture sampling, it is essential that all sampling operations - the transfer of samples to moisture cans and the weighing of the moist samples - be done as rapidly as possible to prevent undue moisture losses. Many difficulties in the use of sampling equipment, whether augers or core samples, may be overcome if all the equipment is kept clean, that is, free of moisture, oil, rust, and dirt.
Sampling equipment can be divided into two groups: that which takes disturbed and that which essentially takes undisturbed samples. A sampler can take point-like samples of the soil layer or long profile samples. The former are taken for determining the soil moisture status and the latter for studying the vertical moisture distribution. Sampling cohesive soils in their natural state is usually not so difficult. For instance, very long (10-30 m) undisturbed profile samples can be removed from clay with special augers. On the other hand, taking undisturbed samples from non-cohesive soils is often difficult, as has already been shown with regard to moraine soils.

Different kinds of sampling methods have been treated in detail in several soil sampling guide books, for example Statens Geotekniska Institut 1970, Tie-ja Vesirakennushallitus 1970, Kairausopas III (Finnish Auger Guidebook) 1972. There is also a recommendation from the Group on Soil Sampling (IGOSS) of the International Society of Soil Mechanics and Foundation Engineering (ISSMFE) in 1967. Only a few of the most important views will be presented here for consideration when sampling is required for the determination of the moisture status of the soil.

4.2.1. The sampling hole approach

A sample is best obtained from the surface layers of the soil by digging a sampling hole to the depth required for driving a sharp-edged, thin-walled metal cylinder into the undisturbed ground. The cylinder is driven into the soil either by hand or by using levers or, for instance, a pneumatic compressor. Use of an external guiding cylinder is recommended to help drive the cylinder straight into the ground without lateral displacements. The cylinder should not be struck, for the sample may then be disturbed. After careful insertion the filled cylinder is dug up from the soil. Figure 4:2 presents some different ways of taking undisturbed samples from a sampling hole.

The wall thickness of the sampling cylinder is generally 2-4 mm and the sharpening angle of the lower edge 45-60°. The walls must be at least so thick that the cylinder will retain its circular form during all normal handling in the sampling operation. The diameter of the cylinder
Figure 4:2  Removal of undisturbed soil samples from a sampling hole (soil pit):
a) and b) with a direction cylinder,
c) and d) by carving or moulding out a sample that is impregnated before removal to jar,
e) by using a jar into which a soil sample is moulded,
f) by means of a pneumatic compressor

must be at least 10 cm. The height of the sample depends on the type of soil. Cylinders that are 15-20 cm high can be used in clay or silt soil. In coarse-textured soils the ratio of height to diameter must be smaller than the measurements given above, and thus the height of the cylinder may only be 5-10 cm.
4.2.2. Sampling with piston drills

Sampling from layers at great depths beneath the surface is best done using different kinds of piston drills, which can produce volumetric samples for calculating moisture content on a volume basis. The sampling team generally consists of 2 or 3 men. Depths of down to 20 m can be sampled. The piston sampler is very useful, especially in sampling through loose or wet materials that tend to slough into the hole.

The piston drill is composed of a piston shaft and a sampler at the base of the shaft. The sampler has an open sampling cylinder with a moving drill. The piston's shaft is composed of a tube (case) as a continuation of the cylinder, and a rod as a continuation of the piston. The operating principle of the piston drill, see Figure 4:3, is that the sampler is driven into the ground to the desired depth with the piston in the lower position (at the base of the cylinder) so that it acts as the point of the drilling system. The piston is then kept in place with the aid of the drill rods, while the cylinder is pressed down. The cylinder then fills with soil from below the piston. When the drill is raised, the soil sample remains in the cylinder because of the friction between the inner surface of the cylinder and the sample and the vacuum between the piston and the sample. During the raising, the piston is locked in the upper position and the drill can be raised mechanically with a lifting jack or a tripod. After the cylinder has been emptied and cleaned, the procedure can be repeated to a greater depth. The case can be extended if desired with a metre long pipe and the piston, in turn, with a drill rod of the same length, so that longer samples can be taken.

The piston drill models generally used in the Nordic countries are standard piston drills St I and St II, Norwegian piston drill MN 54 and the foil piston drill. Figure 4:4 shows the piston drills in question.

St I type piston drill (Figure 4:4 a) works in exactly the manner just described. The St II type piston drill (Figure 4:4 b) differs from the principle of the St I in that no casing pipe is used. Instead, the sampler is driven into the ground with a simple handle, for instance with a rammer drill rod. The sampler is composed of two cylinders, one inside the other, the outermost of these acting as the protective cover.
Figure 4:3  Operating principles of the piston drill (TVI 1970)
   a) drill driven into the soil
   b) filling the drill

Figure 4:4  Different types of piston drills (Kairausopas II 1970)
   a) St I type
   b) St II type
   c) Norwegian type
   d) foil piston drill
The actual sampling process differs from sampling with the St I in that the inner cylinder is driven into the ground by turning the drill rods, while the outer cylinder and the piston remain in place. With both of the St drills the soil sample is driven straight into the sample jars that are contained in the cylinder, measuring 50 mm in diameter and 85 or 170 mm in length.

The Norwegian type of piston sampler (Figure 4:4 c) differs from the Swedish types in that its sample jar consists of only one sample pipe in which the piston moves and which after sampling is removed completely from the sampler. The sample pipe is an unseamed stainless steel pipe, usually 80 cm long and about 50 mm in diameter. The sampler is driven into the ground from a casing pipe. The casing pipe contains rods as extensions of the piston, just as in the St I type drill.

With the above type of ordinary piston drill only relatively short (<1 m) samples can usually be taken. For longer sample lengths the friction between the sample and the sample cylinder becomes so large that the sample is disturbed. For procuring longer uniform soil samples the foil piston drill (Figure 4:4 d) can instead be used. Here the friction between the sample and the cylinder is eliminated by use of metal sheets between the sample and the cylinder. The metal sheets are attached to the locking system at the bottom of the drill in the form of wounded rolls that surround the base of the drill. In sampling, the piston is kept motionless in relation to the soil surface while the piston itself is driven onwards. In this way the sample is pushed into the pipe and at the same time the rolled sheets unwind and surround the sample. When the drill is operated there is no movement between the sheets and the sample. Movement only occurs between the sheets and the interior of the sample cylinder. When the desired sample is in the pipe, the whole drill is raised from the ground and set down in a horizontal position. The sample is pushed out of the pipe with the piston. Even during extraction, the only friction is between the sheets and the interior of the pipe.
4.2.3. Other methods

In addition to sampling from sample holes and with piston drills, there are several other methods which can be used. With the exception of sampling from surface layers, they give disturbed samples which can only give values of water content expressed on a weight basis.

In sampling from surface layers different kinds of cylinder-like core samplers are usually employed. In Finland, for example, a core drill measuring 50 mm in diameter with a removable hard metal point has been used in inorganic soils. When the drill is driven into the ground, the soil sample should fill either the bare cylinder or plastic or fibreglass jars in the cylinder. A metal sampler made of two steel profiles is used for taking profile samples. One of the profiles is a U-profile and the other a sheet blade made of spring steel which with the U-profile forms a case measuring 30 mm x 40 mm x 100 cm (or x 200 cm depending on the length of the profile desired). The U-profile is first driven or struck into the ground to about 3/4 of its length and then the sheet blade with the sharp point first is driven to the same depth along the open edges of the U-profile. The samplers are then lifted up and with them the profile sample, which is usually fairly disturbed in rough inorganic soil, but less disturbed in clay or silt soils.

For sampling from deeper layers of soil, hand augers attached to weight drill rods are the simplest equipment. The samples thus taken are always disturbed, however. Suitable samplers are, for instance, post-hole, helical and spoon-tipped augers. None of these are suitable for sampling in soft or miry layers or under the ground water level.

The tube-like sampling auger of the miniature piston drill, on the other hand, is also suitable for sampling in soft soils. Especially suitable for use in wet and miry soils is the can auger, which is composed of a cylinder-like sampler with handles attached to it. The handles are attached with a special coupling which allows the drill to rotate both clockwise and counter-clock-wise. Samples can also be taken using vibrating, monkey, or spoon type samplers, air pressure and diamond drills and/or by using water and air pressure washing techniques.
4.2.4. Sampling from peat

Sampling from peat poses special problems because of the structure of the peat. For this reason a completely satisfactory sampler is very difficult to develop. Larger samples must be taken from peat than from mineral soils, because of its inhomogeneity and fibrosity. The large water binding ability of peat and its large water permeability have the same effect. Above all, sampling requires effective cutting of the sample from its surrounding soil, because peat fibres can be fairly tough and can easily get packed into the sampler. In this case the sample is apt to be under pressure and its water content will decrease. The water content also changes very easily during other phases of sampling.

Usually, special cylindrical core samplers are used for sampling in peat. For instance, a polyethylene pipe 200 mm in diameter is used in Finland. It has a metal cutting blade with a wavy edge which can be tightened with screws at the lower end and a turning handle attached with wing nuts at the upper end.

A special drill has also been developed especially for peat soils made of a thin-walled cloven steel pipe 120 cm long, with an inside diameter of 94 mm (Figure 4:5). When using the drill, one side is driven into the ground first. When the second side is driven in, the welded guide strips direct it as it presses against the other side. In addition, the blade, which leans inward, helps the two sides to press closely together. This kind of sampling is suitable only for highly decomposed peat, because the point edge of the sampler cuts rather badly since the sampler cannot be turned.

In the newest type of peat samplers special attention has been paid to the construction of the cutting point and to the dimensions of the sampler, as well as to the friction between the peat and the sampler. (Helenelund & Lindqvist & Sundman 1972). Special types of round and square samplers have been developed with various cutting point constructions (Figure 4:6). The round cylinder samplers are 125 cm and 135 cm long, and 150 mm and 250 mm in inside diameter. The square cylinder sampler is 140 cm long and 150 x 150 mm wide. The material
Figure 4:5  A peat-drill made of a cast steel pipe
(Hooli 1971)

Figure 4:6  Different types of cutting edges and pneumatic
pistons for thin-walled samplers used in
sampling tests in fibrous peat
(Helenelund & Lindqvist & Sundman 1972)
usually used is chrome steel 1.5 to 2.5 mm thick. The cylinder's cutting edge is usually narrowed 5 cm so that the outer diameter of the opening is the same as the interior diameter of the upper part of the cylinder. This construction is intended to decrease the friction between the sample and the cylinder. Covering the inner surface with plastic film or a lubricant is also used to decrease friction.

Samples are taken by driving the cylinder evenly into the peat and turning it at the same time to improve the cutting. When the sampler is at the desired depth, a pneumatic piston (Figure 4:6 b, c) is attached to its upper part. The piston is pressed closely against the cylinder's interior surface, using air pressure. The sampler is then raised and the sample with it as a result of the friction and vacuum caused by the piston.

Good results have also been obtained by cutting samples of peat as columns with a long bladed power saw. The columns are cut at the bottom and raised with a special lifting board. Thus a prism-shaped profile sample, which is only very slightly disturbed, is obtained.

4.3 Water content
4.3.1 Gravimetric methods
The gravimetric method is the classic method generally used to calibrate also other methods. (Gardner 1965, Cope & Trickett 1965). Samples are taken and placed in tarred containers. These should be weighed as soon as possible, but careful scaling and cool storage will reduce errors to a minimum. The moisture content is determined by drying the samples at 105°C to a constant weight in a thermostatically controlled oven. Somewhat lower temperatures are used in drying peat samples. The weight of the water lost is expressed as a percentage of the dry soil weight. Sometimes the moisture content is expressed as a percentage of the volume of the original soil sample. The drying and weighing procedures are slow but they may be speeded up by using infra-red or vacuum drying techniques and an automatic balance. Apart from such expensive modifications, the method has the advantage of requiring simple equipment that is commonly available. It is valid for the sample used and, with adequate replication, for the site chosen. The chief disadvantages are those implicit in taking
and handling the samples, and the process is particularly tedious when correction must be made for the variable content of stones and gravel in the samples. The method often involves difficulties connected with following the moisture changes with time and in space. It demands much work and time. Complete drying requires about 24 hours. It is further known that some clay minerals can adsorptively retain water even at temperatures exceeding 105°C.

Other methods have been developed whereby even simpler pieces of apparatus have been used than oven drying and gravimetry in order to speed up operations. Of these, pyknometry and gravimetry based on drying by burning alcohol have been used to some extent. In the pyknometric method oven drying is eliminated by shaking a weighed sample in a flask to remove the air and then weighing the flask in water. The moisture content is calculated by assuming a mean value for the specific gravity of the soil, thus permitting simultaneous determination of the bulk density. On the other hand, the water content of the soil can be determined with an air pyknometer. In this method the soil sample is enclosed in a vessel of known size and the air volume of the sample determined by the combined use of pressure measurements and knowledge of the dry weight of the sample and the specific weight of the grains. Now the volume of water in the sample can be calculated and expressed on mass or volume basis. Water content can be determined with an air pyknometer in all types of soil, although the method is slightly less useful with solid clays, the pores of which are not easily penetrated by compressed air. The method is not very exact.

4.3.2. Nuclear radiation methods

Radiation consisting of neutron particles or gamma energy-quanta can be utilized for the determination of the water content of soil. The main elements required for such a determination are a radiation source and a detector that is sensitive to the radiation used. The flow of signals from the detector is determined with a counting device, i.e. a scaler or a rate meter, see Figure 4:7. A suitable voltage supply, a pulse amplifier and a pulse-height discriminator are other devices that are often needed. The necessary radiation shield and radiation monitoring must also be suitably arranged. An advantage of the radiometric methods
is that the measurement is performed without direct contact between the measuring probe and the soil. Radiometric gauges are relatively expensive but also fairly durable; spare parts and maintenance service are important requisites.

4.3.2.1. Counting rate statistics

The number of pulses obtained with a detector in a certain time interval is a random quantity. When one repeats exactly the same measurement, the number of pulses obtained will vary. Let us assume that the activity of the radiating sources is constant; the random quantity, number of pulses is then Poisson-distributed. The number $N$ of pulses for a certain measurement time deviates from its mean value by more than $2\sqrt{N}$, considering a probability of 5%. The counting rate $R = \frac{N}{t}$ for the same probability has a value outside of error limits given by the expression:

$$\Delta R = \pm 2\sqrt{\frac{R}{t}}$$

where

$t = T$, when a scaler is used; $T$ is the time of the measurement, and $t = 2T$ when a rate meter is used; $T$ is the time constant of the meter.

4.3.2.2. The neutron method

The neutron method is based on the exceptional ability of the hydrogen nucleus to slow down fast neutrons. Other elements existing in the soil merely hinder the outward diffusion of neutrons. Thus the density of the slow-neutron cloud around the fast neutron source mainly depends on the concentration of hydrogen nuclei. In many soils water is the only significant hydrogen compound, so that the concentration of slow neutrons that arise around the fast neutron source will be a function of the water content of the soil.

The radiation shield or some other hydrogenous body can serve as a standard medium on the basis of which a reference counting rate is measured. The ratio of the counting rate in the soil to the counting rate in the standard (or less preferably the counting rate in the soil alone) is related to the water content of soil by means of calibration. See Technical Reports from IAEA 1968-71 and Danfors & Skoglund 1971. Figure 4:7 illustrates a gauge (probe and indication unit) for subsurface measurement, and Figure 4:8 some calibration functions.
Figure 4:7 A typical subsurface gauge for neutron or gamma measurement using a scintillation detector. Some particle tracks are shown.

Figure 4:8 Measured counting rate versus moisture content curves (solid curves and experimental points) for three different Danish soils as compared with the gauge manufacturer's calibration curve (broken curve), from Ølgaard 1969.

Figure 4:9 A typical surface probe for neutron measurement of moisture using ³He- or ¹⁰BF₃ gas tube detectors. (The probe can include a high voltage supply and more electronic operations). Some neutron tracks are shown.
4.3.2.3. Different types of neutron gauges

For many hydrological investigations measurements of water content profiles in soil deposits are required. The subsurface neutron moisture gauge (Figure 4:7) is a reliable device for this purpose. Its disadvantages are poor resolution (4.3.2.8), difficulties of making measurements near the soil surface (4.3.2.9), deformation of the soil matrix and influence on the water profile caused by the access tube installation (4.3.2.7), and its relatively high price. The surface neutron gauge (Figure 4:9) is devised to be set directly on the surface of the soil.

