702

Water in the Unsaturated Zone

NHP-Seminar,
29 - 31 January 1986,
Ås, Norway
Edited by: Sylvi Haldorsen
and Einar J. Berntsen

NORDIC HYDROLOGCAL PROGRAMME
NHP-Report no. 15

The Co-ordinating Committee
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in Norden (KOHYNO)
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List of content

PREFACE

Knowledge about the spatial variability of soil water is of great importance for understanding the hydrology of a certain area. The theories of water flow in the unsaturated zone are rather well established, and many models suit the theories very well. A serious problem today seems to be how to verify the models in the field and to study how well they reflect the spatial variability. Few models used in Norden today are based directly on field measurements of spatial variability. In many cases they represent vertical water movement at one point. This is not because spatial variability cannot be modelled, but merely because there are no field data to be used in the models.

In order to discuss these questions, a NHP-seminar was arranged in the regi of the Norwegian National Committee for Hydrology in January 1986. About 50 individuals, most of them soil scientists and hydrologists from Denmark, Norway, Sweden and Finland, and two from the United Kingdom representing the Institute of Hydrology and Soil Survey of England & Wales, respectively, participated in the seminar.

The aim of the seminar was to discuss "the state of the art" in Norden today. What types of data on spatial variability do exist, in which way are they applied, which data types are most important for future work and how can they be used in modelling? Indeed, the seminar did not give give answers to all these questions. But it hopefully may be an inspiration for further contact and more specific cooperation in the future.

For the
Organizing committee

Einar Berntsen

Sylvi Haldorsen

Einar Berntsen

Sylvi Haldorsen

INTRODUCTION TO THE SEMINAR

Arnor Njøs, Dep. of soil fertility and management,
Agricultural University of Norway

Mr. Chairman

Ladies and gentlemen. Dear collegues.

On behalf of the Agricultural University of Norway, I wish you welcome to this seminar on water in the unsaturated zone. You are now in the community of As which has about 11000 people, and an area of 100 km^2 of which 40 are cultivated.

As may be considered an agricultural science centre. In addition to the Agricultural University there are many institutes, such as the Norwegian Forest Research Institute, the State Seed Testing Station, the State Plant Protection Institute, the Agricultural Engineering Institute, the Norwegian Land Inventory Institute, the administration office of the State Research Stations and others.

At the Agricultural University there are 11 study directions, ranging from General Agriculture to Management of Natural Resources.

The theme of the seminar, Water in the unsaturated zone is a great challenge to several groups of scientists, soil physicists, hydrologists, agronomists, soil technologists, soil surveyors, irrigation and drainage specialists, soil conservationists, soil and water pollution specialists.

Water in the unsaturated zone has at least two important aspects, statics and dynamics, or storage and flow.

Already in ancient times people noticed the problem of no storage systems. In one of the dramas of Aischylos (dating back to around 500 B.C.) fifty daughters of Danaos fled from Egypt to

Greece to find a safe place from aggressive, courting men. But they were forced into marriage. On the night of the wedding the daughters murder their aggressive husbands. The daughters are convicted to forced labour in the underworld, this labour consisting in trying to fill a large strainer by pouring water into it - for eternity. In this case the equation of continuity was very simple: In-Out=0.

With increacing knowledge of the chemical and physical properties of water and of the soil materials, particularly the soil surfaces, the knowledge of moisture storage has been gradually increased. The energy relations of soil water have been the theme for much study. Scientists such as Buckingham (1907) Schofield (1935) Edlefsen and Anderson (1943), Bolt and Frissel (1960), and others have contributed to a deeper understanding of the energetics of soil water. L.A. Richards with his tensiometer (1929) and S.J. Richards (1939) with the pressure membrane apparatus brought the measuring techniques forward.

So much about the statics of soil water. The transport phenomena were gradually systematized and brought under mathematical regulation over more than a century. The first major step forward within the field of transport equations occurred in 1807, when Fourier accompanied Napoleon to the pyramides of Egypt. Fourier developed the equations for heat flux. In 1827 Ohm's equation for transport of elecricity saw daylight. For water transport in capillary tubes we notice that the Hagen-Poiseuille equation dates back to 1844. In 1856 Darcy made some experiments with filtering of water through sand, to clean the drinking water for the town of Dijon. The steady state equation for transport of water through porous media was a fact. This equation applied to the saturated condition.

Work with transport of water in unsatorated soil gained momentum in the 1930-ies. The simple case of transport upwards from shallow water tables was studied by Moore (1939). Childs & Collis-George (1950) worked with transient state transport, and introduced the term water diffusivity analogous to the term thermal diffusivity. It was defined as the ratio between water conductivity and specific water capacity. J.R. Philip in the late 1950-ies worked with the infiltration process. Hundreds of papers appeared on water transport through unsaturated soils, corresponding to different geometries and boundary conditions.

The need to understand the leaching of salts from saline soils in dry regions, or the effect of acid atmospheric inputs by dry and wet deposition, or the need for downwords transport of Calcium in acid tropical soils led to development of combined equations for solutes and water. The unsaturated condition was shown to be the more important condition for effective leaching.

In soil physics the methods of study and the size and number of samples have not received enough attention. We think in micro, meso and macro viewpoints. We try to use the analytical methods from pure physics, suitable for simple media, such as bundles of capillaries or glass beads, on complicated biological - mineralogical systems called soils. We use the conservation type of analysis for whole watersheds of several km^2, we take volumetric samples of 100 cm^3, and we study the pore structure with electron microscope. Then we try to extract useful information from these different approaches. But soil variability is often unknown, thus the conclusions reached may not be scientifically justified.

While much of the early work in water transport was centered on idealized porous media, such as glass beads or uniform sands, the adaption of the transport equations to field conditions gradually brought forward the tremendous field variation in time and space of hydraulic conductivity. It may be stated that during the period 1920 - 1970 very few soil physicists were really concerned with soil variation. It was considered a noise, which should be avoided by avoiding replication.

However, Don Nielsen and collegues in California, started out to study the variation of soil physical parametres in the field. The only parameter with reasonable small variation was dry bulk density or total porosity, while the hydraulic conductivity could differ several orders of magnitude between replicated measurements. An arithmetic average was meaningsless for hydraulic conductivity.

A sample volume or area is often chosen without critical considerations with regard to the application of resulting numbers from experiments or measurements. For bulk density, available water, air porosity, and grain size parametres the volume of sample may not be that important if sufficient replicates are taken, and provided the error variance is properly calculated. We may perhaps state that the 100 cm^3 sample is valuable as an approximation for some static parametres, for instance in connection with soil survey or in soil treatment experiments. The double-ring infiltrometer method of measuring water infiltration in the field may be a much more dubious method, as may be conductivity measurements on 100 cm^3 samples. The error increases tremenduously when the largest pores are engaged, such as in saturated water flow.

Evaluation of the soil variability will be an important task for soil physicists in the near future. The tools of covariance and variance component analysis as well as scaling, autocorrelation and other methods may help us on the route forward.

It should be in our minds that variability may be considered in several ways. On the one hand it is a noise disturbing the base for conclusions. On the other hand it is a tremendous source of information which may help to clarify many causal relationships, if systematized. We see easily the usefulness of such work for the boundary question in soil survey, for the management of fields in agriculture and for tactic considerations in military field operations. We must learn to live with the variability, even to love it.

Mr. chairman! I wish you success with your seminar and hereby declare it opened. Thank you for your attention.

Winter Rain Acceptance Potential: Soil and Spatial Aspects

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The need for better estimates of flood flows and levels in streams is well established. For example the estimation of flood flows, following extreme storms, is invaluable for sizing culverts or designing reservoir spillways. In Britain, as elsewhere, the long-term records of flood flows are too few for reliable nation-wide estimates to be made. On the other hand long-term rainfall records are relatively widespread. Thus for low probability floods, that is major events with a long return period, a way of estimating floods from the more extensive rainfall statistics is desirable. Conversely low flow studies are needed to assess dilution of sewage effluent in dry summers.

The combinations of soil properties controlling winter run-off and in particular run-off likely to cause flooding from rivers can be summarized by the expression "Winter Rain Acceptance Potential" (WRAP). The term winter is used to exclude periods when an appreciable soil moisture deficit exists. It is assumed that the storm is falling on soil which is already at field capacity. The rain cannot be retained within the soil profile, but must be disposed of either vertically or laterally. As the potential to accept rain increases, rapid response run-off for a given rainfall event decreases and conversely as the potential decreases rapid response run-off increases.

Many river catchments are monitored for their river flow but, for numerous reasons, records are of varying periods. Even so these records enable us to calculate, for example, the amounts of various storm rainfalls which flow from specific monitered catchments within a short period of the rainfall events; the percentage run-off.

It then becomes necessary in any flood study to predict this percentage for ungauged catchments as a step towards converting rainfall to river flow. In order to produce a model of catchment hydrology a knowledge of the distribution of soil types and an understanding of their properties is fundamental. The Soil Survey of England and Wales (SSEW) have attempted to quantify the effects of soil type on catchment hydrology.

In an ideal world where costs were unimportant large numbers of accurate physical measurements could be made over each catchment. Measurements could include vertical and horizontal conductivity at various depths together with water retention properties of individual horizons. In the real world, this is too expensive and too time consuming. In the early 1970's an ordinal classifica-tion of run-off potential, was constructed using five classes, and based on readily observable soil properties (Farquharson et al. 1978). It was decided to integrate the following four soil and site properties.

(a) Soil water regime
(b) Depth to an impermeable layer
(c) Permeability of soil to 80 cm depth or above an impermeable layer
(d) Slope

Soil water regime

A three-fold system was established based on the Wetness Classes defined in the Soil Survey Field Handbook (Hodgson 1976). Guidelines to the field identification of Wetness Class are given in the Field Handbook. The gley morphology of the profile, in particular the depth to layers mottled with greyish colours, in conjunction with experience of field moisture conditions at different times of the year and a knowledge of dip-well levels is used to allocate the Wetness Classes. Climatic data to assess the length of the field capacity period is also needed (Jones and Thomasson 1985).

Soil water regime class exerts a strong influence on the acceptance of rain. A water-table at shallow depth reduces the soil's ability to accept heavy rain simply because storage space above the water-table is restricted. Percolating water also has a short pathway to a saturated layer and later, lateral movements in the saturated zone are likely to be rapid.

Table 1 Soil water regime class

Soil water regime class	**Depth and duration of waterlogging**	**Wetness Class** (Hodgson 1976)
1	Rarely waterlogged within 40 cm Waterlogged for <90 days within 70 cm	I + II
2	Commonly waterlogged within 40 cm Waterlogged for <335 days within 70 cm	III + IV
3	Waterlogged for >180 days within 40 cm Waterlogged for >335 days within 70 cm	V + VI

Depth to an impermeable horizon

The soil water regime and the proximity of impeding layers to the surface are often closely related but, because of exceptions, it is desirable to separate these two properties. Definition of an impermeable horizon in exact physical terms is not easy. Commonly, a layer with a hydraulic conductivity of 1/10th that of the layer above is considered to be impermeable (Luthin 1957). When conductivity approaches 10 cm d^{-1} its contribution to water movement towards the drain or watercourse is negligible. In this situation the hydraulic conductivity usually declines with depth, and at some stage, particularly where the water-table is shallow, horizontal movement will be the controlling factor in the disposal of rainfall. Thus, it is unlikely that great error results if the impermeable horizons are defined as having a horizontal conductivity of less than 10 cm d^{-1}. It is still necessary, however, to identify such horizons in the field. Comparative studies of conductivity, texture, structure and pore-size distribution indicate that soils with a conductivity of less than 10 cm d^{-1} are mainly subsoils with less than 5% macropores (drainable pores >60μm diameter) and high packing density (Thomasson 1975, Hall et al. 1977).

The final step is to express the concept of an impermeable horizon in terms of particle-size class and soil structure for easy field identification (Table 2). It is assumed that such horizons must be more than 20 cm thick, and that topsoils and humose or peaty soils are excluded. Unfissured rock and thin ironpans are interpreted as impermeable.

Table 2 Criteria for impermeable horizons

Structure	Particle-size class
Massive, platy	All classes, excluding sandy loam, loamy sand and sand
Moderately and weakly developed, medium coarse or very coarse, angular blocky or prismatic	Clay, sandy clay, silty clay
Unfissured rock	
Thin ironpan	

When the packing density class (Table 3) is known, identification of impermeable horizons is greatly simplified. Packing density (Lagerungsdichte) combines bulk density and clay content in the form:

$$L_d = D_b + 0.009\ (\%\ \text{clay})\ \text{g cm}^{-3}.$$

Table 3 Packing density classes

Class	Packing density ($g\ cm^{-3}$)
Low	<1.40
Medium	1.40 - 1.75
High	>1.75

Most soil horizons having high L_d and more than 35% clay can be considered impermeable. For WRAP purposes 3 depth classes for impermeable horizons are used (Table 4).

Table 4 Classes of depth to impermeable horizon

Class	Depth to impermeable horizon (cm)
1	>80
2	40 - 80
3	<40

Permeability above impermeable horizons

The site properties already discussed probably account for most of the variation in rapid response run-off from soil to soil. Nevertheless some attempt must be made to differentiate soil profiles which lack an impermeable horizon, or where the properties of soil above such horizons are very different. The chosen model assumes (as before) that saturated flow at or near the soil surface is the main component of rapid response stream discharge. Deep infiltration and horizontal movement at depth 'buffer' the system and reduce the immediate response of streams to a rainfall event. This delay normally reduces the subsequent peak discharge. A second assumption is that, whatever its order of magnitude, saturated hydraulic conductivity decreases with depth in most soil profiles. Exceptions to this occur but such profiles are not extensive.

Three broad classes of permeability were chosen. Owing to the lack of measured values in the early 1970's they are identified purely in terms of structure and particle-size class (Table 5).

Table 5 Relationships between structure, particle-size and permeability classes

Structure	Particle-size class	Permeability class
Moderate or strong granular; all sizes	All classes	1
Strong, medium, fine sub-angular; weak granular, all sizes	All classes, excluding clays;	1
	Clay, sandy clay, silty clay;	2
Moderate or weak, sub-angular, all sizes; strong fine, medium, coarse and very coarse angular; strong fine prismatic	Sand, loamy sand, sandy loam;	1
	Sandy silt loam, silt loam, sandy clay loam, clay loam, silty clay loam;	2
	Clay, sandy clay, silty clay;	3
Moderate and weak angular, all sizes; all remaining prismatic	Sand, loamy sand;	1
	Sandy loam;	2
	Sandy silt loam, silt loam, sandy clay loam, clay loam, silty clay loam;	3
Massive, structureless; all platy structures	Sand, loamy sand;	2
	Sandy loam;	3
Chalk, limestone, sandstone and other well fissured rock		1

Permeability classes for layers within 80 cm or above an impermeable horizon: 1 - rapid, 2 - medium, 3 - slow; structure terms and particle-size classes are defined in the Soil Survey Field Handbook (Hodgson 1976).

Notes:
Lowland humose peat soils are classified as for sandy silt loam.
Ploughed topsoils (Ap) are impermeable, irrespective of structure.
Combinations of structure and particle-size class not listed here are classed as impermeable (see Table 2).
Hill peat is allocated to permeability class 3.

Soils with rapid permeability give slow stream response to storms, while those with slow permeability cause rapid run-off. In the field one profile type may have two or more permeability classes either within 80 cm or above an impermeable layer, for example

Depth	Permeability class
0 - 30 cm	1
30 - 60 cm	3
60 cm +	Impermeable layer

Profiles of this type are placed in group 3 by weighting in favour of the slow horizon. Similarly the three profiles with the following patterns of permeability

Depth	Permeability class		
0 - 30 cm	2	1	2
30 - 60 cm	3	3	3
60 cm +	3	2	2

are all placed in permeability group 3. Restricting layers of medium or slow permeability that are less than 20 cm thick can be discounted.

Slope

In an area of medium or slow permeability, slope is the prime variable affecting run-off. Although it is not strictly a soil property, slope interacts with soil properties to accentuate run-off where the water-table is shallow, impermeable layers are present, or permeability is slow. Where the three main soil parameters favour a large WRAP value (Classes 1 or 2) the effect of slope is negligible. The following three classes were used because information on their distribution had already been collected for a national soil map.

1. $<2^{o}$
2. $2 - 8^{o}$
3. $>8^{o}$

Allocation of WRAP class to soil profile

Although the directional effects of the four main parameters are reasonably clear, their relative magnitude is a matter of judgment. In constructing a national WRAP map, the chief requirement was that many individual soil scientists, often considering this type of problem for the first time, should use the system in a uniform manner. Figure 1 was constructed with this aim in mind.

Broadly the factors are given the priority order:

(1) Soil water regime >
(2) depth to impermeable horizon >
(3) permeability above 2.

Slope (4) accentuates the effect of the first three.

This order of priority was based on the known behaviour of a number of gauged catchments for which soil surveys had been completed and various soil physical properties already known particularly the river Dee above Erbistock (Rudeforth and Thomasson 1970). There was also more general information on the behaviour of streams in Chalk, hill peat catchments and Bunter Sandstone and Pebble Beds.

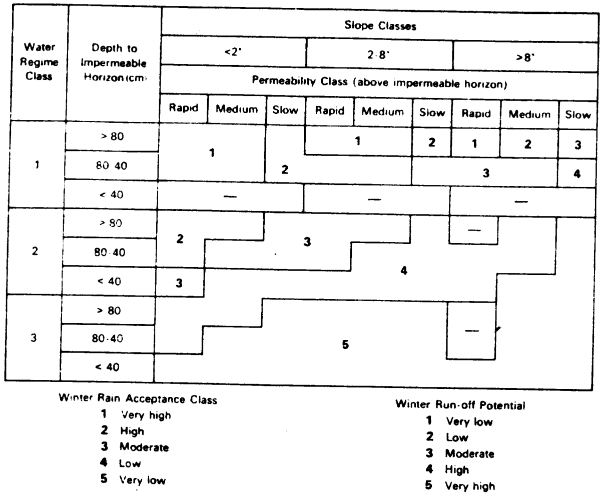

Fig. 1. Winter rain acceptance potential class in relation to soil and site properties

Figure 1 is a four-way chart from which the WRAP class can be obtained given appropriate soil and site data. Certain combinations of properties are unlikely to occur in Northern or Western Europe and so are left unclassified, for example, well drained soil (water regime class 1) with an impermeable horizon at less than 40 cm depth. The methods of allocating WRAP classes to individual profiles is therefore clear. The National 1:1,000,000 soil map (Avery et al. 1975) which was used as the basis for the WRAP map indicates the distribution of soil associations (71 in all) showing broad groupings of soil profiles on distinctive parent material and containing distinctive soil patterns. After allocating the WRAP class to each of the dominant soil series within the 71 map units, the relative location of soils, throughout the geographic range of the association, was taken into account especially where this may affect the relative yields of surface and subsurface water. A composite WRAP assessment was therefore achieved for each occurrence of all 71 soil associations.

Occasionally final WRAP assessment had to be adjusted to arrive at decisions which matched experience. For example soils in clayey drift over chalk which cover much of the Chiltern Hills would be classified as WRAP 2, on the basis of soil properties and slope. However the appropriate soil association has been placed in WRAP 1 as there is little surface flow, even at high rainfall intensities, in the dry valleys of the Chilterns. The lateral flow from the plateau ceases where bare chalk or shallow soil over chalk are encountered on valley sides. A distinction was also made between drained and undrained lowland peat soils. Intensively drained peats, used for arable cropping and horticulture, were placed in Classes 1 and 2 depending on the degree of water-table control. Where such soils have not been drained the map separates have been placed in WRAP class 5.

Recent developments in WRAP

Since the 1:1,000,000 WRAP map was published (Mackney and Thomasson 1977) the SSEW has refined and improved the knowledge of the soil properties involved in the assessment of WRAP is available. In 1983 a set of soil maps at 1:250,000 scale covering the whole of Britain was published a culmination of a 5-year project (Ragg et al. 1983). These maps portray the soils in far more detail and give much greater accuracy than the earlier 1:1,000,000 map. Work is at present underway in collaboration with the Institute of Hydrology to produce an improved WRAP map at 1:250,000 scale. The preparation of this map will involve two computerised databases; the Soil Information System which includes a digitized 1:250,000 soil map and the Agroclimatic database. In the latter, the most useful parameter is the Field Capacity dataset. Field capacity period is defined as a duration in days and is the period of zero soil moisture deficit. As such, it is a meteorological parameter describing climatic wetness. In England and Wales the field capacity period ranges from a minimum of about 100 days in the eastern lowlands to 200-250 days in the west to >300 days in the uplands of Wales, northern England and Scotland.

For the assessment of the duration of waterlogging in soils, the field capacity period calculated from meteorological data is a better measure of climatic wetness than average rainfall. This is partly because it has an evaporation component, and partly because, unlike rainfall, it is a duration property.

A soil type which occurs in both dry East Anglia and the wetter western lowlands will straddle a broad range of field capacity days (approximately 125-225). Within this range of climatic wetness the depth and duration of waterlogging in this soil will vary from east to west. Following extensive monitoring of soil water levels, Wetness Class (and hence water regime) can be broadly predicted from soil profile and field capacity dates (Table 6). Hence the Wallasea series in the driest parts of East Anglia where the field capacity period is <100 days is Wetness Class II (soil water regime 1). In areas between 100 and 200 field capacity days it is Wetness Class III and IV (soil water regime 2). While in coastal Lancashire where the field capacity period exceeds 200 days it is Wetness Class V (soil water regime 3). Table 6 indicates the variation of Wetness Class for individual soil series depending on the climatic conditions prevailing.

Table 6 Field capacity days and Wetness Class

Soil series	**Field capacity days**					
	<100	100-125	125-150	150-175	175-200	>200
Evesham			III	III	IV	IV
Salwick			II	III	III	IV
Wallasea	II	III	III	III	IV	V
Wyre		II	II	III	III	IV

During the last 15 years water retention and porosity properties have been measured for over 3,500 horizons by the SSEW at its Soil Physical and Engineering Laboratory at Shardlow, near Derby. The routine measurement of packing density and air capacity has led to a better understanding of both the depth to impermeable horizons and the permeability above them. Although there have been relatively few measurements of hydraulic conductivity serviceable estimates can be made from the physical measurements undertaken at Shardlow (Talsma 1985, Puckett et al. 1985).

Data held by the Institute of Hydrology on Base Flow Index (BFI) is much more widespread than in the 1970's. Numerical relationships have been worked out between BFI and WRAP and will be used to refine the new WRAP map. In catchments giving similar BFI data it will be possible to test the consistency of the WRAP classifications. Where the WRAP classes appear anomolous a check will be made on the accuracy of both soil and WRAP classifications. Where both these assessments appear satisfactory it may be possible to determine any factors, other than soil physical properties, markedly affecting the BFI.

Conclusions

By the end of 1987 a revised WRAP map of Britain will be available at 1:250,000 scale based on the recently published 1:250,000 soil maps. Allied to this marked improvement in the understanding of the spatial relationships of British soils will be the better assessment of the soil properties incorporated in the WRAP model brought about by the large number of measurements made of soil physical parameters in the last 10 years by the SSEW. The 5 km datasets held in the Soil Information System including the digitized national map boundaries, 1:250,000 map units and field capacity days will add precision. The soil profile data on water retention and porosity properties extrapolated for soil series is particularly helpful when refining the assessments of depth to, and the permeability above, an impermeable horizon. In conjunction with the soils input from the SSEW the continued collaboration with the Institute of Hydrology will result in the use of many hundreds of BFI measurements to check WRAP assessments for catchments of varying sizes across Britain.

Acknowledgments

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THE USE OF REGIONAL SOIL SURVEY INFORMATION FOR FLOOD AND LOW FLOW DESIGN

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Abstract

Flow estimates are required at ungauged sites for a wide range of water resource and flood alleviation schemes in the UK. Underpinning any national design technique is the need for information on soil properties in gauged basins to calibrate hydrological models and of equal importance, the need for national maps of these same properties to enable models to be applied to ungauged basins. The Winter Rainfall Acceptance Potential map meets these requirements and its application in flood and low flow design is described. It is proposed that hydrological data can be used to estimate the characteristics of soil types at the catchment scale and also in conjunction with traditional regional soil survey data to interpret the spatial variability of soil types.

1. Introduction

Information on extreme discharges is required for a number of applications in both water resource and flood design. For example, predicting the low flow behaviour of rivers is necessary for designing river abstraction schemes, for assessing the dilution of sewage effluent and estimating the reservoir storage required to supply a given yield. In the area of flood hydrology data are needed for flood alleviation schemes in rural and urban areas, for sizing culverts and bridges and designing reservoir spillways. Where long discharge records are available then these can be used directly for hydrological design. However this is not possible where recorded flow data at the site of interest is either absent or too short for estimating extreme events. In these situations it is necessary to model catchment behaviour and these models must incorporate the important role of soil in determining rainfall-runoff relationships.

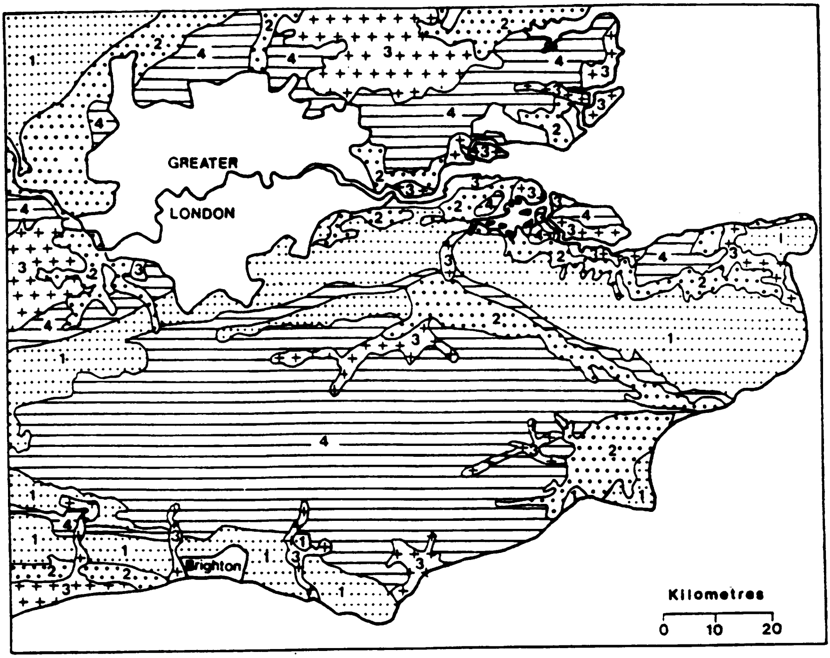
2. Flood estimation in the UK

The Flood Studies Report (NERC 1975) describes two approaches to flood estimation at the ungauged site. The first is based on a statistical relationship derived from 500 catchments between the recorded mean annual flood and single number basin characteristics which include indices of size, climate, slope and soil type. The second approach uses a rainfall-runoff model and the more reliable statistics of rainfall data. Both models take account of the soil properties of the catchment using a five class Winter Rainfall Acceptance Potential map (Farquharson et al 1978, Palmer 1986) which was produced by the Soil Survey of England and Wales. Figure 1 shows an example of this WRAP map which was used to calibrate the flood estimation procedures enabling floods to be determined at any location in the United Kingdom. Although the map was developed from the physical properties of soil associations it does not provide numerical information on the differences in flood response between soil classes. This was achieved by estimating the percentage runoff (the proportion of rainfall which becomes rapid response runoff) under standard conditions of catchment wetness and storm depth for 150 gauged catchments. Standard Percentage Runoff (SPR) was then related to the proportion of different soil classes in each catchment to provide a national equation for estimating SPR.