Both the subsurface and surface neutron gauges have the property that the influence of the moisture content on gauge readings diminishes rapidly with distance from the probe, Kristensen 1971.

In one type of gauge the distance between a strong-activity source and a neutron detector is over 30 cm. The source and detector can then be placed in separate tubes, see IAEA Technical Report No 126.

4.3.2.4. Neutron gauge components

Thorough tests of performance have been made on most of the commercially available commercial gauges, see IAEA Technical Report No 130. Danfors & Skoglund 1971 have described the design features of subsurface meters and their use in the Nordic countries.

The californium-252 spontaneous fission source, which, in spite of its short half life (2.6 y), is an important future neutron source for moisture measurements because of its low mean energy (2.35 MeV), and low (and decreasing) price, Corey et al 1970. Tritium-target deuterium bombarded neutron (14 MeV) generators are adequate for large-volume single or dual-tube measurements, IAEA Technical Report No 126. A non-operating generator poses no great hazard.

A proportional counter filled with helium-3 or boron-10-trifluoride gas, or a scintillation counter, are mostly used for detection. The Geiger-counter is cheapest, but the detector using a Geiger-counter can be too ineffective, IAEA Technical Report No 130. When the detector of thermal neutrons is completely covered by a thin cadmium foil, only epithermal neutrons are detected.
In a good gauge the repeated number of pulses (e.g. in the standard) varies according to the Poisson distribution (4.3.2.1), Danfors & Skoglund 1971, Kristensen 1971; in a weak gauge it varies more.

The effect of temperature on the electronics of the gauge is significant in many devices and the instabilities of some devices under extreme temperature conditions can make the measurements futile, IAEA Technical Report No 130. Repeated counting may reveal such instabilities.

In recent years relatively light and small electronic units have been offered for sale. High voltage supply and many electronic operations have also been incorporated in the probe to improve field use, Danfors & Skoglund 1971.

The shield can serve as a standard, or another standard sufficiently large can, to insure the appropriate temperature conditions, be inserted in the soil near the measuring site.

4.3.2.5. Soil matrix effects

For mineral soils especially, the following measurement aspects can be distinguished:

1) Since the effect of hydrogen in the soil matrix is similar to that of hydrogen in water, the hydrogen content of the dry soil matrix should be determined in the soil profile. This will improve calibration precision.

2) In many soils the effect of other elements, apart from hydrogen, can be summarized as the effect of dry density (Kasi 1971). Since the soil matrix can often be considered as stable, one measurement of the dry density profile combined with some control measurements is all that is required. A gamma gauge is generally used for the density measurement (see 4.3.2.11). The dry density is determined by making simultaneous measurements of density by gamma radiation and of moisture by the neutron method.

3) Absorbers of thermal neutrons, elements such as boron, gadolin-
ium, samarium, etc. can strongly affect the measurement of thermal neutrons, Ølgaard 1969, Kasi 1971. Their effect is eliminated in epithermal measurements, IAEA Technical Report No 112, Kasi 1971.

All kinds of moisture measurements are affected by the heterogeneity of the soil medium (variety of matrix, stones, holes, etc).

4.3.2.6. Measurement of moisture changes

For the measurement of moisture changes the effects of factors such as bulk density, matrix hydrogen content, etc. can be neglected and it becomes sufficient for the calibration curve to have the form:

\[ w = f(R) + B \]  \hspace{1cm} (1)

provided that the disturbances only appear in the value of the constant B, \( f(R) \) is a monotone function; \( w \) is the moisture content, and \( R \) is the counting rate. The range in which the calibration functions obey (1) has not been very thoroughly investigated. The form (1) may be violated especially by the occurrence of low mineral densities or high organic matter densities in the soil matrix, as can be the case when measuring peat and snow (Figure 4:8), IAEA Technical Report No 91, Kasi 1973, or by the occurrence of large concentrations of elements that strongly absorb thermal neutrons, Ølgaard 1969.

4.3.2.7. Access tube, probe lowering

For full instrumental sensitivity and accuracy, the access tube diameter and soil disturbance from the tube installation must be reduced to a minimum. The counting rate varies inversely with the diameter of the access tube and in cases in which the tube diameter is much wider than probe diameter the counting rate is also influenced by the position of the probe in the tube due to possible variability in the measurement geometry. Thin-walled aluminium tubes are ideal, but unfortunately lack sufficient strength and durability in many situations.

Stainless steel tubes and galvanized iron tubes are commonly used. (Corrosion of the iron tubes may be extensive and hence obstructive.) They are inserted into the soil by simultaneous augering, or knocked
into the soil with a vibration hammer machine. IAEA Technical Report No 126, Myhre et al 1969. The Danish manufacturer recommends the use of 44.5/42.0 mm aluminium tubes with a conical metal head, which are driven into an augered hole in the soil by means of a steel rod that fits into the access tube with a sleeve support at the open end. A motorized portable auger is used for making the hole prior to tube installation. Distortion effects on soil structure are probably greatest for the vibration hammer method, but on the other hand this method is fairly reproducible and the compression, etc. caused may not be so serious in the case of studies involving changes in soil moisture.

Some difficulties can be encountered in the field in defining the correct depth of measurement, partly due to the lack of distinctness in the reference plane (the soil surface has a complex microtopography) and partly due to the difficulties involved in adjusting the correct probe depth. The important point to remember here is not to overlook the reproducibility of the depth adjustment. A definite reference level should be assigned to each measurement station.

4.3.2.8. Measured volume and resolution ability

Haahr & Ølgaard (IAEA Technical Report No 112) have introduced the concept of the sphere of importance, which signifies the volume of soil required to give 95% of the counting rate value obtained if the same soil is present in an infinite amount around the probe. The radius of the sphere of importance of the subsurface probe ranges from about 110 cm at zero water content to about 20 cm for 0.50 g H₂O/cm³. The large measurement volume of the neutron probe naturally reduces its power of resolution - i.e. its ability to resolve changes in water content with depth. Experience has shown that layers thinner than 30 cm can hardly be distinguished correctly and that water content profiles cannot be obtained with a better accuracy than that given by measurements at intervals of 15 cm, IAEA Report No 112. However, although the water content profile obtained with the neutron probe is very much smoothed out, the integrated water content over the entire profile is fairly correctly indicated.
For surface measurement, the depth corresponding to the radius of the sphere of importance in the subsurface measurement is shorter than the latter because of neutron leakage into the air. IAEA Report No 112.

4.3.2.9. Measurements near the soil surface

For measurements of moisture just under the soil surface the surface probe has been developed. Measurements with the subsurface probe in the top 0-30 cm layer involves many difficulties due to the interference of the atmosphere. A reflector can be used to reduce this interference, IAEA Report No 112. Also a special calibration function can be derived for the uppermost soil layers, Kristensen 1971; in some cases soil sampling may be used in the top 0-20 cm.

Both surface and subsurface neutron measurement have a rather poor resolution ability. A gamma method is presented in 4.3.2.13, which can achieve a resolution of 0.5 cm.

4.3.2.10. Calibration procedures

Factors which affect the counting rate v. water content relationship are as earlier discussed due to certain matrix effects, the nature of the access tube and the probe itself.

The best method to evaluate measurements from mineral soils is to use calibration curves for non-hydrogen soil. The water equivalent of the hydrogen content in the soil matrix is determined and subtracted from the measurement result to give the right moisture content, IAEA Report No 112, Kasi 1971.

Special calibrations are required, especially for organic soil matter, (e.g. humus, peat, etc) and for snow, Figure 4:8, Kasi 1973.

Commonly, users of the neutron probe arrive at adequate calibration functions on the basis of three experimental types of information:

a) the manufacturer's calibration curve(s)
b) their own laboratory measurements in drums with known materials
c) their own field calibration, which involves soil sampling.
Computer programmes have been developed to give theoretically correct calibration functions for different soils, IAEA Report No 112, Kasi 1973. The input data for these theoretical calibrations include elemental composition, bulk density, probe characteristics, and access tube dimensions and material. However, some experimental results are needed for the absolute values of the counting rates.

A relatively good establishment of an experimental calibration function is then necessary, whilst computer programmes can be used to present and to set down for consideration the effects of hydrogen content, bulk density, absorbing elements, etc. More details on the calibration procedures are found in IAEA Report No 112.

4.3.2.11. The gamma-ray method

Nuclear gamma radiation has long been used for measurement of bulk density and for density correction in the neutron moisture measurement. However, in many soils the bulk density changes are due to moisture changes, so that density gauges can be used for direct moisture measurement. The most usual gamma sources are given in Table 4:1.

Table 4:1. Some sources of gamma radiation.

<table>
<thead>
<tr>
<th>Isotope</th>
<th>Main energies of gamma-quanta</th>
<th>Half life (years)</th>
<th>Coefficient L ( \frac{\text{rem m}^2}{\text{h Ci}} )</th>
</tr>
</thead>
<tbody>
<tr>
<td>60(^{Co})</td>
<td>1.17, 1.33 MeV</td>
<td>5.26</td>
<td>1.35</td>
</tr>
<tr>
<td>137(^{Cs})</td>
<td>662 keV</td>
<td>30</td>
<td>0.33</td>
</tr>
<tr>
<td>226(^{Ra})</td>
<td>1620</td>
<td>0.83</td>
<td></td>
</tr>
<tr>
<td>241(^{Am})</td>
<td>60 keV</td>
<td>458</td>
<td>0.0925</td>
</tr>
</tbody>
</table>

The detector is generally a Geiger-tube or a scintillator.
4.3.2.12. The conventional type of single-well density gauge

This type of gauge has the same principle construction as the neutron gauge, Figure 4:7. It is not used much for moisture measurement, but is used more for determining the density correction in connection with neutron measurement of moisture content. This measurement has approximately the same resolution ability as the neutron measurement of moisture. 4.3.2.8. The effect of the access tube on the calibration curves for mineral soils is presented in Figure 4:10. The character of the gamma measurement is very similar to that of neutron measurement in the determination of peat moisture, Kasi 1973.

4.3.2.13. Dual-tube arrangement

An interesting and important device for measurement of water content profiles is a special type of meter based on the gamma-ray attenuation between two tubes, de Vries 1969, Giesel et al. 1970, Figure 4:11. Cesium-137 is a suitable emitter of gamma radiation. The lower discriminating voltage level (L.D.L. in Figure 4:12) for pulses from the scintillation detector must be just below the voltage level of the $^{137}$Cs single gamma-peak at the energy of 662 keV. Only gamma-rays are counted which have not been scattered or been very little changed in their direction via scattering. The collimators in Figure 4:11 are not always necessary, but they can improve the resolution ability, especially near the soil surface, de Vries 1969.

The counting rate produced by gamma radiation is given by:

$$R = (R_o e^{\sigma u_o x} e^{-w u_w x})$$

where \(\sigma\) is the dry bulk density, \(w\) is the water content in g/cm$^3$ and \(u_o\) and \(u_w\) are the corresponding mass attenuation coefficients dependent on the energies of gamma-rays and on the collimation-counting system used. \(R_o\) is the radiation intensity when there is no medium between the tubes. \(x\) is the distance between the tube walls. In order to obtain a good accuracy of measurement, \(x\) should be selected according to:

$$1 \leq (\sigma u_o + w u_w)x \leq 3$$
The values of $u_p$ and $u_w$ are around 0.08 cm$^2$/g for gamma radiation of $^{137}$Cs.

The accurate measurement of the water content profile presupposes that $\varphi$ remains constant. Giesel et al. 1970 achieved an accuracy of approximately 0.0015 g/cm$^3$ by using a strong source of 50 mCi. A resolution of the order of 0.5 cm can be achieved. The tubes must be installed in the soil as parallel as possible and the source and the detector must be lowered steadily to exactly the same depth.

The serious source of errors in this measurement is the fluctuation and drift of the pulse amplitude caused by temperature variations, etc., Giesel et al. 1970. This effect can be controlled with a standard measurement procedure, in which temperature conditions in the standard is made to correspond to those of the soil measurement. Reference measurements can also be made below the water table. Standard measurements slightly increase the approximate upper limit, 3, in formula (3). The temperature effects can also be corrected by means of an extra source of light or gamma radiation being used to directly affect the photocathode or the scintillation crystal, thereby achieving electronic compensation.

4.3.2.14. Measurements near the surface

The gamma-ray device described in 4.3.2.13. can be used for measurements just below the surface of the soil, especially when collimators are used. Figure 4:11. Another commercial transmission device has the source of gamma-rays at the end of a thin rod inserted in the soil and the detector placed at the surface of the soil. The rod can be inserted into the soil to different depths. Measurements can be made without much disturbance to the soil.

4.3.2.15. Safety

IAEA Reports No 112 and 126 include considerations of radiation safety. The dose is the dose rate multiplied by the time of exposure. IAEA Safety series No 9 gives maximum permissible doses for professional workers.
Figure 4:10 Some typical calibration curves of gamma backscatter as affected by the density of a mineral soil and by use of different aluminium and iron tubes.

Figure 4:11 Dual-tube gamma transmission measurement. Field gamma beam positioning system and geometry of source, detector and their collimators with respect to undisturbed soil, from de Vries 1969.

Figure 4:12 Pulse height spectrum of $^{137}$Cs, from de Vries 1969.
The gamma dose rate obtained in air from a point source is given by:

\[ D_u = \frac{LA}{r^2} \]

where

- \( D_u \) = gamma dose rate, \((\text{rem} = 10^3 \text{mrem} = \text{mr})\)
- \( A \) = activity of gamma radiation source
- \( r \) = distance from the source
- \( L \) = coefficient for the isotope emitting gamma-rays, Table 4:1

The neutron dose rate \( D_n \) (mrem/h) from an isotope point source in air is correspondingly obtained from:

\[ D_n = 1.15 \times 10^{-6} \frac{S}{r^2} \]  \( (5) \)

where \( S \) is the source strength (n/s), and \( r \) is the distance from the source (in units m).

The gamma and neutron doses from an Am-Be source are rather equal in air. The dose rate can be considerably diminished with a material shield, such as lead or other metal shielding against gamma radiation, and a hydrogenous material such as paraffin, wax, etc., against neutron radiation. As seen in the formulae above, a large distance from the source acts as a shield.

The manufacturer should give information on the maximum dose rate and the dose rate distribution on the surface of the meter, and of the attenuation of the dose rate. The operator of a gauge should be most careful when handling the probe, even when it is situated in the shield. The hazard is generally much below the permissible doses, when operations are carried out in a proper way.

4.3.3. Other methods

For certain purposes, penetration and chemical methods have been used to establish the water content of soil layers (Cope & Trickett 1965, Geary 1971). Many of these are rapid, single-structure methods, but their accuracy is fairly low and they have not been used extensively in hydrological studies.
4.4. Potential of Soil Water

4.4.1. Tensiometric Methods

A tensiometer consists of a water-filled porous cup buried in the soil and connected to a manometer in a vacuum gauge, Figure 4.13. The water in the cup reaches pressure equilibrium with the soil, and this becomes a measure of the matric suction. Before equilibrium is reached, water in the tensiometer flows out into the surrounding soil and thus the column of mercury in the attached manometer sinks. The drier the soil, the stronger the resulting pressure reduction. With equilibrium, water no longer flows away from the tensiometer and the suction remains constant. Tensiometers can be used to measure a maximum suction corresponding to one atmosphere. If this value is exceeded air will enter the porous cup and the instrument can no longer be used. The tensiometer cannot detect osmotic potential as the cup is permeable to solutes. For most soils in the Nordic region osmotic potential can be neglected.

Fine-textured soils may clog the porous cup, but this can be delayed by setting the tensiometer in sand. Diurnal variations in temperature will affect exposed uninsulated parts. This can be remedied in different ways, e.g. by using plastic non-conducting materials, by installing the instruments completely below the soil surface or by reading each instrument at the same time each day.

Tensiometers are cheap and easy to construct, calibrate, and install and they permit adequate replication. Matric suction, which is a most useful measurement in field experiments, is recorded directly in situ and requires no mathematical transformations. The translation of matric suction into soil moisture content involves individual calibration for each soil and is not recommended.