A continuation of the flood event data collection has enabled the original rainfall-runoff equations to be revised on a larger data set with improved quality control of both rainfall and discharge data (Boorman, 1985). The main development which relates to the WRAP map is in the following revised equation for estimating SPR, from the proportion of the five WRAP classes in the catchment.

$$SPR = 10S_1 + 30S_2 + 37S_3 + 47S_4 + 53S_5$$

This gives an estimated SPR of 10% for a catchment with exclusively WRAP class 1 soils and 53% for the most impermeable class 5 soils. The value of SPR for class 1 soils is lower than in the original equation, because many catchments with these soils also have significant urban development and this produced a bias in the original coefficient of S_1. The revised equation has allowed for the influence of urbanisation and hence the estimates of SPR from WRAP class 1 soils are lower.



Key.

WRAP class.

1 very high potential.

2 high potential.

3 moderate.

4 low.

5 very low.

Figure 1 **Example of Winter Rainfall Acceptance Potential (WRAP) Map for south east England (Farquharson et al 1978).**

3. <u>Revision to WRAP map</u>

With the advances in regional soil survey and the experience of relating hydrological response to soil types it has been possible to recommend changes to the original WRAP map. Of particular importance has been the completion of 1:250,000 scale soil maps in England and Wales and in Scotland which has provided a national coverage of soil survey information at a scale more appropriate for regional hydrology. Revisions to the WRAP map have been prompted both by the pedologist, following more detailed soil survey information, and by the hydrologist unable to reconcile the observed rainfall runoff relationship with that predicted by the WRAP map.

This disparity is illustrated by the results from a small research catchment in north west England where the percentage runoff ranged from 50% to 70%, typical of the most impermeable WRAP class 5. However the soil was classified as type 1 which, although in accordance with the published classification scheme (very permeable soils with no impermeable layer), did not support the hydrological response. This was a result of the dominant influence of the Carboniferous Limestone ; a fissure permeable rock with very thin, but permeable soils which resulted in a very low soil moisture storage capacity and very rapid response through the fissures in the limestone. These soil characteristics gave very high percentage runoff values from rainfall events. Although the results from one small catchment would not normally justify a reclassification of the WRAP map, in this particular example local field evidence (Gustard 1981) and the results from other catchments supported the change from class 1 to class 5.

4. Low Flow Studies

Although the Winter Rainfall Acceptance Potential map was developed for flood applications it has also been used for low flow estimation at the ungauged site. The flow duration curve and flow frequency curve are two of several summary statistics which are used for describing the low flow behaviour of rivers (NERC 1980). The flow duration curve is obtained by re-assembling the daily mean flows in reducing order of magnitude from which the proportion of time that a given discharge is exceeded can be identified. The 95% exceedance flow is commonly derived from the curve and used to assess the availability of dilution for effluent discharge, or the likelihood that a given level of abstraction can be maintained. If the discharge ordinate is standardized by the mean flow then curves can be compared from catchments of different sizes and with a different mean annual precipitation. Figure 2 illustrates the results from three flow records each draining a catchment of predominantly one soil class. The more impermeable WRAP class 5 soils result in low dry weather flows; in contrast the more permeable soils have higher low flows and reduced flood discharges.

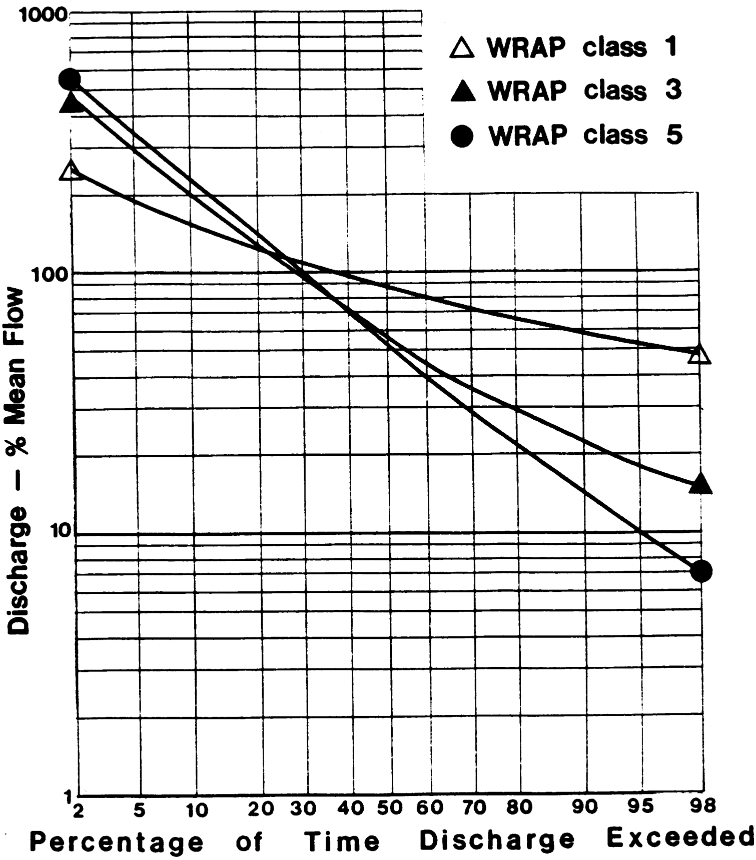


Figure 2 Flow duration curves from catchments with different soil classes.

The annual minimum series is more suited to estimating the severity of historic drought events or conversely for estimating the drought discharge of a given return period. Figure 3 shows plots of the annual minimum 10 day discharge for 3 catchments in different WRAP classes. Discharge is expressed as a percentage of the mean flow and a clear trend is evident

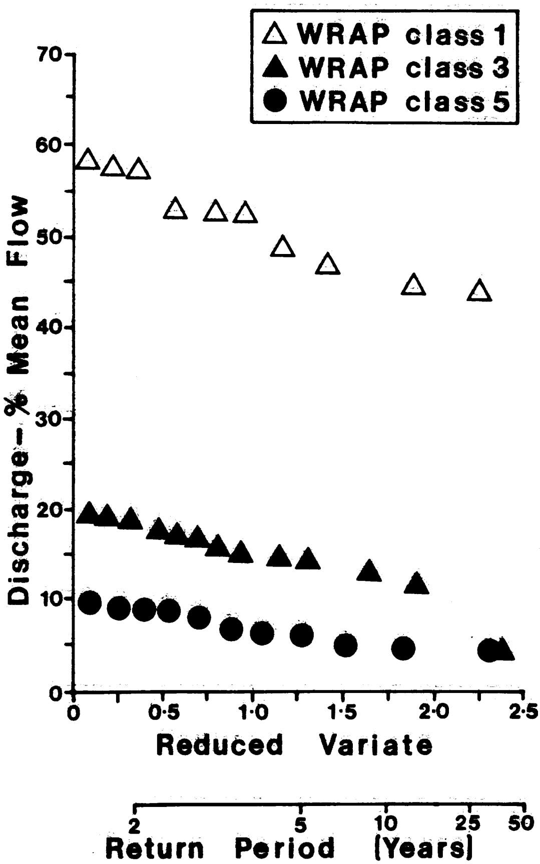


Figure 3 **Annual minimum series from catchments with different soil classes.**

with annual minimum discharges being higher on the class 1 soils than on the class 5 soils. This is a result of the greater storage capacity of the lower WRAP class soils which allow low flows to be sustained during dry weather and because of the close relationship between the WRAP class and underlying geology. For example class 1 soils often develop on aquifers which will maintain dry weather flows.

To avoid deriving a number of estimation techniques for each low flow statistic a Base Flow Index (BFI) incorporating the influence of soil and geology was developed. A computer program applies simple smoothing and separation rules to the flow hydrograph, as shown on Figure 4, from which the BFI is calculated as the ratio between the mean flow under the separated hydrograph to the recorded mean flow. Values of the index range from 0.89 for a permeable chalk catchment with a very stable flow regime (Figure 4) to below 0.20 for an impermeable catchment with a flashy flow regime. This index has been used as a key variable for estimating a range of low flow statistics at ungauged sites by using relationships between the index calculated from flow data and catchment geology (NERC, 1980).

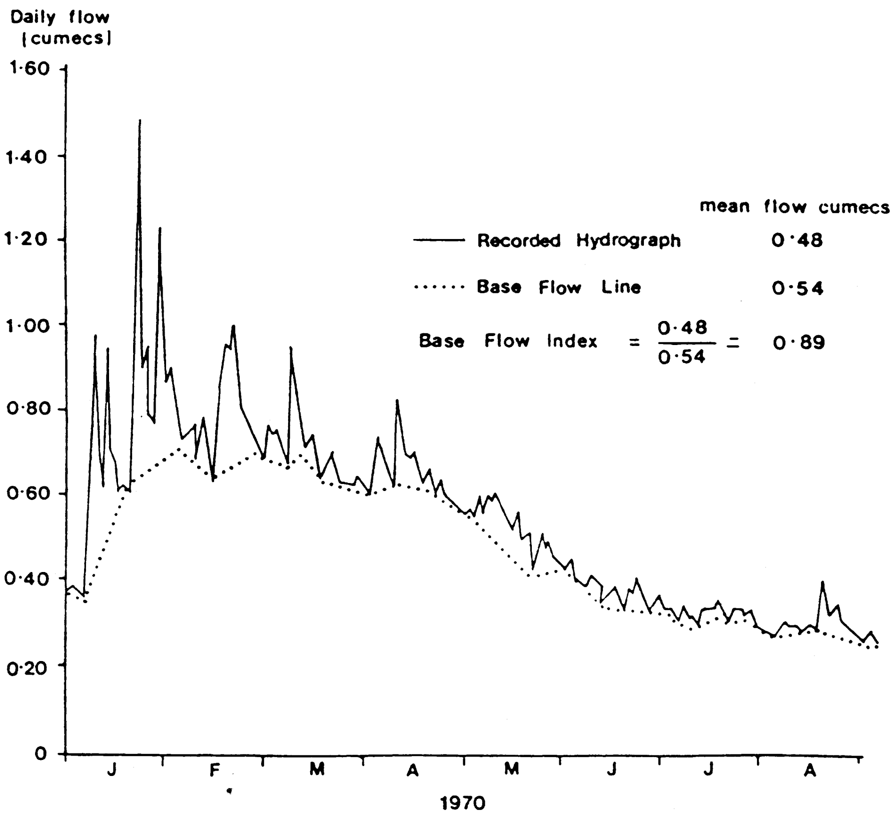


Figure 4 **Base flow separation for WRAP class 1 catchment with well sustained dry weather flow.**

5. Comparison of WRAP classification

It is important where different organisations are responsible for interpreting the WRAP soil classes that consistency of classification is maintained. In the application to rainfall runoff modelling this can be carried out by comparing the mean SPR of one soil type in one region with the mean from the same soil type in another region. However this approach is not always possible because of insufficient data. For example, of the 210 catchments used to revise the Flood Studies rainfall-runoff equations only 12 catchments are in Scotland and none are in Northern Ireland. The main reasons for the lack of suitable catchments are associated with the difficulties of accurate rainfall and flood discharge estimation in these areas. However, many sites have a good record of mean daily discharge and this has been used to compare soil classifications in different areas.

It has been shown by Boorman (1985) that the Base Flow Index (IH 1980) can also be used to predict SPR at sites which have only a mean daily flow record. The Index is closely related to the WRAP soil class with high BFI values from WRAP class 1 soils and low values from WRAP class 5 soils. Furthermore the index has been calculated for over 1000 catchments in the British Isles and provides a useful description of the spatial variability of soil types at the catchment scale. The following relationship with the WRAP soil classes was based on an analysis of over 600 catchments in the U.K.

$$BFI = .60 + .23S_1 - .03S_2 - .12S_3 - .17S_4 - .21S_5 \qquad se = .13$$
$$R^2 = 43\%$$

By deriving similar equations for different areas it has been possible to check the consistency of the WRAP map. For example Scotland was divided into three geological zones - the Southern Uplands, Midland Valley and Highlands. The absence of some WRAP classes in some regions prevented a comparison of all five WRAP classes and only the result for three soil classes are shown in Table 1.

	Class 3	Class 4	Class 5
United Kingdom	.48	.43	.39
Highlands	.50	.35	.40
Midland Valley	.60	.38	.30
Southern Uplands	.53	.44	.30
Northern Ireland	-	.40	.30

Table 1 Estimated BFI for 100% coverage of given WRAP class.

As a result of difficulties in regional soil surveying, in the interpretation of soil profiles and in relating simple hydrological indices to soil class, perfect consistency between regions cannot be expected. However if the mean UK results are used as a bench mark from which to compare regional variations then some inconsistencies in assigning WRAP classes are apparent. Most noteworthy are the high BFI values of class 3 soils in the Midland Valley, which from the hydrological evidence may be more appropriately mapped as class 2. There are also some inconsistencies in mapping class 4 soils many of which would be more appropriately mapped as class 5. Figure 5 shows the wide range in BFI values for those Scottish catchments with predominantly WRAP class 5 soils. It can be seen that the spread encompasses a wide range of different flow regimes from very impermeable (BFI 0.2 - 0.3) to relatively permeable (BFI 0.5 - 0.6) catchments. Inconsistent BFI values from one catchment should not be used to override the pedologist's judgement in WRAP classification. However a group of catchments with similar BFI values within an inappropriate WRAP class would suggest that a review of the WRAP classification in that area is justified. This approach has also been used to check the consistencies of a hydrological classification of soils carried out by different soil survey organisations in Europe. This work has enabled a draft WRAP map of northern EC countries to be produced for a European Flood Study (Gustard 1983). The analysis of hydrological data thus provides a useful integrated measure of the behaviour of soils at the catchment scale. This may assist in the interpretation of soil survey data and in mapping soil characteristics in gauged basins.

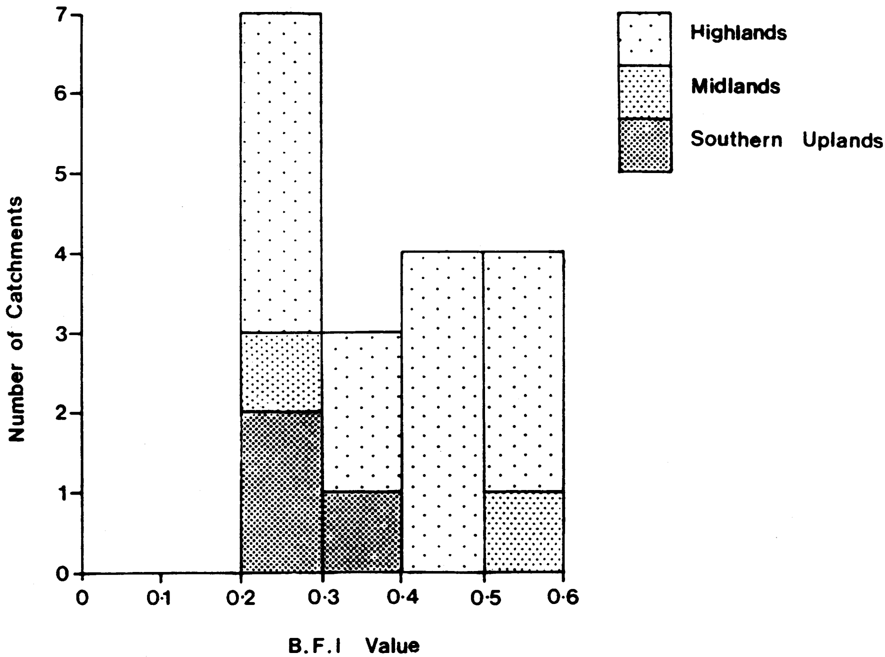


Figure 5 Frequency of BFI for WRAP class 5 soils in Scotland.

6. Conclusion

Regional flood and low flow estimation at the ungauged site requires an index of the hydrological characteristics of soils. To provide national design standards this index must be available for all locations irrespective of whether hydrometric or detailed process studies have been carried out in the area. The hydrological interpretation of regional soil survey information in the form of a WRAP map has provided a useful index to meet this specific requirement. A joint research programme between the Soil Survey of England and Wales and the Institute of Hydrology will develop a new hydrological response classification of the soils of the UK. This will make full use of the ability to overlay hydrological and soil survey data bases in order to combine hydrological information at the catchment scale with more detailed results from national soil surveys.

Improvements in interpreting the spatial variability of soils can best be achieved by close cooperation between soil scientists and hydrologists.

Acknowledgements

I would like to thank the organizing committee for the invitation to the NHP seminar on Water in the Unsaturated Zone. The Ministry of Agriculture, Fisheries and Food and the Department of the Environment have funded the work on flood and low flows respectively. Thanks are due to David Boorman of the Institute of Hydrology who was responsible for the work on flood estimation. This paper is a contribution to a study of Flow Regimes from Experimental and Network Data (FREND) which is part of IHP III, Project 6.1.

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THE USE OF DATA FROM THE DANISH SOIL CLASSIFICATION FOR ELABORATION OF ROOT ZONE CAPACITY MAPS.

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Summary

In 1975-80 a nation-wide Danish soil mapping was carried out dividing the agricultural land into eight classes according to the texture and organic matter in the topsoil. In order to improve this classification nationwide computerized landscape-, subsoil-, and wetland maps have been elaborated and pedological investigations carried out. Based on the different maps and analytical data, root zone capacity maps can be elaborated according to the method described in this paper.
The method used was first to combine the four maps in mapping units with nearly similar root zone capacity. Within each unit a typical soil profile was constructed, and based on root- and water retention data from pedological investigations the root zone capacity was calculated. Furthermore, areas with high possibility of capillary rise of water into the root zone was outlined.

Introduction.

In the period 1975-84 soil maps in scale 1:50000 were elaborated covering Denmark. They were mainly based on texture analyses of the topsoil and according to these, the agricultural land was divided into 8 classes. Furthermore, the soil maps give information about location of forest areas and urban zones (Mathiesen, 1980).

The soil maps have been widely used in the planning of rural land, e.g. for protection of valuable agricultural land around expanding urban settlements. Furthermore, the soil maps have been used for calculating the amount of water needed for irrigation (Madsen and Platou, 1983) and the actual and potential need for drainage (Holst and Madsen, 1986).

In order to improve the soil maps the following nationwide maps have been constructed

- landscape maps
- wetland maps
- clayey and sandy subsoil maps

All maps have been digitized in order to facilitate the production of computer-drawn maps at different map scales and with different combinations of parameters. This has also facilitated statistical treatment of the data and calculation of the area distribution within specific regions such as counties and catchment areas, to mention some.

The establishment of the new nation-wide databases combined with computerized pedological data makes it possible to construct maps showing the root zone capacity for different crops. The principles are:

- Construction of a combination map whose mapping units are based on a combination of existing computerized maps.
- Construction of a typical soil profile within the different mapping units.
- Calculation of root zone capacities for wetland- and non-wetland areas.

This paper describes a rapid method for mapping the root zone capacity by combining computerized soil maps and pedological data on Danish soils.

The construction of combination maps.

The combination map is based on four computerized maps covering the entire country. These four maps are briefly discribed.

The Danish soil mapping.

The soil maps (scale 1:50000) are based on texture analyses of soil samples from approximately 35000 sites, on slope classes constructed from topographic maps, and on the geological origin of the soil to 1-m depth, where this information is available. Samples for texture analyses were taken at all sites from a depth of 0-20 cm and at selected sites from a depth of 35-55 cm. In the laboratory texture, organic matter, and calcium carbonate were determined in all samples, and the results stored in a computer system (Mathiesen, 1980).

The agricultural land was classified into 8 soil types (mapping units) according to texture in 0-20 cm depth, table 1. Each soil type has been given a map colour code (1-8). The remaining areas are divided into urban- and forest areas.

Table 1: Definition of soil types and map colour codes for the soil classification of Denmark.

Map Colour Code	SOIL TYPE	JB-nr.	Percentage by weight				
			Clay < 2 µm	Silt 2-20 µm	Fine Sand 20-200 µm	Total Sand 20-2000 µm	Humus 58,7 % C
1	Coarse Sand	1	0-5	0-20	0-50	75-100	≤10
2	Fine Sand	2			50-100		
3	Clayey Sand	3	5-10	0-25	0-40	65-95	
		4			40-95		
4	Sandy Clay	5	10-15	0-30	0-40	55-90	
		6			40-90		
5	Clay	7	15-25	0-35		40-85	
6	Heavy Clay or Silt	8	25-45	0-45		10-75	
		9	45-100	0-50		0-55	
		10	0-50	20-100		0-80	
7	Organic Soils	11					> 10
8	Atypic Soils	12					

Landscape map.

For subdividing the map colour codes, a nation-wide landscape map has been elaborated in scale 1:100000. This map is based on previous landscape maps (Schou 1949, Smed 1979), topographic maps, and geological surveys published at scale 1:100000. The country is divided into 9 different landforms:

Old morainic areas from the Saale glaciation
Young morainic areas from the Weichsel glaciation
Outwash plains from the Weichsel glaciation
Dune sand areas
Old marine deposits (Yoldia).
Marsh areas
Littorina deposits and younger marine forelands
Reclaimed areas
Rock

Moreover some mixed areas were outlined where it was not possible to separate the landforms.

Maps showing the location of clayey and sandy subsoil.

From a geological map in scale 1:500000 (Bornebusch & Milthers, 1935) and from geological maps in scale 1:100000 the clayey and sandy subsoils have been outlined.

Maps showing the location of wetlands.

The wetlands were outlined from old topographic maps (1:20000) showing the extent of wetlands 60-80 years ago and from the landscape map. From the latter marsh areas, younger marine deposits and recent marine foreland together with areas below sea-level are considered wetlands, while in the other landscapes all areas with wetland signature on the old topographic maps are considered wetlands. These old maps were preferred to younger ones because of the recent decrease in wetlands due to drainage. The wetlands cover roughly 20% of the country.

Fig 1a shows a combination map from an area of roughly 20 km^2 in the young morainic area east of Gadbjerg, eastern Jutland.

The construction of typical soil profiles for the different mapping units.

A typical soil profile has been constructed for each mapping unit. These profiles are described according to texture, organic matter and calcium carbonate at three depths, 0-30 cm, 30-60 cm, and 60-120 cm. The following method is used:

From the soil map and the subsoil map the depth 0-60 cm and 60-120 cm are classified according to the MCC-system. The MCC-value from the soil map gives the information of the uppermost 60 cm of the profile, while the sandy or clayey subsoil determine the MCC-value within the depth interval 60-120 cm. The principle for the MCC-assessment of the subsoil is

- If the topsoil is MCC 1-3, the sandy subsoil will have the same MCC-value as the topsoil.
- If the topsoil is MCC 4-6 or 8, the sandy subsoil is as clayey as possible that means MCC 3.
- If the topsoil is MCC 7, the sandy subsoil is MCC 1.
- If the subsoil is clayey, it will in all cases be MCC 5, that means between 15-25% clay.

The mean textural composition of the different soil layers within a given area is then calculated from the texture-database established in connection with the Danish soil mapping The value for 0-30 cm depth is the topsoil value while the value for 30-60 cm depth is the subsoil value. The value for 60-120 cm depth is the 30-60 cm value, where the content of organic matter is reduced to 0.5%.

The calculation of the root zone capacity in non-wetland areas.

<u>The available water content</u>

The available water content in the different layers is calculated by regression equations combining the water content at field capacity (pF 2.0) and permanent wilting point (pF 4.2) with texture and content of organic matter. The regression equations based on data from e.g. Northern Jutland are as followed (Madsen & Platou 1983):

$$\text{pF } 2.0\ (\text{vol\%}) = 2.888 * \%\text{organic matter} + 0.490 * \%\text{clay} + 0.455 * \%\text{silt} + 0.164 * \%\text{fine sand} + 2.376 \quad r = 0.894, \quad s = 4.32 \tag{1}$$

$$\text{pF } 4.2\ (\text{vol\%}) = 0.758 * \%\text{organic matter} + 0.520 * \%\text{clay} + 0.075 * \%\text{silt} + 0.42 \quad r = 0.970, \quad s = 1.63 \tag{2}$$

This means that the available water content can be calculated from this equation

$$\text{AWC } (\text{vol\%}) = 2.13 * \%\text{organic matter} - 0.03 * \%\text{clay} + 0.38 * \%\text{silt} + 0.164 * \%\text{fine sand} + 1.96 \tag{3}$$

There is a clearly positive correlation between available water content and the content of organic matter, silt and fine sand, while there is none between the clay fraction and the available water content. In table 2 the mean available water content in the different soil layers is shown. The values are for the country as total.

Table 2. The mean available water content for the different map colour codes. The values are for the country as total.

depth	Available water (vol%)					
(cm)	MCC1	MCC2	MCC3	MCC4	MCC5+6	MCC7+8
0- 30	15.9	21.1	19.8	20.0	21.2	-
30- 60	11.7	17.3	16.4	17.2	17.0	-
60-120	9.0	14.8	13.4	14.9	15.1	-

The effective root depth

The effective root depth is defined as the depth of the soil in which AWC is equal to the amount of soil water utilized by the plants until wilting occurs due to lack of water.

When knowing the effective root depth for different crops in relation to soil type, it will be possible to calculate and map the root zone capacity. Based on investigations of Madsen (1979, 1983, and 1985) the following mean effective root depth can be established, table 3.

Based on the available water content in table 2 and the mean effective root depth in table 3, the root zone capacity can be calculated. Table 4 shows for the country as total the root zone capacity for grass, spring-sown cereals, and winter-sown cereals in relation to mapping units.

Table 3: The mean effective root depth in cm for different crops in relation to texture of top- and subsoil.

	grass		spring-sown cereals		winter-sown cereals	
	clayey	sandy	clayey	sandy	clayey	sandy
MCC 1	50	50	50	50	50	50
MCC 2	60	55	80	60	90	60
MCC 3	60	55	90	60	110	60
MCC 4	60	60	90	80	110	100
MCC 5,6	60	60	90	80	110	100

Table 4: The mean root zone capacity in mm for different crops in relation to texture of top- and subsoil.

	grass		spring-sown cereals		winter-sown cereals	
	clayey	sandy	clayey	sandy	clayey	sandy
MCC 1	71	71	71	71	71	71
MCC 2	115	107	145	115	163	115
MCC 3	109	100	149	109	176	109
MCC 4	112	112	156	141	186	171
MCC 5,6,8	115	115	160	145	190	175
MCC 7	>200	>200	>200	>200	>200	>200

Calculation of the root zone capacity in wetland areas

The root zone capacity in non-peaty wetlands is calculated as for non-wetland areas. The calculated root zone capacity will be the minimum available water content for plant production because capillary rise of water into the root zone from shallow ground water might be pronounced. Thus the wet-

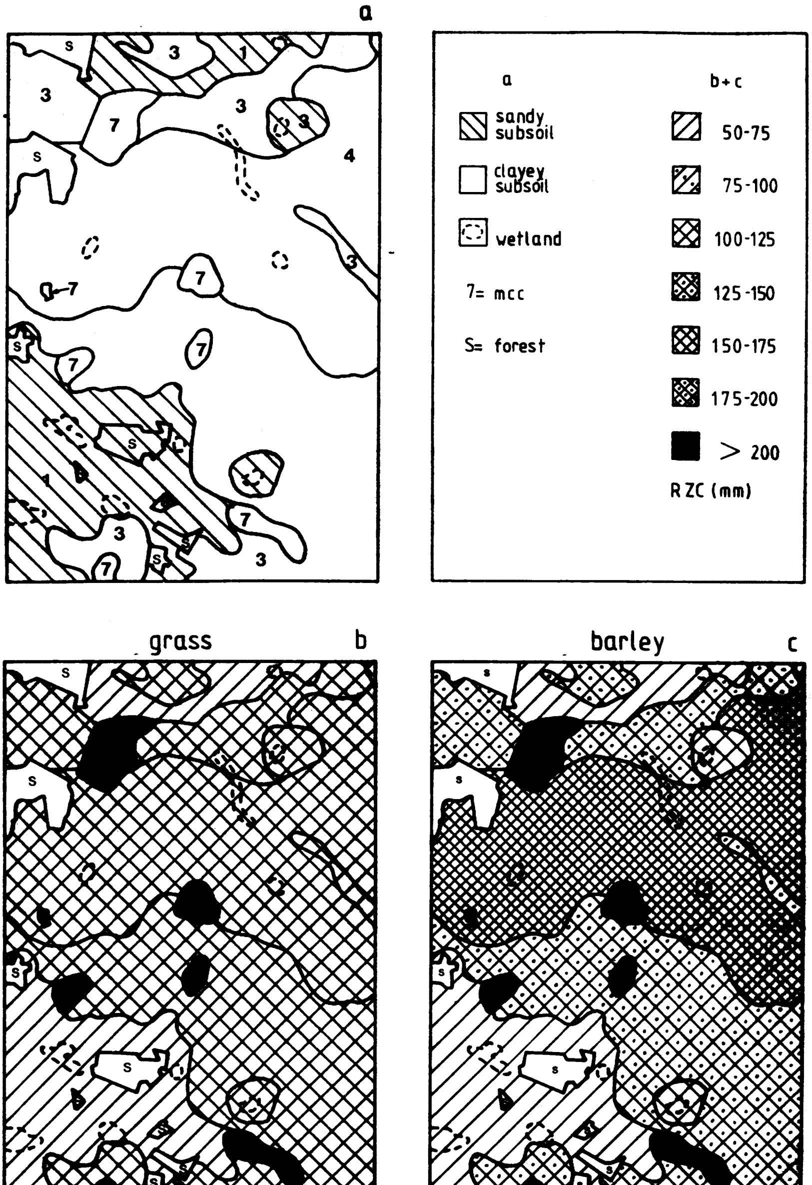

Fig 1. Combination map showing an area in eastern Jutland
and root zone capacity maps for grass and barley.

land areas are encircled on the map, and within these areas the total available water content for plant production might be much greater than the calculated root zone capacity.

The peaty soils are given the highest root zone capacity that means more than 200 mm.

Example of root zone capacity maps.

Fig 1 shows the combination map for an area just east of Gadbjerg in eastern Jutland and root zone capacity maps for grass and spring-sown barley. The areas is situated in the Weichsel glaciation landscape. The major part is clayey till, but in the north and southeast dune sand and melt water sand dominate.

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LAND AND SOIL MAPPING PROGRAM FOR SOIL MOISTURE EVALUATION

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ABSTRACT

Land, soil and vegetation mapping programs of interest for soil moisture are introduced. The land type classification is based on land conditions and productivity and is a part of the Economic Map System, wich covers approximately 180 000 km^2. For soil moisture evaluation land types can be grouped into soil depth and soil moisture classes.

Soil mapping of agricultural land will give more informations on soil characteristics and moisture contitions. The basic unit of the classification is soil series, which is based on soil genesis, drainage and other soil properties significant to plant growth. On the maps, soil series are subdivided into soil phases based on slope, stone content and rock outcrop.

A program for vegetation mapping for the entire country has been initiated. The natural vegetation gives informations on ecological contitions, and reflects variations in soil moisture. Vegetation types can be grouped according to the drainage classes of the soil maps.

I INTRODUCTION

Soil development and natural vegetation depend closely of the environmental moisture condition, and can therfore serve as useful indicators on soil moisture. Thus, soil and vegetation maps will give important information about the variability of soil moisture regimes.

In this lecture, I will give a general view of planned programs for land, soil and vegetation mapping of interest for soil moisture. I will introduce the products of The Norwegian Institute of Land Inventory, wich is the national institution for inventory of agricultural land, forests and natural vegetation.

II LAND TYPE CLASSIFICATION

The land type classification is a part of the Economic Map System, which is a combined topographic and thematic map. Land types, property bounderies and identification and ancient monuments represent the thematic content of the map.

The Economic Map System will cover approximately 180 000 km^2. The covering of the map distributed on different region types is schematically illustrated in figure 1. Field inventories has been carried out for approximately 170 000 km^2. The remaining area will be mapped within the 1980s.

Urban areas	3 000 km^2	Covering of the Economic Map System: 180 000 km^2
Intensive agricultural land	7 000 km^2	
Productive forest and scattered agricultural land	100 000 km^2	
Other areas below the timber line	30 000 km^2	
Productive areas above the timber line	40 000 km^2	
Mountain areas	128 000 km^2	
Permanent ice and snow	16 000 km^2	
Total area	324 000 km^2	

Figure 1. The Economic Map System - mapped area distributed on different region types.

The land type classification is a complex classification based on land condition, management condition for agriculture, forest production potential and soil properties of potential agricultural land. Land types are grouped into 10 major classes based on land condition. Agricultural land is grouped into 3 classes based on management conditions for agriculture. Forest land is grouped into 5 site classes based on forest production potential. The classification of potential agricultural land is based on soil properties:

1. Stone content - 3 classes
2. Drainage - 3 classes
3. Depth of peat - 2 classes
4. Peat decomposition - 3 classes

Peat depth and decomposition are also determined on bog soil capable for afforestation.

For soil moisture evaluation, land types can be grouped into soil depth and soil moisture classes.

Soil depth classes:

Class 1. Deep soils

Fully cultivated land
Other areas covered by mineral soil
Bog soils (can include forest)

Class 2. Soils with variable depth

Surface cultivaded land
Pasture
Forest land - dependent of site class

Class 3. Shallow soils

Shallow soil (can include forest)
Bare bedrock

Soil moisture classes:

Class 1. Permanent or frequently water saturated areas

Bog soils
Waterlogged mineral soils

Class 2. Periodically water saturated areas

Agricultural land
Potential agricultural land, not selfdrained mineral soil.
Forest land, dependent of site class
Other areas covered of mineral soil

Class 3. Infrequently or never water saturated areas

Selvdrained and excessively drained potential agricultural land
Shallow soil
Bare bedrock

Land types are published in Ecomomic Maps at the scale 1:5000, and for a minor part of the country at 1:10 000. Economic Maps are also published at the scale 1:20 000 where land types are printed in colour.

The land types and property boundaries and identification are digitized and stored on magnetic tape in a land register. Every map units (land type within a property parcel and map sheet) are identified by municipal, map sheet number, coordinates, altitude, property identification, land type and area. From the land register, derivid data and printouts can be made according to a great number of data combinations.

The land register covers approximately 60 000 km^2. The scheduled program is 8 000 km^2 pr. year.

III SOIL MAPS

The land type classification gives imperfectly information on soil characteristics, especially on existing agricultural land. A system for detailed soil mapping of agricultural land has therefore been outlined.

The soil classification is based on soil genesis and physical and chemical characteristics. The classification according to soil genesis is based on the Canadian System of Soil Classification, because of geological and climatic similarities between Norway and Canada. This system is a hierarchnical one in which the highest categorical level is <u>order</u>, which reflects the dominant soil forming processes:

Cryosolic order	- soils that have permafrost
Organic order	- organic matter accumulation
Podzolic order	- podzolization (vertical translocation of humus, iron and aluminium)
Gleysolic order	- oxidation-reduction influenced of water
Solonetzic order	- sodium salt accumulation
Luvisolic order	- clay eluviation
Brunicolic order	- weathering
Regosolic order	- weakely developed soils

Soils of the Chernozemic order are not expected to be found in Norway.

Soils of an order are subdivided into <u>great groups,</u> in which taxa are based on properties that reflect differences in the strength of dominant processes or a major contribution of a process in addition to the dominant one. Soils of a great group are differentiated into <u>subgroups</u>, based on kind and arrangement of horizons.

The basic unit of the Norwegian system is <u>soil series</u>. These units are formed by a subdivisjon of subgroups in the Canadian system. Soil series within a subgroup are differentiated on basis of geological deposits, depth to bedrock, texture, drainage,degree of decomposition of organic soil and climatic region.

The deposit classes are identical to those used i geological mapping in Norway. Soil textural classes are illustrated in figure 2.

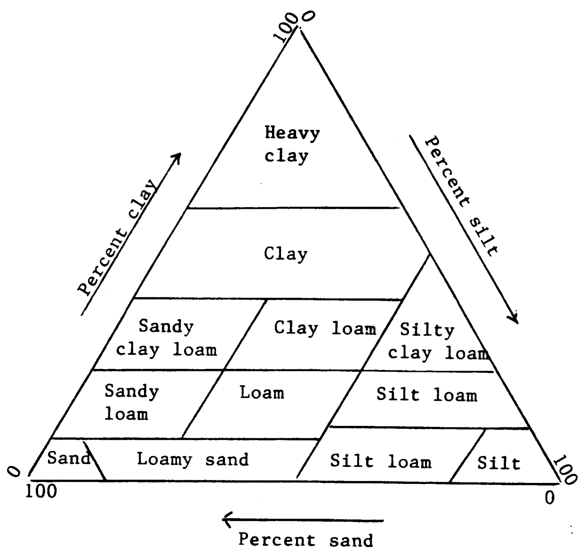

Figure 2. Soil texture chart.

The drainage classification is based on 6 classes:

1. <u>Excessively drained</u> - water is removed from the soil rapidly or very rapidly. The soil is sandy and very porous.
2. <u>Well drained</u> - water is removed from the soil readily. The soils commonly retaines optimum amounts of moisture for plant growth after rains or irrigation.
3. <u>Moderately well drained</u> - water is removed from the soil somewhat slowly, so that the profile is wet for a small but significant part of the time. The soils commonly have a slowly permeable layer,

a relatively high water table, additions of water through seepage, or combinations of these conditions.

4. Imperfectly drained - water is removed from the soil slowly enough to keep it wet for significant periods but not all of the time. The soils commonly have a slowly permeable layer within the profile, a high water table, additions through seepage, or a combinations of these conditions.
5. Poorly drained - water is removed so slowly that the soils remains wet for a large part og the time. Poorly drained conditions are due to a high water table, to a slowly permeable layer within the profile, to seepage or to some combinations of these conditions.
6. Very poorly drained - water is removed from the soil so slowly that the water table remains at or on the surface the greater part of the time. Soils of this drainage class usually occupy depressions and are frequently ponded.

Soil phases are units based on landscape properties that are not used as criteria in soil taxonomi. The major phases are listed below:

Slope class	Percent slope	Stone class	Percent stone in the soil	Rockiness class	Percent bedrock exposure
A	< 2	1	< 0,1	a	< 0,1
B	2 - 6	2	0,1 - 2	b	0,1 - 2
C	6 - 12	3	2 - 5	c	2 - 5
D	12 - 20	4	5 - 10	d	5 - 10
E	12 - 25	5	10 - 20	e	10 - 25
F	25 - 33	6	20 - 40	f	25 - 50
G	> 33	7	> 40	g	> 50

Soil series subdivided into soil phases are usually the smallest units on a detailed soil map.

Each soil series within a surveying area is object of a detailed soil profile description. Following properties are determined:

1. Information on the site - profile number, name of classification unit, location, elevation, land form, slope, vegetation or land use and climate.
2. General informations on the soil - parent material, drainage, actual moisture, depth to ground water table, presence of surface stones and rock outcrops, evidence of erosion and human influence.
3. Description of individual soil horizons - horizon symbol, depth of

top and bottom of horizon, colour, colour mottling, texture, structure, consistence, pores and contents of roots.

Samples for physical and chemical analysis are taken from each horizon. Physical analyses of textural fractions, porosity and pF should be of special interest for soil moisture evaluation.

Data sampled by soil mapping and soil description will be stored in a soil data bank, which is a computer based information system for storage, management and presentation of soil data.

Detailed soil maps at scale of 1:5 000 cover only minor areas of Norway.

The aim of The Norwegian Institute of land Inventory is to map 3-4000 km^2 agricultural land for the next 10 year, and 10 000 km^2 for the next 20-25 years. The planned mapping area is illustrated in figure 3.

IV VEGETATION MAPPING

A comprehensive vegetation mapping program has recently been initiated i Norway. The objective of this program is to complete a 1:50 000 scale vegetation mapping of the entire country within 25-30 years (figure 3). This will come in addition to the detailed vegetation mapping which has been conducted more arbitrarily for a number of years.

The natural vegetation is regarded as the best informant on ecological conditions. Plant communities reveal certain conditions in their environment through a natural selection of species that prefer these conditions. Thorough knowlegde of the ecological preferance of individual plants as well as those of plant communities enables us to read environmental characteristics through the vegetation.

A number of ecological factors, as well as interrelationships between these, contribute to the formation of a pattern of various plant communities throughout a landscape. A map of the occurance and distribution of thise plant communities provides information on ecological conditions. This is the core idea behind the vegetation map.

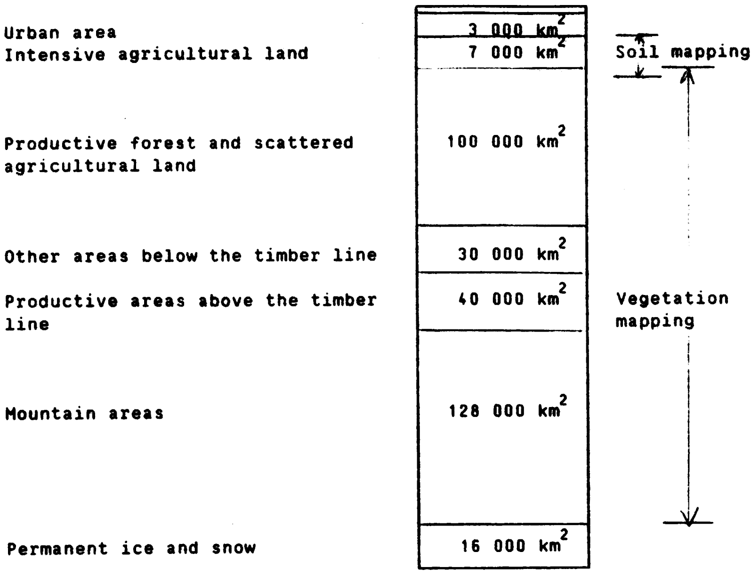

Figure 3. Planned soil and vegetation mapping.

A major ecological factor is soil moisture. Variations in soil moisture have a pronounced effect on the vegetation. Certain plant species and communities are adapted to wet conditions, their roots may obtain oxygen in saturated soil, or even completely submerged. On the other extreme, we find species and communities that may survive on a very limited amount of water supply. Between these two extremes we find a gradient of soil moisture. The amount of water available in the soil is reflected by the vegetation cover on any one site. The picture is, however, comlicated by the presence of other ecological factors. Also, in order to be able to handle the data, we must classify. This represents a simplification and leaves room for dangerous short-cuts to conlusions.

A tentative grouping of vegetation types according to the drainage classification of the soil maps, may come out as follows:

<u>Exsessively drained</u>

Lichen and dry heath communities, rock and scree communities, sand dunes (10 vegetation types)

<u>Well drained</u>

Heath communities (16 vegetation types)

Moderately well drained

Moderate snow beds, blueberry heath communities, low herb meadow communities (18 vegetation types)

Imperfectly drained

Extreme snow beds, moist heath communities, tall herbs and coastal meadow communities (16 vegetation types)

Poorly drained

Swamp forest and dry bogs and fens, wet heath and meadow communities (9 vegetation types)

Very poorly drained

Wet bogs and fens, sedge and rush communities on lake shores (4 vegetation types)

USE OF SOIL PROFILE ATTRIBUTES TO STRATIFY DRAINAGE AND SOIL MOISTURE REGIMES OF A CATCHMENT

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Summary

The key to developing a more sophisticated understanding of catchment hydrology is to use the soil profile as an hydrologic indicator and devise soil survey techniques that can adequately interpret and evaluate the attributes observed. The hydrological classification of soil moisture regimes must be based on water storage and water flow properties of the soil. Collaboration between hydrologists and soil scientists has been poor in the past and therefore proposed systems of classification are very generalized and therefore not very useful. Available knowledge indicates, however, that many possibilities can be afforded to relate specific soil moisture measurements with respective profile properties, so that water balance studies and other hydrologic investigations and evaluations can be greatly facilitated. The paper gives examples of interesting work that has been done in the past.

Introduction

The determination of soil characteristics such as water content, water storage capacity, residence time and flow terms such as infiltration capacity and permeability belong to the most difficult and tedious of hydrological investigation problems. Apart from differences in meteorological elements and vegetation, there are other factors, such as the heterogeniety of the soil both in the vertical and horizontal direction. For large areas it is not possible to achieve a network of measuring stations of sufficient density to cover the mosaic structure of soils and their water regimes, see Kutilek 1971. Instead methods must be developed to coordinate field measurements with attempts to stratify the catchment into more or less homogeneous zones, where soil characteristics can be generalized and classified in a satisfactory manner and but a limited number of observations need be made to arrive at good appreciations of the moisture and flow conditions that prevail. Soil is the interface not only between rainfall and runoff (river flow), but also between rainfall and evaporation and between soil water and groundwater. The key to developing a more sophisticated understanding of catchment hydrology is to utilize the soil profile as an hydrological indicator and to devise soil survey techniques that can adequately interpret and evaluate the attributes observed. This paper will discuss these aspects. Attention will be focused on the glaciated landscape of the boreal zone of Scandinavia.

Hydrological Classification of Soil Moisture Regimes

The hydrological classification of soils may essentially be limited to two main components - water storage and water flow properties.

Storage is a measure of the water volume that can be and is stored in the soil (potential and actual water volume content). It will depend on the soil depth available for water exchange and also on the character of the pore system that is to hold the water.

The available soil depth will depend on the occurrence or not of an impermeable soil layer or of a groundwater table, below which the moisture content remains constant either at a level near to or equivalent to full saturation. In depressions soils will generally exhibit limited soil depths for water storage due to high water tables. In other cases a water table may be absent, but still the water content remains fairly constant throughout the year at a level at field capacity which is close to full saturation and within a meter or less of the soil surface, therefore limiting water storage for excess rainfall, see Figure 1. In other cases an impermeable layer may be present, beyond which moisture conditions will not be affected by infiltration of rainwater. Finally there are situations where soil depth almost seems to be limitless (more than 5 and 10 meters). Such conditions can be found on thick deposits of coarse textured soils (sands and gravels) where drainage is such that no groundwater is formed and water can flow down freely.

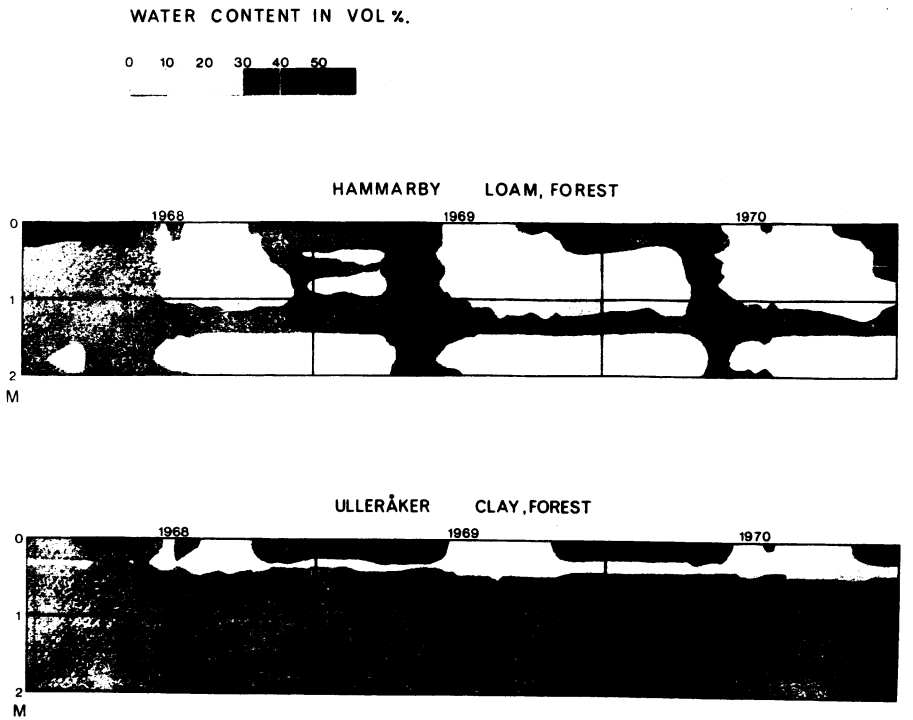

FIGURE 1 Chronoisopleths of soil moisture for 3 years 1968-70. Hammarby shows moisture variations down to more than 2 meters. Ulleråker indicates storage changes during the 3 year period limited to the upper 0.5 meters of the profile.

Character of pore system involves both the total volume and the pore size distribution of the soil medium throughout the profile. The main factors that affect the pore system character are the texture of the soil and the structure, ie state of aggregation, compaction, development of cracks, root and animal channels, etc. The finer the pores the greater will be the adsorption of water through capillary forces and thereby the less volume of air-filled pores which are still available for further water storage. Thus clay soils are always moist and plastic below the top soil layer exhibiting relatively small moisture variations both with depth and time. Coarser textured soils show greater variability, since their pore systems bind water more weakly, see Figure 1. The presence of larger pores is important for the air volume of the soil and their continuity for the exchange of gases, ie removal of carbon dioxide and renewal of oxygen. Where the pore system is very fine as in non-aggregated clay soils, water will generally fill the pore volume to such an extent that anaerobic conditions develop whereby certain chemical substances will be reduced thereby changing the colour of the soil. Thus poorly drained soils show "cool" colours whilst well drained and aerated soils show "warm" reddish colours.

Flow properties characterize the ease with which water enters and moves through the soil. They are composed of two factors, the permeability of the soil to water and the driving forces which act, ie on the combined effects of the gravitational force and on the capillary forces of the soil matrix.

Permeability to water can and will vary throughout the soil body, depending on the nature of the pore system and on the moisture conditions that prevail. The value will be higher the coarser the pore system is and the more water that fills this system. Sands therefore exhibit higher permeabilities than do clays and saturated sands higher than do dry sands. The nature of the pore system is determined by the texture and the structure of the soil, whereas the drainage and climatic conditions of the area will affect the moisture level of the soil.

The driving forces will depend on the slope of the land in the area (slope gives the gravitational force for drainage of the area) and on the capillary forces as they are governed by the texture, structure and degree of water saturation of the soil. Surface water runoff will be governed by land slope, groundwater flow by the change in water table elevation and soil water flow by the capillary potential gradient as measured by tensiometers along the direction of flow. In homogeneous soils there is a close relationship between water content of the soil and the equivalent capillary potential, such that approximations of the capillay potential can be obtained from measurements of the water content of the soil.

The hydrologic soil water regime is reflected by the morphology of the soil profile and the genetic soil forming processes leading to the formation of the genetical soil type, see Figure 2. Thus the classification and distinction of soil water regimes may be derived from mapped data and from the results of soil surveys, if these are evaluated from maps of genetical soil units. The possibilities of utilizing aerial photographs is evident. Another important source of information for larger areas is climatic maps. A further significant source is vegetation maps. The most suitable form in which to distinguish soil moisture regimes is one based on ecological considerations to define the levels of moisture conditions, ie

1. Wet state = the soil at full saturation
2. Moist state = moisture level from full saturation to field capacity.
3. Semi-moist = moisture between field capacity and point of decreased plant availability of moisture.
4. Semi-dry = level between point of decreased availability to wilting point.
5. Dry state = moisture level below the wilting point.

In the climatic zones of Scandinavia, the top soil layer may in many cases during the year pass through all of the above 5 moisture levels. Deeper horizons seldom reach the dry state due to the fact that precipitation exceeds evaporation most of the year.

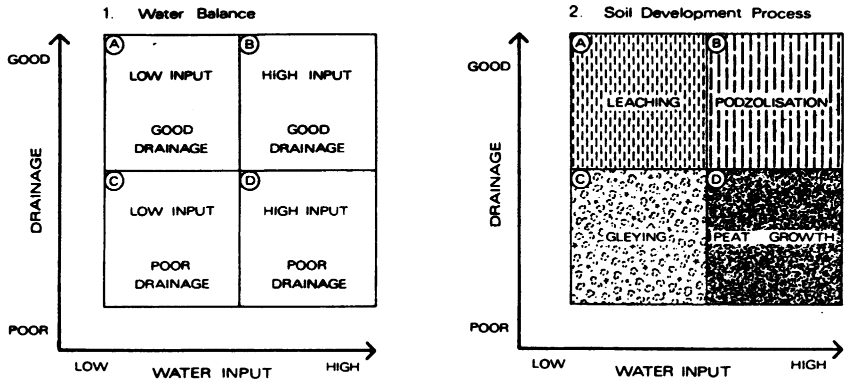

FIGURE 2 Diagram showing the effects of different soil moisture regimes on profile development when soil depth is not limited. When drainage is efficient (good) low rainfall (water input) will cause leaching whilst increased rainfall will finally result in the development of a podzolised profile. When drainage is poor equivalent water inputs will produce gleying respectively peat formation. Thus the soil profile reflects the moisture status, that prevails.

Using the above listed 5 levels of moisture contents of the soil, moisture regime classification can be achieved by further taking into account the duration of these storage levels and how the moisture is stratified throughout the profile, see Kutilek 1971. The classification involves an attempt to generalize the amount of water stored in the soil profile with regard to the variation in moisture level that prevails with depth and time. To solve this problem Kutilek has suggested a classification where distinctions are made concerning the following hierarchical taxa:

1. Class: based on the moisture level which predominates for the whole profile and for most of the year (6 or more months).

2. Order: based on the degree of variability of the moisture levels throughout the year:

A. Permanent order with a dominating moisture level in the profile for more than 10 months of the year and non-dominating levels for less than 2 months.

B. Temporary order having a dominating moisture level for more than 6 months and non-dominating levels for a time of between 2-6 months.

C. Indifferent order for which no level has a duration equal to or more than 6 months and does not contain any marked subsiduary levels.

3. Suborder: determined according to the season, whereby the order is further broken down into sub-taxa.

4. Type: distinguishes how the moisture is stratified vertically, here simplified by dividing the profile into topsoil denoted T and subsoil denoted S. The following types are distinguished:

a. Permanently decreasing: S has a lower moisture level than T and not even a short inversion takes place.

b. Temporarily decreasing: same as a) except that a short inversion takes place for up to 2 months.

c. Fluctuating: S can be lower or higher than T as regards moisture level, but inversion sets in for more than 2 months.

d. Constant:S and T show similar class and order for the same interval of the same duration.

e. Temporarily increasing: S is higher than T and an inversion takes place up to 2 months.

f. Permanently increasing: as for e), but no inversion takes place.

The classification presented above is not meant to be rigid, but subject to adjustments to suit local conditions of the region where it is to be applied. Applying the system to the two soils in Figure 1, a likely designation would be:

Hammarby: T3B (2,4) - S3B (4,2) d, which would be interpreted to mean that the topsoil layer T has a moisture level class 3 (semi-moist) and a moisture variability order B (temporary) showing in parenthesis that moisture levels 2 (moist) and 4 (semi-dry) prevail for between 2-6 months of the year. The subsoil S has a similar classification except that for periods between 2-6 months level 4 is more prevailent than 2. Thus for the profile as a whole the moisture stratification type is d (constant)

Ulleråker: T3B (4,2) - S2A (3) c, which means that the topsoil T has the same classification as Hammarby subsoil, while the subsoil S is wetter with class 2 (moist) and variability order A (permanent), resulting in a profile stratification type of c (fluctuating).

Unfortunately there has in the past been rather insignificant liaison between pedologists/ecologists on the one hand and soil physicists/hydrologists on the other. Thus where profile and site descriptions have been made as to soil genesis and vegetation descriptions and classification, specific and year round measurements of soil moisture have seldom been made. Conversely soil water stations have seldom been subject to thorough soil genetical and vegetational descriptions. Thus material is lacking to work with and develop soil moisture regime classifications useful for both categories of scientists. This situation must be remedied.

Relationships between soil moisture regimes and soil profile attributes

The downward movement of water is one of the prime factors in the transformation of a parent material into a soil with characteristic horizon differentiation. Alluvial deposits, moraines, loess, etc are not necessarily soils in the strict sense if they lack properties which reflect the effects of the soil-formning factors (climate, organisms, parent material, topography and time). According to the regional concept of soil formation, all parent materials within a given environment should produce soils that ultimately possess similar morphological features. The rate at which the soil climax is reached varies, however, considerably depending on the constitution of the soil matrix, particularly with its capacity for water percolation. Thus if a sandy and a silty deposit are placed in a podsol environment and inspected after say 10 000 years, the former is likely to show a welldeveloped orstein layer, whereas the latter may show only very weak podsolozation. The two profiles probably would be classified as belonging to different soil series. Carrying this argument still further, it is conceivable that variations in water permeability not only affect the rate of development of the regional soil type, but also in extreme cases may give rise to the genesis of different climatic soil types and even aclimatic profiles. Thus the interpretation of soil attributes must be made with much care and with thorough knowledge of the environment and its history, see Jenny 1941.