4.4.2. Electrical Methods

Electrical resistance units, see Figure 4.14, are used to measure the soil moisture status in situ. (Gardner 1965, Cope & Trickett 1965, Aslyng 1968, WMO 1968, Geary 1971). The resistance of absorbent blocks that are in moisture equilibrium with the soil is a
Figure 4:13  Principle of two types of soil moisture tensiometers, from Aslyng 1968

Figure 4:14  Electrical resistance method
function of the moisture content of the block, and can be used as a measure of also soil suction by means of suitable calibration.

The resistance of the units must be measured with an a.c. Wheatstone bridge (about 1000 cycles/second) because electrolysis and polarization occur if direct current is used. The resistance units are made of various materials, but usually consist of two carefully spaced electrodes surrounded by water-absorbent material. The materials most widely used are gypsum, nylon and fibreglass, either alone or in various combinations. Such units will measure suctions within the pF-range of 2 to 4.2. The measuring range of gypsum blocks alone, however, is considerably smaller, see Figure 4:15.

All resistance blocks must have very low geometric tolerance values and printed circuits have therefore been used; even so, individual calibration is necessary for each unit. Calibration for each soil type is also necessary and may be done in the field or in the laboratory. It is usually on the drying curve that blocks display marked hysteresis, and the resulting curves do not necessarily become stable with time. Changes due to possible dissolution of the calcium sulphate (applies to gypsum blocks) cannot be checked once the block is installed. However, this does give some degree of buffering against the effect of soluble salts - a major disadvantage of the method. Many blocks are temperature-sensitive, and corrections must be made either manually or from built-in thermocouples.

The resistance block method has a number of advantages for in situ measurements. Blocks are cheap, and a number can be installed in the field and read at intervals without disturbance to the site. The leads may be led to a central point, and the method lends itself to automatic recording. Blocks and measuring bridges for field use are commercially available. However, the method is still lacking in reliability and precision and will require considerable refinement in block-manufacture and in calibration techniques, before it receives wide acceptance.
Figure 4.5: Possible uses for the tensiometer and certain
lysimetric blocks in different moisture ranges.
For soil moisture determinations, the change in the capacitance of a condenser unit that is buried in the soil has also been tried (Gardner 1965, Cope & Trickett 1965, Geary 1971). Multiple-sensing heads are needed with the capacitance method, as with the resistance block. The former is more expensive, since screened cables of stable shunt capacitance are needed. Temperature and hysteresis effects as well as power loss that occurs in a wet medium add to the complications involved.

A micro-wave meter, comprised of an oscillator, receiver, and a calibrated attenuator, is commercially available for determining the moisture content of samples (Cope & Trickett 1965, Geary 1971). The sample is placed between the receiving and the transmitting antennae. The advantage of the method lies in the speed of the reading. The heterogeneous nature of soils is the biggest limitation to in situ use of the micro- or radio-wave methods. One method based on the infrared absorption of water has the disadvantage that the penetration of the rays in the soil is shallow, but it is useful especially for aerial measurements of large tracts of bare soil.

4.4.3 Thermal methods

Attempts have been made to exploit the thermal conductivity of the soil or that of moisture absorbent thermal blocks in close contact with the soil to determine soil moisture in situ (Cope & Trickett 1965, Geary 1971). Thermal conductivity is independent of salt content and this gives the method an advantage over the resistance block method, particularly for saline soils. Direct measurements depend on very close contact between the soil and the probe, which is hard to achieve.

The heat source is an electrically heated wire in a needle-shaped probe 13 cm long. A thermo-junction near the centre of the heated wire gives a continuous measurement of the temperature change. Under laboratory conditions, soil samples can be heated by a 300-watt lamp mounted above the samples. Measurements are made with a metal plate on top of the soil and thermo-couples located at various, relatively shallow depths.
The temperature dependence of resistance has also been used, e.g., one arm of a wheatstone bridge is made of enamelled copper wire wound around glass tubing that is set in the soil to follow moisture changes in time. The other three bridge resistance units are made of manganin (temperature independent). In one method a mercury thermometer with half the bulb wound round with electrically heated wire is set in the soil. The time required to attain a constant temperature rise is said to be dependent on the soil moisture content.

4.4. Methods used in the laboratory

When soil water potential is measured in laboratory studies, pressure methods are usually used, most often the air-pressure membrane apparatus, but sometimes also vapour-pressure methods. These methods can be used to measure the uptake and release of moisture from samples of soil over a wide range of suction values. In addition temperature measurements of freezing-point depression have been used for determination of soil water potential (Aslyng 1968).

4.5. Conclusions

In determining the soil moisture status, the goal is to get as undisturbed a sample as possible and also to keep the soil at the measuring spot as undisturbed as possible. In hydrology, especially, non-disturbance is considered important, because the soil water content is usually stated as volume percent water. Undisturbed samples can usually best be obtained from sampling holes or with various kinds of piston drills. When undisturbed samples cannot be obtained other sampling methods can be used, which means that the water content usually has to be expressed as percentage weights. Different kinds of sampling methods are treated more in detail in soil sampling guidebooks.

All the present methods of soil moisture measurement have different ranges of applicability.

1) Measurement of water content per volume unit

Gravimetric determination of moisture is the natural choice for small research tasks and for calibration of other methods. However, for large scale projects it often becomes more rational to invest in meth-
ods with greater measurement capacities, such as the neutron method. The neutron method is recommended for long term in situ measurements. The neutron method is the best nuclear method for mineral soils, but for pure peats gamma methods may be as applicable. With the dual tube gamma method a very good depth resolution is obtained and therefore also soil layers just below the soil surface can be measured as well. The single tube probe may be used for measurements below 15 cm depth.

2) Measurements of soil water potential

The measurement of soil water potential in the wet range will often supplement, but not replace, that of water content measurements. For in situ measurements the tensiometer, with adequate replication to overcome its lack of precision, is usually considered best suited for the purpose of measuring water potential.

Resistance and capacitance blocks provide rather imprecise measurements over a wide range, providing the site is not too wet or saline. Blocks are cheap, but not long-lived, and must be well replicated. These methods are also applicable for in situ measurements over the dry range in the topsoil. Figure 4:15 shows the possible uses of the tensiometer and certain gypsum blocks in different soil moisture ranges.
5. DESIGN OF MEASUREMENTS AND PROCESSING OF DATA

5.1 Purpose of soil water studies within the IHD-programmes

According to resolutions and suggestions made by the Co-ordinating Council of the UNESCO IHD programme, soil water studies are considered important in many phases of the IHD. Studies on this reservoir parameter are expected to be concentrated within representative and experimental basins with objectives:

- to establish laws governing increases and decreases in soil water
- to determine its influence on changes in the water balance
- to produce data necessary for calculating runoff due to thaw and rain water
- to evaluate economic aspects of water available for crops and for establishing irrigation and drainage requirements, etc.

Suggestions have also been made concerning supplementary observations that should be made where applicable, such as: depth of water table, precipitation, freezing and thawing in the soil, evaporation, water equivalent of snow packs, type and stage of development of vegetation. These observations should be made at the same time and at the same location as routine soil water measurements are made for water balance purposes. Soil water values should be expressed in a form compatible with other hydrological data on precipitation, evaporation, etc., that is in mm or inches depth of water. Furthermore it is suggested that also the following soil hydrological properties be determined for each depth increment of soil that is studied:

- minimum moisture capacity (field capacity)
- capillary moisture capacity (the soil water content at various distances above the water table to the top of the capillary fringe. This will vary with changing depth of the water table in the soil)
- maximum water capacity (soil water content at different depths when the water table is at the ground surface)
- wilting point
- bulk density or volume weight of soil.
Recommendations regarding soil water measurements during the IHD are thus comprehensive and ambitious. Data existing before the IHD is generally scarce, of short term character and usually inadequate for calculations of the water storage of even fairly limited areas. The extremely high variability of soil water has created an awareness of the need for extensive sampling and processing techniques to be used during the IHD. Proper design of measurements to satisfy adequate coverage in time and space has also been recognized as important.

5.2 Design and assessment of soil water networks in Norden

The purpose of the soil water networks in the Nordic representative basins is essentially twofold - water balance studies and research on more specified processes. Here only the former objective will be discussed. It is concerned with quantitative estimates of the soil water changes within well defined catchment areas. The measurement techniques and network designs that have been evolved have been reviewed by Danfors 1972, see also figure 5:1 and table 5:1.

The design of the networks has hardly been scientific, but rather pragmatic in nature. Desire for geographical spread, measurement accessibility, labour economy, administrative convenience, etc have been the main factors governing the network development. In many cases the stations have been limited to one or two small sub-basins within a representative basin. The station density is generally sparse, with but 1-15 stations per 100 square kilometers, see table 5:1.

The neutron moisture gauge was early recognized as the basic equipment for data acquisition in representative basins. One man can handle this equipment and scan an access tube to a depth of 2-3 meters in less than 30 minutes. Taking into account also the transport and equipment handling time that is involved it becomes clear that one man can cover only about 10-20 tubes per day, depending on the density and accessibility of the access tubes across the basin. Since water balance calculations are recommended on a monthly basis, measurements should preferably be made on the very last day of the month and at any rate no longer than a day or so on either side of the transition between
<table>
<thead>
<tr>
<th>Code</th>
<th>Basin name</th>
<th>Area in km²</th>
<th>Total number of stations</th>
<th>Network density stations/km²</th>
<th>Average profile depth cm</th>
<th>Year of installation</th>
<th>Measurement intensity Time interval, days</th>
<th>Depth interval, cm</th>
<th>Network design based on</th>
</tr>
</thead>
<tbody>
<tr>
<td>Kp</td>
<td>Karup (D)</td>
<td>514 km²</td>
<td>7</td>
<td>0.01</td>
<td>870</td>
<td>1966 and 1968</td>
<td>30</td>
<td>20</td>
<td>Topographic-geometric spacing</td>
</tr>
<tr>
<td>St</td>
<td>Stevna (D)</td>
<td>295 km²</td>
<td>6</td>
<td>0.02</td>
<td>150</td>
<td>1968</td>
<td>7 summer 30 winter</td>
<td></td>
<td>- &quot; &quot;</td>
</tr>
<tr>
<td>Ps</td>
<td>Pajärvi (F)</td>
<td>244 km²</td>
<td>4</td>
<td>0.02</td>
<td>200</td>
<td>1969-70</td>
<td>30</td>
<td>10</td>
<td></td>
</tr>
<tr>
<td>Fi</td>
<td>Filefjell (N)</td>
<td>155 km²</td>
<td>3</td>
<td>0.02</td>
<td>100</td>
<td>1968-70</td>
<td>irregular</td>
<td>10-20</td>
<td>Transport aspects and soil conditions</td>
</tr>
<tr>
<td>Ro</td>
<td>Romerike (N)</td>
<td>200 km²</td>
<td>8</td>
<td>0.04</td>
<td>225</td>
<td>1967-71</td>
<td>7-30</td>
<td>20-25</td>
<td>Soil, topography and vegetation aspects, certain combinations with agricultural experiments</td>
</tr>
<tr>
<td>La</td>
<td>Lappträsket (S)</td>
<td>1008 km²</td>
<td>11</td>
<td>0.01</td>
<td>175</td>
<td>1968</td>
<td>30 summer 7 winter</td>
<td>10</td>
<td>Transport and soil aspects</td>
</tr>
<tr>
<td>Ks</td>
<td>Kassjön (S)</td>
<td>160 km²</td>
<td>10</td>
<td>0.06</td>
<td>155</td>
<td>1968</td>
<td>14 for 5 tubs irregular in other tub.</td>
<td>10</td>
<td>- &quot; &quot;</td>
</tr>
<tr>
<td>Ve</td>
<td>Velen (S)</td>
<td>65 km²</td>
<td>7</td>
<td>0.15</td>
<td>170</td>
<td>1966-67 and 1971</td>
<td>7 for 3 tub.</td>
<td>10</td>
<td>- &quot; &quot;</td>
</tr>
<tr>
<td>Va</td>
<td>Verkån (S)</td>
<td>117 km²</td>
<td>8</td>
<td>0.07</td>
<td>300</td>
<td>1967-69</td>
<td>30</td>
<td>10</td>
<td>Topographic, soil and vegetation aspects, limited coverage</td>
</tr>
<tr>
<td>Va</td>
<td>Väringe (S)</td>
<td>3 km²</td>
<td>4</td>
<td>1.23</td>
<td>250</td>
<td>1970</td>
<td>14-21</td>
<td>10-20</td>
<td>Central location</td>
</tr>
</tbody>
</table>
Most representative basins lack the resources to cope with more than 1-3 man days per month to collect soil water data and so the networks become naturally maximised to 10-40 tubes. The need for replications will then further reduce the number of actual sites (zones) involved to some 5-20. This is the explanation for the inherently low density of the Nordic networks presented in table 5:1.

The most common neutron moisture gauge in use in Norden is the Danish BASC meter, the manufacturer of which has established certain maintenance facilities for the benefit of INDO-scientists (gauge substitutes in cases of emergency, individualised calibration procedures, etc), see figure 6:1. Primary measurement values are obtained as time readings corresponding to the accumulation of a preset count number (10-30 000 counts). Shallowest readings are made at 10 cm below the surface for which special corrections are made. Consideration is otherwise taken to the bulk density and in certain cases also to the chemical composition of the soil when readings are evaluated. Calibrations are made either theoretically on the basis of soil analyses (the manufacturer has applied the model of Øgaard 1967) or by field and laboratory trials using gravimetric techniques to establish the "true" water content. Bulk density determination is made by means of soil sampling or by means of a special gamma probe unit that is attached to the end of the neutron probe. The most common type of access tube used is the 1 1/2 inch galvanised iron tube. These tubes are made water tight at the bottom end by means of a pointed steel cap. The tubes are installed by hammering with or without preaugering of the soil.

A typical soil water station consists of 1-3 access tubes located within a radius of less than 20 meters. The total depth of the tubes varies between 1-3 meters, the measurement interval between 10-20 cm in depth and 7-30 days in time. The establishment of the stations has generally occurred 1967/68, which means that records of 7-8 years length exist at present (1975). Auxiliary background information concerning the geology texture, porosity, water retention, groundwater and plant root depth characteristics is rather lacking, but in process of investigation.
In an endeavour to test the precision of soil water data collected from Swedish representative basins, Eriksson 1970 later followed by Bergström 1971 and Rydén 1974 carried out analysis of variance. The analysis was made primarily on data from the Velen basin and involved monthly values of the soil water changes from six to seven stations that are fairly evenly distributed across the basin. The soil depth chosen for analysis was 100 cm, the number of intervals for each year was about 10 with rather wide differences in length (from 7-40 days). The total area of the basin is 45 km², the soil and vegetation cover at the chosen stations is glacial till respectively coniferous forest, and broadly speaking the basin may be regarded as fairly uniform (for location see figure 5:1).

The statistical model used assumes that there is no interaction between any special station and time interval. The standard deviation may be divided into three parts - one dependent on the variance between stations, another on the variance between intervals and the third on residual variance. Here the residual variance may be interpreted as expressing the uncertainty that is attached to a single soil water change value. Results for 1968 are given in table 5:2.


<table>
<thead>
<tr>
<th>Sources of variance</th>
<th>Sum of squares</th>
<th>Degrees of freedom</th>
<th>Variance</th>
</tr>
</thead>
<tbody>
<tr>
<td>Between stations</td>
<td>144.5</td>
<td>6</td>
<td>24.1</td>
</tr>
<tr>
<td>Between intervals</td>
<td>39,089.1</td>
<td>10</td>
<td>3,908.9</td>
</tr>
<tr>
<td>Residual</td>
<td>20,817.8</td>
<td>50</td>
<td>416.4</td>
</tr>
<tr>
<td>Total</td>
<td>60,048.4</td>
<td>66</td>
<td>909.8</td>
</tr>
</tbody>
</table>

The results show that the variance between stations is much smaller than that of the residual, indicating that the stations react differently with regard to water changes during the year even though the mean reaction is fairly constant. The variance between intervals is high, indicating significant differences between these. Bergström was able
Figure 5.1. Soil water networks in Norden. Symbols:

- Kp - abbreviated name of basin (here Karup, see Table 5.1)
- 7 = number of stations

Legend:
- Soil water station symbol

ICELAND
SWEDEN
NORWAY
FINLAND
PA
KS
VA
RO
FI
VE
ST
KP

Table 5.1: Soil water networks in Norden.
to reduce this variance by omitting intervals of extreme length in a
renewed analysis, indicating that the choice of interval length is a
critical factor. The uncertainty attached to a single water change
estimate from a particular station can be found by taking the square
root of the variance of the residual (i.e. \(\sqrt{416.4} = 20.4\) mm).