Tamm 1931 has early declared his belief that the multitude of the profile differentiations that may be found in the forest environment of Sweden is caused by the variable character of the vegetation, which, in turn, is conditioned by the properties of the parent material, particularly its water permeability and its relation to the groundwater table. Where the subsoil is sand or light textured moraine, the groundwater level is low and the vegetation consists of spruce forest with Vaccinium. The soils are iron podsols. On less permeable material, with higher groundwater, Sphagnum plant associations cover the soil and humus podsols prevail. Where the groundwater table reaches the forest floor and downward movement of water is nil, the soils are of a peaty nature or belonging to the gray-blue swamp group.

Lundmark 1974 treated several thousands of plots collected by the Swedish National Forest Survey to rank with respect to their height growth effects on pine and spruce various site characteristics, such as occurence of superficial water flow, thickness of humus layer, texture, moisture index, etc. Of the forest types between which he differentiated, he found that those which were rich in lichens gave very clear indications of the moisture status of the site. On the whole his studies indicate that classification of the moisture regime of the soil can with quite good confidence be judged on the basis of the forest vegetation characteristics.

Many pedologists have placed great emphasis on the effects of groundwater and drainage on the characteristics of soil profiles. In fact groundwater has even been discussed to be treated as a major soil forming factor since it can be made to vary independently of the otherwise recognized factors (climate, organisms, topography, parent material and time). The well known "hydrologic podsol series" presented by Mattson and Lönnemark 1939 can be used to illustrate the general effects of groundwater, see Figure 3. A sequence of soils, often termed catena, are here depicted within a short range, indicating the effects of topography on groundwater and profile development. At the upper end, the soil is not affected by groundwater while at the lower end there is a wet depression covered with water-loving mosses. Soil attributes which are affected by groundwater level and the general moisture regime of the site are organic matter content (also nitrogen content), acidity (pH, base status) and colour (redox status of iron, manganese and organic compounds). It is not possible to discuss in detail these attributes, just to point out the value of attempting to relate these in more quantitative detail with hydrological observations of the storage and flow conditions of water in specific environments.

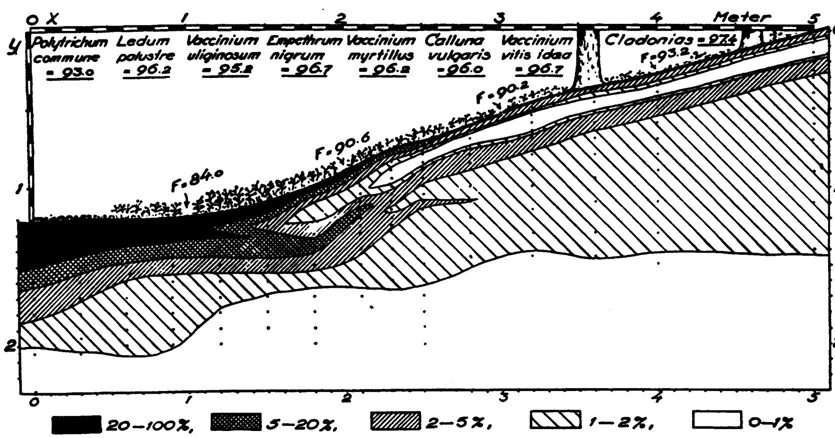

FIGURE 3. Hydrologic podsol series showing sequence of forest floor vegetation, distribution and content of organic matter and profile changes as a function of slope and drainage. From left to right the sequence of soils is peat podsol (0-1 along X-axis), Humus podsol (1-2) and iron podsol (from 2 and onwards). Figures below plant species and above the F-layer (F= litter) denote the percentage loss on ignition. The diagram serves to demonstrate the close inter-relationship between various environmental factors. From Mattson and Lönnermark.

Concerning colour, Kowalik 1972, has demonstrated the relationships with oxygen status **when he used oxygen diffusion** electrodes in Gniew clay in Poland. The oxygen diffusion rates measured were found to be related to both moisture content and to the presence and not of "cool colours " at the site of measurement, see Figure 4. Gley morphology and cool colours (Munsell designations 7.5Y, 10Y, N) appeared first att oxygen diffusion rates below about 30 $x10^{-8}$ $gcm^{-2}min^{-1}$. Furthermore brown soils were found to be much better oxidized than black earths developed from the same Gniew clay, showing that also the organic matter material is affected by the level of soil aeration.

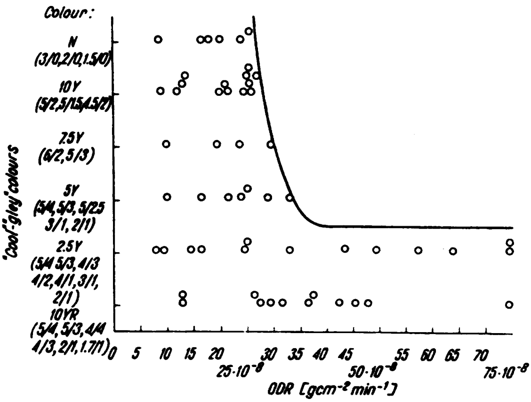

FIGURE 4 The relationship between oxygen diffusion rate and soil colour designation for Gniew clay. For ODR above ca 30 note absence of cool colours. From Kowalik 1972.

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SUMMARY OF DISCUSSION 1

The question was raised if the representatives from UK and Denmark feel that the scale chosen for the soil mapping gives a satisfactory background for the modelling, and if a more detailed mapping would give any important improvements in the modelling.

The experience from the UK is that new soil maps in scale 1:250000 will be very useful for the new WRAP-maps which will now be published in the same scale. For physically based models, the data will be unsuitable in many cases, and a more detailed soil mapping is needed. However, even in these applications, soil mapping in a scale of 1:250000 may assist in transferring model parameters to ungauged areas. After the period with application of soil maps in 1:250000 it is probably possible to make better use of more detailed soil series in scale 1:50000. This would, however, only be cost effective in some areas.

Appropriate scale depends upon the desired application. There must also be potential customers for the maps. Single parameter maps are made very cheaply in Denmark, because the data are stored in a database. Examples of applications in Denmark are

1) Urban land use planning
2) Water-use planning
3) Nitrate runoff evaluation
4) Acidification risk in Jutland
   Four classes are defined according to pyrite content.
5) Erosion risk, based upon calculated K-values, slope etc.
6) Drainage requirement. Four classes are defined.
   Good agreement is found with farm survey data.

Denmark is well covered by soil maps as opposed to Sweden and Norway. Lack of contact between different disciplines has hindered such advances in Sweden. However, the relative economic importance of agriculture is much greater in Denmark than in Sweden and Norway. It seems quite unrealistic to propose a programme in which these countries soils would be mapped. Quaternary maps, which are much more abundant than soil maps in Sweden and Norway, may be of help in large-scale spatial variability estimates.

We have soil data for geologically different sediment types that may be used in connection with geological maps. In Denmark there is a close correlation between soil maps and geological maps. In Northern Ireland the WRAP-maps were successfully based on drift geology maps.

The importance of the soil resources should decide whether soil mapping is needed. In northern Scandinavia, the hydrology is greatly controlled by spring melt of snow and ice, and the soil properties are a less important. In relation to water quality studies, however, soil properties are of importance, even in connection with the spring floods.

PHOTOGRAMMETRIC METHOD FOR STUDYING THE PHYSICAL ENVIRONMENT OF COARSE HETEROGENEOUS SOILS

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Summary

A new tehnique in describing the micro-topography and determining the density and the porosity of each horizon of a till soil in a forested slope is described. The technique uses the photogrammetric way of analysing stereo images. A camera base is put up approximately 2 meters above ground. 35 mm cameras are used to photograph a marked square meter under the camera base. The first stereo image will show the soil surface with the ground vegetation. After the first image investigation is started inside the square. The first horizon is then taken away and weighed on a balance. Then the next stereo pair is taken. The investigation and the exposure of stereo pairs continues in succession, i.e. the ground inside the square can be treated in a differentiated way, by alternatively removing soil and other objects (roots and stones, etc) and then taking stereo pair images. During this investigation pF-cylinders and soil samples have also been taken from each horizon to make pF-, permeability-, texture- and water content analysis. The project is financed by the Swedish Council for Forestry and Agricultural Research, SJFR.

Introduction

The method is part of a project financed by the Swedish National Science Research Council, NFR, "Studies on the relationships between drainage and soil physical properties". Attention is focused on the factors that rule moisture variations in shallow forest soils, especially glacial till which is the most common soil type in Sweden. The aim of the project is to interpret the interaction between soil physical properties (including pedological features) and the drainage characteristics and from this to derive the flow paths and rates of water in different parts of the terrain.

Since the physical environment in Swedish forestland is generally very heterogenous (undulating terrain even in the micro-topograhy, with effects due to vegetation and profile developement) a photogrammetric method was regarded worth trying, in order to analyse large samples of the physical environment. Studies of the same kind have been done by Andersson and Håkansson (Andersson and Håkansson, 1963) on agricultural soils.

Study area

The investigation has been located to a forest slope situated on till. The area, LUND, is situated 35 km northwest of central Stockholm, see figure 1.

The slope has an elevation of approximate 10%. The vegetation consists mainly of coniferous trees, approximate age about 50-60 years.

Seismic investigation has shown that the thickness of the whole soil profile down to the bedrock is about 2 - 3 meter (perhaps even less). So far no geotechnical drilling has been done, because of difficulties in penetrating the till.

Figure 1. The investigated area,"LUND", catchment is marked out by the white line. Aerial photo from 600 meters height. 35 km from NW of Stockholm

0 50 100m

Ditch which drains the catchment

Scale
1:2500

Photogrammetric Method

In a forest slope a camera base was put up approximately 2 meters above ground. The equipment consisted of two single reflex Canon Fl cameras for 35 mm film with motors mounted on a 70 cm wide base. The motors ensured filmdrive during the investigation. Each camera was equipped with a 28 mm lense. The lenses were focused on 3 meter in order to give the best possible depth of sharpness. Below the base a square of 100cm x 100cm was marked out, within which physical investigations could be permitted. Around this square control points were marked out so that the different stereo pairs could later be connected for the volume analysis. See Figure 2.

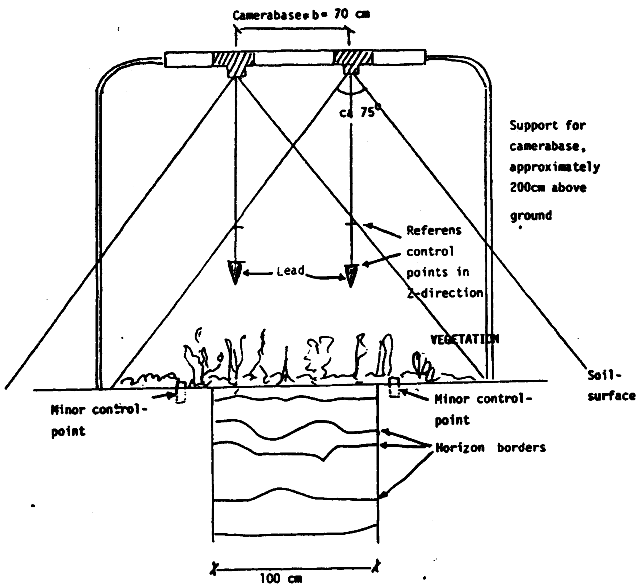

Figure 2. Camera support

The photogrammetric technique starts with an analysis of the vegetation inside the square. With each camera a photograph is taken, forming the first stereo pair. The L-horizon (litter) is then uncovered and the litter is collected and weighed on a balance. The next stereo pair is then taken. The investigation and the exposure of stereo pairs continues in succession, i.e. the ground inside the square can be treated in a differentiated way, by alternatively removing soil and other objects (roots and stones, etc) and then taking stereo pair images. Any heterogeneity can be studied, even if it concerns only part of the observation square, as long as the detail is photographed before and after removal of the studied objects. See Figure 3.

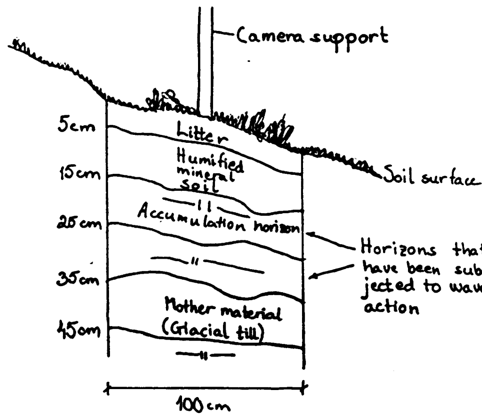

Figure 3. Principal stratification of the soil at study area LUND.

By treating the stereopair in a Wild STK1-stereocomparator each horizon can be analyzed topographically in relation to the ground surface (zero-level). The accuracy with which the measurements is done is around 10 microns (10E-6 mm), which makes it possible to determine the volume between the horizons very accurately. The soil between each layer is weighed and representative soil samples are taken to determine various physical attributes (water content, texture analysis, pF-analysis, etc).

With the help of the photoimages any heterogeneous feature can be documented, such as distribution of plants, roots, boulders and stones. Even differences in topography that cannot be seen during the investigation can be documented through the photogrammetric treatment of the stereo pairs.

With this method there is consequently a possibility to indicate probable pathways for the water by looking at the physical and pedological features. Soil physical variability such as density and consequently the porosity can also be determined. By developing the technique there will be possibilities to determine the density variations around and below boulders and roots, etc.

Film

The films that are used are Ectachrome 400 ASA and black and white film, 400 ASA. The black and white film is the best film to use with which to make the stereo measurements, but the colorfilm is superior in determination of the characteristics of the objects.

Measurement in Wild STK1-Comparator

The stereo treatment in the comparator was made at the Department of Photogrammetry, KTH. In each pair of the stereo images the co-ordinates were measured in a 1 mm grid (corresponding to approximately a 70 mm grid on the ground). Approximately 14 x 14 measurement points were measured inside each square. The comparator connected to a microcomputer collected and stored, the X and Y-co-ordinates (on floppy disc) of the left image and the X and Y-parallaxes between the images. See Figure 4.

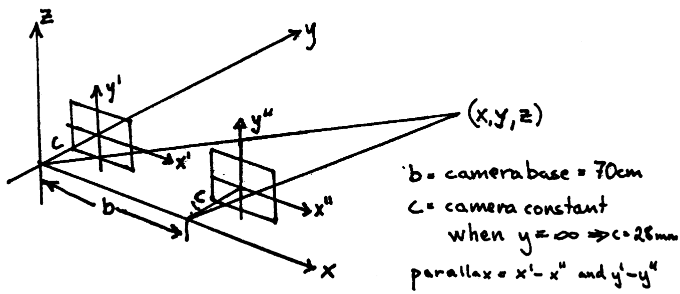

Figure 4. Optical geometry for how a point on the ground (X,Y,Z) is registered in each camera (from Torlegård, 1982).

To get the ground co-ordinates in X, Y and Z from which the volumes could be computed (co-ordinates for the surface and for each soil horizon) the data will be treated in a computer, which also makes it possible to make accurate calibrations of the camera constant and the radial distortion of the lenses (camera constant, see figure 4). The camera constant, C, refers to the length of the camera lense.

At the moment no computer calculation has been made on how the bulk volume varies from top to bottom of the pit that has been dug. A vital program for the calculation of each soil surface co-ordinate has had an undetectable error embedded which has prolonged the calculations. It is however possible to use other programs on micro-computers. This procedure is more time consuming.

The pF-analysis has shown that the porosity decreases from top to bottom. This means that the bulk density is increasing with the depth. This corresponds to other investigations made in till soils, Lundin (1982).

Accuaracy of the Photogrammetric Technique

To verify the accuracy in the measurements, one stereopair showing a horizon of boulders was selected. These boulders can verify the technique when the particle mass density is assumed to be 2.65 g/cm3. The boulders were regarded as a layer and were weighed on a balance. The volume is gained through the formula, V= mass/density.The volume measurement in the Wild STK1-comparator will then actually show how fine the technique is.

The accuracy in the volume measurement depends partly on the performance of the stereooperator, but mainly on how well each parallax measurement represents its volume element, see Figure 5, i.e. the profile of the surface. No computer calculations have been made so far but, estimations of the accuracy by hand , considering an error in the measurement of the Z-co-ordinate of about 130 microns in the image (representing 10 mm on the ground), have shown a maximum error of the estimated volume of approximately 2 per mille.

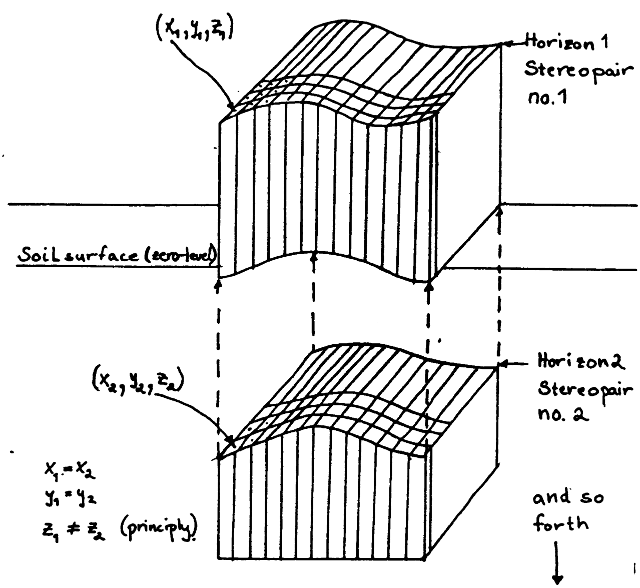

Figure 5. Principal figure on how each stereo pair represents each horizon border.

Investigation of the soil profile

The first time this method was tried out, the investigation was done in an archeologic way. This was done principally to learn the stratigraphy of the soil profile; how the roots were situated, frequency of macropores, stones, cracks and fissures in each horizon. This knowledge is very important since the aim of the project is to interpret if the water is moving in a differentiated way or like a water front in this soil.

The soil that has been taken away during the photogrammetric technique is weighed on a Statmos balance that is self horizonting. This means that it can cope with small differences in elevation without giving the wrong result. The maximum error in each measurement is about +/- 0.5 kg. This is very little conserning the total amount soil that is taken away can weigh anything from 500 to 1 000 kg for one pit.

The boulder horizon which is taken away to account for the accuracy of the technique will also be accurately measured in volume and weighed in the laboratory. This is done to verify the technique even more. Similar studies of investigating boulder frequency have been made by Bethlahmy (1952), Crocker and Major (1955), Donahue (1940), and Lyford (1964).

pF- and permebility-analysis

From each horizon 6 pF-cylinders (Ø 7,2 cm), three of 5 cm-type and three of 10 cm-type were taken to analyse the water tension caracteristics. The cylinders were randomly spread and carefully put down into the soil. Since it is till there were difficulties especially in the horizons with coarse material, especially the horizons which have been subjected to wave action and the dence basal till. The analysis show that the relationship water content / tension can vary a lot at the same level (in the same horizon). This is visible especially at lower tensions. At higher tensions the different samples show very similar behaviour. In the litter horizon (pF-cylinders covering the range from L-horizon through to the upper A-horizon) the differance can be accounted to the presence of macropores (see Beven et al, 1982). In the coarse horizon (mentioned above) it can be accounted to its larger structure.

Permeability-analysis have shown that presence of the above mentioned macropores have a greate influence on the saturated hydraulic conductivity. pF-cylinders taken from the same horizon show variability in the saturated hydraulic conductivity.

Results

The method will give very descriptive images of each soil horizon from above. These images will describe the situation of certain phenomena; e.g.where boulders are, how roots are finding their way through the soil etc.

If the pit is documented with colorfilms it is also possible to detect moist areas.

Eventually the technique will result in describing the microtopography of each soil horizon, dry and wet bulk density and porosity of the soil at different depths.

The investigation of the slope at three different levels has shown what significance of macropores must have. At the lower levels of the slope, especially in the humified mineral soil (see figure 3) , macropores are found in a great number. The appearance of the macropores are especially old roots (only a cylindric shell remains intact, with no inner material) with an inner diameter of more than 1,5 cm. One macropore crossing the investigated pit laterally was found at one level in the slope.

Discussion

The technique has great future views since it has very good accuracy in determining the microtopography. The conventional way to take pictures of a soil profile is from the side. This gives a very nice stratification at one point in the terrain. By taking stereo images of a limited area (for instance one square meter as in this case) the images will show a microtopography of each border to a horizon. The technique will eventually show in differences in density and porosity. In this case this is very interesting, since the aim is to describe how the water of the unsaturated zone in a till soil of this kind is moving.

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VARIATION OF THE SATURATED HYDRAULIC CONDUCTIVITY IN THE UNSATURATED ZONE OF NORWEGIAN SEDIMENTS

By

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Summary

Variation of the hydraulic conductivity in Norwegian sediments classified by genesis and lithofacies was studied. The discussion is based on 103 measurements with the inverse auger hole method , all made in the C-horizon.

The data give a rough outline of the variation in different sediment groups. However, the range of variation would be larger if more samples were obtained. The range of variation is larger in the genetic than in the lithofacies groups. Hydraulic conductivity values, measured by the inverse auger hole method, are often higher and may show a lower range of variation than laboratory-derived data. This is most distinct in the tills.

The data can be used for a very rough prediction of the hydraulic conductivity in some groups. This is a valuable supplement to quarternary maps and can rationalize field investigations in connection with onsite wastewater disposal.

Introduction

The saturated hydraulic conductivity (K-value) measured above the groundwater table is an essential parameter when evaluating soil suitability for disposal of wastewater. However, direct measurements of the hydraulic conductivity in the field or laboratory can be timeconsuming and expensive. Simpler means of predicting the hydraulic conductivity are therefore of interest. If the hydraulic properties of soils or sediments could be assessed from maps or a subjective field evaluation of soil or sediment characteristics, field investigations could be rationalized. The possibility of predicting hydraulic properties by the above approaches depends upon the accuracy needed in the prediction and the range of variation or confidence limits within which the hydraulic properties can be predicted. The accuracy of prediction depends upon the variability within a soil or sediment unit.

The determination of hydraulic properties, and subsequently the suitability for wastewater disposal, from field evaluation of soil type or soil class has shown to be difficult due to the large variation within each group and small differences between the groups (Derr et al. 1969, Hill 1966, Bouma 1977). However, there are indications that a better distinction between the hydraulic properties of soil or sediment groups is possible when the sedimentological properties of the subsoil are the basis for classification (Hill 1966, Derr et al. 1969, Lindblad 1981, Lundin 1982).

In this article the variation of the hydraulic conductivity in the C-horizon of sediments classified by genesis and lithofacies is discussed, and the practical consequences of the results in connection with onsite wastewater disposal are evaluated.

Materials and methods

All the measurements were made with the inverse auger hole method (Kessler & Oosterbaan 1974), using standardized procedures and equipment (Miljøverndepartementet 1980, Jenssen 1982). The method measures the flow through a soil surface of at least 1625 cm^2 and more than 0.5 m^3 of undisturbed soil is affected by the measurement. The variability due to method is therefore low because the affected soil volume often exceeds the hypothetical representative elementary volume of the soil (Bouma 1983).

103 measurements make the basis for the analysis. The measurements were made in connection with consultant work or in connection with testing of different methods for measurement of the saturated hydraulic conductivity. The sampling procedure is therefore biased and has not been designed to characterize the variability of different Norwegian sediments. 1-5 measurements were used from each location. The individual measurements at one location are treated as independent because of the distance between the measurements (up to several hundred meters) and the sediment differences observed; however, some dependency cannot be excluded. This is a weakness statistically. Because of the low number of samples extensive statistical treatment of the data has been avoided. All the measurements were made in the C-horizon of the soil profile and at least 60 cm below the soil surface. The individual measurements are accompanied by a detailed soil profile description and an analysis of the grain size distribution of the fractions smaller than 32 mm.

Results and discussion

Variation and frequency distribution of, genetic groups.

The frequency distribution of the genetic groups are shown in Fig.1. The total number of samples is only 103 and the number in the different groups varies from 6 to 26. Therefore, no effort was made to determine the type of distribution. The marine and shore groups contain less than 10 samples, and are therefore not mentioned in the discussion.

Because of the low number of samples and a biased sampling procedure it can be questioned whether any of the distributions shown in Fig.1 are realistic. To elucidate this the sedimentology of the groups is discussed and the range of variation is compared to data from the literature.

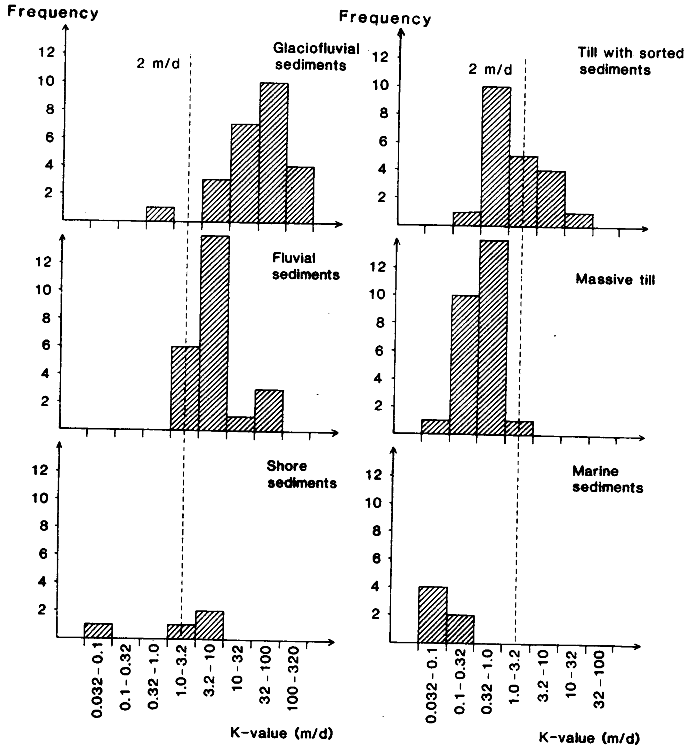


Fig.1 Frequency distributions in the genetic groups. The dotted line indicate a K-value of 2 m/d which is a limit in the Norwegian regulations for soil treatment of seawage effulent.

A Wilcoxon test showed that the distributions of the four largest genetic groups were statistically different. A difference between the till groups can be expected from sedimentological considerations. The group "massive till" contains compact homogeneous tills which also are referred to as basal tills or lodgement tills (Haldorsen et al.1983). The other till group contain sediments with lenses or bands of sorted material. These sediments were generally less conslidated than the massive till. Comparison of the analyzed grain size distributions (Fig.2) indicates that the tills with sorted sediments were slightly coarser and on the average better sorted. The difference in hydraulic conductivity is therefore a consequence of the differences in structure, grain size distribution and density.

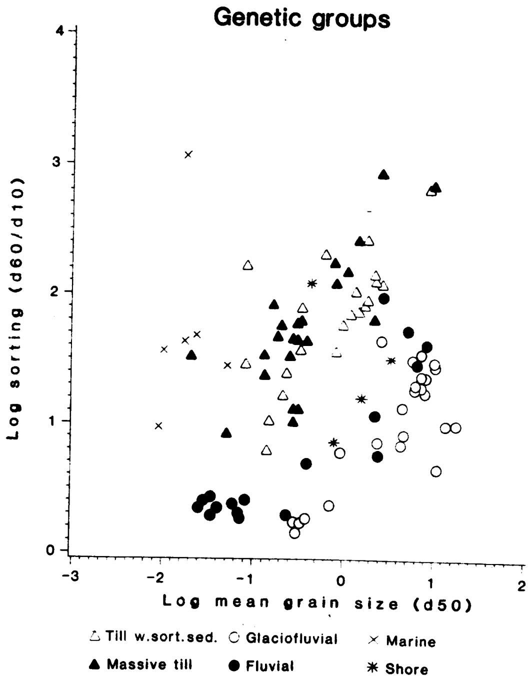


Fig.2 The samples grouped according to mean grain size, sorting and genesis.