The uncertainty involved in an estimate of the mean soil water change
for all seven stations can then similarly be found to be \(20.4 \times 7 = 77\) mm.
If the number of stations is doubled (i.e. to 14 stations), the error will
be \(7.7 = \sqrt{5.4}\) mm. If it is desired to reduce the error of monthly
estimates to around 2 mm, which essentially applies to rainfall
estimates, then the number of stations would have to be increased to
satisfy the equation: \(20.4 \times n = 2\), that is to \(n = 104\) stations. Such
a large increase in station number is hardly possible or practicable,
see figure 5:2.

The application of analysis of variation to small series of soil water
data is questionable since it assumes complete randomness of soil
water conditions. It does, however, throw some light on the problems
of judging representativeness of individual values and of optimizing
the design of networks and measurement programmes.

5.3 Discussion of network design procedures
It is apparent that soil water is a most difficult storage term to evaluate.
Not only is it highly dispersed in a pore complex, but also its bound-
daries to ground and surface water are diffuse and continuously
changing. Its very existence is in fact dependent on the extent of these
other two storage entities. Thus, during wet periods, when the ground
water table rises, the zone of soil water will decrease in extent,
whilst in dry periods it will increase, quite oppositely to what is
expected from a water balance point of view. Thus soil water is not an
independent but an open residual type of storage term. Hysteresis
caused by the irregularities of the porous system of this water zone,
will furthermore make every distribution of soil water an unique
and irreproducible one. Such conditions obviously inflict special
measurement and data handling problems. It is evident from what has
earlier been said concerning the measurement capacity of one man
Figure 5.2 Error in mean soil-water change estimation in the Velen basin as a function of the confidence limits and number of stations (n) used. From Bergström 1971.
using the neutron moisture gauge, that soil water inventories must be carefully planned, if collected data is to have validity for water balance calculations.

The international literature on procedures for soil water studies in catchment hydrology is very limited and general. Radda 1969 has given a review of the status of hydrological networks with considerations regarding possible improvements in procedures. Kutilek 1971 has proposed a scheme for soil water assessment based on an ecological classification of soil water domains in a catchment. The representivity of measured point values is determined through the interpretation of aerial photos combined with certain field controls. The method is believed to make possible satisfactory areal assessments of soil water storage from a limited number of point values. No test results are presented, however. WMO 1972 gives a review on work concerning hydrological network design practice, but the treatment of soil water involves here but one paper, that by Kvesel 1972, which concerns the error of distributed point values as a function of the number of measurement points.

If the basin that is studied is completely uniform with a homogeneous soil and vegetation cover, then statistical randomization of measurement sites will obviously give the best design and best check of variability through ordinary analysis of variance. If, however, the basin is heterogeneous, which is commonly the case, then some form of systematic siting of measurement points will undoubtedly provide a better design and produce a better estimate of mean areal storage. Precision will be more difficult to judge, however, since analysis of variance will tend to underestimate the error. In the case where the basin consists of a mixture of fairly homogeneous strata, a combination of systematic and random station siting will be expected to give the most appropriate design of the measurement network.

Important features to consider in stratifying a basin into more homogeneous soil water regimes are altitude, slope, soil type, soil depth, vegetation (land use) and drainage. For each landscape type it becomes necessary to find a suitable ranking system to emphasize the relative importance of the various factors involved. Gustafsson 1968 has demon-
strated how topography affects drainage flow for completely saturated conditions and how a drainage basin can be separated into inflow and outflow areas on a topographic basis. Holtan et al. 1970 in developing a watershed model also differentiate between uplands, hillsides and bottomlands. These categories form elevation and hydrologic sequences, which influence the dynamics of the basin, see figure 5:3. Following up these thoughts it would seem expedient to base soil water inventories on a division of the basin into inflow, outflow and transition areas.

The **inflow areas** are located on the uplands, where the soil cover in Norden due to glaciation is relatively thin and coarse textured. The vegetation is generally forest with well developed root systems. The drainage potential is high as also is evaporation due to the favorable height and exposure location. Soil water conditions will here be characterized by rapid and wide changes in concentration which will necessitate intense measurement input to achieve the same precision in assessment as for the **outflow areas**, which are more stable and homogeneous. The outflow areas are located in the bottom lands, where the soil cover is deeper and much finer textured. The vegetation here

![Figure 5:3](image-url)

Figure 5:3 Drainage basin characteristics with emphasis on runoff dynamics. In the upland zone there is throughflow, on the hillsides interflow and ephemeral streamflow and in the bottom lands perennial streamflow. From Gregory and Walling 1973.
generally consists of farm crops which have exposure and drainage conditions that are poorer than in inflow areas. The pattern of lateral flow of these areas is one of convergence as against divergent type of flow imposed by inflow areas. Soil water conditions tend to be much more stable and homogeneous due to the homogeneity of the soils and the lesser influence of drainage and evaporation. This leads to reduced measurement requirements and a basis for a more randomized type of network design pattern.

The transition areas are located along the hillsides where drainage is favourable and the soils are variable in character due to greater erosion susceptibility, see figure 5:4.

Skoglund & Danfors 1971 have studied the variability of soil water distribution in a basin in Central Sweden and found seasonal and spatial variations that rather well fit the kind of zoning just depicted, see table 5:3.

Figure 5:4 Diagrammatic representation of flow areas of a drainage basin. $P$ is precipitation, $E$ evapotranspiration and $R$ runoff or interflow.
Table 5.3: Measured characteristics of flow zones at Ultuna basin, Sweden. From Skoglund & Danfors 1971

<table>
<thead>
<tr>
<th>Measurement site</th>
<th>Type of flow area</th>
<th>Height above sea level m</th>
<th>Depth to groundwater m</th>
<th>Vegetation</th>
<th>Soil texture</th>
<th>Soil water characteristics of upper 100 cm</th>
<th>Field capacity mm</th>
<th>Wilting point mm</th>
<th>Max. water content change (range), mm</th>
</tr>
</thead>
<tbody>
<tr>
<td>Ultuna outflow</td>
<td>bottom-land</td>
<td>10</td>
<td>&lt; 3</td>
<td>farm crops</td>
<td>heavy clay</td>
<td>Porosity</td>
<td>1050</td>
<td>840</td>
<td>600, 65</td>
</tr>
<tr>
<td>Ulleråker transition</td>
<td>hillside</td>
<td>20</td>
<td>&gt; 2.0</td>
<td>coniferous forest</td>
<td>clay with sand lenses</td>
<td>Field capacity</td>
<td>950</td>
<td>740</td>
<td>420, 130</td>
</tr>
<tr>
<td>Hammarby inflow</td>
<td>upland</td>
<td>25</td>
<td>&gt; 3</td>
<td>deciduous forest</td>
<td>sand-loam</td>
<td>Wilting point</td>
<td>890</td>
<td>700</td>
<td>280, 270</td>
</tr>
</tbody>
</table>
The flow type zones just described constitutes a basis for designing networks. The influences of atmospheric processes, particularly precipitation, show a much greater variation in time than do those due to groundwater flow. The territorial variations in soil water content therefore decrease, as one goes from inflow via transition to outflow areas. To receive the same amount of information in all areas, the station density should therefore decrease in accordance with the sequence stated.

A basin may be primarily differentiated into the three flow areas described by means of air photo interpretation of the topography and vegetation cover depicted. After some field checks the boundaries of the zones should be fairly well established. Figure 5:5 shows an idealized drainage basin divided into the 3 different flow type areas. Outflow areas exist in the valleys occasionally in saddle points and depressions of the undulating landscape. Transition areas are widespread along the hillside slopes, over high saddle type plains and in smoothly inclined forested areas. Inflow areas are found along the ridges, over open high plateaus and over the mountain slopes of weathered rocks.

Figure 5:5: Pattern of inflow, transition and outflow areas in an idealized drainage basin.
Having completed this first order type of division of the basin, the next step is to consider the usefulness of further sub-divisions based on suitable lower order parameters. One such parameter is that suggested by Kutilak 1971. It is based on various duration characteristics of defined moisture levels in the soil (wet, moist and dry states etc., see table 3:5). The strata thus obtained have ecological significance and can be identified through interpretation of the vegetation and soil from suitably taken aerial photographs. Unfortunately this scheme presupposes a considerable knowledge of the biohydrology of the region, which is often lacking in Norden, see Appendix 2.

In conclusion it should be clear that some type of bio-geo-hydrological zoning is necessary for proper distribution of measurement points. Having achieved this, a stratified network is designed to give a basic coverage of the basin. More detailed and more precise data can then possibly be achieved by introducing auxiliary networks, which are managed on a rotation basis, whereby labour economy can still be achieved. As experience is gained of the complexity of the basin, the network is improved through suitable adjustments. The aim is to find the interrelationships which exist between natural landscape parameters and soil water so that representivity of stations can be found and thereby correct interpolations made between point values for calculations of valid mean areal values.

5.4 Automatic processing of soil water data
A remarkable development of techniques and methods for the collection and processing of data has occurred in the last decade or so in most fields of science. As concerns soil water hydrology, various types of radiometric techniques (neutron scattering, gamma absorption, infra red radiation, etc) have created very sophisticated ways of collecting mass information concerning soil water conditions in time and space and under both natural and artificial conditions. Mechanical and electronic devices for data processing are also in a state of accelerated development, whereby new yet unimagined possibilities of handling data are being created. Examples of such new developments in progress are the automatic weather station and the autoprobe that have
been constructed at the Institute of Hydrology, Britain. Their usefulness is especially forthcoming in remote and difficultly accessible areas.

The trend is towards achievement of a comprehensive integrated hydrological measurement and recording system, where not only water content of the soil, but also many other related events (precipitation, moisture flux, etc) are measured simultaneously. The data may be stored in various ways - punch cards, paper tape, magnetic tape or microfilm depending on facilities and conditions at hand. The processing involves both routine and long-term developmental endeavours, where basic values are plotted in various ways and checks are made of precision, variabilities in time and space, interrelationships with other factors, etc. see Institute of Hydrology 1967-74.

A computer programme for processing raw data from soil water measurements with the neutron gauge was first presented in Norden by Danfors & Thunvik 1969. The programme formed a basis for later modifications in the other Nordic countries. The programme transforms primary gauge readings into water values expressed in volume percentages and millimeters and displays these in tables and figures showing distributions in time and space (chrono- and topoisopleths, vertical and horizontal profiles, etc), see Thunvik et al 1972.

Johansson 1974 has developed a model for estimating and displaying the soil water regime in various agricultural systems - cultivated fields with different crops and farming practices. The model is based on standard meteorological observations and is computerized to give symbol lines for each growing season as shown in figure 5-6. The year is divided into 5 day periods (pentads) and three horizontal dashes implies non-vegetation period, a star (asterix) means that at least 3 of the 5 days in the pentad have soil water contents above field capacity, a zero sign is similarly used to show that at least 3 days of the pentad have been below the wilting point and a single dash implies that the water content has been between wilting point and 33.3 % of the field capacity. No sign means that the pentad has had soil water conditions between 33.3 % of field capacity and full saturation. Similar displays
could be applied for characterizing more specifically water balance conditions based on hydrological significance.

Modelling work is developing rapidly. The effort is towards expressing watershed hydrology as a continuum and to finding a predictable pattern of physical probabilities that will account for the dispersion of water and its subsequent collection in aquifers and channel systems. A foundation will then be laid for safe and proper hydrological manipulations in fields of engineering and community planning, see Holman & Lopes 1970.

Figure 5:6 Soil water balance for a clay soil at Uppsala during vegetation season. Upper part shows degree of saturation by means of symbols whereby years can be compared. Lower part gives an example of what the symbols imply as regards plant root availability of the soil water.
6. SOIL WATER STUDIES IN THE NORDIC REGION

6.1 Hydrologic cooperation between the Nordic countries

Already at the start of the IHD-programme it became natural that the 5 Nordic countries, Denmark, Finland, Iceland, Norway and Sweden established cooperation. A joint Nordic coordinating panel, called NUTSAM, was set up to link together the IHD-work in Norden. The panel was composed of the secretaries of the 5 national IHD-committees and was given the task of planning Nordic hydrologic meetings of coordinating activities of the Nordic working groups that were set up in various facets of the IHD-programme and to promote policies of all Nordic interest. It was considered expedient to treat Norden as a natural region especially as regards the global aspects of the IHD-programme and to pool available resources for obtaining optimum information and research results.

Among the Nordic working groups that were established, one was designated to soil water (NAG 2). Its mandate was to handle IHD problems concerned with the unsaturated zone and to elucidate the general hydrologic role of soil water. Its primary tasks were to find the most convenient techniques to obtain representative values of the areal monthly means of soil water changes in representative basins and to study the possibilities of establishing a Nordic soil water network for regional coverage. The group was composed of two soil water specialists from each country and has met regularly about once a year. The activities of the group during 1966-73 may be summarized as follows:

1. Recommendations regarding soil water measurements in representative basins have been formulated in the reports from the working group. These recommendations include various particulars regarding the use of the neutron method such as choice of access tube material, method of installation of access tubes, number and distance between replicate tubes, procedure of measurement with the depth meter, measurement intensity with respect to time and depth, method of reporting and processing of the primary data, various ways of checking and calibrating the neutron gauge etc.
Since the BAJC combined moisture/density gauge is the most commonly used equipment for soil water measurements in representative basins in Norden, a special service agreement has been negotiated with the Danish manufacturer to guarantee measurement functions in cases of emergency.

2. Exchange of information and ideas have been promoted through special meetings and seminars - a course on soil water measurement was arranged in Denmark 1970, a symposium on soil water was held in Norway 1971 and a field meeting on special problems of soil water observations in representative basins was organized in Finland 1971. Through questionnaires the development and status of soil water studies in the Nordic countries has been reviewed. A list of about 350 soil water terms has been compiled for inclusion in the Nordic Glossary on Hydrology.

3. The hydrologic significance of soil water has received much attention. According to an agreement made at the start of the IHD work, as concerns water balance studies in representative basins, Denmark was to concentrate on the hydrology of arable land, Norway on mountainous land, Finland and Sweden on forests and Iceland was to deal with glaciers. This diversification in the IHD effort has helped broaden the regional outlook, but also led to certain difficulties in collaboration on specific problems and goals of the working group. Thus the group has had difficulties in establishing a regional soil water network and in making up general directions for evaluating areal monthly means of soil water changes. On the other hand the group's promotion of soil water measurements in the representative basins has led to very valuable and unique long term series of data information, now successively being published in the Nordic publication, "Hydrological Data - Norden". Prior to the IHD, there existed very few such records of soil water measurement.

6.2 Soil water studies in the Nordic countries

A short summary will now follow on the separate activities and developments in soil water hydrology of the 5 Nordic countries as a consequence of the IHD-programme. The presentation will be of a very general nature and the IHD projects in progress in Norden will be dealt with only in a very concise tabular form at the end of the chapter, see table 6.1.
Nevertheless it is hoped that the presentation will suffice to give an adequate state of the art impression and that the reader will excuse eventual emissions and simplifications, keeping in mind the difficulties involved in properly appraising scientific achievements in reviews of the kind attempted here.

6.2.1 Denmark

Prominence in agriculture probably made Denmark the best prepared of the Nordic countries to take on soil water studies within the IHD-programme in 1965. Water balance studies of the type intended by UNESCO were in many respects already in progress here and much effort had already been made to investigate the hydrological characteristics of various agricultural regions of the country.

What the IHD-programme essentially has achieved in Denmark is better coordination of hydrological endeavours and a broader outlook on water problems of more general type - urban as well as rural. Many different organisations have become engaged in IHD-work and taken on measurement and service tasks to fulfil set goals. Special efforts have been made to establish and to concentrate studies to two representative basins - Karup and Stevne. Other noticeable developments of interest for soil water hydrology are those that have taken place in hydrochemistry and geohydrology. Thus routine chemical analyses are now being made in many areas on rainfall and on surface, soil and ground waters. Roughly 100 groundwater measuring sites have been set up in each of the representative basins and at some of these also soil water is determined to considerable depths. Radiometric and tracer type techniques have been developed and here mention should be made of the BASC combined moisture and density gauge, which is the most widely used gauge in the Nordic region for registering soil water changes. see Figure 6.1. The Danish Isotope Center, which has developed the BASC gauge has also contributed with many important developments in the application of tracer techniques to hydrological problems.

Denmark is endowed with a fairly medium and less variable rainfall than the other Nordic countries. The average runoff is 5-13 l/s and km². The interaction between surface water and groundwater is complicated and a key problem in Danish hydrology. The Danish Meteorological Institute-
Figure 6:1 The BASC Depth Moisture Gauge, designed and manufactured in Denmark and used extensively for soil water measurements in the Nordic representative basins. The equipment consists of a scaler (1), a neutron moisture probe (2) which is housed in a shield and depth measurement device (3), a gamma density extension unit (4) with special housing (5). The complete neutron-gamma (moisture-density) probe is seen at the centre (6). Photo by courtesy of the manufacturer, Nea-Lindberg A/S, Denmark.
is responsible for the national network of precipitation stations, the Geological Survey for groundwater studies and the Danish Health Society maintains the national network of streamflow stations. Routine soil water measurements are carried out by the Danish State Experiments and Research in Plant Culture, by the Hydrotechnical Laboratory of the Veterinary and Agricultural College and by the Danish Geological Survey. The IHD-programme has led to a coordinated effort between the above named organisations in establishing the 2 representative basins - Karup and Stevns, where Karup represents the extramarginal areas of West Jutland and Stevns a moraine plain landscape typical of eastern Denmark.

Amongst research activities of interest in soil water hydrology, mention should be made of the study in progress at the Agricultural Experimental Station at St. Jyndevad in southern Denmark on actual evapotranspiration and the interaction between soil moisture status and crop growth. The studies are correlated with observations on precipitation, air temperature, wind velocity and radiation, making possible extrapolations to other areas through meteorological comparisons. Similar studies are also being made by the Hydrotechnical Laboratory at Højbakkegård near Copenhagen, where there is available an advanced scheme of lysimeters and evapotranspirometers. As regards evaporation studies these have attained a far more advanced state of development in Denmark than in any of the other Nordic countries.

Other studies which have aroused interest in Norden are the geohydrological studies made by the Geological Survey in Karup, where the depth of the unsaturated zone in some places exceeds 20 m and is composed of fairly homogeneous sandy soil. Here it has been found possible to evaluate and separate the velocities of wetting fronts and individual water molecules by means of combined neutron scattering and tritium content measurements. The wetting front has been found to move at the rate of 2-3 meters per month whereas corresponding water molecules have been found to move at only about 2-3 meters per year. Thus where the depth of the unsaturated zone is around 20 meters there is a lag time between the annual precipitation maximum and the groundwater level maximum of 10-12 months, while for depths of the unsaturated zone around 1-2 meters the corresponding lag time is but one month in the investigated area.
IHD-work in Finland began as a spontaneous voluntary effort between different organisations involved in water problems. Although stimulating in the start, the work was soon found to be seriously handicapped by lack of external funds to fulfil ambitions in accordance with the IHD-programme. The number of projects that could be initiated were therefore fewer than in other Nordic countries.

Only one representative basin, Pääjärvi, has been set up and here soil water measurements are made in only 3 places to a depth of 2 m and with a measurement interval of 30 days. The stations were installed in 1969/70 and data records from 1970 are reported in the Nordic publication on Hydrological Data. The measurements are made by the Hydrological Office.

In the latter part of the decade some reorganisations have taken place. Newly reorganized is the National Board of Waters, which among other activities has commenced an intensified regional study of groundwater resources and recharge conditions throughout Finland. Soil water stations are being set up in coordination with these investigations and very soon Finland will have a rather extensive network of hydrological base stations with potentialities of resolving important elements of the water cycle in different hydrological districts.

Soil water investigations are being made by 3 institutions within the IHD-programme, namely Department of Peatland Forestry of the University of Helsinki, Department of Civil Engineering of the Technical University at Otaniemi and the Hydrological Office. Also agricultural scientists have carried out soil water studies concerning soil water retention characteristics and uptake of water by different crops. The Dept of Peatland Forestry has been engaged in interesting studies on the basic hydrological properties of peat soils and the interaction of these with forest growth and drainage techniques. The Finnish research on peat soils has a leading position in Norden and considering the large proportion of organic soils covering this region, it is understandable that developments in Finland are closely followed and highly esteemed. The research programme of the Dept of Peatland Forestry also includes studies on evapotranspiration...
from woodlands growing on peat. The measurements are based on a special method that has been developed and which involves diurnal variations in groundwater level as a means of evaluating evaporation losses. Efforts are also being made to develop adequate methods of characterizing and differentiating organic soils. See figure 6.2.

Figure 6.2 Rectilinear correlation between depth of groundwater table and water content of different peat layers in drained forest swamps in Finland. A is a top layer (0-15 cm), B is a middle layer (15-25 cm) and C a bottom layer (25-40 cm) sphagnum peat, where the degree of humification of the peat increases with depth. From Heikurainen et al 1964.
The Departments of Civil Engineering and Technical Physics of Helsinki University of Technology have been engaged in refinement of basic instruments in soil physics for sampling, for density, water retention and water flow measurements. Basic studies have been made on the principles and functions of the neutron and gamma methods for determining density and moisture of soils. A statistical study of regional character has also been performed on the effect of weather factors on water economy and crop yields. Variations in crop growth have been correlated with various available expressions for depicting air temperature, humidity, rainfall, cloudiness, hours of sunshine and radiation. It has not been until the last decade or so and more especially during the IHD, that regular observations have been made in Finland on soil temperature and soil water content and on pertinent quantities governing evaporation. Much of this is rapidly changing now, due to the present rapid development of hydrology in Finland. A substantial investment has recently been made in the construction of a modern lysimeter field, see figure 6:3.

6.2.3 Iceland

Long term soil water measurements were first initiated in Iceland in 1964 when the Meteorological Institute established an agro-meteorological survey. Four measuring sites were set up, of which three were located at state research stations. The measurements are based on sampling of soil cores at 10 cm levels to a depth of 60 cm. The water content is determined by the gravimetric method. The work is limited to the non-frost period of the year.

Iceland has only recently set up a representative basin (Ellidaá near Reykjavik) and therefore lacks records of soil water measurements in this connection. Interesting results of regional character have been obtained, however, from deuterium studies. Thus by comparing deuterium contents of "young" water from snow and rain with "older" waters from springs and glaciers, it has been possible to elucidate the origin of water in various types of groundwaters. For example water from a borehole near Reykjavik has been found to originate from a glacier 100 km away (Langjökull glacier). The technique also allows estimations of the annual snow precipitation that is stored in the glaciers and will also be of significance for studies of the soil water regime.
Figure 6.3  New lysimeter field at the Helsinki University of Technology. The field consist of 12 concrete lysimeters arranged in 2 rows, see lower cross section. The surface area of each lysimeter is 10 m² and the depth 1.5 m. Five different soils are tested, some lysimeters have grass vegetation, others are bare. The surface and groundwater runoff and the soil temperatures at 6 levels are registered automatically as also are meteorological elements such as precipitation, wind speed and direction, air temperature and humidity, etc. From Kaitera 1970.
6.2.4 Norway

According to an early Nordic IHD agreement, Norway was given responsibility for water balance studies in mountainous areas. Such areas afford very special problems of study, quite different from those confronting the other Nordic countries, and this has given Norway a considerable challenge within the IHD-programme. Three representative basins have been organized - Romerike, Filefjell and Sagelva. Soil water is studied only in the former two basins.

The average runoff for Norway as a whole is about 40 l/s, while in some westerly regions the average runoff may amount to 5 times this amount, depending on rainfall of about 6000 mm/year. The groundwater resources are small. Studies of surface water, river, ice and glaciers are made by the Norwegian Water Resources and Electricity Board, whereas major geohydrological activities are carried out by the Geological Survey of Norway.

Responsibility for soil water measurements in Romerike and Filefjell is held by the Department of Agricultural Hydrotechnique of the Agricultural College of Norway. Measurements were started in 1968 using the Danish neutron gauge with intervals ranging from 7-30 days in Romerike and more irregularly and limited to the summer season in Filefjell. Results from two areas within and one area just outside Romerike are reported in Hydrological Data - Norden. In contrast to most other representative basins in Norden, application of tensiometer measurements have been made and the Department of Agricultural Hydrotechnique has designed its own equipment. Results from the studies are in progress of analyses.

Norway excels in snow pack studies and has gained much insight into thawing processes and runoff phenomena. Also erosion studies are lively. In one area in Romerike studies are being made of the chemical changes occurring during infiltration and percolation of rain water through the unsaturated zone. A considerable change has been found to occur in the chemical composition of the rain water in the upper 20 cm layer with a very evident exclusion of anions and cations at about 10 cm depth. The pH of the rain is just below 5 and remains low in the upper soil layer. Below 20 cm depth the pH reaches 6 and the chemical composition remains
Figure 6:4 Moisture oscillations in the surface layer (a) of a forest soil as a consequence of high temperature changes in an open clear cut forest area in Norway. The vegetation cover is lichens and mosses which have a small water consumption from the humus (a) and mineral soil (b). The arrows show the direction of water diffusion into and out of layer a, which has a thickness of about 20 mm and maintains a water content around 15%. From Bjor 1965.
and climate of the area. A special study has been started to determine the effective porosity of till, using radiometric and direct measurement techniques.

At the Dept of Land Improvement and Drainage of the Institute of Technology, Stockholm, systematic studies have been made on the use of the neutron gauge for soil water assessment. A computer programme has been developed to handle field data. This programme has been adopted by some other neutron gauge users in Norden and provides transformation of count rate values into equivalent water contents taking into account particulars regarding the differences in soil type and soil density. The results are displayed in a series of tables and graphs, which give information on soil water distribution with time and in space and on statistical parameters such as means, standard deviations, maximums and minimums, etc. Studies have also been made on the geomorphological dependence of soil water contents and changes. These studies have permitted stratification of hydrological areas into more defined soil water regimes.

The Swedish Meteorological and Hydrological Institute carries out soil water measurements in three representative basins, Velen, Kassjön and Lapptrasket and has applied the data to monthly water budget calculations. Results show that soil water changes occur very abruptly especially during thawing in spring and that they can exceed precipitation values manifold for certain drier months of the year. There is generally much difficulty in deriving reliable mean monthly areal changes due to the large heterogeneity of the soil cover and its environment. Statistical analyses of the present network has shown that the error in the areal mean estimation of the monthly soil water changes is around 10 mm, which is too high for a satisfactory evaluation of the actual monthly evaporation.

For the top meter of soil cover the annual amplitude in soil water content is about 50 mm for a normal year, the field capacity is about 290 mm, the wilting point 150 mm, effective porosity 2-5 % and the groundwater level about 1 meter in depth. Some improvement in the reduction of the variability of soil water values has been obtained by expressing water contents
W as saturation deficits so called R-values, where

\[ R = \frac{W - W_{\text{wp}}}{W_{\text{fc}} - W_{\text{wp}}} \]

and \( W_{\text{wp}} \) is the wilting point and \( W_{\text{fc}} \) the field capacity content. Estimation of mean areal values can therefore be improved by using R-values instead of W as a basis for estimation.

A study has been made at the Soils Dept of the College of Agriculture regarding the movement and storage of soil water in the root zone. Tritium has been used as a tracer in irrigation water applied in various amounts to plots about 6 m² area and located in various types of land (pine blueberry forest, deciduous forest, arable land, in fallow and in grass pasture). The average rate of downward movement of percolating water (wetting front) was 1-3 mm per day giving a groundwater recharge corresponding to 0.3-0.7 mm per day. The dissipation of applied tritium water could be divided into a biologically governed phase, involving evapotranspiration, and a hydrophysical phase, involving percolation and groundwater recharge. Losses of water through evaporation and transpiration were very rapid immediately after irrigation and the general trend in the dynamics of the biological phase was one of exponential decay with a duration of about 100 days. The hydrophysical phase was governed by the intensity of irrigation. Results have yet to be published.

Considerable basic work has also been done at the Soils Department on the physical structure of arable soils. These investigations involve soils from all parts of Sweden and most physical parameters. Various instruments have been designed to permit routine measurements of aggregate size distribution, water retention characteristics, water, air and heat movement, etc. Also chemical studies have been made on rain, soil and drainage waters in certain agricultural regions to give information on pollution effects and on the mobility and transformation of various nutrients and elements of interest in both agriculture and nature conservation, see figure 6:5.
Figure 6.5  Water retention curves for Swedish arable soils with clay contents from 0-43% clay (numbered 1-10 rain). From Andersson & Wiklert 1972.
Table 6.1 IHD-projects in soil water hydrology 1965-74

1. **BASIC DATA**

   **SOIL MOISTURE STATIONS (I)**
   Icelandic Meteorological Service

   **SOIL MOISTURE STATIONS (S)**
   Swedish Meteorological and Hydrological Institute

   Regular measurements of soil water at 2 locations, Reykjavik in south-western and Akureyri in northern Iceland

   A network is being set up based on specimen stations within IHD representative basins

2. **REPRESENTATIVE AND EXPERIMENTAL BASINS**

   **SOIL WATER INVENTORIES**
   Karup (D) Geologic Survey of Denmark and Hydrotechnical Laboratory of Royal Veterinary and Agricultural College

   Stevns (D) Hydrotechnical Laboratory of Royal Veterinary and Agricultural College

   Päijärvi (F) Hydrological Office

   Filefjell (N) Inst. of Agricultural Hydrotechnical Institute of Agricultural College

   Romerike (N) Dept of Land Improvement and Drainage of Institute of Technology

   Lapplång (S) Swedish Meteorological and Hydrological Institute

   Kassjö (S) Dept of Hydraulics, Lund Institute of Technology

3. **RESEARCH PROJECTS**

   **RUNOFF FROM SMALL BASINS (F)**
   National Board of Waters

   Study at several small basins of water balance with regard to meteorological, hydrological and physiographic factors of the basins
INfiltration and Water Balance in Small Catchments of Varying Slope (N)

Inst. of Agriculture Hydrotechnique of Agricultural College, Norwegian Water Resources and Electricity Board

Measurement of Actual Evapotranspiration from Various Farm Corps (D)

St. Jynderad station of the Danish State Experiments and Research in Plant Culture

Energy Exchange Between Soil and Atmosphere (F)

Finnish Meteorological Inst.