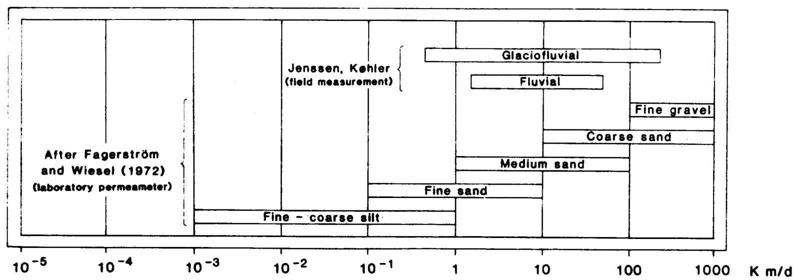
A difference between the tills and the fluvial groups could also be expected, because the fluvial sediments are better sorted (Fig.2). The samples presented also show a difference between the fluvial and glaciofluvial groups. This may be an effect of the biased sampling procedure, as a cluster of the fluvial sediments was found in the fine sand-silt fractions (Fig.2). However, the results of Selmer Olsen (1954) and Derr et al. (1969) also indicate that glaciofluvial material is coarser, more poorly sorted and generally has a higher water conducting ability than fluvial sediments.

The difference in K-distributions between the groups is supported by the sedimentological differences. This indicates that the samples presented represent a rough outline of the real areal distribution of K-values in the genetic groups.

In Fig.3A the range of variation in the fluvial and glaciofluvial groups is compared to the range of variation in different well sorted sediment fractions suggested by Fagerstrøm & Wiesel (1972). In Fig.3B the till groups are compared to data from Fagerstrøm & Wiesel and Dahl et.al (1981). Both figures indicate that our measurements yield higher conductivities and show a lower range of variation than the measurments by Fagerstrøm & Wiesel and Dahl et.al.

A

RANGES K-VALUES



B

RANGES K-VALUES

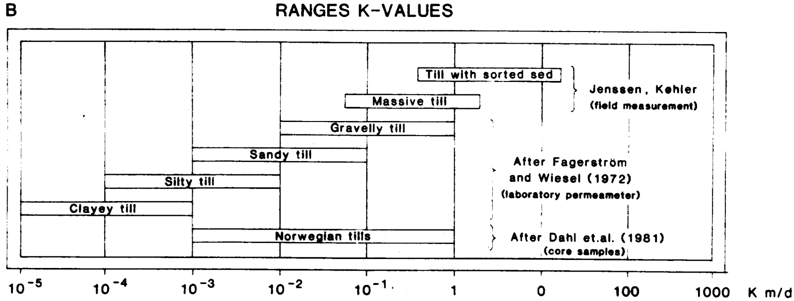


Fig.3 Ranges of K-values in different sediment groups and measured by different methods.

Our measurements were made in situ. The K-values of Fagerstrøm & Wiesel refer to measurements on disturbed samples compacted in a laboratory permeameter. In laboratory permeameters and the method used by Dahl et al. the sedimentary and soil structure present in situ is not fully accounted for. This structure can have a dominating effect on the hydraulic conductivity, especially in fine-grained and poorly sorted sediments. In situ measurements can therefore exhibit K-values 1-2 orders of magnitude higher than measurements in a laboratory permeameter (Healy and Laak 1973, Haldorsen et al. 1983). This fully agrees with the observed values, and shows the importance of structures. In tills the primary structure was shown to be essential for the hydraulic conductivity (Haldorsen et.al. 1983). It is also possible that structures caused by soil forming processes have influenced the measurements, because the depth of measurement seldom exceeded 1.2 m.

The measurements with the inverse auger hole method may overestimate the hydraulic conductivity in certain fine-grained sediments. This problem, which is currently being investigated by the author, has probably not significantly affected the measurements in tills. A possible error due to overestimation is probably less than one order of magnitude.

The range of variation in the glaciofluvial group is about 3 orders of magnitude. We think the same range can be expected in the fluvial group. The range in both groups would probably increase if more samples were present and a better sampling procedure was used. The range in tills would also increase if more samples were presented and the clayey tills included. However, for sandy and silty tills we feel that the measured values give a fairly good picture of the range of variation that can be expected. The data show that the range of variation is lower when measured in situ than in a laboratory permeameter. However, the range of variation found in the genetic groups was so large that prediction of the K-value from genesis alone becomes very rough.

Classification of sediments by lithofacies

The system for classifying sediments by lithofacies proposed by Miall (1978) and expanded by Eyles et al. (1983) is rapidly gaining recognition. The lithofacies classification is based on a subjective description of grain size, sorting and sedimentary structures, all features which may be essential to the hydraulic conductivity.

The lithofacies classification contains more groups than classification by genesis, but each group is distinctly characterized by the features mentioned above. Each lithofacies group therefore defines a specific sediment population which again may exhibit specific hydraulic properties. However, the classification by Miall and Eyles was not developed to distinguish populations with different hydraulic properties. Characters such as grain size and sorting, which are of extreme importance to the hydraulic conductivity, are only roughly described, while description of sedimentary structures is stressed. The classification of Miall and Eyles has therefore been modified to better describe hydraulic properties and to suit the purpose of classifying sediments for onsite wastewater disposal (Table 1). Sediments classified genetically as tills are usually covered by the lithofacies term diamict.

Table 1. Suggested lithofacies code. The code is preliminary and a revised edition is in progress. Example: Sih=horizontally laminated, well sorted medium sand.

MAIN GROUP Grain size	SUBGROUPS Sorting	Sedimentary structures
G - gravel	g - well-sorted b - poorly sorted.	n - massive, no structures h - horizontal bedding imbrication t - trough crossbeds p - planar crossbeds
S - sand	c - coarse well sorted i - medium (intermediate) well sorted f - fine well sorted b - poorly sorted	t - trough crossbeds p - planar crossbeds h - horizontal lamination
F - fine	s - coarse to medium silt well sorted y - clay well sorted b - poorly sorted mud (loam)	n - massive no structures l - horizontal lamination a - aggregated*
D - diamict	m - matrix-suported very poorly sorted or bimodal	n - massive, no structure often high consolidation l - loose, often low consolidation may be stratified, graded and can contain lenses or laminae of sorted material.

* not a sedimentary but a soil structure.

Frequency distribution, lithofacies groups

The lithofacies grouping (Fig.4) is mainly according to grain size and sorting, because the sedimentary structures were not known for all the samples. In the tills the suggested lithofacies codes, Dmn and Dml, and the genetic classification gave indentical groups. These groups have been discussed above. Fig.4 and Fig.5 indicate that the

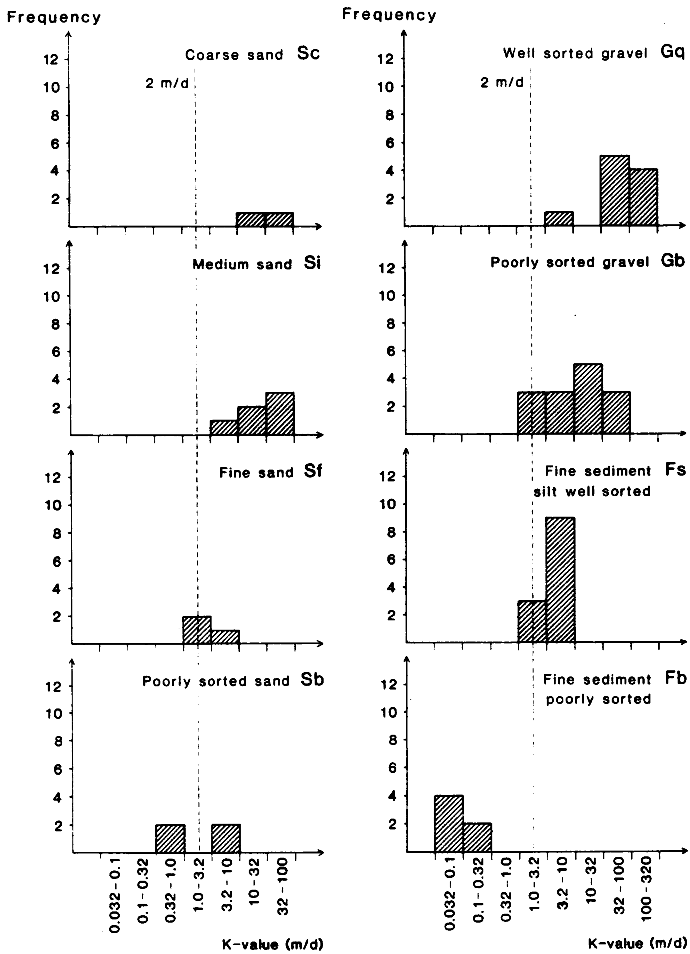


Fig.4 Frequency distributions in the lithofacies groups. The dotted line indicate a K-value of 2 m/d which is a limit in the Norwegian regulations for soil treatment of sewage effulent.

lithofacies groups are narrower and more distinct than the genetic groups. The range of variation in K-value is two orders of magnitude or less. The results of Fagerstrøm & Wiesel (1972) show that a range of two decades can be expected in sediment fractions similar to some of the lithofacies groups. A prediction of the K-values within a range of two orders of magnitude may therefore be possible from subjective field evaluation of lithofacies.

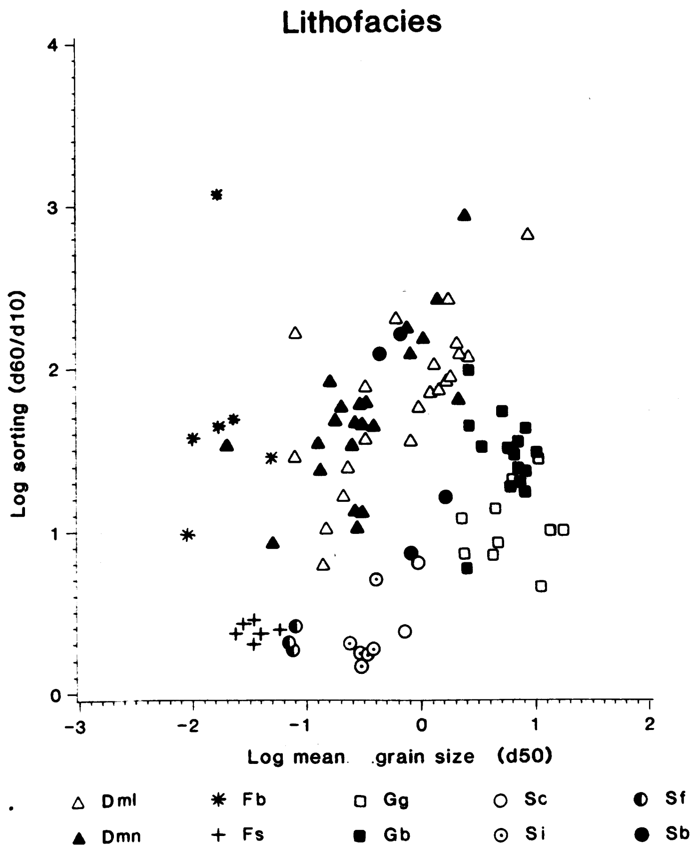


Fig.5 The samples grouped according to mean grain size, sorting and lithofacies. The lithofacies codes are according to Table 1.

Practical application

An interesting comparison may be made of the range of K-values in the different sediment groups and the current Norwegian regulations for small-scale wastewater infiltration systems (Miljøverndepartementet 1985). Infiltration of wastewater by conventional methods is not recommended if the hydraulic conductivity is below 2.0 m/d.

The presented data, (Fig.1) indicate that in marine and massive till deposits soil absorption of wastewater by conventional methods will be difficult. The areas of highest conductivity in massive tills are only suitable if special technology is used. In the till with sorted sediments the possibility of finding suitable sites for conventinal systems is higher than in the massive till. However, the conductivities can be expected to lie in the critical range around 2.0 m/d. Careful field investigations are therefore necessary whenever the genesis of the sediment is classified as till. In the fluvial and glaciofluvial sediments, conventional infiltration will be possible on the majority of sites. Similar information can be obtained by knowing the lithofacies groups (Fig.4). Because the lithofacies groups are more distinct than the genetic groups it may be possible, in some cases, to size infiltration systems for wastewater from knowledge of the lithofacies group alone.

Knowledge of suitability for onsite wastewater disposal in different genetic groups can, in combination with quaternary maps, aid land use planning and help consultants in selecting site and investigative procedure. Furthermore we feel that the informative value of the quarternary maps would increase if they were supplied with lithofacies codes.

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SOME ASPECTS ON SEDIMENT STRUCTURES AND HYDRAULIC CONDUCTIVITY IN TILL

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ABSTRACT

This paper refers to a project with the main purpose to determine the influence of different sediment structures on hydraulic conductivity in till. The project is sponsored by the Swedish Council for Building Research and carried out within the Urban Geohydrology Research Group at Chalmers University of Technology. The project started in 1 July, 1984 and is planned to be finished during July, 1987.

Methods for collecting, analysis and description of undisturbed till samples have been worked out. Preliminary results show that there is a notable difference in hydraulic conductivity due to fabric variations. Besides fabric, there are many other factors affecting the hydraulic conductivity. Although very little is known about these factors an attempt has been made to quantify the influence. The structural factors listed below are closely interrelated in such a way that the factors of influence cannot be added. Nevertheless, it is possible to state that sedimentary structures have a significant influence on the hydraulic conductivity.

Sediment structure factor Note, grain-size and sorting are not considered	Magnitude of influence on hydraulic conductivity (very preliminary values)
Grain shape (sphericity)	≈ 5
Pore shape (incl. fractures)	strongly varying
Roundness	?
Grain surface structure	≈ 5?
Aggregate occurrence	≈ 5
Fabric	≈ 5-10
Porosity-pore orientation - spatial pore-size variations	?
Spatial grain-size variations	significant
Mineralogy	significant

Introduction

For a long time it has been a general opinion among hydrogeologists that, besides the texture, the structure has an influence on the soil moisture conditions, as well as hydraulic conductivity in soft sediments, (cf Fair & Hatch 1933, Fraser 1935, Knutsson 1971, New York DOT 1973, Currie 1979, Dahl et al 1981, Prudic 1982, Lundin 1982, Williams et al 1983, Haldorsen et al 1983). The problem: structure vs. hydraulic conductivity is also touched upon in engineering geology concerning dams of diamicton sediments (cf Bernell 1976, Kjellberg et al 1985). However, the sediment structure is built up of many separate factores and the importance of each of these factores is very little known. It is clear that the sediment properties often are the weakest link in the imput data for hydrogeological calculations.

Empirical methods for calculation of intrinsic permeability and hydraulic conductivity in soft sediments have so far been based on the relationship between pore volyme and conductivity. Hydraulic conductivity, K, is defined by the equation

$$K(m/s) = C \cdot d^2 \cdot (p \cdot g \;/\; \mu)$$

where C is a dimensionless empirical constant, d (mm) the characteristic flow space (pore size), g (m/s^2) the gravity acceleration and p (kg/m^3) and $\mu (Ns/m^2)$ the density and the dynamic viscosity of the fluid.

However, Gustafson (1983), among others, has shown that within certain grain size intervals there is a direct relationship between the pore size and the grain size. This means that instead of the flow space the pore size could be used. This is, for instance, the case in the well known Hazens formula

$$K \ (m/s) = 1{,}157 \cdot 10^{-4} \cdot d_{10}^2 \ , \qquad d_{60}/d_{10} \leqslant 5$$

where d_{10}(mm) is the grain size for 10 weight-% passing quantity according to grain size analysis and d_{60} is the grain size for 60 % passing material. In general form and irrespective of the fluid properties the intrinsic permeability, k, can be expressed as

$$k = C \cdot d^2$$

If d in this equation is a measurement of the representative grain size diameter, then the constant C is related to all other factores that influence the degree of flow through the media; those are:

* Grain shape (sphericity)
* Pore shape (incl. fractures)
* Roundness
* Grain surface structure
* Occurrence of aggregates
* Fabric
* Porosity - pore orientation - spatial pore size variations
* Spatial grain size variations
* Mineralogy

Most of the empirical equations for hydraulic conductivity calculations, perhaps with exception for the Fair & Hatch (1933) formula, do not separate and take into consideration asymmetry in the factors above. Instead the empirical methods are defined and valid for well sorted and isotropically porous media.

It is well known that till often has a well developed structure, considering foliation, bedding and fabric. It it often possible to identify these or other specific structures within separate till facies.

Methods

Sample preparation

Samples of undisturbed till have been taken from a drumlin-shaped accumulation at Tahult, 10 km east of Gothenburg. On the top of the moraine two 5 m long and 3 m deep shafts, parallel and perpendicular to the moraine were excavated. A detailed study of the till was carried out and it was regarded as a lodgement till.

At a depth of 2 - 3 m, lumps of undisturbed till were carefully excavated from the shaft walls. The till contains about 6% clay and is quite easy to handle. The lumps were formed with a knife to samples 70 x 80 mm and placed in 90 mm high PVC-cylinders with diameters of 95 mm. The samples are taken with known vertical and horizontal orientation.

In the laboratory weight and moisture for each sample were measured and dried weight was calculated.

Permeameter tests

For sediments with hydraulic conductivity (K), 10^{-2}-10^{-8} m/s different types of permeameters can be used for conductivity measurements. In this project a permeameter was made for each sample by pouring epoxi-resin over the undisturbed samples in the PVC-cylinders. Before this, the surface (1 mm) were dried with warm air to get better contact with the resin. The natural moisture content of c. 8 vol% was kept in the sample. Openings with diameters of 70 mm were made at the top and the bottom by a slow-turning lathe and the permeameters were preparated as shown in fig 1. The hydraulic conductivity was measured by a constant head permeameter with the pressure height H = 0.2 m (fig. 2). Tests have shown that some difficulties can occur when higher pressures are used. For instance can successively decreasing hydraulic conductivity be a consequence, probably due to grain migration.

Chemical analyses on the water were carried out before and after passing the permeameters.

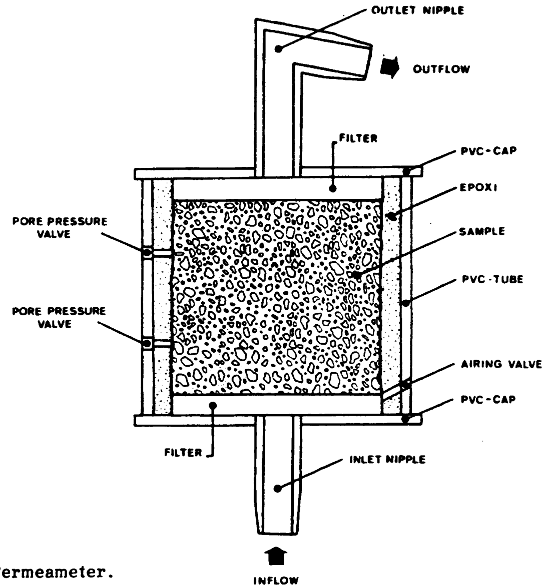

Figure 1 Permeameter.

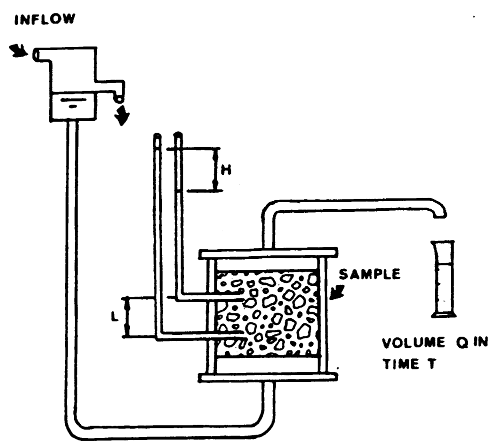

Figure 2 Constant head permeameter.

Sample analyses

After the hydraulic measuring the permeameters are cut at the top and bottom and placed in an equipment for water retention analysis (see Holm 1981). Pore distribution is calculated from the water retention curves. The suction steps used are 0.15 m (pF 1.18), 0.35 m (pF 1.54), 0.95 m (pF 1.98), 2.05 m (pF 2.31). A Suction with higher driving head is not used because of the risk for disturbing microstructures which are intended to be analysed as thin sections.

After the pF-analysis about 10% of the sample (50 g) is taken for analysis of grain size distribution. After replacement of pore water with increasing degrees of concentration of acetone, the rest of the sample, still in the PVC-cylinder, is impregnated with Epoxi resin mixture, (**fig** 3). Thin sections for 3-dimensional analysis are preparated according to Evenson (1970). Microscopic analysis of the thin sections is carried out in accordance with the analysis-form below (**fig** 4). The impregnated samples are also studied macroscopically with reference to fissures and visible structures.

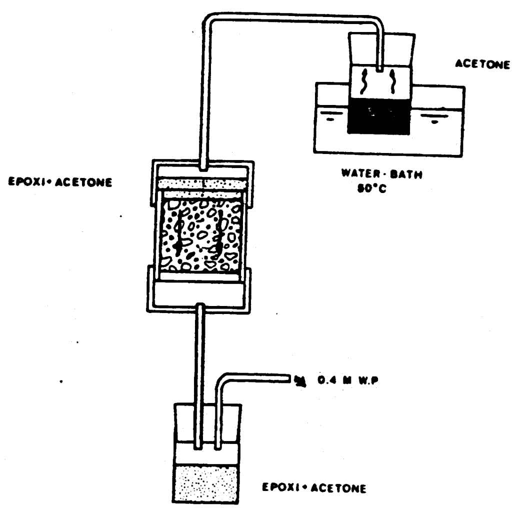

Figure 3 Impregnation of samples containing acetone with Epoxi resin mixture.

(Thin-section analysis)

TUNNSLIPSANALYS

Nr........ *Förstoring*..........

Struktur	*Typ*	*Frekvens*	*Anm.*
Kornform Grain shape	4 Klasser		
Porform Pore shape	5 Klasser		1 2 3 4 5
Rundning Roundness	5 Klasser		
Ytstruktur Surface structure	4 Klasser		1 2 3 4
Aggregat Aggregate occurrence	2 Klasser		1=ja 2=nej
Fabric			
Mineralogy *Mineralogi*			
Kornfördelning Spatial grain-size variations	4 Klasser		1 bandad 2 homogen 3 grupper 4 zonerad

Figure 4 Form for thin-section analyses.

Results

About 30 permeameter-tests have been carried out so far. The ambition is to make 100 tests.

The results from the permeameter-tests are shown in fig 5. Note that there is differences in hydraulic conductivity due to orientation, but also due to different pressure heights. Permeameter samples with grains larger than 1/5 of the sample-diameter (that is, more than 14 mm) have not been considered in the proceeding work.

The water retention curves are all very similar and steep in the measured interval (see fig 6). The effective porosity n_e, is concentrated to pores 100-10μ,the total porosity is about 22%.

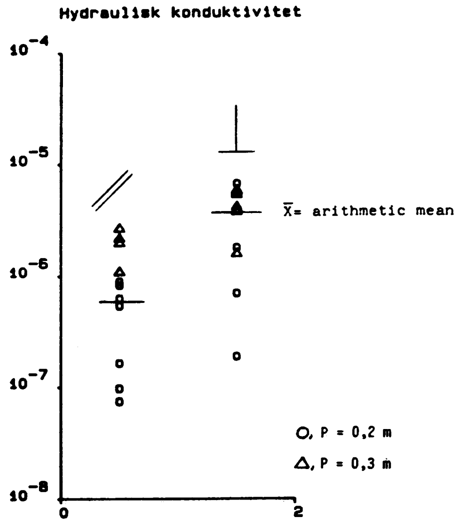

Figure 5 Hydraulic conductivity

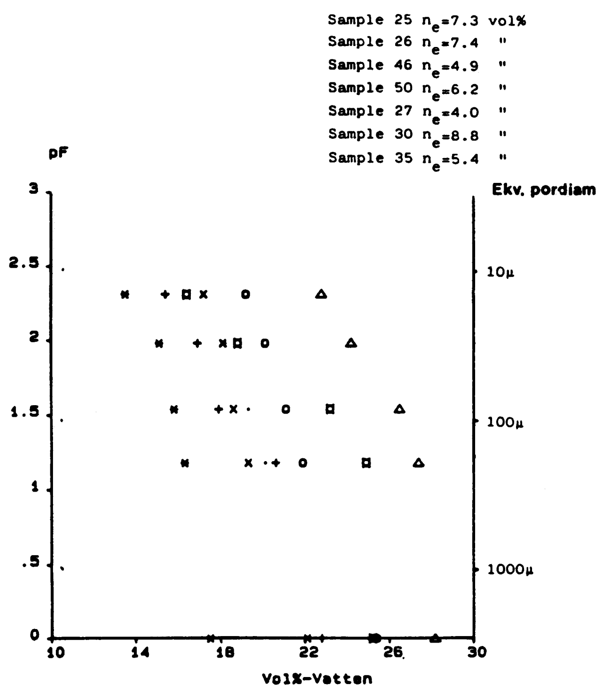

Figure 6. Water retention curves

In thin sections about 200 elongated grains have been counted, in each the horizontal and vertical plane. The mean direction and its significance are calculated from the equation:

$$F(\theta) = \sum_{i=1}^{N} \cos^2 (\theta - \theta_i), \; F(\theta) = F(\theta)_{max}$$

where θ = mean direction

θ_i = direction of separate particles

N = number of particles

The fabric analysis is presented in a 360 unfolded circle, see fig 7-8.

An initial hypothesis stated that there should be a positive correlation between the fabric, expressed as mean orientation of grains, and the hydraulic conductivity. This hypothesis has been verified, although there is a weaker fabric orientation than expected in the till.

Field investigations have shown that great parts of the till can be more or less unsaturated even below the ground water level. The till is very homogeneous and no cracks have been noted neither micro- nor macroscopically. It is evident that a gas pore pressure exists in parts of the till. This is supposed to be a result from either a fast rising ground water level that captures the pore gas, or from wethering of organic material. Organic concretions of the sizes of 0,5- 2 cm have been found at 2- 3 m depths in the till.

The permeameter tests are carried out with water of another chemical composition than that of the natural ground water. It is of interest to study the interaction between the exchangeble cations in the clay minerals. So far this study has just begun.

Conclusions

It is possible to state that sedimentary structures have a significant influence on the hydraulic conductivity. The

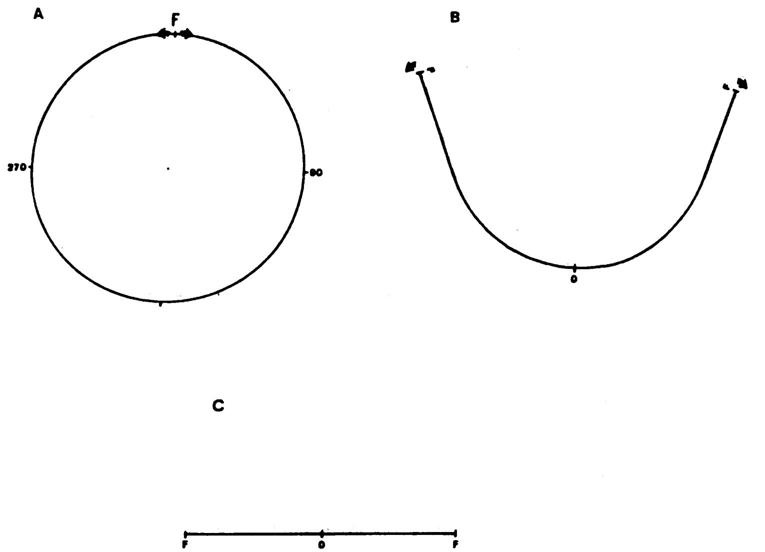

Figure 7 Presentation of fabric analysis.