Lysimetric Measurements of Evapotranspiration (F)

Lab of Water Resources Engineering of Helsinki University of Technology, Otanemi

Evaporation of Tree Stands (F)

Dept of Forestry of University of Helsinki

Movement of Moisture in the Unsaturated Zone (D)

Geological Survey of Denmark

Variations of Soil Moisture in Small Basins (F)

Hydrological Office

Shelterbelts and Water Balance in Arid Region (N)

Inst. of Agricultural Hydrotechnique of Agricultural College, Norwegian Meteorological Inst. Agricultural Research Council

Water balance studies in 2 small areas within Romarvik's representative basin with special emphasis on infiltration and erosion as affected by slope, soil type and vegetation cover

Actual evapotranspiration is studied by following changes in soil water content of the root zone of various farm crops with a neutron gauge. Correlations are made with meteorological factors such as temperature, saturation deficit and wind speed of the air and radiation parameters

Micrometeorological observations are made of various parameters to determine evaporation based on energy and vapour flux calculations

Lysimeter studies to determine effect of different soil types and groundwater levels on water balance of surface layer of soil

Study of evapotranspiration of different tree stands based on method of diurnal fluctuation of groundwater level

Study of soil water changes using neutron gauge in 22 m deep tubes at Karup representative basin. Velocity of wetting front is traced in relation to amount of precipitation. Information on recharge to groundwater and loss through evapotranspiration is anticipated

Monthly measurements of soil water with neutron method yield possibilities of studying water balance of the unsaturated zone in limited areas

Effects of shelterbelts on temperature and water balance of soil in an arid region, Leeja. Also study of plant reactions to soil water and temperature variations
CHEMICAL CHANGE OF SOIL MOISTURE (N)
Geological Survey of Norway

VERTICAL MOVEMENT AND STORAGE OF SOIL MOISTURE (S)
Dept of Pedology of Agricultural College

RELATIONSHIP BETWEEN SOIL MOISTURE AND TEXTURE (S)
Dept of Land Improvement and Drainage of Inst. of Technology

RUNOFF FROM SMALL BASINS (S)
Dept of Agricultural Hydrotechnics of Agricultural College

WATER QUALITY IN AGRICULTURAL DISTRICTS (N)
Inst. of Agricultural Hydrotechnique of Agricultural College

THE EFFECT OF TREE STANDS ON SOIL MOISTURE AND GROUND WATER TABLE (F)
Dept of Peatland Forestry of University of Helsinki

OPTIMAL DRAINAGE OF PEAT LANDS (F)
Dept of Peatland Forestry of University of Helsinki

STUDIES ON THE EFFECT OF HYDROMETEOROLOGICAL FACTORS ON CROP YIELDS (F)
Lab of Water Resources Engineering of Helsinki University of Technology, Otaniemi

THE EFFECT OF URBANIZATION ON WATER BALANCE (S)
Division of Hydraulics of Inst. of Technology, Lund

Study of the chemical changes in percolating water through soil to groundwater at Romerike representative basin

Movement and dissipation of percolating water through different soil-plant environments is studied using tritium as a tracer

Study based on statistical analyses of long term measurements of soil water in different soils, of the effect of soil texture on the retention and exchange of water in the unsaturated zone

Study in small basins to get scientific basis for the design of drainage schemes. Study involves considerations to topographic soil vegetation and climatic conditions

Quantitative and qualitative assessment of nutrients in drainage water from different agricultural districts

Study of different kinds of silvicultural treatments on soil and groundwater in drained peat lands.

Relationship is studied between ground water level and water storage and tree growth of peat soils

Statistical analyses of certain meteorological observations to ascertain response of crops to different moisture conditions

Study of changes occurring in a small virgin area as a consequence of successive urbanization. Soil water is measured with a neutron gauge
MOVEMENT OF WATER IN THE UNSATURATED ZONE, GRØNHØS, JUTLAND (D)
Geological Survey of Denmark
Danish Isotope Centre

NEUTRON METHOD FOR SOIL WATER ASSESSMENT (F)
Dept of Technical Physics of Technical University, Otaniemi

NEUTRON TECHNIQUES FOR DETERMINATION OF WATER CONTENT OF SOILS (S)
Dept of Land Improvement and Drainage, Institute of Technology, Stockholm

SOIL WATER BALANCE AT ULVSUNDA, STOCKHOLM (S)
Swedish Meteorological and Hydrological Inst.

HYDROGEOLOGICAL STUDIES IN TILL (S)
Geological Survey of Sweden

SOIL WATER DISTRIBUTION AS A FUNCTION OF LANDSCAPE STRUCTURE (S)
Dept of Land Improvement and Drainage, Institute of Technology, Stockholm.

STUDY OF THE PRECIPITATION - GROUNDWATER RESPONSE IN AQUIFERS OF VARIOUS GEOLOGIC COMPOSITION (S)
Geologic Survey of Sweden

INFILTRATION AND PERCOLATION OF WATER IN SOILS (S)
Hydrological Division, University of Uppsala

STUDIES ON SOIL WATER AND GROUNDWATER (S)
Swedish Meteorological and Hydrological Institute

Tritium profiles of water in the unsaturated zone to 22 m depth have been determined with intervals of 2 years. The results give indications of the rate of recharge and evapotranspiration.

Basic principles and application techniques underlying the neutron method for determining water in porous media are studied.

Development of suitable routines for areal determination of water storage in the unsaturated zone using the neutron method.

Water balance studies through intermittent determinations of soil water using the neutron method.

Effective porosity of till has been studied at Läpptrasket and Velen representative basins. Direct gravimetric and indirect gamma-neutron measurements have been used.

Relationship between variability in soil water and landscape attributes such as geology, topography, vegetation, etc are studied.

Purpose is to characterize the precipitation - groundwater relationship for different geological features.

Existing information on infiltration and percolation in different soils in Norden has been compiled for purposes of evaluating surface runoff.

A detailed study of the infiltration and percolation processes under natural conditions in a subbasin of Kassjöån representative basin.
APPENDIX I

TEXTURAL CLASSIFICATION OF SOILS

There is at the present no international agreement regarding the textural classification of soils. Probably the most well-known classification system is the American system (USDA Soil Survey Manual, handbook No 18), which has also been recommended by the FAO for use in developing countries (Guidelines for Soil profile description - FAO publ. 1970). It will be presented here together with systems in use in the Nordic countries as a reference.

The textural classification is based on the relative proportions of various particle size fractions that make up the soil. These fractions have been defined in various ways as shown in Table A:1 below. The most systematic division of particle size fractions is probably that of the International Society of Soil Science (ISSS) also called the Atterberg system. For many purposes it is too detailed and in the more applied sciences, e.g. soil mechanics (geotechnical field) it has been simplified. The US Department of Agriculture (USDA) has a separate system. It is important when comparing soil textures to know according to which system they have been defined and named.

1. Systems used in the Nordic countries

All the Nordic countries have acknowledged the ISSS-system for particle size classification. The classification used for soil texture varies, however.

In Denmark a strict system of textural classification has not been established. The nomenclature most commonly used is closely related to the USDA system with a three fraction division. The word "muld" is used here as a translation of the American term "loam".

Sweden and Norway have essentially the same system, which is based on the content of clay, while Finland has a separate system. The relationships between the systems used in Norden to denote the texture of soils is shown below:

<table>
<thead>
<tr>
<th>Country</th>
<th>Clay content</th>
<th>Soil name</th>
</tr>
</thead>
<tbody>
<tr>
<td>Denmark</td>
<td>below ca 25 %</td>
<td>sands, silts and loams</td>
</tr>
<tr>
<td></td>
<td>ca 25 - 40 %</td>
<td>clay loams</td>
</tr>
<tr>
<td></td>
<td>above ca 40 %</td>
<td>clays</td>
</tr>
<tr>
<td>Finland</td>
<td>below 30 %</td>
<td>sands, fine sands and silts</td>
</tr>
<tr>
<td></td>
<td>30 - 60 %</td>
<td>fine sand, clays and silty clays</td>
</tr>
<tr>
<td></td>
<td>above 60 %</td>
<td>very fine (heavy) clays</td>
</tr>
<tr>
<td>Norway and</td>
<td>below 5 %</td>
<td>non-clayey soils</td>
</tr>
<tr>
<td>Sweden</td>
<td>5 - 25 %</td>
<td>coarse textured clays</td>
</tr>
<tr>
<td></td>
<td>above 25 %</td>
<td>fine textured clays</td>
</tr>
</tbody>
</table>

For a more detailed subdivision of the soil textural types used in Norway and Sweden see Table 3:2.
<table>
<thead>
<tr>
<th>System</th>
<th>0.005</th>
<th>0.05</th>
<th>0.1</th>
<th>0.25</th>
<th>1.0</th>
<th>2.0</th>
<th>10</th>
</tr>
</thead>
<tbody>
<tr>
<td>ISSS</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Clay</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Silt</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Fine sand</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Coarse sand</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Gravel</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Stones Pebbles</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Boulders</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>DENMARK</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Ler</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Silt</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Fine sand</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Grov sand</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Soil Science</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>SWEDEN</td>
<td>Mjälla</td>
<td>Mo</td>
<td>Sand</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Geotechnical</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>GEOTECHNICAL</td>
<td>Silt</td>
<td>Sand</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Geological</td>
<td>Siltti</td>
<td>Hiekka</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>FINLAND</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Geotechnical</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Geological</td>
<td>Sora</td>
<td>Kivet</td>
<td>Lohkarsett</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>General</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>NORWAY</td>
<td>Leir</td>
<td>Silt</td>
<td>Sand</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Soil Science</td>
<td>Grov Leir</td>
<td>Fine sand</td>
<td>Grov sand</td>
<td>Grus</td>
<td>Stein</td>
<td>Blokker</td>
<td></td>
</tr>
<tr>
<td>US DEPT OF AGRICULTURE</td>
<td>Clay</td>
<td>Silt</td>
<td>Very fine</td>
<td>Fine sand</td>
<td>Medium</td>
<td>Coarse sand</td>
<td>Very coarse</td>
</tr>
</tbody>
</table>

Table 1:1. Comparison between different systems of classifying soil particle fractions
2. **The American system (USDA-system)**

The American nomenclature adopted to express the texture (i.e., the relative proportions of sand, silt, and clay) of a soil is based on a subdivision by means of a triangular diagram. The class names consist basically of the terms sand, silt, clay, and loam used either as nouns or adjectives or both, see 1:2.

![American soil textural diagram](image)

**Figure 1:2. American soil textural diagram**

a. **Loamy sand**: Loamy coarse sand—25% or more very coarse and coarse sand, and less than 50% any other one grade of sand. Loamy sand—25% or more very coarse, coarse, and medium, and less than 50% fine or very fine sand. Loamy fine sand—50% or more fine sand (or) less than 25% very coarse, coarse, and medium sand and less than 25% very fine sand. Loamy very fine sand—50% or more very fine sand.

b. **Sands**: Coarse sand—25% or more very coarse and coarse sand, and less than 50% of any other material or sand separate. Sand—25% or more of very coarse, coarse, and medium sand, and less than 50% of fine or very fine sand. Fine sand—50% or more of fine sand or less than 25% very coarse, coarse, and medium sand and less than 50% very fine sand. Very fine sand—50% or more very fine sand.

c. **Sandy loam**: Coarse sandy loam—25% or more very coarse and coarse sand and less than 50% of any other one grade of sand.
3. Classification of the organic fraction

A commonly used classification in Norden for the organic soil fraction is the following:

I. Mineral soils < 20 % humus
   - < 3 % humus poor soils
   - 3 - 6 % moderately humus containing soils
   - 6 - 12 % humus rich soils
   - 12 - 20 % very humus rich soils

II. Organic soils having mineral fractions
    20 - 40 % humus (Mull)

III. Organic soils essentially devoid of mineral fractions
     above 40 % humus (mainly Carex and Sphagnum peats)

4. Geological aspects

Besides distinguishing the textural composition of soils, it is common to also make distinctions as regards the geological origin in naming soils. The following geological classification is then generally followed:

I. Mineral soil types:
   - Unassorted - glacial tills (moraine soils)
     - weathered material (scree)
   - Assorted - glaciﬂuvial deposits
     - ﬂuvial deposits (post glacial)
     - shore deposits
     - deep water sediments (varved glacial clays, post glacial clays and silts)
     - wind deposits

II. Organic soil types:
    - Mull - humus
    - Mor - humus
    - Peat soils
    - Gyttja
    - Mire

(III. Chemical deposits: bog-ore, ochre, alum, kaolin, etc)

Concerning the differentiation of peat soils (organic), ﬂoristic composition and the nature of the environment are used as the basis for classification - see Heikurainen 1973
APPENDIX II

Vegetation as an Index of Environmental Factors

1. The major plant regions of Norden

Within the Nordic region five major plant belts may be differentiated, as shown in Table II:2.

Table II:2 Plant geographical regions of Norden

<table>
<thead>
<tr>
<th>Plant region</th>
<th>Latitude</th>
</tr>
</thead>
<tbody>
<tr>
<td>I Subartic and Arctic</td>
<td>Above 68</td>
</tr>
<tr>
<td>II Alpine- and Subalpine</td>
<td>59-70, 500-1000 m.a.s.l</td>
</tr>
<tr>
<td>III Boreal coniferous forest</td>
<td>60-66, excluding the Scandes</td>
</tr>
<tr>
<td>IV North European mixed forest</td>
<td>56-60</td>
</tr>
<tr>
<td>V North European deciduous forest</td>
<td>Below 58</td>
</tr>
</tbody>
</table>

North of the timber line. Heath and tundra with scattered dwarfed birch species. Includes all of Iceland and northernmost part of Fennoscandia.

The Scandes Mountain range. Altitudes above timber line. Similar to I.

Most of Finland and 2/3 of Sweden. Forest dominated by Scotch pine and Norway spruce mixed with birch. Also bogs and mires cover substantial areas of the region.

Southern 1/3 of Sweden, southern coast belt of Finland and west coast strip of Norway. Conifers mixed with deciduous species, but not beech.

All of Denmark and some narrow southern and western coastal belts of Sweden and Norway. Natural forests dominated by beech.

2. Ordination of the Boreal coniferous forest region according to forest site-types

In the Boreal coniferous forest region it has been found possible to classify the forest ecosystems according to the interrelationship between vegetation and habitat. Systems have been developed in both Finland and Sweden, where forest site-types have been identified and classified according to gradients of moisture and nutrients of
The habitat. The system developed by Arnborg 1953 for northern Sweden is shown in the following table.

**Table II:3 Forest types of northern Sweden, arranged against theoretical ordinates of moisture and nutrients (Arnborg 1953). Each type is defined by its vegetation.**

<table>
<thead>
<tr>
<th>Degree of moisture</th>
<th>Degree of nutrient supply</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Poor (Dwarf-shrub series)</td>
</tr>
<tr>
<td>Very dry</td>
<td>Very dry dwarf-shrub type</td>
</tr>
<tr>
<td>Dry</td>
<td>Dry dwarf-shrub type</td>
</tr>
<tr>
<td>Fresh</td>
<td>Fresh dwarf-shrub type</td>
</tr>
<tr>
<td>Moist</td>
<td>Moist dwarf-shrub type</td>
</tr>
<tr>
<td>Wet</td>
<td>Wet dwarf-shrub type</td>
</tr>
</tbody>
</table>

The indications used to separate the various levels of moisture and nutrition have been based on the composition of the bottom respectively the field vegetation layer of the grown forest stand. In Table A:4 the indications of site moisture conditions portrayed by the bottom (ground) layer of vegetation are given, and in Table A:5 the response characteristics of the field layer to different nutrient levels are shown.
Table II:4 Relation between moisture status of the habitat and the character of the bottom layer of the fully grown stand of Northern Swedish coniferous forest region

<table>
<thead>
<tr>
<th>Bottom layer of grown forest stand</th>
<th>Moisture status of stand site</th>
</tr>
</thead>
<tbody>
<tr>
<td>Lichens form a complete cover even under the trees</td>
<td>Very dry</td>
</tr>
<tr>
<td>Lichens form the main cover, mosses occur under the trees</td>
<td>Dry</td>
</tr>
<tr>
<td>Bottom layer is lacking or characterized by mosses other than Sphagnum</td>
<td>Fresh</td>
</tr>
<tr>
<td>Sphagnum mosses occur in scattered clusters</td>
<td>Moist</td>
</tr>
<tr>
<td>Sphagnum mosses form a complete cover</td>
<td>Wet</td>
</tr>
</tbody>
</table>

Table II:5 Relation between nutrient status of habitat and the character of field layer of the fully grown stand of Northern Swedish coniferous forest region

<table>
<thead>
<tr>
<th>Field layer of the grown forest stand</th>
<th>Nutrient status of the stand site</th>
</tr>
</thead>
<tbody>
<tr>
<td>Dwarf-shrub series</td>
<td>Poor</td>
</tr>
<tr>
<td>composed mainly of Vaccinium species and some less demanding herbs. Lastrea bracken is not present</td>
<td></td>
</tr>
<tr>
<td>Lastrea-dwarf shrub series</td>
<td>Good</td>
</tr>
<tr>
<td>The small Lastrea dryopterae bracken characterizes the field layer. Vaccinium myrtillus dominates the dwarf-shrubs and is more thriving here than in the previous series</td>
<td></td>
</tr>
<tr>
<td>Forb-dwarf shrub series</td>
<td>Rich</td>
</tr>
<tr>
<td>Dominating is the herb, Geranium silvaticum. Vaccinium myrtillus is abundant and thriving</td>
<td></td>
</tr>
<tr>
<td>Forb-series</td>
<td>Abundant</td>
</tr>
<tr>
<td>This is the most fertile site type with a dominance of herbs, high growing bracken and grasses. The dwarf-shrubs are generally set back and lacking</td>
<td></td>
</tr>
</tbody>
</table>
3. Ordination of the North European mixed and deciduous forest regions

As progression is made from the coniferous forest region to the south towards mixed and deciduous type forests, vegetation classification based on clear environmental gradients becomes more complex. Theoretically it is possible to map synecological or environmental ordinates by entering point values on a map and drawing isolines on a grid basis. However, gradient discontinuities and reversals will generally occur with such frequency that the meaningfulness of patterns obtained will be questionable. Simplified systems of ordination based on vegetation and environment characteristics can of course be utilized to broadly classify landscapes and their components. Applications to hydrological investigations have unfortunately been very lacking, but are in progress, see 3.4.5. For the southern mixed and deciduous forest regions it is common to look at the complete vegetation stand and not only the bottom and field layers. The type of synecological systems that can thus be derived is shown in Table II: 6 below from the work by Kielland-Lund.

<table>
<thead>
<tr>
<th>Table II: 6 Scandinavian forest vegetation arranged according to environmental ordinates, from Kielland-Lund</th>
</tr>
</thead>
<tbody>
<tr>
<td>Dry</td>
</tr>
<tr>
<td>---</td>
</tr>
<tr>
<td>Pine-lichen forest</td>
</tr>
<tr>
<td>Pine bog forest</td>
</tr>
</tbody>
</table>

Low fertility | High fertility

(Kielland-Lund)
As can be seen, Table II:6 does not include the beach and oak forest type ecosystems, which are common in Denmark and the southernmost part of Fennoscandia. However, even for these, classification systems based on environmental ordinates are available.

4. **Root surveys**

Root characteristics, may in the same way as branch and leaf characteristics, reveal important hydrological features of the landscape. The presence of living roots defines the volume of soil influenced by transpiration. Root distribution is affected by various environmental factors and retarded by such conditions as texture resistance (compacted layers), oxygen deficiency (e.g. high water table), temperature excesses (e.g. permafrost), poor nutrient and moisture status (e.g. coarse textured droughty soils). Root development affects soil structure through increased porosity and aggregate development and stability. The roots of forest stands have been found to increase porosity by 10-20% which implies 40-50 mm of increased water holding capacity and significant effects on infiltration and runoff.

In a special technical publication by UNESCO 1973 - "Root Survey Methods for Hydrological Experiments" - methods for root surveys have been presented. Direct field methods include:

1) Excavation techniques that permit charting the lateral and vertical extent of roots.

2) Removal of soil monoliths for extraction of root patterns by washing away the soil matrix.

3) Various kinds of core sampling techniques.

4) Dyes, trace elements and isotopes for indirect detection of root distribution.

5) Repeated soil moisture measurements for estimating the spheres of root influence.
APPENDIX III

Glossary of Soil Water Terms

An important and integral part of editing a state of the art report is to consider terminology. A list of terms with definitions generally widens the scope and usefulness of such reports very considerably. The preparation of a glossary also helps to check the quality and comprehensiveness of the material presented and is in this way a kind of measure in itself of the state of the art of the field in question.

The glossary that is here presented is based on the Nordic Glossary of Hydrology (first draft of terms published 1973) and on a careful consideration of the contents of the report and of developments by IASH regarding hydrological investigations of the unsaturated zone. The Nordic Glossary of Hydrology is a multilingual list of terms with English, Danish, Finnish, Icelandic, Norwegian and Swedish equivalents, but lacks definitions. The list has been assembled through the joint efforts of several Nordic specialist groups. Members of the IHD Nordic working group on soil water (NAG 2) formed such a group and compiled 1970 a list of about 350 English terms with Nordic language equivalents concerning soil water hydrology. The next step is now the preparation of definitions and is a tedious one. It seems uncertain at present whether such work can be completed for publication before 1976. The present glossary can in this context be considered to be an intermediate step along the way, helping to awaken interest and discussion for a more refined terminology.

In addition to definitions, the glossary here contains references to pertinent sections of the report, thereby also acting as an index of the contents.
Adhesive water - water retained in the soil against gravity as thin films covering the walls of the pores (or surfaces of the soil particles). See 3.2.2.

Aeration zone - see Unsaturated zone, 2.2

Aggregate structure - the cementing or binding together of soil particles into secondary units, aggregates or granules. Water held in aggregates, which will not disintegrate easily, are, in contrast to single grain structure, of special importance to soil structure. See Soil structure, 3.1.2.

A-horizon - see Soil horizon, Eluvial horizon, 3.2.1.

Air capacity - commonly considered to be the larger pores filled with air when the soil is at field capacity. The bulk volume of pores that are unable to hold water when subjected to tension equivalent to free drainage (50-100 cm). See Gravitational porosity, Non-capillary porosity, Effective porosity, 3.1.2.

Air-filled porosity - the proportion of the bulk volume of the soil that is filled with air. See Air capacity, 3.1.2.

Available water - the portion of water in a soil that can be absorbed by plant roots, usually considered to be that water held in the soil against a tension between field capacity and wilting point. See Field capacity, Wilting point, 3.2.2.

E-horizon - horizon affected by accumulation (illuviation) of clay, iron and organic matter. See Soil horizon, Illuvial horizon.

Bulk density (Soil) - the mass of dry soil per unit bulk volume expressed as g cm\(^{-3}\). The bulk volume is determined in the natural state before the sample is dried to constant weight at 105°C. The value is dependent on texture, water content and pressure burden. It ranges from 1.2 - 1.8 for arable soils. See shrinkage, 3.1.2.

Capillary fringe zone - zone just above the water table (zero gauge pressure) that remains saturated. The extent and the degree of definition of the capillary fringe depends upon the size distribution of pores.

Capillary porosity - the small pores or the bulk volume of small pores that hold water in soils against a tension usually greater than 60 cm (2 ft) of water. These pores are commonly filled with water when the soil is at field capacity, See Porosity, 3.1.2.

Capillary potential - see Matrix potential, 3.2.2.

Capillary water - the water held in the "capillary" or small pores of a soil, usually with a tension greater than 60 cm (2 ft) of water. Much of this water is considered to be readily available to plants. See Field capacity, Unsaturated zone, 3.2.2.

C-horizon - the layer of weathered parent material in a soil profile. See Soil horizon.
Coarse texture - the texture exhibited by soils containing large proportion of fine and coarse sand fractions (particle diameters greater than 0.02 mm). See Soil texture, Appendix I, 3.1.1.

Condensation - transition from the vapour to the liquid phase. See Evaporation, Sublimation, Vapour pressure.

Consistency (Soil) - 1: the resistance of a material to deformation or rupture. 2: the degree of cohesion or adhesion of the soil mass. Terms used for describing consistency of soil materials at various soil moisture contents and degrees of cementation are:
- Wet: Nonsticky, slightly sticky, sticky, very sticky, nonplastic, slightly plastic, plastic, and very plastic
- Moist: Loose, very friable, friable, firm, very firm, and extremely firm
- Dry: Loose, soft, slightly hard, hard, very hard, and extremely hard
Cementation: Weakly cemented, strongly cemented, and indurated. See Plasticity, Appendix I, 3.1.1.

Consumptive use - the quantity of water used and transpired by vegetation plus that evaporated. See Evapotranspiration, 3.4.3.

Density - the mass per unit volume material. See Bulk density, Particle density, 3.1.2.

Depth of water table - distance between the surface of soil and upper boundary of groundwater. See Unsaturated zone, 2.2.

Dew point - the temperature attained upon cooling moist air at which its water vapour starts to condensate, i.e. when the actual vapour pressure equals the saturation vapour pressure. See Condensation, Vapour pressure.

Drainage (Soil) - when the soil is free of saturation; for example, in well-drained soils the water is removed readily but not rapidly; in poorly drained soils the root zone is waterlogged for long periods unless artificially drained, and the roots of ordinary crop plants cannot get enough oxygen; in excessively drained soils the water is removed so completely that most crop plants suffer from lack of water. Strictly speaking, excessively drained soils are a result of excessive runoff due to steep slopes or low available water-holding capacity due to small amounts of silt and clay in the soil material. See 2.3, 3.4.4.

Dry state (Soil) - the condition of a soil after heating at 105°C during at least 24 hours to remove by vapourisation all physically bound water. In this state the soil has maximum bulk density due to shrinkage. See Dry weight, shrinkage, 3.2.1.

Dry weight (Soil) - the equilibrium weight of the solid soil particles after the water has been vapourised by heating to 105°C. See 3.2.1.

Ecosystem - a community of organisms and its controlling environment. See Appendix II, 3.4.5.
Eluvial horizon - a soil horizon formed by the process of eluviation (leaching). See Eluviation, Illuvial horizon, Soil horizon.

Eluviation - the removal of soil material in suspension (or in solution) from a layer or layers of a soil. (Usually, the loss of material in solution is described by the term "leaching"). See Leaching, Soil horizon.

Evaporation - physical process which transforms solid or liquid media into vapour (gas). See Evapotranspiration, Condensation, Sublimation, Vapour pressure, 3.4.3.

Evapotranspiration - water transpired by vegetation plus that evaporated from the soil. See Consumptive use, 2.2, 2.3, 3.4.3.

Evaporation zone - the layer of the unsaturated zone between the soil surface and thermocline level. See 2.2.

Effective porosity - the ratio of the volume of interconnected pore-space available for fluid transmission to bulk volume of the soil. See Air capacity, 3.1.2.

Field capacity - the greatest amount of water that the soil will hold under conditions of free drainage (i.e. against gravity), usually expressed as a percentage of the oven-dry weight of soil. See Air capacity, Available water, 3.2.2.

Fine texture - the texture exhibited by soils containing large proportions of silt and clay (particle diameters smaller than 0.02 mm). See Soil texture, Appendix I, 3.1.1.

Free water - water not adsorbed to the wall of the pores and therefore free to move under the force of gravity. See Gravitational water, Saturation, Groundwater zone, 2.2, 3.2.2.

Gravitational water - water that moves into, through, or out of the soil under the influence of gravity. See Free water, 3.3.2.

Groundwater - subsurface water in the zone of saturation, having atmospheric and hydrostatic pressure. See 2.2.

Groundwater table or level - the upper surface of the groundwater zone, where the pressure is atmospheric. See 2.2.

Groundwater zone - zone beneath the groundwater level (table) where water has atmospheric and hydrostatic pressure. See Free water, Groundwater, 2.2.

Humus - that more or less stable fraction of the soil organic matter remaining after the major portion of added plant and animal residues have decomposed, usually amorphous and dark coloured. See Organic matter, Appendix I, 3.1.2.

Hydraulic conductivity - property of soil to transmit water under pressure and in a saturated condition. Generally expressed as the flux through a unit section of soil perpendicular to the direction of flow governed by a gradient of unity. See Permeability, Transmissivity, 3.3.1.
Hygroscopic coefficient - amount of adsorbed water on the surface of soil particles in an atmosphere of 50% relative humidity at 25°C. For classifying soil water conditions in the field on an ecological basis other definitions have been supported, see 3.2.2.

Hygroscopic water - water so tightly held by the attraction of soil particles that it cannot be removed except as a vapour by raising the temperature above the boiling point of water. It is unavailable to plants. See Adhesive water, 3.2.2.

Hysteresis - the irreversible effects of drying and wetting on the water retention properties of a soil. It causes the water retention curve to describe a loop, where the higher position applies to the drying process and the lower position to the wetting process. See Moisture retention curve, 3.2.2.

Iluvial horizon - a soil layer or horizon in which material carried from an overlying layer has been precipitated from solution or deposited from suspension. The layer of accumulation in contrast to eluvial horizon. See Soil horizon.

Illuviation - the process of deposition of soil material removed from one horizon to another in the soil, usually from an upper to a lower horizon in the soil profile. See Soil horizon.

Impervious soil - a soil through which water, air, or roots cannot penetrate.

Infiltrability - ease of infiltration. See Infiltration capacity, 3.3.2.

Infiltration - flow of water from the soil surface into the underground. See 3.3.2.

Infiltration capacity - the maximum rate at which a soil can absorb rain or irrigation water during unit time. See Infiltrability, 3.3.2.

Infiltration coefficient - the ratio of infiltration rate to rainfall intensity. See 3.3.2.

Infiltrometer - a device for measuring the rate of entry of fluid into a porous body, for example, water into soil. See 3.3.2.

Interception - the process by which precipitation is caught and held by foliage, twigs, and branches of trees, shrubs, and other vegetation. Often used for "interception loss" or the amount of water evaporated from the intercepted precipitation. See 3.4.1.

Intermediate zone - that part of the soil water zone that lies between the root and capillary zones. See Unsaturated zone, 2.2.

Internal soil drainage - the downward movement of water through the soil profile. The rate of movement is determined by the texture, structure, and other characteristics of the soil profile and underlying layers and by the height of the water table, either permanent or perched. Relative terms for expressing internal drainage are
none, very slow, slow, medium, rapid, and very rapid. See Drainage, Percolation, 3.3.2.

**Intrinsic permeability** - ability of a soil to transmit any fluid, irrespective of the nature of the fluid or the flow conditions. It is proportional to the square of a characteristic length parameter of the soil (pore diameter). See Hydraulic conductivity, Permeability, 3.3.1.

**Irrigation** - the artificial application of water to cultivated land. See 2.3.

**Leaching** - the removal of materials in solution from the soil. See Eluviation, Soil horizon.

**Lower plastic limit** - see Plasticity.

**Lysimeter** - device to measure the quantity or rate of water movement through a block of soil, usually undisturbed or in situ; or to collect such percolated water for analysis.

**Matrix** - natural material in which larger particles are embedded.

**Matrix potential** - the suction caused by adhesion or capillarity expressed as the height of an equivalent water column in relation to the water table. See Moisture tension, Capillary potential. For more details, see 3.2.2.

**Moisture (Soil)** - water in the unsaturated zone. It has negative hydrostatic pressure, due to adhesive and capillary forces and occurs as thin films covering the pore walls. See Adhesive water, 3.2.2.

**Moisture equivalent** - water retained in the soil when the latter is subject to a centrifugal force of 1,000 times the force of gravity for 30 minutes. A laboratory equivalent to field capacity. See Field capacity, 3.2.2.

**Moisture profile** - a curve representing the distribution of soil moisture versus depth. See 3.2.2.

**Moisture retention curve** - a curve representing the relationship between suction and moisture volume percentage.

**Moisture tension or suction** - the negative pressure occurring in soil water as a result of adhesion or capillarity. See Matric potential.

**Moisture volume percentage** - the ratio of the volume of water to the total bulk volume of the soil. See 3.2.1.

**Moisture weight percentage** - the water content expressed as a percentage of the oven-dry weight of soil. See 3.2.1.

**Neutron moisture gauge** - an instrument with a fast-neutron source and a slow (thermal or epithermal) neutron detector which measures the moisture content of soil in terms of the detector count rate, after relating this count rate to the moisture content of the soil by calibration. It consists of a probe containing the source and detector and an electronic unit indicating the count rate. See 4.3.2.
Non-capillary porosity - See Air capacity, 3.1.2.

Organic matter - the organic fraction of the soil that includes plant and animal residues at various stages of decomposition, cells and tissues of soil organisms, and substances synthesized by the soil population. Commonly determined as the amount of organic material contained in a soil sample passed through a 2 mm sieve. See Humus, Appendix I, 3.1.1, 3.1.2.

Organic soils - soils containing more than 20% organic matter. See Organic matter, Appendix I, 3.1.2.

Particle density - the ratio of the dry mass of the soil particles collectively and their respective volume excluding pore spaces between the particles. See 3.1.2.

Particle size - the effective diameter of a particle measure by sedimentation, sieving, or micrometric methods. Various particle size fractions have been defined for the purpose of grouping soils into textural classes. See Particle size distribution, Texture, Appendix I, 3.1.1.

Pedon - the smallest volume that can be called "a soil". It has three dimensions. It extends downward to the depth of plant roots or to the lower limit of the genetic soil horizons. Its lateral cross section is roughly hexagonal and ranges from 1 to 10 square metres in size depending on the variability in the horizons. See Soil classification.

Perched groundwater - groundwater supported by a zone of impermeable material separated from the main body of groundwater by unsaturated material. See Impervious soil, Groundwater.