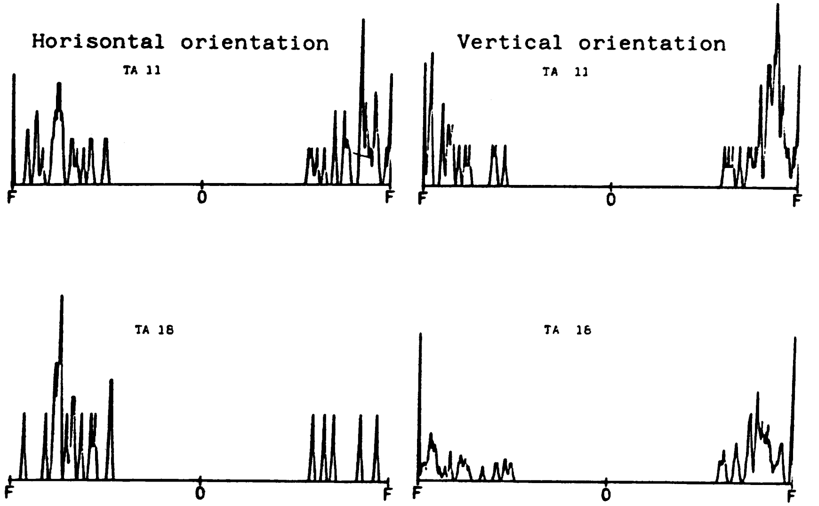

Figure 8 Example of fabric analysis.
The data has been smoothed.

permeameter tests show that the hydraulic conductivity of the investigated till is higher in one horizontal direction than in another. The difference is in the magnitude of a factor 5- 10. From analysis of thin sections it is evident that the grain orientation is responsible for a great deal of the anisotrophy. Other factors of influence can be spatial grain-size variations and spatial pore-size variations.

Many authors have concluded that the saturated water movement in till is mainly concentrated along fractures and fissures (cf Prudic 1982, Haldorsen et al 1983). In this study it has not so far been possible to detect any fissures in the C-horizon of the investigated till. The studies indicate that the water movement in the C-horizon mainly is concentrated to zones with sorted sediments. However, the possibility of water movement in small-scale fissures shall not be denied.
It is supposed that the vertical hydraulic connections in the profile are located to certain narrow zones with coarse sediment or roots.

Besides fabric there are many other factors having influence on the hydraulic conductivity in till. Although very little is known about these factors an attempt has been made to quantify the influence. The structural factors listed below are closely interrelated in such a way that their magnitude of influence cannot be added. The table should be looked upon as a hypothesis for proceeding work.

Sediment structure factor Note, grain-size and sorting are not considered	Magnitude of influence on hydraulic conductivity (very pre- liminary values)
Grain shape (sphericity)	≈ 5
Pore shape (incl. fractures)	strongly varying
Roundness	?
Grain surface structure	≈ 5?
Aggregate occurrence	≈ 5
Fabric	≈ 5-10
Porosity-pore orientation - spatial pore-size variations	?
Spatial grain-size variations	significant
Mineralogy	significant

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VARIATIONS IN TIME AND SPACE OF SOIL MOISTURE CONTENT WITHIN AND BETWEEN TILL OF DIFFERENT SOIL TYPES.

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SUMMARY

In a humid region of Sweden with till soil and a variety of soil types, soil moisture content was studied by neutron scattering technique. The moisture conditions were determined and differences both in moisture content and of changes in moisture content with time were investigated with regard to distance between measuring stations and soil types.

It was concluded that the largest soil moisture variation expected at the driest soil type and the smallest at the wet one did not fully agree with the natural situation. This assumption applies in special circumstances but normally the variation is largest at mesic and moist soil types since in these soil types there is a high moisture content in the beginning of the growing period which later often decreases.

There is often fairly good agreement between soil moisture changes in all soil types but the best agreement is achieved within the same or at least closely related soil types. Similar soil properties at the sites increases the agreement further.

INTRODUCTION

The development of a soil profile depends greatly on the soil moisture conditions (Tamm, 1931). These, in turn, depend on the distance to the groundwater. Over the years the moisture conditions have participated with other pedogenic processes in developing different soil profiles. These can be used to describe the moisture conditions in different soil types. Within a geographical region similar soil properties and hydrological conditions often develop similar soil profiles.

The topography of an area often consists of hills, valleys and slopes. Due to these landscape elements the area is divided into subareas of different soil types with characteristic soil profiles. In a podzolic region such as the Kloten region iron podzols are found in the dry and mesic uphill areas while humus podzols and waterlogged soils are found in the wet lowlands. In between these areas we find a variety of podzols between iron and humuc podzols - iron-humic podzols.

Since there is a relationship between the soil profiles and the moisture conditions, the variations in soil profile should also reflect variations in soil moisture. This implies not only a similarity in soil moisture variations in space in the same soil type at different distances to the measuring stations but also that moisture variations in time were similar at locations with the same soil type. The variable pattern of moisture conditions would be different in different soil types. To illustrate these implications the moisture conditions were studied in soil types at a variety of distances from each other within a comparably uniform area.

INVESTIGATION AREA

The investigation area is situated in central Sweden in the Bergslagen region ($59^{0}54'$ N, $15^{0}50'$ E) above the highest marine transgression (Fig 1). The granitic bedrock is covered by glacial till of 0-20 m thickness, but often less than 2 m. Peatlands are scattered throughout the area.

The area belongs to the northern boreal coniferous region and due to a comparably high altitude the climate is cold and humic with a mean precipitation of about 900 mm and a mean annual temperature of $+5.0^{0}C$.

METHODS

Soil moisture and groundwater levels were determined in an area of totally 50 km^2. Moisture and groundwater stations were scattered throughout the area. The distances between the stations varied. Within small areas, about 100 m^2, of homogeneous soil types there were three to four stations. In somewhat larger areas (100-500 m) single stations were located on different soil types. This distribution of stations was repeated on different hillslopes in large areas, 10 km.

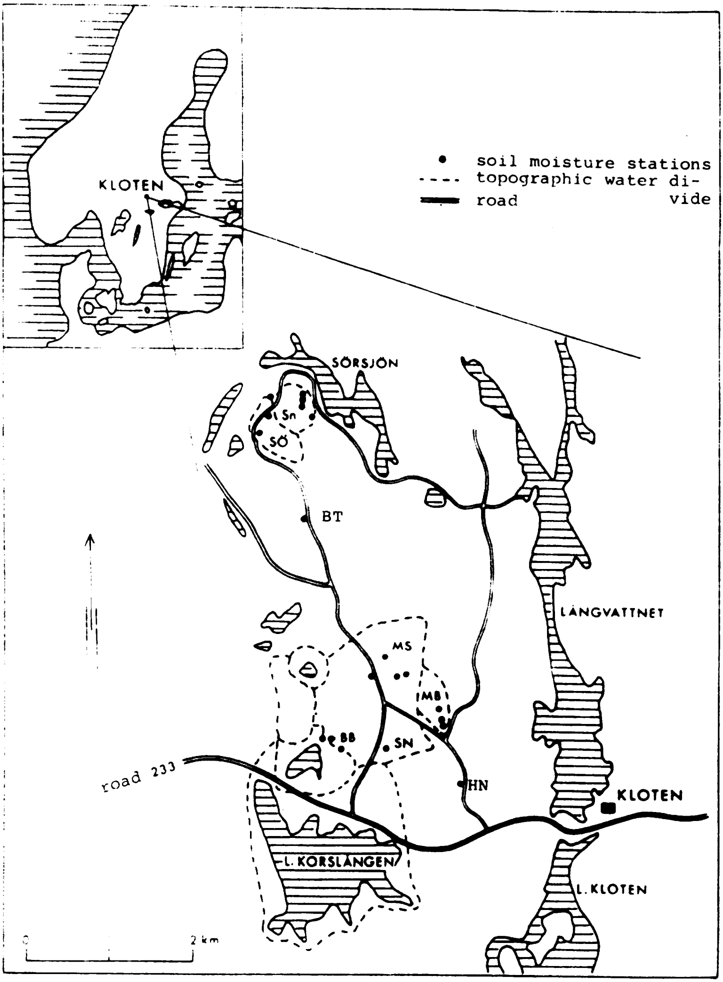

Figure 1. The investigation area and its location.

The measurements of soil moisture content were made using the neutron scattering technique in two-inche iron access tubes inserted in the soil without previous augerhole drilling. The instruments used were the Danish Basc (Jensen & Somer, 1967) and the Nuclear Chicago P19 probe (Nuclear Chicago). The Basc was used to determine soil densities and absolute moisture values whereas the P19 was used for routine measurements.

The groundwater levels were measured manually in open pipes.

Measurements of both soil moisture and groundwater were carried out at weekly intervals during the months of May-September during 1973-1977. Unfortunately, changes in site conditions caused moisture conditions to change, resulting in usually only short comparable periods.

The soil properties determined in the laboratory were texture and water retention. The latter, together with porosity, was calculated from volumetric samples in steel cylinders 5 cm high and with a width of 7.2 cm.

RESULTS

Soil types, soil profiles and groundwater levels

The soil types are characterized by moisture conditions, from dry to wet. The wet class is actually a simplification of the two classes slightly waterlogged and waterlogged. The simplification is not made because of no need for two classes but because there were too few moisture stations in the two classes, both of which are needed to describe the moisture variations in nature.

Moisture conditions were related to the soil profile in a topographic scheme with iron podzols, deep groundwater and a vegetation cover composed of lichens, some mosses and dwarf shrubs in the uphill areas, whereas downhill or lowland areas showed humic podzols or waterlogged soil, shallow groundwater and mosses of species *Sphagnum sp.* and *Polytricum sp.* On the hillslope areas between these dry and wet sites, mesic and moist areas were found with iron to iron-humic podzols, a groundwater 1-2 m deep and mesic mosses together with a varying occurrence of wet mosses (Lundin, 1982)(Table 1).

Table 1. An example of variation of some elements due to different moisture classes.

Moisture class		Dry	Mesic	Moist	Wet
Soil	A_0	0 - 10	5 - 20	10 - 25	>25
horizons	A_2	0 - 10	5 - 20	10 - 25	>25
	B	20 - 40	20 - 40	30 - 40	-
Groundwater level, median		deep	>1.5 m	1.0 m	0.6 m

Soil properties: Texture, water retention and porosity

There was a variation in texture in the investigated area but as a whole it could be characterized as a sandy till with 25% silt, 18% fine sand, 12% medium sand, 8% coarse sand and 37% of gravel to boulders, and having a porosity of 40 volume-% as a mean for the upper C-horizon. In the horizons close to the ground surface, as well as in uphill areas, the texture was somewhat coarser. Porosity decreased with depth and this, together with an increasing proportion of fine material, decreased the total porosity but increased the water retention (Fig 2). This implies that only small amounts of water were drained from the C-horizon when the groundwater was lowered. The main amount of water drained derives from the upper soil horizons.

Difficulties occur when determining porosity and true soil moisture content values, all of which are dependent on the determination of soil bulk densities. Different techniques give somewhat different values and there are also differences resulting from the instrument used, as can be seen in Figure 2b. There is also considerable spatial variation even at short distances between measuring stations (Fig 3), which complicates field calibration. Variations in bulk densities also lead to variations in moisture values (Fig 3b).

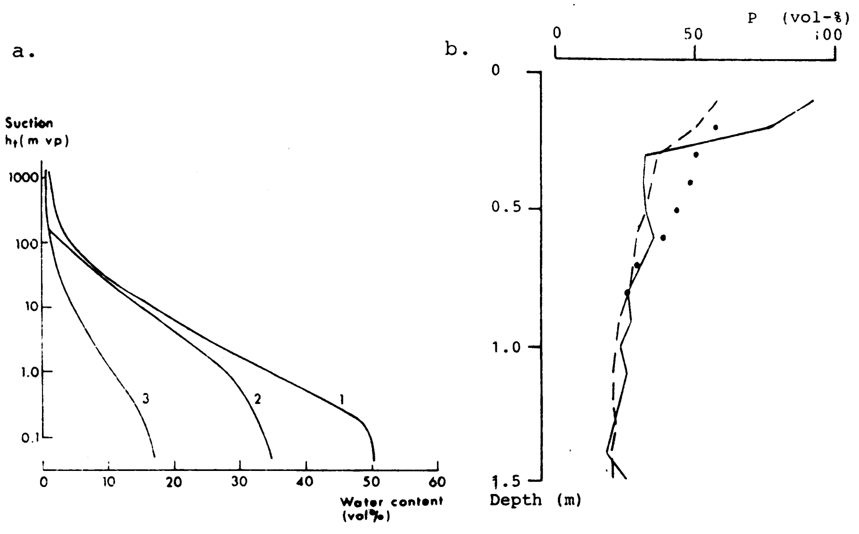

Fig 2. a. Typical water retention relationships for different soil levels. Shallow - 1, medium deep (0.7 m) - 2 and deep - 3 layers.
b. Porosity, P. The porosity determined from density and bulk density ——, neutron and gamma probe 1973-74 --- and 1985 • •.

Variations in soil moisture within short distances between stations.

In a small area (100-400 m^2), with one single soil type, differences between different points mainly depend on the heterogeneity of the soil properties. Primarilly this is seen, as mentioned earlier, in the soil bulk densities (Fig 3a). These differences are mainly of importance locally but also have some influence on the total soil moisture content. The differences between points within a small area are mostly less than 10% of the mean soil moisture content. At specific levels the differences might be considerably greater and then often amount to 5-7 vol-%, i.e. 20% of the mean soil moisture content.

The variation in soil moisture content with depth in the soil is dependent on the porosity and in wet situations the moisture content follows the porosity distribution (Fig 2b). During the growing period the moisture content increases with soil depth in the topsoil down to a depth of 0.2-0.4 m. Below this depth, moisture content follows the pattern of the porosity (Fig 3b).

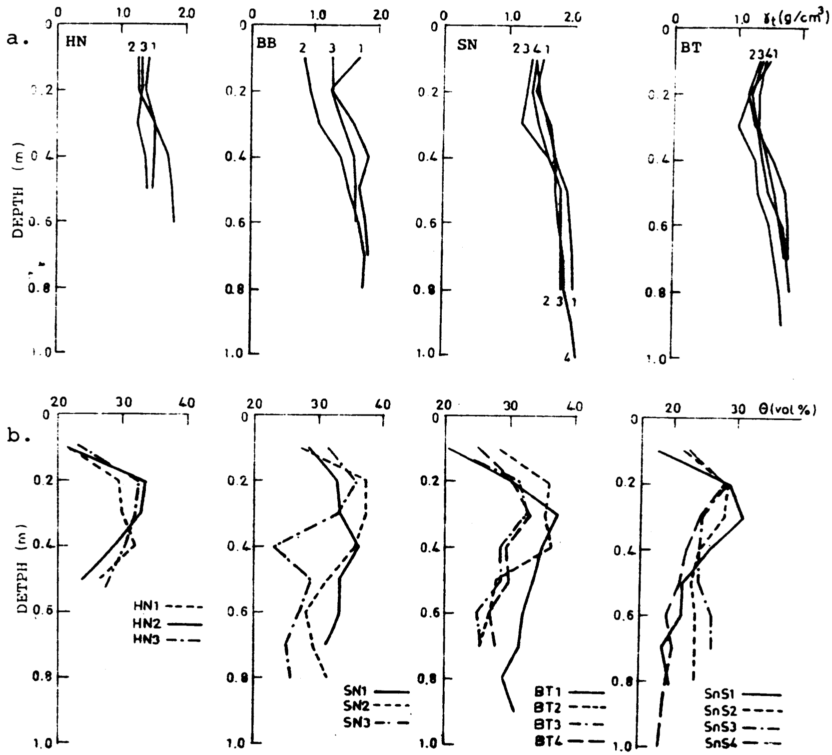

Fig 3. a. The bulk densities at different depths within four small areas with three to four stations.
b. The mean soil moisture content at the four areas in a.

In time the main moisture variations occur in the 0.5 m topsoil, below which changes in moisture content only contribute little to the total moisture variation. The moisture content changes considerably with time.

A total amplitude of 40-50 mm during a season is not unusual (Fig 4). This variation is similar at different stations. It can be compared with the differences of time-variations between the stations within the area during one and the same season. These differences vary between 2 and 15 mm, i.e. 20% of the total variation as a mean.

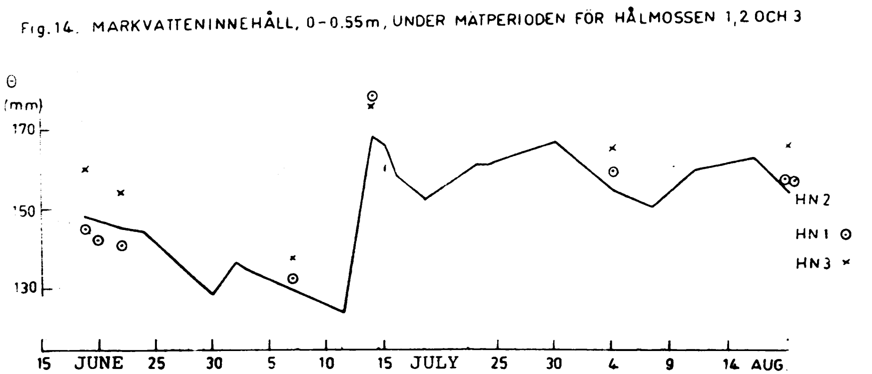

Fig 4. The changes in soil moisture content (Θ), 0-0.5 m depth, at three stations within a small area during two months.

Thus, the soil moisture content at different stations within a small area varies in a similar way and shows good agreement with determination coefficients of about 0.90 (Fig 5).

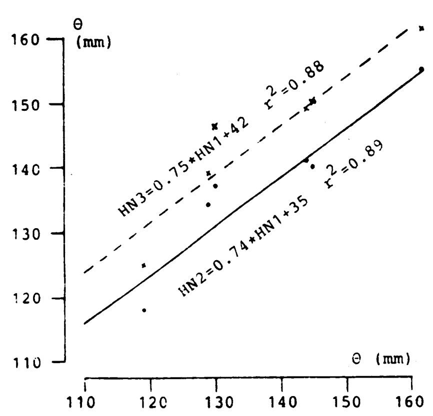

Fig 5.

Two comparisons of soil moisture contents, 0-0.5 m depth, between two stations within a small area of one soil type.

Soil moisture content in different soil types.

Differences in variations of soil moisture content between different soil types could be expected to be found in a larger time variation in the drier soil type. This could be due to a greater decrease in moisture content during dryperiods. However, this is not entirely the case since both total moisture content and moisture status at the start of the period influence the change in moisture content.

In the Kloten area, dry sites have about 50 mm less moisture content than mesic and moist ones and about 100 mm less than wet sites (Table 2). There are also large differences in moisture variations at different depths. At dry and mesic sites, in depths below the topsoil, the moisture variations are larger than at wet sites (Fig 6).

Table 2. The soil moisture contents and variations at different sites in the Kloten area.

Soil types	Soil moisture content in the levels 0-0.55 m		
	Mean (mm)	Range (mm)	Δ (mm)
Dry	127	102 - 146	44
Mesic	159	116 - 182	66
Moist	198	151 - 211	60
Wet	212	188 - 224	36

At the wet sites the dominating moisture variations occur in the topsoil but even in these levels is not necessarily larger than in mesic or moist sites.

The porosity in the topsoil is often larger in moist and wet sites than in dry and mesic sites, especially in places with a thick organic layer. This results in a larger total moisture content in the moist and wet soil types than in dry and mesic ones. A large moisture content also causes large changes. In the wet sites these changes are often retained

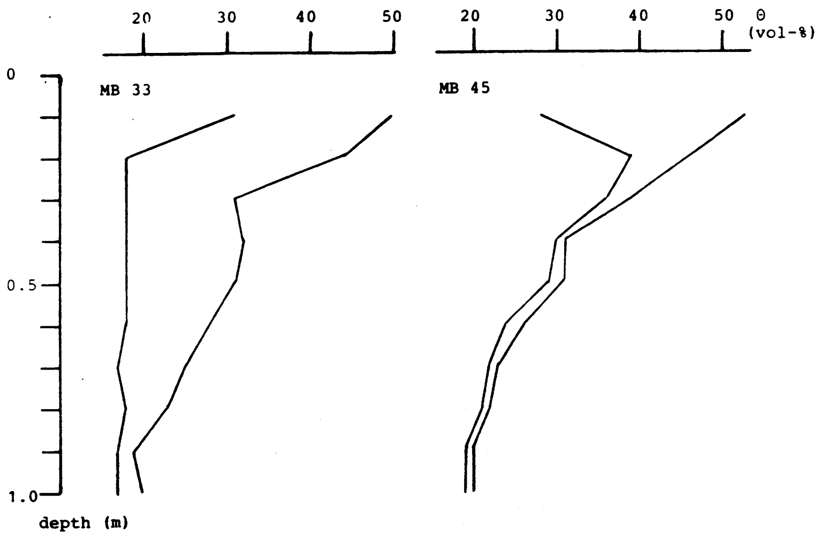

Fig 6. The changes in the lowest and highest soil moisture contents (Θ) with depth at one mesic (MB 33) and one wet (MB 45) soil type.

by a large inflow of water from upslope. As can be seen in Figure 7a, the moisture content in the moist soil type varies more than in the wet one but less than in the mesic one. This might be the ideal case but in nature it is far from the rule.

Not only differences in soil properties and moisture content influence the soil moisture variations. These also change with the current moisture status. The largest changes take place at the moist and wet sites during dry periods due to a large soil moisture content whereas during wet periods the changes are larger at the mesic and dry sites (Fig 7b). The latter sites presumably then have a large water content. This, in turn, can easily be drained and then largly decrease the water content. At the moist and wet sites the inflow of water will continuously provide a high moisture content during wet periods.

The normal natural conditions result in larger variations in mesic and moist soil types than in dry and wet. These circumstances, together with

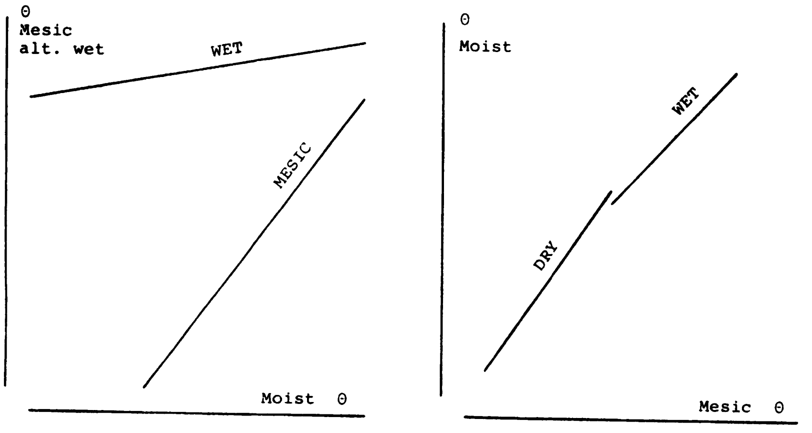

Fig 7. a. The relationships between soil moisture contents (Θ) of soil types of different moisture status (moist soil type compared to mesic alternatively wet).

b. The relationships between soil moisture contents (Θ) in one mesic and one moist soil type during a dry and a wet year.

the soil properties, influence the moisture variations in time. The best agreement of soil moisture variation in time is found between soil types of similar character (Table 3).

Table 3. Comparison of differences in time-variation of soil moisture content between soil types. The differences between changes of moisture content (mm) at different stations as 95% confidence intervals.

Soil types	Dry	Mesic	Moist	Wet
Dry	8	11	11	17
Mesic	11	8	13	21
Moist	11	13	10	17
Wet	17	21	17	14

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SPATIAL CORRELATIONS OF THE SOIL QUANTITIES RELATED WITH THE NEUTRON GAGE OF SOIL MOISTURE

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Determination of the soil moisture content quantities (parameters), those that absolute and permanent (porosity, field capacity, wilting point) especially, require the accurate calibration of their gages. The soil moisture content (the loose water per volume unit) can be determined accurately by nuclear gages. — In addition of soil moisture parameters we have the neutron gage soil parameters. We are using three in a good calibration. The spatial variability of a quantity is proposed to be analyzed by its correlation (or autocorrelation) function. This measure of the spatial dependence between the values of the parameter is useful.

1. Introduction

Soil is a porous matter. It varies considerably in porosity, in texture and chemically. Except mineral as rock it may be organic, or it has an organic matter content. Soils usually comprise water. It may be even strongly chemically bound or loose.

Soil moisture, movable water of soil, has parameters:

PARAMETER	ITS FIELD GA(U)GE
Moisture content	Neutron gage
Binding of water	Tensiometer

The second column above tells the means by which we can measure the parameters accurately in field. Sometimes it is argued that these gages can be replaced with each others. However, dependence of water content on the binding (presented, e.g., as the pF value) is not at all unique.

Table 2. **Permanent parameters in soil profile. Moisture contents, measured by neutron gaging**

Parameter	Auxiliary gaging
Porosity	
Field capacity	
Wilting point	tensiometer use

The parameter porosity is the moisture content of saturated soil. The field capacity and wilting point often mean the moisture contents of these soil moisture states (the binding — a pressure quantity — is then fixed). These three quantities of soil moisture: porosity, field capacity and wilting point moisture contents are all permanent soil parameters. They are also all absolute values of soil moisture.

The soil moisture contents of table 2 are necessary in order to determine proper actual soil moisture quantities (e.g. SMD) from temporally varying soil profile moisture content data.

2. Good calibrations for n and γ gages

The author has material for these both, but a lot of works is to be done in this important task. For both the gages I have theoretical models. The gage parameters of the models are to be determined. They are probe and access tube dependent. A n method and fitting to data, that of *Kasi, Immonen and Saikku, 1983*, has been described in the FAO/IAEA symposium. The γ calibration curve given by a Nordic factory is a falling straight line. The response function of *Czubek, 1983* still accepted, has been used to fit to the results of water, a sand and clay. The calibration curve obtained is clearly curved. The soil moisture and density values by weighing were determined from nondestructive, but not very representative samples. In Appendix, there we have some substances to use in accurate measurements. When doing those we are gaging moisture (or H content) in chemically known and, should be, homogeneous matters.

For the use of the good calibrations we should have the following neutron gage soil parameters (*Hooli and Kasi, 1975*):

1. bound hydrogen content,

2. thermal n absorption cross section Σ_a — *Couchat et al., 1975*, *McCulloch and Wall, 1976*, and *Czubek et al., 1983*, show three different methods for its determination,

3. soil density.

The two first should be measured from soil samples extracted when the access tube is inserted in soil. Σ_d by *Couchat, 1983*, is less preliminary a quantity than the hydrogen content. The density can be gaged by the γ probe. When a parameter is lacking, it is to be estimated. The estimation may be done by using results from nearby places or known value of the same type soil.

3. Spatial correlation of parameter

How does a quantity vary in field? E.g. the water content of field capacity is different in clay, sandy soils and peat. *Webster, 1983*, has represented a review of the correlation measurements in soil, a lot of which he has been measuring with. We know that the soil types make combined areas, but also vertical layers. The soil types have subtypes. The hydrological nature of soil is described by means of adequate parameters. The soil moisture contents of table 2 are such.

Set x to be a soil parameter. The spatial fluctuations of x can be analyzed by the correlation C and variation V functions, *Journel and Huijbregts, 1978*, *Yevjevich, 1972*, illustrating them with the correlograms and variograms. Along a straight line, having equally separated sites of samples, for the correlation function C we have the estimate

$$C_k = \frac{s_k^2}{\sigma_E^2}, \quad s_k^2 = \frac{1}{n-k-1}\sum_{i=1}^{n-k}(x_i-\bar{x})(x_{i+k}-\bar{x}), \quad \bar{x} = \frac{1}{n}\sum_{i=1}^{n} x_i. \quad C_0 = 1,$$

where n is the number of samples, taken or direct gaged. $\sigma_E^2 = s_0^2$ is the value of the nonbiased estimate of the variance as well as $\bar{x}$ is the estimate of the mean value of the statistical distribution of x. The function V is estimated by

$$V_k = \frac{1}{2(n-k)} \sum_{i=1}^{n-k} (x_{i+k} - x_i)^2.$$

x is a continuous variable on its line. It can then, too, be integrated over its line when we know the distribution.