Percolation (Soil water) - the downward movement of water through soil, especially the downward flow of water in saturated soil at hydraulic gradients of the order of 1.0 or less. See Permeability, 3.3.1, 3.3.2.

Permeability (Soil) - ability of soil to transmit fluids (liquids or gases) under pressure. It is the rate at which a fluid of standard viscosity can move through the soil in a given interval of time under a given hydraulic gradient. Quantitatively it can be expressed either as intrinsic permeability involving only the characteristics of the soil itself, or as hydraulic conductivity involving the characteristics of both the soil and the actual fluid (water). See Transmissivity, 3.3.1.

pF - parameter designating the matric potential and equal to the decimal logarithm of the potential expressed in cm of water column. See 3.2.2.

pF-curve - see Moisture retention curve, 3.2.2.

Piezometer - a tube for measuring the pressure of groundwater.

Piezometric surface - the imaginary surface to which water in a well will rise above an aquifer. See Groundwater level, 2.2.
Plasticity (Soil) - capability of a soil to be moulded or deformed. It is a property that is limited to cohesive soils (i.e., soils containing more than 10% clay) and to a limited moisture range. The lower plastic limit is the minimum moisture weight percentage that permits a soil to be deformed without rupture and the upper plastic limit is the minimum moisture weight percentage at which the soil will barely flow as a result of stress. The plasticity index is the numerical difference between the moisture weight percentages of the upper and lower plastic limits. This index is dependent on the clay content and mineralogy of the soil.

Pore - interstice, void in rock or soil. May be classified into different size classes. See Soil structure, 3.1.2.

Pore size distribution - the amount of the various soil-pore fractions in a soil. Equivalent to the derivative of the moisture relation curve. Characterizes soil structure. See Soil structure, 3.1.2.

Pore space or volume - see Porosity (total).

Porosity (total) - total space not occupied by soil particles in a bulk volume of soil, commonly expressed as a volume percentage. It may be estimated from the bulk and particle densities of the soil according to the following expression.

\[
100 - \frac{\text{Bulk density}}{\text{Particle density}} \times 100
\]

See Air-filled porosity, Capillary porosity, Saturation, 3.1.2.

Potential - specific energy of unit material in relation to some reference state. See Matrix potential, 3.2.2.

Potential evaporation - highest possible amount of evaporation permitted by the vapour receiving capacity of the atmosphere as determined by prevailing meteorological conditions. See Evaporation, 2.3, 3.4.3.

Potential evapotranspiration - maximum quantity of water lost to the atmosphere from vegetation and soil, where the latter is well supplied with water. See Evaporation, Transpiration, 3.4.3.

Reaction (Soil) - the degree of acidity or alkalinity of a soil usually expressed as a pH value. Descriptive terms commonly associated with certain ranges in pH are: extremely acid, less than 4.5; very strongly acid, 4.5-5.0; strongly acid, 5.1-5.5; medium acid, 5.6-6.0; slightly acid, 6.1-6.5; neutral, 6.6-7.3; mildly alkaline, 7.4-7.8; moderately alkaline, 7.9-8.4; strongly alkaline, 8.5-9.0; and very strongly alkaline, more than 9.0.

Recharge (Water) - process by which water is added from the outside to the zone of saturation. See Percolation, 3.3.2.

Regime (Soil water) - seasonal pattern of water distribution and movement in the soil. See 2.2, 2.3.
Relative humidity - see Vapour pressure.

Retention curve - see Moisture retention curve.

R-horizon - stratum underlying the soil profile. See Soil horizon.

Root depth - depth to which a soil profile is penetrated by water extracting roots. See Root zone, 2.2.

Root zone - the upper part of the soil water zone, where the pores contain roots that can extract water. The depth of the root zone depends on the species of vegetation, the growing phase of the plants and on the water content of the soil. See 2.2.

Runoff - outflow of water toward the streams along or underneath the ground surface.

Saturated zone - that part of the lithosphere where the soil pores are completely filled with water at atmospheric and hydrostatic pressure. The zone includes both the groundwater zone and the capillary fringe zone. See 2.2.

Saturation - state when all the soil pores are filled with water, and the suction is zero. See Groundwater, 2.2, 3.2.2.

Saturation vapour pressure - see Vapour pressure.

Seepage - the slow movement of gravitational water through the soil. Water emerging from the ground along a line or surface. See Percolation, 3.3.1, 3.3.2, 3.4.4.

Shrinkage (Soil) - the decrease in porosity of the soil due to decrease in water content as shown in figure below. See Swelling, Plasticity, Bulk density, 3.1.2., figure III:1.

Figure III:1 Shrinkage of clay soil due to successive loss of water.
Single-grain structure - physical state of soil, implying that there is no aggregation of primary particles. See Aggregate structure, 3.1.2.

Soil - the unconsolidated mineral and organic material on the immediate surface of the earth that serves as a natural medium for the growth of land plants. 2: The unconsolidated mineral matter on the surface of the earth that has been subjected to and influenced by genetic and environmental factors of parent material, climate (including moisture and temperature effects), macro- and micro-organisms, and topography, all acting over a period of time and producing a product - soil - that differs from the material from which it is derived in many physical, chemical, biological, and morphological properties and characteristics. 3: A kind of soil is the collection of soils that are alike in specified combinations of characteristics. Kinds of soil are given names in the system of soil classification. The terms "the soil" and "soil" are collective terms used for all soils, equivalent to the word "vegetation" for all plants.

See Pedon, Soil classification, Soil profile, Appendix I, 3.1.1.

Soil air - the gaseous phase of the soil. See Air-filled porosity, 3.1.2.

Soil auger - a tool for boring into the soil and withdrawing a small sample for field or laboratory observation.

Soil classification - the systematic arrangement of soils into classes in one or more categories or levels of classification for a specific objective. Broad groupings are made on the basis of general characteristics and subdivisions on the basis of more detailed differences in specific properties. See Table 1 beneath.

Soil-formation factors - the variables that are responsible for the formation of soil. The factors are usually grouped as follows: parent material, climate, organisms, topography, and time. Many people believe that activities of man in his use and manipulation of soil becomes such an important influence on soil formation that he should be added as a sixth variable. Others consider man as an organism.

Soil horizon - a layer of soil material approximately parallel to the land surface and differing from adjacent genetically related layers in physical, chemical, and biological properties or characteristics, such as color, structure, texture, consistence amount of organic matter, and degree of acidity or alkalinity. Table 2 lists the designations and descriptions of the major soil horizons. Few if any soils have all of these horizons well developed, but every soil has some of them, see Table III:2 beneath.
Table III: A comparison of the new United States soil classification system adopted in 1965 with the approximate equivalents under the soil classification system in use before 1965.

<table>
<thead>
<tr>
<th>Soil Order (Adopted in 1965)</th>
<th>Approximate Equivalents (In Use Before 1965)</th>
</tr>
</thead>
<tbody>
<tr>
<td>2. Aridisols</td>
<td>Desert, Reddish Desert, Sieirozem, Solonchak, some Brown and Reddish Brown soils, and associated Solonetz</td>
</tr>
<tr>
<td>3. Entisols</td>
<td>Azonal soils and some Low-Humic Gley soils</td>
</tr>
<tr>
<td>4. Histosols</td>
<td>Bog soils</td>
</tr>
<tr>
<td>5. Inceptisols</td>
<td>Ando, Sol Brun Acidé, some Brown Forest, Low-Humic Gley, and Humic Gley soils</td>
</tr>
<tr>
<td>6. Mollisols</td>
<td>Chestnut, Chernozem, Brunizem (Prairie), Randzina, some Brown, Brown Forest, and associated Solonetz and Humic Gley soils</td>
</tr>
<tr>
<td>7. Oxisols</td>
<td>Laterite soils, Latosols</td>
</tr>
<tr>
<td>8. Spodosols</td>
<td>Podzols, Brown Podzolic soils, and Ground-water Podzols</td>
</tr>
<tr>
<td>10. Vertisols</td>
<td>Gruviosols</td>
</tr>
</tbody>
</table>
Table III:2 Hypothetical soil profile, showing principal horizons

| Organic debris | O           | 0₁           | Undecomposed fresh debris (leaves etc) |
|               | 0₂          | Partially decomposed |
|               | A₁          | Darker coloured horizon, where organic matter is mixed with mineral grains |
|               | A₂          | Lighter coloured horizon with maximum eluviation |
|               | A₃          | Transitional to B. |
| Eluvial horizon | B₁        | Transitional to A. |
|               | B₂          | Maximum accumulation (illuviation) of clay, iron and organic matter |
|               | B₃          | Transitional to C. |
| The Solu       | C           | Parent material. Can contain gleyed layers, accumulations of calcium carbonate and sulphate. |
| (The genetic soil developed by soil forming factors) |               |               |
| The weathered parent material |               |               |
| Underlying stratum (bedrock or soil) |               | Material from which the overlying horizon was formed or material unrelated to this altogether (lithologic discontinuity) |
Soil map - a map showing the distribution of soil types or other soil mapping units in relation to the prominent physical and cultural features of the earth's surface. The following kinds of soil maps are recognized in the U.S.: detailed, detailed reconnaissance, reconnaissance, generalized, and schematic.

Soil monolith - a vertical section of a soil profile removed and mounted for display or study. See Soil horizon.

Soil morphology - the constitution of the soil, including the texture, structure, consistence, color, and other physical, chemical, and biological properties of the various soil horizons that make up the soil profile. See 3.1.

Soil organic matter - see organic matter.

Soil profile - a vertical section of the soil from the surface through all its horizons, including the C-horizons. See Soil horizon.


Soil solution - soil water with its dissolved materials.

Soil structure - the combination or arrangement of primary soil particles into secondary particles, units, or pedes. The secondary units are characterized and classified on the basis of size, shape, and degree of distinctness into classes, types, and grades, respectively. See Aggregate structure, Single-grain structure, 3.1.2.

Soil texture - an expression of the relative proportions of soil separates - clay, silt, sand, etc - that compose the soil material. See Appendix I, 3.1.1.

Soil water - see Moisture.

Soil water potential - see Matrix potential.

Soil water zone - that part of the lithosphere between the surface and the water table where the water is under negative pressure (less than atmospheric). It consist of the unsaturated zone (pores contain air and water) and the capillary fringe zone (pores contain only water). See 2.1, 2.2.

Solum - the upper part of a soil profile, above the parent material, in which the processes of soil formation are active. The solum in mature soils includes the A and B horizons. Usually the characteristics of the material in these horizons are quite unlike those of the underlying parent material. The living roots and other plant and animal life characteristics of the soil are largely confined to the solum. See Soil, Soil classification, Soil horizon.
Storage capacity (of Soil water zone) - the amount of soil water which can be stored between the surface and the water table. A dynamic concept dependent on the characteristics of climate and soil and on time dependent variables such as moisture content and depth of water table. Complexity is increased by hysteresis. See Field capacity, 3.2.2, 3.3.2.

Sublimation - direct transfer of water from the vapour phase to the solid phase, or vice versa. See Condensation, Evaporation, Vapour pressure.

Subsoil - the B horizons of soils with distinct profiles. In soils with weak profile development, the subsoil can be defined as the soil below the plowed layer (or its equivalent of surface soil), in which roots normally grow. Although a common term, it cannot be defined accurately. It has been carried over from early days when "soil" was conceived only as the plowed soil and that under it as the "subsoil". See Soil horizon.

Substratum - any layer lying beneath the solum. See R-horizon, Soil horizon.

Subsurface water - water that occurs beneath the surface of the earth, see Soil water, Groundwater, 2.1, 2.2.

Surface soil - the uppermost part of the soil ordinarily moved in tillage or its equivalent in uncultivated soils, ranging in depth from about 10 to 30 cm. Frequently designated as the plow layer, the Ap-horizon, see Soil horizon, Top soil.

Surface water - water which flows over or is stored on the earth surface.

Swelling (Soil) - the increase in porosity and thereby bulk volume of the soil due to the uptake of water. Is particularly marked in soils containing sodium (alkali soils). See Shrinkage (opposite concept), Bulk density, Plasticity.

Tensiometer - instrument used for measuring the suction or negative pressure of soil water. Generally a porous device filled with water that is made to reach suction equilibrium in-situ with the desired soil horizon. The suction is registered on some type of vacuum meter and is limited to a maximum tension of about 0.85 bar. See 4.4.1.

Tension - see Moisture tension or suction, 3.2.2.

Texture - see Soil texture, Appendix I, 3.1.1.

Throughfall - that part of the precipitation which falls to the ground in a vegetated area. See Interception (opposite term), 3.4.1.

Topsoil - the surface plow layer of a soil, the original or present dark-colored upper soil or the present A horizon, varying widely in depth and nature among different kinds of soil. See Soil horizon.

Total porosity - see Porosity (total).
Transmissivity - product of hydraulic conductivity and the cross-sectional area of flow. The hydraulic conductivity decreases in accordance with the degree of unsaturation of the soil. See Hydraulic conductivity, Permeability and Unsaturated conductivity, 3.3.1.

Transpiration - process by which water is transferred from vegetation to atmosphere in the form of vapour. See Evapotranspiration, 3.4.3.

Transpiration zone - larger of soil between the thermocline and root depth levels. See Root depth, 2.2.

Unsaturated conductivity - ability of unsaturated soil to transmit water. Decreases in accordance with degree of unsaturation. See Permeability, 3.3.1, 3.3.2.

Unsaturated flow - water flow through the unsaturated soil. See 3.3.1, 3.3.2.

Unsaturated zone - upper part of soil water zone where water is held under negative pressure and where the pores contain some air. See Aeration zone, Soil water zone, 2.2.

Upper plastic limit - see Plasticity.

Vadose water - see Gravitational water.

Vapour pressure - that part of the atmospheric pressure that is exerted by the water vapour. The maximum vapour pressure at a given temperature is the saturation vapour pressure over water. The saturation vapour pressure with respect to ice is less than that with respect to water. The percentage of actual vapour pressure to saturation vapour pressure is by definition the relative humidity. See figure 3.2.

Viscosity - that property of a fluid, due to the cohesiveness of its molecules, which resists relative motion and deformation in real fluids. It creates shear forces between fluid elements and gives rise to fluid friction. Under ordinary conditions of pressure, viscosity has been found to vary only with temperature. See Permeability.

Volume weight - see Bulk density.

Water content - see Moisture, Moisture volume percentage, Moisture weight percentage, 3.2.1.

Water film - thin layer of water held by pore walls through the force of adhesion. See Adhesive water, 3.2.2.

Water holding capacity (Soil) - see Storage capacity, Available water, Field capacity, 3.2.2.
Schematic phase diagramme for water vapour in the neighbourhood of the triple point (T), where three phases may exist in equilibrium - gasous, liquid and solid. The triple point occurs here at a pressure of 6.11 mb and at a temperature of +0.0098°C.

The curves represent sublimation (AT), evaporation-condensation (TD) and melting-frezing (CT) respectively. The sectors represent ice (ATC), water (CTD) and vapour (DTA). Along the evaporation curve TD there exists equilibrium between water and vapour, along the sublimation curve AT equilibrium between ice and vapour.

It is seen that saturation with respect to super-cooled water means supersaturation with respect to ice.

See Vapour pressure, Sublimation, Evaporation, Condensation.
Water logged soil - soil saturated with water. See Saturation.

Water table or level - see Groundwater table, 2.2.

Water table, Perched - See Perched groundwater.

Water year - instead of calendar year, a period starting with the unsaturated zone at its maximum extent, and consequently other storages of water (surface and groundwater) reduced to a minimum.

Weathering - the group of processes such as chemical action of air and rainwater and of plants and bacteria, and the mechanical action of changes in temperature, whereby rocks, on exposure to the weather, change in character, disintegrate, decay, and finally crumble in the process of making parent material of soils. See Soil.

Wetting front - the air-water interface in wetting. See 3.3.2.

Wilting point - the water content of soil on an oven-dry basis at which plants, specifically sunflower plants, wilt and fail to recover their turgidity when placed in a dark humid atmosphere. It is approximately the moisture content at 15-bar tension. See Available water, 3.2.2.

Zero flux plane - the level which separates upward and downward movement of water.

Zone (Soil water) - see Soil water zone.
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