In the literature concerning the unsaturated horizon I have found few examples of use of the C_k formular above: the analysis by *Warrick and Nielsen, 1980*, from the data of Gajem and Warrick, and the application of *Moutonnet, Perrochet and Couchat, 1983*. In the former $C_k \neq 0$ when k covers not more than 4 m. In the latter example three hosts in Aix find no horizontal correlations in the two *n* gage soil parameters investigated. These were Σ_a and Σ_d. Their soil was too homogeneous and the accuracy of the neutron data device of *Couchat, et al., 1975*, too poor. It is so poor and the device so slow, that this result does not recommend dense determinations of the *n* gage soil parameters. Their measurements, in the 100 m $\times$ 100 m square area along two crossing lines, only faintly tell that in the surface layer 0...30 cm the bound hydrogen content is larger than in the layer 30...60 cm. This is seen from the values of Σ_d.

The parameters can vary in large. E.g., the diffusion cross section Σ_d of peat is about 3 times bigger than in the Cadarache arable land we circulated. In Cadarache the hasty French should have measured the neutron soil parameters more widely, — or perhaps more densely than at 2 m intervals.

Below I have a fictive example of a sample line in Nordic uncultivated soils. I have selected as a parameter the loss of mass when the soil piece is heated to 900 °C and hold there since its mass remains constant. Sample places have the interval of 10 m (and because n = 35, the line does not extend, e.g., over our countries).

clay sand peat sand

7 9 8 10 7 4 5 3 4 7 50 75 80 90 82 50 6 3 7 4 0 100 10 4 3 3 0 5 2 3 4 11 15 60 2

This is an exercise, please; the unit is % by weight. The correlation function, you then obtain, has both positive and negative values intervals. The nugget value of the function V is large.

The simple variation function V has the advantage that the unexplained zero variance the nugget is found out. Also the range of the positive correlation is clearly seen. The correlation function C tells well at which distances the positive or negative correlations occur and which are their strengths.

Instead of the preceding presentation, where the formulas are along a line, we generally, and often, have a quantity $x(\mathbf{r})$, where $\mathbf{r}$ is the space vector, two- or three-dimensional. x is a scaler. $C = C(\mathbf{r})$ (and $V = V(\mathbf{r})$) may be at best a tensor, but its vector components seem to be most informative. In many cases the tensor is isotropic, but mostly in horizontal planes. A horizontal plane is the "projection" of the unsaturated soil layer. We have the area quantities and the spatial ones.

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Appendix. Substances for calibration

Substance	Density kg/m^3	Phase changes °C
$MgSO_4 \cdot 7\,H_2O$	1677	150 ($\Rightarrow MgSO_4 \cdot H_2O$)
$Na_2SO_4 \cdot 10\,H_2O$	1464	32,4 and 100 ($\Rightarrow Na_2SO_4$)
$KAl(SO_4)_2 \cdot 12\,H_2O$	1757	92,5 ($\Rightarrow KAl(SO_4)_2 \cdot 3\,H_2O$)
$NaAl(SO_4)_2 \cdot 12\,H_2O$	1675	61
$Na_2CO_3 \cdot 10\,H_2O$	1440	33.5 ($\Rightarrow Na_2CO_3 \cdot 9\,H_2O$)

Except measuring in these substances, when pure, we can combine:

CRYSTAL WATER SUBSTANCE + QUARTZ
(or)+ any FELDSPAR

And we can also make mixtures:

SUGAR + pure QUARTZ (or FELDSPAR)

MOISTURE RETENTION PROPERTIES OF SOME CULTIVATED SOILS IN SOUTH-EAST NORWAY

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ABSTRACT

Porosity and water retention properties of some common Norwegian soil types have been studied in relation to texture, organic matter content and bulk density. Satisfactory prediction using these variables is possible in most cases. Effects of organic matter and bulk density on the above properties were studied on individual soil types. Organic matter mainly affected water retention, whilst bulk density was more closely related to air capacity. The variability of various pore size fractions in loam, silt and clay soils is discussed. Variation in porosity was low, whilst it was moderate in available water capacity and high in air capacity and non-available water. Subsoil variation was generally higher than that of topsoils.

INTRODUCTION

Soil productivity and irrigation requirements under dry climatic conditions are closely related to soil moisture holding capacity. The amount of water theoretically available to plants is affected by soil texture, soil structure and soil organic matter content, whilst the amount which plants actually utilize is limited by rooting intensity and rooting depth. This paper shows the effects of soil physical composition on water availability in some Norwegian soils, and discusses their variability at particular locations.

THE PREDICTION OF MOISTURE RETENTION FROM SOIL PROPERTIES

Moisture retention from 62 soil profiles in SE Norway, mainly from loamy, sandy and silty soils (fig. 1), were used in a regression analysis with mechanical analysis, organic matter content and bulk density as independent variables (Riley 1979). The main trends that were found are shown in Table 1. Generally similar effects were found in both topsoil and subsoil.

Moisture retention curves

Total porosity, besides being closely related to bulk density, showed a tendency to increase with the content of coarse silt. The volume of freely draining pores (air capacity) declined with bulk density, silt and organic matter contents. However, all of these factors had positive effects on the quantities of water retained at higher suctions. Non-available water (pF>4.2) was also markedly affected by clay content. It increased by about 0.4% per unit clay percentage, in close agreement with the finding of Hansen (1976) in Denmark. Examples of moisture retention curves, calculated from regression equations for different soil types, are shown in fig. 2.

Total capacity for available water

The choice of pF 2 as an appropriate value for field capacity has been confirmed in a separate study at SF Kise (fig.3). Equations containing terms for various combinations of silt, clay, gravel, organic matter and bulk density, accounted for about 85% of the variance in total available water capacity (AWC). An example of measured AWC versus that calculated using such an equation, is shown fig.4. An estimation accuraccy of ±5% available water was obtained. Approximate AWC ranges are given in table 2 for the soil textural classes studied, assuming average topsoil values of bulk density and contents of organic matter and gravel.

Table 1. Factors affecting pore size distribution in some mineral soils of SE Norway (data from 62 soil profiles mainly in Hedmark and Oppland).

Parametre	Positive effects	Negative effects	Variance accounted for
Total porosity	Coarse silt	Bulk density	83-89%
Air capacity at pF 2	-	Bulk density Org.matter All silt fractions Clay	78-87%
Water retention at pF 2	Coarse silt Medium silt Clay Org.matter Bulk density	Gravel	80-89%
Readily available water (pF 2-3)	Coarse silt Fine sand	Clay	45-62%
Strongly-held available water (pF 3-4.2)	Org.matter Coarse silt Medium silt Bulk density Clay	Gravel	67-78%
Total available water (pF 2-4.2)	Org.matter Coarse silt Medium silt Bulk density	Gravel	84-87%
Non-available water (pF>4.2)	Clay Org.matter Bulk density	-	79-88%

Table 2. Typical ranges of available water capacity calculated from regression equations for some Norwegian soil types.

Norwegian name	Approx.equivalent	Silt range (%)	AWC range (%)
Sand	Sand/loamy sand	0 - 15	7 - 15
Siltig sand	Sandy loam	15 - 45	15 - 25
Lettleire	Loam/silt loam	25 - 47	20 - 30
Sandig silt	Sandy silt loam	45 - 80	25 - 40
Siltig lettl.	Silty loam	47 - 92	28 - 42
Silt	Silt	80 - 100	38 - 45

Different fractions of available water

The fraction of water most readily available to plants (pF 2-3) is of particular importance in agriculture, since, under conditions of high evaporative demand, moisture stress commonly occurs in plants when this fraction is used up. Considerably less of the variance for this fraction was explained by equations than was the case for total AWC and strongly held water. It may be assumed that factors other than soil texture are also of importance here, for example those associated with aggregate formation and crumb structure (organic matter type, microbial activity?).

Readily available water content increased with silt content and declined with clay content. Calculated values for some typical soil types are given in table 3. Though silty clay loams and silt soils were poorly represented in the data for this study, later investigations have confirmed the predictions for these soil types (Riley 1983).

The strongly bound available water fraction (pF 3-4.2) increased almost equally with both silt and clay contents. Whilst in silt soils the proportion in this fraction was about 50-60% of total AWC, it was approximately 60-70% in sandy soils, 70-80% in light

Table 3. Contents of readily available water (pF 2 - 3) calculated for some typical soil types.

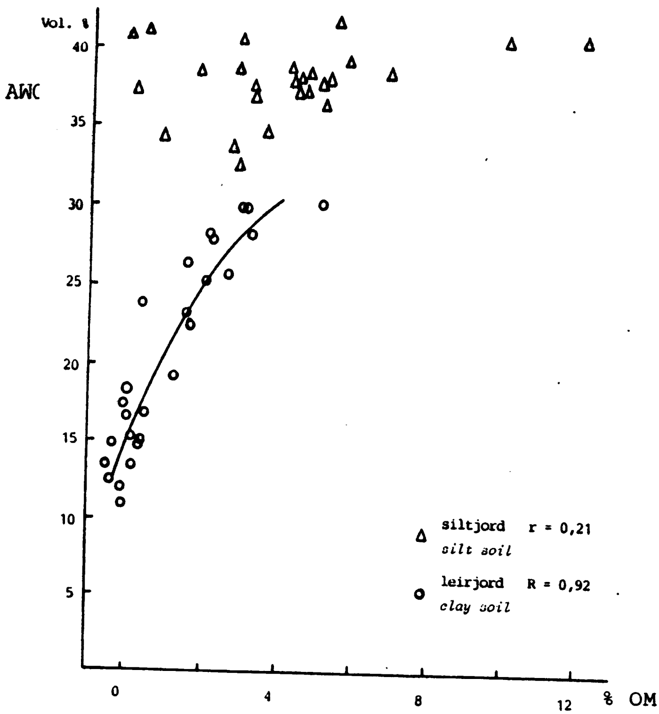
Norwegian name	Silt (%)	Clay (%)	Readily avail.water (%)
Silt	90	5	18
Sandig silt	75	5	15
Siltig lettleire	70	10	13
" "	70	20	10
Siltig sand	25	5	6
Lettleire	35	15	5
Siltig mellomleire	65	35	4

loams and 80-90% in heavier loams. There were also clear positive relations of strongly bound water with soil organic matter and bulk density. The effects of these parametres are discussed below in connection with separate studies on various soil types.

EFFECTS OF ORGANIC MATTER AND BULK DENSITY ON DIFFERENT SOILS

The overall increase in total AWC in the above-mentiomed study (Riley 1979) was 1.0% per unit increase in organic matter content. Loam soils predominated in this study, and this soil type on its own showed a similar relationship for topsoil samples, but a somewhat greater effect in subsoil samples (1.7%). The latter were lower in organic matter content, and considerably denser than topsoil samples.

In a group of 26 silty clay loam soils with generally low organic matter content, an increase of 4.0% AWC was found per unit organic matter increase (Riley 1983), whilst for a similar group of silt soils there was little relationship betwee AWC and organic matter content (fig.5).



Figur 5. Available water capacity (pF 2-4.2) in relation to soil organic matter (loss on ignition corrected for clay content) in silty clay loam and silt soils.

The effects on AWC of organic matter variation in topsoils has also been examined in conjunction with field trials where many samples were taken at the same sites. Such data are less likely to contain undesirable intercorrelation of soil organic matter with textural or other parametres. The results of such studies confirm the finding that the effects are greater on clay soils than on other soil types, particularly at low organic matter levels. A greater effect on aggregation probably occurs in such soils.

The effects of organic matter may sometimes be confounded with those of soil density in such studies (Heinonen 1954). An attempt to distinguish between the relative importance of these factors has been made using the results of multiple regression for some of the above trials. The independent effect of each factor on pore

Table 4. The influence of soil organic matter on available water capacity. Samples taken within field trials on various soil types.

Soil type	Location	No.samples	OM range	%AWC per %OM	Corr.coeff.
Silty	Nannestad	72	1.2-4.9	+4.6	0.87
clay	Nes på Rom.	96	3.3-6.4	+3.2	0.74
soils	Frogner	72	2.1-6.7	+3.0	0.82
	Rakkestad	96	5.4-8.0	+2.4	0.68
Silty	Sør-Østerdal	96	2.4-4.6	+1.6	0.41
soils	Brandval	72	3.2-5.1	+0.7	n.s.
	Namnå	72	3.6-9.4	+0.5	0.39
Loam	Toten	96	5.7-7.2	+0.3	n.s.
soils	Toten	96	4.5-8.2	+0.3	n.s.
	Nes på Hedm.	48	8.7-16.5	+0.5	0.31
	Nes på Hedm.	144	7.5-20.2	+0.4	0.42

size distribution has been calculated for their mean values ± one standard deviation in four trials (fig.6). No correction was made for the reduction in soil particle density with increasing organic matter content, but this effect was considered to be slight.

The greatest effect of bulk density variation was on air capacity. It had little effect on AWC except on the silt soil, but there was a slight positive effect on the amount of non-available water on all soil types. The fact that increased organic matter also appeared to reduce air capacity, may be understood in relation to its positive effect on water holding capacity.

The latter effect was greatest on the strongly held fraction of AWC in clay and silt soil. An equal effect was found in both fractions of AWC in loam soil. In silt soil the increase in strongly held water took place at the expense of the readily held fraction, in agreement with previous findings for such soil (Riley 1983). In general AWC was affected more by organic matter than by bulk density in all of the soils studied.

The effect of organic matter on non-available water varied between soil types. Increases in organic matter reduced this fraction in both clay soils, whilst on silt and loam soils the reverse was true. A likely explanation of this is that organic matter enhances aggregation of clay soil, thereby making it more "open". In soil types with less pronounced aggregation, on the other hand, increases in organic matter probably cause a gradual in-filling of the smaller soil pores. (Organic matter has itself a high proportion of pores in the non-available range). These findings are also in accordance with previous experience.

VARIABILITY OF WATER HOLDING CAPACITY WITHIN SOIL TYPES

Coefficients of variation (CV) for some pore size fractions of various topsoils are shown in table 5. Data are given for the variation both between sites with similar soil and between trial blocks at the same site. Both groups of coefficients are of similar magnitude. Total porosity showed relatively small variation, especially on silt and clay soils. Total AWC showed very little variation on silt soil, and moderate variation on clay and loam soils. Both air capacity and the non-available water fraction displayed high variation in most cases. It may be mentioned that permeability, which is closely related to air capacity, commonly shows even greater CV values.

The variability of water-holding properties for the soils shown in fig.1 was considerably higher in subsoil samples than in topsoil (table 6). Many of these soils were of morainic origin. Similar variation with depth was also found within a small area (0.5 ha) at SF Kise. Such high variability may be associated with high variation in clay and gravel content. Soils derived from sedimentary deposits probably exhibit somewhat less variability at depth than at the surface.

Table 5. Relative variability (CV) of some pore size fractions in various topsoils.

	Between sites			Between plots		
	Loam	Clay	Silt	Loam	Clay	Silt
Total porosity	13	10	6	16	5	7
Air capacity	45	63	86	42	31	35
Available water	17	32	6	19	22	6
Non-available water	28	28	40	43	24	15
No. samples	32	26	26	12	12	12

Table 6. Relative variability (CV) of some pore size fractions in topsoils and subsoils.

Depth (cm)	Porosity	Air cap.	pF 2-3	pF 3-4.2	pF 2-4.2	pF>4.2
I. 62 profiles on loamy, silty and sandy soils.						
0 - 25	13	43	54	30	27	41
25 - 50	19	33	88	48	46	61
I. 12 profiles on morainic loam at the same site.						
20 - 30	7	23	12	12	11	13
30 - 40	10	50	25	31	23	52
40 - 50	13	44	40	23	21	40

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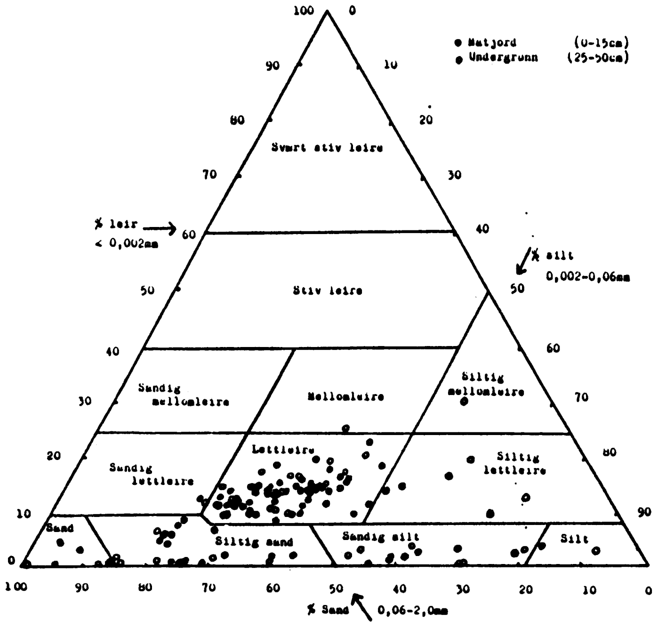


Figure 1. Distribution of soil samples used in multiple regression analysis of water holding capacity in relation to texture etc.

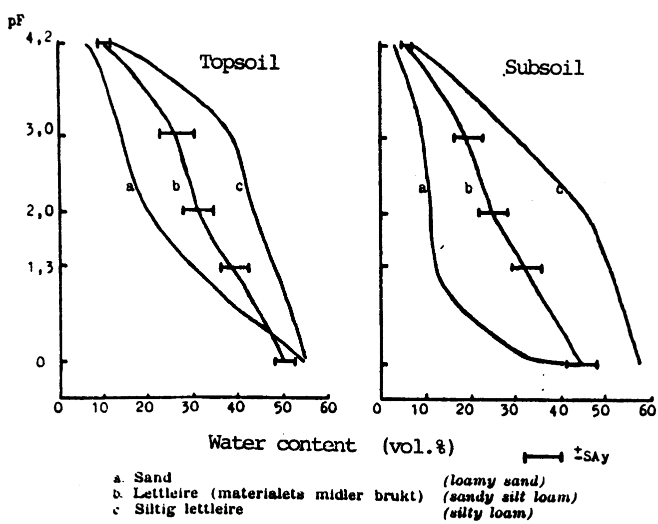


Figure 2. Moisture retention curves calculated for different soil types from regression equations on texture etc.

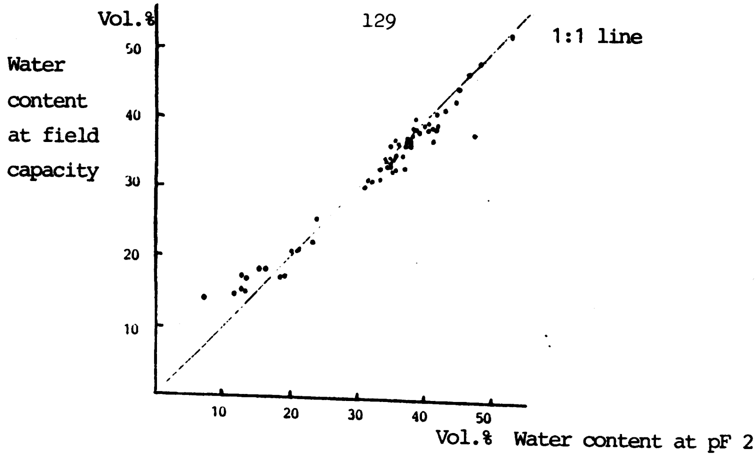


Figure 3. Comparison of water content measured in lab. at pF 2 with that in samples taken in field from plots covered with plastic following irrigation to full saturation.

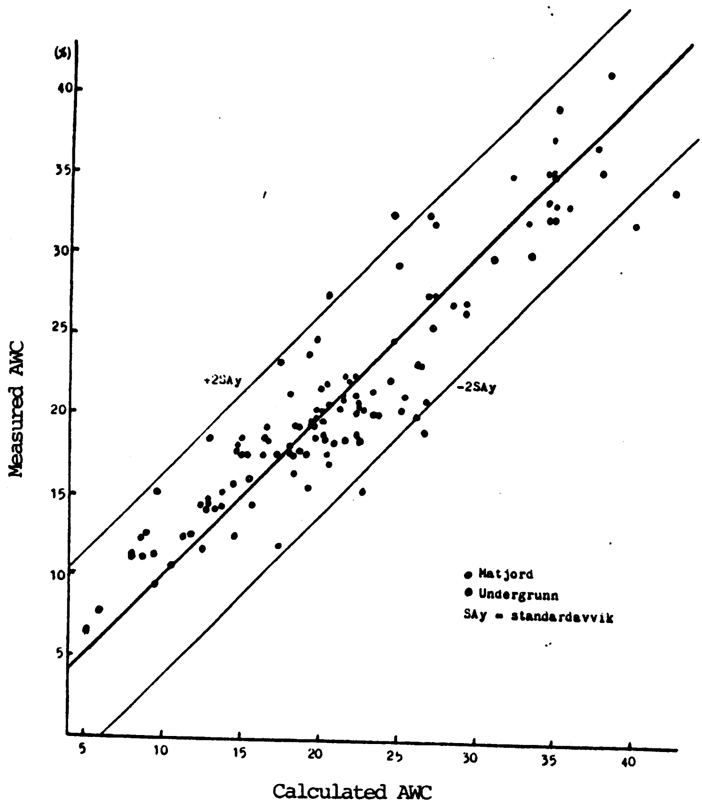
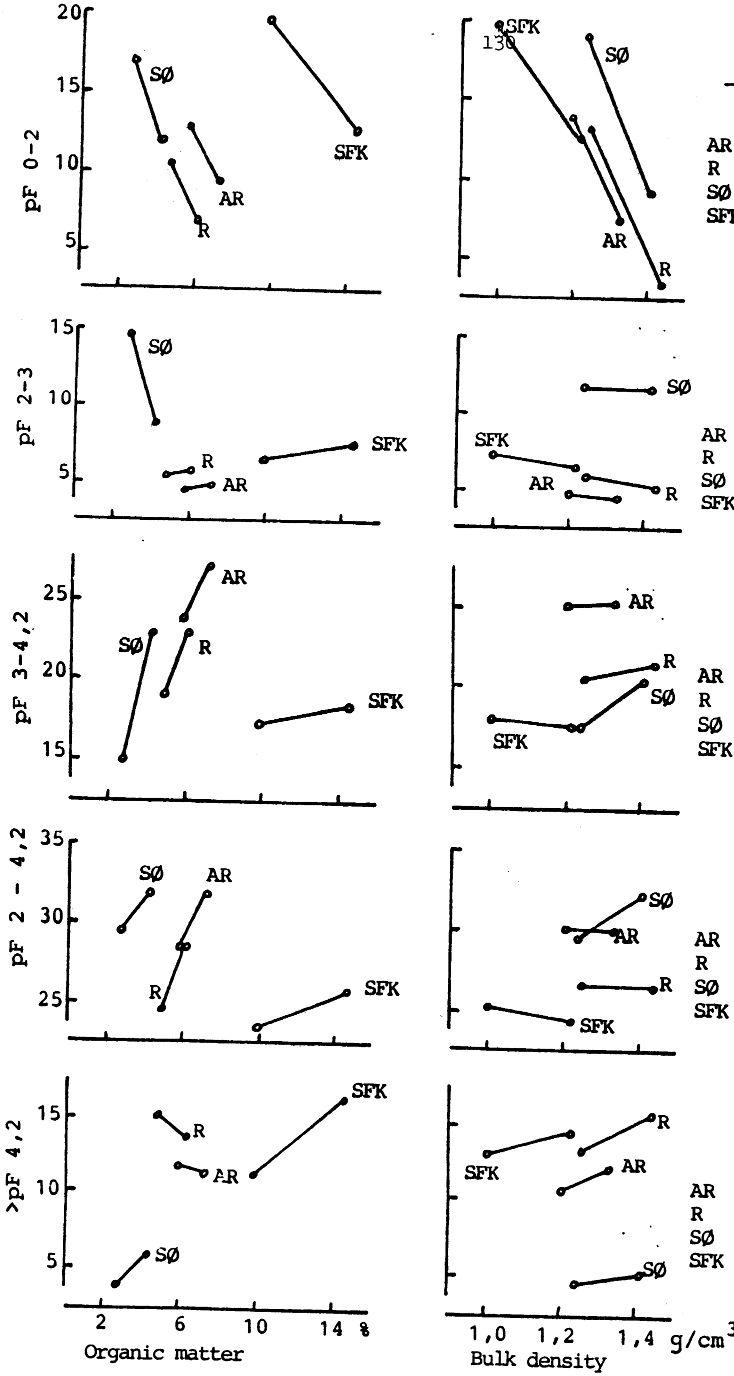


Figure 4. Comparison of measured total available water (AWC) with values calculated from regression equations .



	Partial correl. O.M.	B.D.	CD
AR	0,70	0,88	77
R	0,62	0,90	83
SØ	0,73	0,91	86
SFK	0,65	0,66	48
AR	0,25	0,27	19
R	0,24	0,35	36
SØ	0,79	0,00	62
SFK	0,42	0,32	43
AR	0,63	0,14	41
R	0,71	0,18	58
SØ	0,81	0,47	71
SFK	0,13	0,05	4
AR	0,66	0,00	47
R	0,65	0,00	55
SØ	0,44	0,54	41
SFK	0,20	0,10	10
AR	0,33	0,68	59
R	0,65	0,84	89
SØ	0,95	0,74	92
SFK	0,73	0,31	56

Figure 6. Independent effects of changes in organic matter content and soil density on pore volume at various pressure intervals. Calculated for data on different soil types (AR = clay loam R = clay loam SØ = silt SFK = loam) using mean values + one standard deviation.

PLANT AVAILABLE WATER IN THE ROOT SONE-REGIONAL AND VERTICAL VARIATIONS.

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Summary

Data for root zone capacity are needed when using agro-hydrological models in the unsaturated zone. The distribution of 8 root-zone capacity classes ranging from 50 to 200 mm plant available water capacity was calculated for the avable land in the catchment of Vigga river, Lunner and Gran kommune, Oppland county.

The method for calculating the distribution is discussed. A total of 400 sampling

sites were dug out in order to investigate the distribution of effective root depth and water-holding capacity of the different soil Layers within the area.

The sampling sites were located along lines across the valley, at intervals of 1000 m and 200 m distance between the sites in the line.

The root-zone capacity distribution was also calculated based on longer distance between the sampling sites. The results showed that for an area of this size, over 5000 ha, with mainly till and weathering material, the presision was good also with 1000 x 1000 m between the sampling sites.

INTRODUCTION

This work is a result of an investigation made to get input-data for using agro-hydrological root-zone models. In this special case the purpose was to calculate the potential demand for irrigation-water in the catchment area of the Vigga-river in Gran and Lunner kommune, Oppland County.

In the investigated region there is 5 000 ha of arable land. Spring cereals, most barley, is covering 63 % of the areals, grass-production 27 % and potatoes and row-crops 10 %.

The annual precipitation in the growing season is approximately 350 mm, and the potential evapotranspiration is 345 mm (Penman).

Method

The root-zone capacity (see fig. 1) is here defined as the product of volume % plant available water and effective root-depht. The effective root depht is defined as the depht to the soil layer where there is 0,1 cm root per cm^3 of soil. (Madsen 1979).

Observations of root-distribution were carried out in each profile. On most sites the root-distribution was just described as seen in the profile wall.

The following description was used:

Many roots
Some roots
Few roots
Very few roots.

On some locations cylinder-samples (ca 500 cm^3) were taken for laboratory-analysis of root lenght per soil-volume. The analysis was done after a method described by Breuning Madsen. (Madsen 1979)

Results from the laboratory-analysis showed that the layer with 0,1 cm root per cm^3 of soil was lying somewhere between "few roots" and "very few roots".
Soil texture and structure, stoniness and drainage-conditions were abserved on each location.

About 400 sites were investigated in this manner. The location of each site was determined in the following way: (see fig. 2).

Lines were drawn on the map (M 1:20 000) across the Vigga-valley, at intervals of 1 000 m. As the valley is not a stright line, all the lines were not parallel. The sites were located along the lines at intervals of 200 m. (In one part of the area the location was at intervals of 100 m).

Geological maps (m 1:20 000) were used for the field-work. Till and weathering material is covering most at the area. Close to the river there is fluvial and laustrine deposits and organic matter. Appr. 5-10 % of the investigated sites were on these deposits. Plant available water capacity was higher here than on the till and weathering material -sites.

The statistical analysis of the data was therefore made individual for these two groups.

Results. Plantavailable water.

Plant available water capasity determined where the roots have their maximal distribution ranged from appr. 45 to 240 mm.

The profiles were grouped in 8 classes as follows:

50 mm	plant	available	water	capasity
75 mm	"	"	"	"
100 mm	"	"	"	"
125 mm	"	"	"	"
150 mm	"	"	"	"
175 mm	"	"	"	"
200 mm	"	"	"	"

The 200, 175 and 150 mm classes are drought resistant soil, often high in organic matter and silt. This classes are dominant on the deposits along the river and in lower parts of the till area. Deep root distribution is necessery to obtain such high water capacity.

The 125 and 100 mm classes are soil profiles where the plants may be exposed to drought. These profiles are predominant in the till area where the organic matter content is low. On weathering material one might find the 100 mm class where the root-distribution is good.

The 75 and 50 mm classes are soil profiles where the plants often are exposed to drought. These profiles are predominant in the weathereing material area, where the soil often is very shallow. In the till area one might find there classes where the root distribution is shallow or the texture is course.

The capacities of plant aveilable water refers to cereals. The root-investigation showed that the capacity for grass is not far from that of cereales. The capacities for potatoes is, espescially on the high-capasity classes, lower, as a result of shallower root-depht.

Regional variations.

The acreage of each root-zone capasity class is shown in table 1.

Table 1. Root-zone capacity distribution for cereals in the Vigga-catchment area. Oppland County. Norway.
Sampling sitedistance 1 000 m x 200 m

			Lunner		Gran	
Root-zone capacity mm	Area ha	% of total area	area ha	% of tot.area	area ha	% of tot.area
200	485	9,5	220	12	265	8
175	690	13.5	235	13	445	13
150	815	16	370	20	455	14
125	1275	25	490	26	785	24
100	1325	26	435	24	895	28
75	355	7	70	4	285	9
50	155	3	20	1	135	4

The table is showing more drought-resitant soil (200+175+150 mm) in Lunner Kommune (45 %) than in Gran Kommune (35 %). Drought exposed soil (75+50 mm) covers 13 % of the area in Gran and just 5 % in Lunner.

Importance of sampling-density.

Instead of using 200 m intervals between the sampling-sites in the lines, as done in table 1, distances of 400 m and 1 000 m were tried.

Table 2. Root-zone capacity distribution calculated with lower sampling density.

Root-zone capacity mm	% of total area			
	Sampling site distance 1000 m x 400 m		Sampling site distance 1000 m x 1000 m	
200	11,7		11,1	
175	16,6	40,5	7,9	39,5
150	12,2		20,5	
125	28,6		39.6	
100	19,7	48,3	12,3	51,9
75	7,2		4,3	
50	4,0	11,2	4,3	8,6

Discussion.

Comparing table 1 and 2, it is obvions that for the total area of 5 150 ha, a distance between sampling sites of 1000 m x 1000 m will give adequate precision for estimating the need of irrigation.

For comparing smaller parts of the area, the precision will be not so good with the long smapling site distance.

Literature

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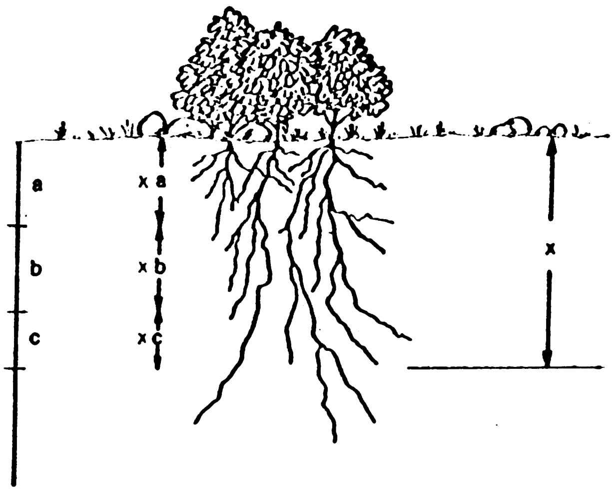

Fig. 1. Illustration of root-zone.

x = effective root depth, (dm)

a,b and c - specific water holding capacity in layers a, b and c (vol %)

x_a, x_b, x_c - thickness of layers a, b and c (dm)

R - root zone capacity (mm)

$$x = x_a + x_b + x_c$$

$$R = a\,x_a + b\,x_b + c\,x_c$$

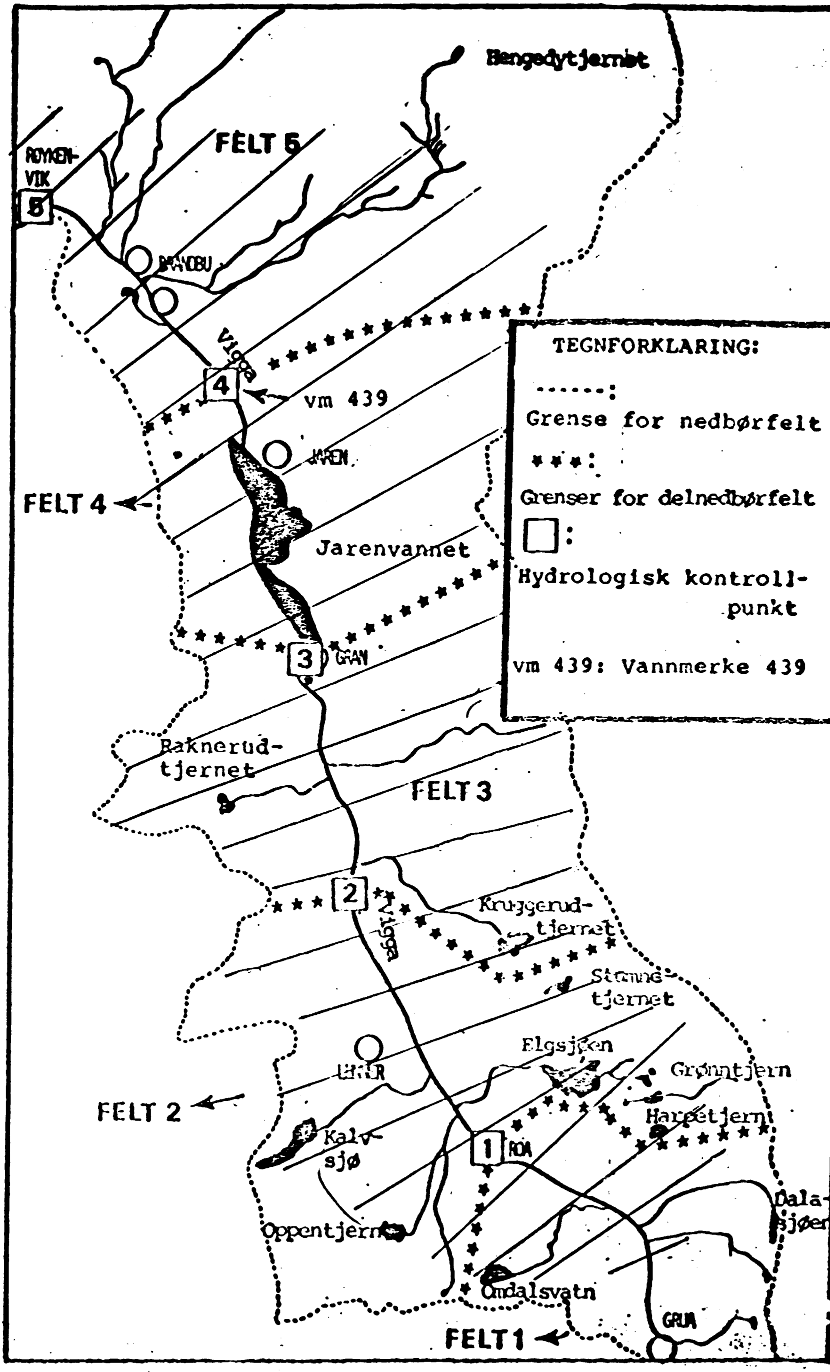

Fig. 2. Map showing the catchment of Vigga river. Sampling sites is located

along the cross yellow lines.

INFILTRATION AND VARIATION OF SOIL MOISTURE IN A SANDY AREA

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Abstract

The variations of soil moisture content, infiltration, cumulative percolation and wetting front movement in a sandy area in southern Finland are presented. The soil moisture measurements were made in the upper 3m layer. The groundwater table lay at a depth of 6.4... 7.8 m. On the basis of seventeen years of observations (1968-1984) the maximum April soil water storage was on the average 55 mm greater than the corresponding December value. The annual variation of the soil water storage was about 185 mm (6.2 vol- %). The typical velocity of the wetting front movement during the melting period was about 10 cm/day. The presence of soil frost decreased the hydraulic conductivity, but did not totally prevent the infiltration during the snowmelt period. During the frozen period the cumulative infiltration was more dependent on the initial soil moisture content in the frost layer than on the depth of the frost layer. The mean annual uncorrected precipitation at the field was 613 mm, of which over 60 per cent percolated through the one meter layer. Mean cumulative percolated water amount during winter-spring months was about 95 per cent, during summer 20 per cent and during autumn 60 per cent of uncorrected precipitation.

Key words: soil moisture storage, infiltration, wetting front movement, percolation, soil frost.

1. Introduction

The Hyrylä experimental field station (60°23'N, 25°02'E) is located in southern Finland in an open area surrounded by pine forest. The vegetation consists of lichen and heather. Among others, the following measurements are made at the station:

vertical soil moisture profile, infiltration, soil frost, ground water level, water equivalent of snow and precipitation. The textural composition of the soil layers was analysed in the laboratory. Soil types are sand and silt, sand being dominating. The soil layers are nearly horizontal. The ground water table varied between depths of 6.4 and 7.8 m during the course of the investigation.

Soil moisture measurements using a neutron scattering method have been made at the experimental field station at Hyrylä since 1968. The method is an indirect way of determining soil moisture content; the calibration curve of counts versus volumetric water content was made using gravimetric techniques (Lemmelä, 1970a).

The soil water content was measured along the vertical profile with a neutron probe at 10 cm intervals, usually once a month. During the melting period the soil moisture measurements were made about every fifth day.

In this study, infiltration during the frozen and unfrozen periods and data on soil water contents in the field station are presented. The seasonal variation of soil moisture in a layered sandy profile is studied. In addition, the wetting front movement in a vertical profile in 1969 and 1981 is presented.

2. Variation of soil moisture

On the basis of seventeen years of observations, the maximum soil moisture storage in April was on the average about 55 mm greater than the corresponding value in December (Figure 1). The difference between the annual maximum and minimum soil water storage varied from 65 mm (2.2 vol-%) to 350 mm (11.7 vol-%), the average difference being 185 mm (6.2 vol-%). The soil moisture storage was exceptionally small in 1975 and 1980. The wettest years were 1970, 1974 and 1981.

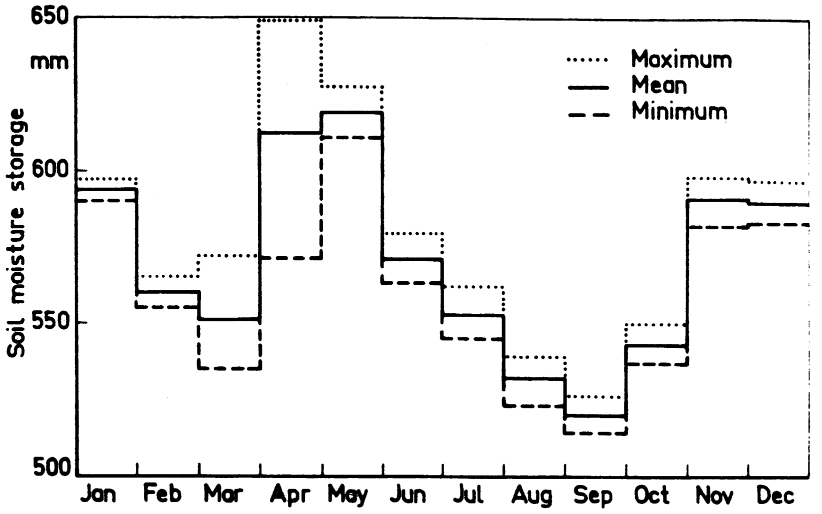

Figure 1. The average monthly mean, minimum and maximum soil water storages during the seventeen years 1969-1984 period at the experimental field station of Hyrylä.

The typical seasonal variation of the soil moisture storage is presented in Figure 2. The soil water storage in 1969 reached its maximum value, 670 mm (22.3 vol-%) on April 19th and decreased only slightly during the next four weeks. Thereafter the storage started to decrease continuously. The rainfall events in July caused only a slight increase, and the storage reached its minimum, 490 mm (16.3 vol-%) in the middle of September. Thereafter the storage increased, because of the decrease of evapotranspiration and because of the rainy periods in September and November. A small decrease in the soil water storage occurred in late October, after a relatively dry period. The field capacity value (Lemmelä 1970b), 605 mm (20.2 vol-%), was exceeded on

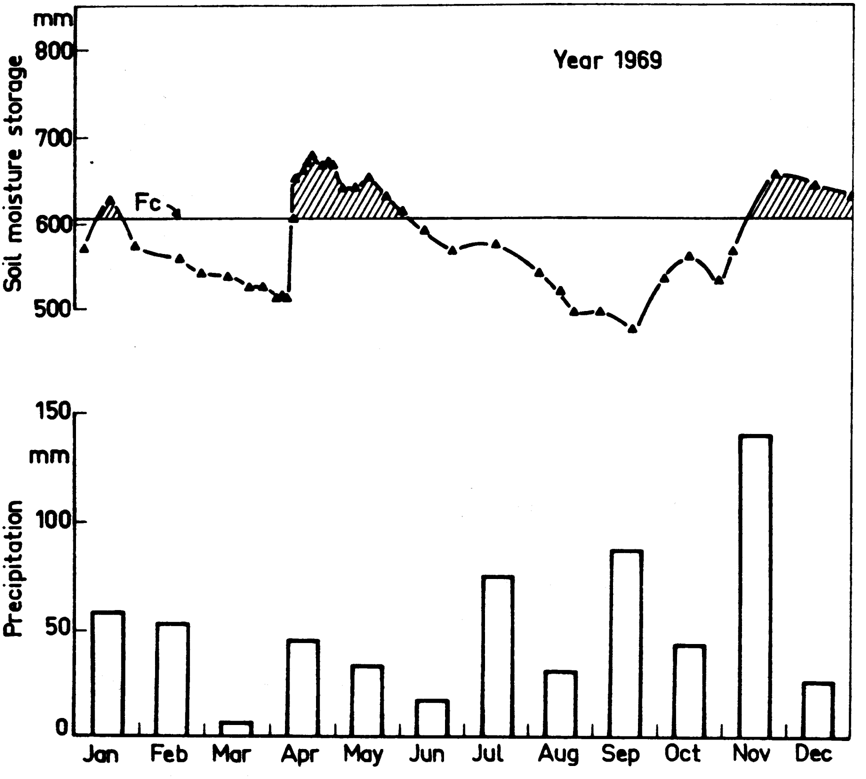

Figure 2 . Variation of soil moisture storage in a 3 m soil column and monthly precipitation in 1969 (Fc is field capacity).

three occasions, during the melting period (April-May), during the rainy late autumn months (November-December), and also after a January thaw.

3. Infiltration during frozen and unfrozen periods

Infiltration measurements at the area were made with a double ring infiltrometer. The diameters of the inner and outer cylinders were 225 mm and 350 mm, respectively. The height of the cylinders was 350 mm. When measuring the infiltration the cylinders were pressed into the ground to a depth of 10 cm. Water was added to the double ring infiltrometer in such a way that the water table was about two millimeters above the soil surface and at the same level in both of the rings in order to obtain a small and equal water pressure on both cylinders. A burette with valves was installed above the inner cylinder as shown in Figure 3, in order to maintain a constant water table during the measuring experiment. The infiltration was calculated from the amount of water needed to maintain the water level at a constant level in the inner cylinder. The function of the outer cylinder was to prevent the water within the inner space from spreading in horizontal directions.

Examples of the mean infiltration capacity both during an unfrozen period and during a frozen period are shown in Figure 4. The approximate mean value of the final infiltration capacity during the frost period (frost depth 53 cm) was about 3 mm/min and during the unfrozen period between 5 and 6 mm/min. The infiltration reached its final capacity between 1 and 1.5 hours after the start of each experiment both during the frost and in the unfrozen period. The temperature of the water used in these experiments varied between 0.0 and 3.2°C according to the climatological conditions and soil temperature. Because the changes in water temperature were small, no viscosity corrections were made.

According to the field measurements made in different spring seasons during the frost period, the depth of frost did not influence infiltration capacity as much as the initial water content of the frost layer. Figure 5 shows the final infiltration capacity as a function of initial moisture content of the soil frost layer.

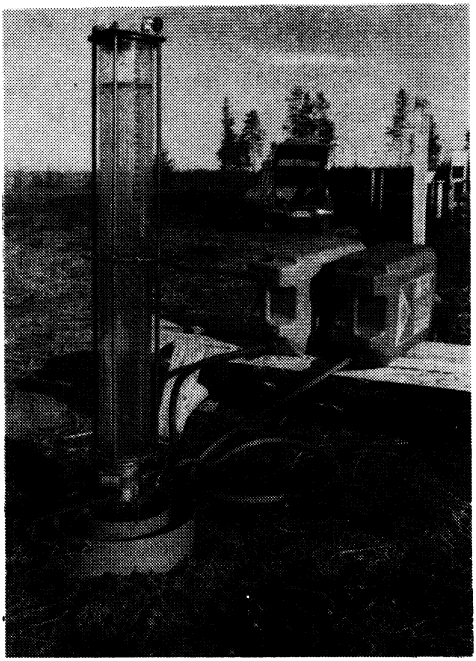

Figure 3. The double ring infiltrometer.

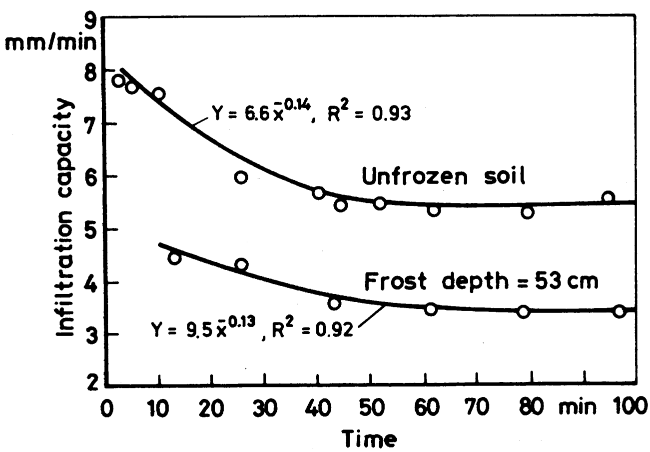

Figure 4. The mean infiltration capacity during the frozen and unfrozen periods.

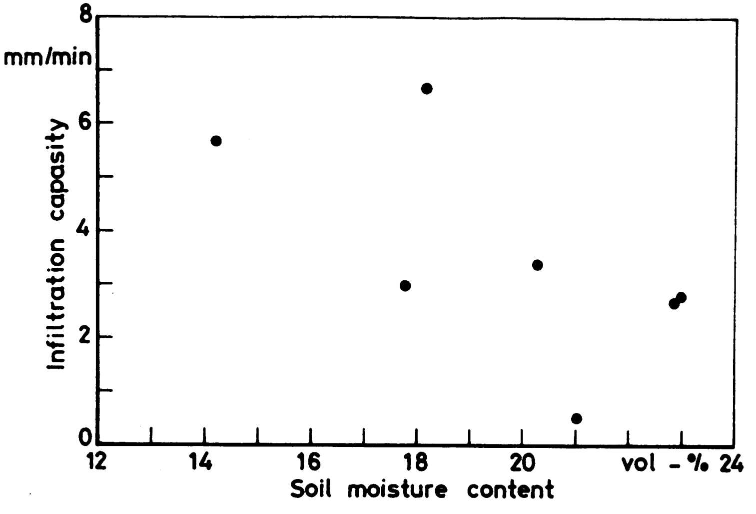

Figure 5. The final infiltration capacity as a function of initial moisture content of the soil frost layer.

At the experimental field station the maximum daily melting as registered with a snow pillow and drip-pans varied during the melting periods 1968...1973 from 8.4 mm (1971) to 29.3 mm (1968) (Lemmelä and Kuusisto, 1974). Although the results concerning the infiltration capacity do not necessarily correspond with those occurring under natural conditions, it can be concluded that in springtime the infiltration capacity was more than two orders of magnitude greater than the maximum snowmelt intensity. Therefore all the meltwaters could easily infiltrate into the soil, without the occurrence of any surface flow.

A cubical lysimeter with a volume of 1.0 m^3 was also used at the Hyrylä experimental station. Figure 6 shows the cumulative outflow from this lysimeter in the spring of 1969. When the 56 cm frost layer started to melt from the surface (on April 13th), the cumulative outflow from the lysimeter already amounted to 74 mm. Within the same period the average water content of the uppermost 1 m soil layer increased about 130 mm.

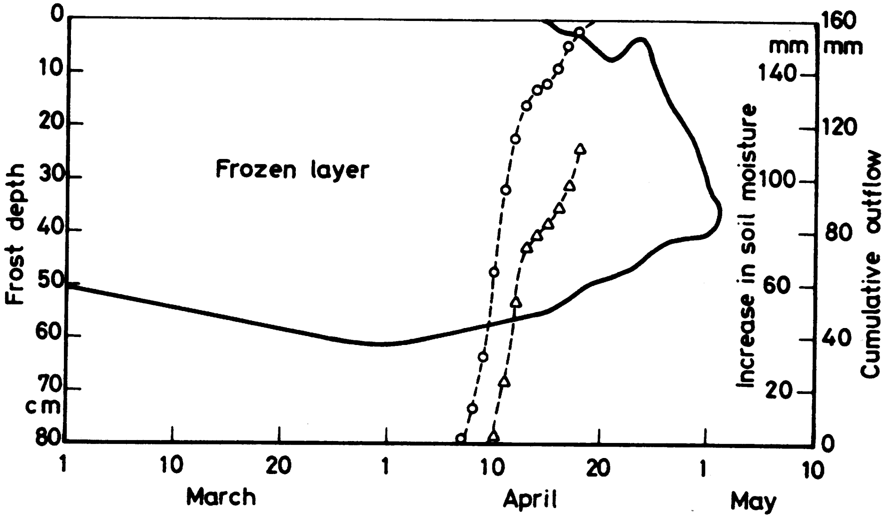

Figure 6. Frost depth (——), increase in soil moisture (o---o) and cumulative outflow (Δ---Δ) from the uppermost 1 meter layer in spring 1969.

4. Cumulative percolation

The percolation from the lysimeter was analysed separately for winter-spring, summer and autumn periods (Fig.7). The mean cumulative values of percolation during these periods were 220, 40 and 123 mm, respectively. These amounted to 94, 22 and 61 per cent of the corresponding totals of uncorrected precipitation. From the mean winter-spring value (220 mm) 182 mm percolated during the snowmelt period. On an annual basis, the percolation was 62 per cent of the uncorrected precipitation.

In early winter, rainfall and melting events rather often caused some percolation to occur. In January-March percolation was a rare phenomenon. About 50 per cent of the annual percolation occurred in April-May during and after the snowmelt period.

In summer the percolation usually occurred only when rainfall exceeded 25 mm per a single event. In autumn the decrease of evaporation and a smaller soil moisture deficit increased percolation considerably. The total average duration of percolation was 31 days in winter-spring, 14 days in summer and 24 days in autumn.

When calculating the real percentages of percolation from precipitation, the corrections for the precipitation according to 1968-1969 values were 7 per cent in the case of rainfall and 23 per cent of snowfall (Lemmelä, 1970a). The figures coincide with the results mentioned in the WMO publication, 1982.

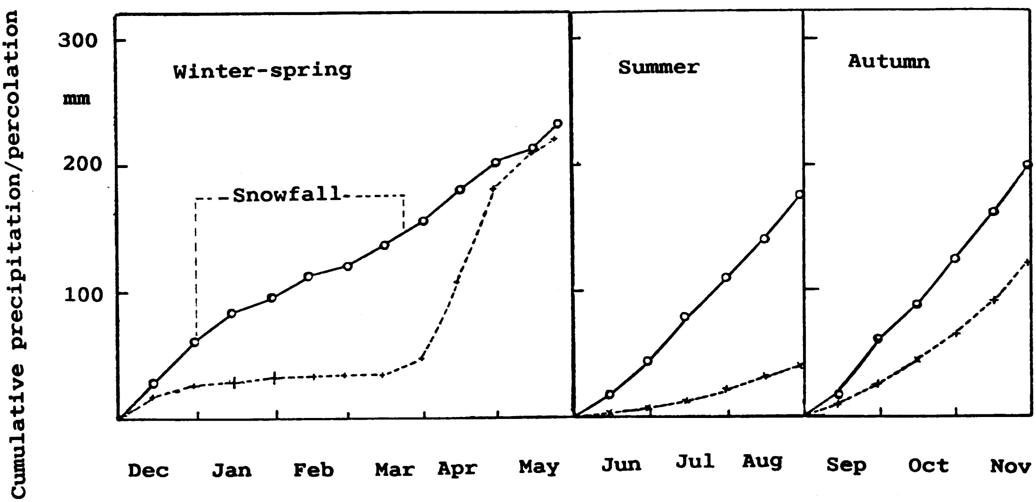

Figure 7. Cumulative percolation (----) and uncorrected precipitation (——) curves during winter-spring, summer and autumn months.

5. Wetting front movement

In soils which have restricting layers, such as those in Hyrylä, the infiltration process is complex. When the wetting front reaches the restricting layer the water content above this layer starts to increase. The time taken for the wetting front to reach the groundwater table depends mainly on the melting rate and on the soil structure. When the soil is frozen, the conductivity of water is less than when it is unfrozen.

The variation of soil moisture storage in three consecutive layers in Hyrylä during the melting periods of 1969 and 1981 is presented in Figure 8. The thicknesses of the layers were 100, 80 and 100 cm, respectively. The textural compositions of the soil layers are also shown. The soil moisture content reached its maximum value in the upper layer 8 days after the wetting front started to move (April 11th, 1969). In the second layer, the corresponding maximum was reached after 15 days and in the bottom layer after about 35 days. Thus a clear delay between the layers was observed. The fine grained silty layer within the layer 2 did not considerably affect the delay.

The increments in the soil water contents were 12, 9 and 6 volume per cent, respectively. In the first layer the soil moisture storage increased very rapidly, whereas in the second layer the increase was slower. In the third layer the storage remained approximately constant for about one week and then started to increase. The average daily velocity of the wetting front for the whole profile in 1969 was 8 cm/day.

In 1981 the wetting front started to move on April 1st. Due to the occurrence of two melting seasons, two water storage maxima could also be distinquished in the two uppermost soil layers. They occurred almost simultaneously in both layers. In the third layer only one maximum occurred. In this case the increments of water contents above the winter averages were 13, 12 and 8 vol-%, respectively. The average daily velocity of the wetting front for the whole profile was 10 cm/day.

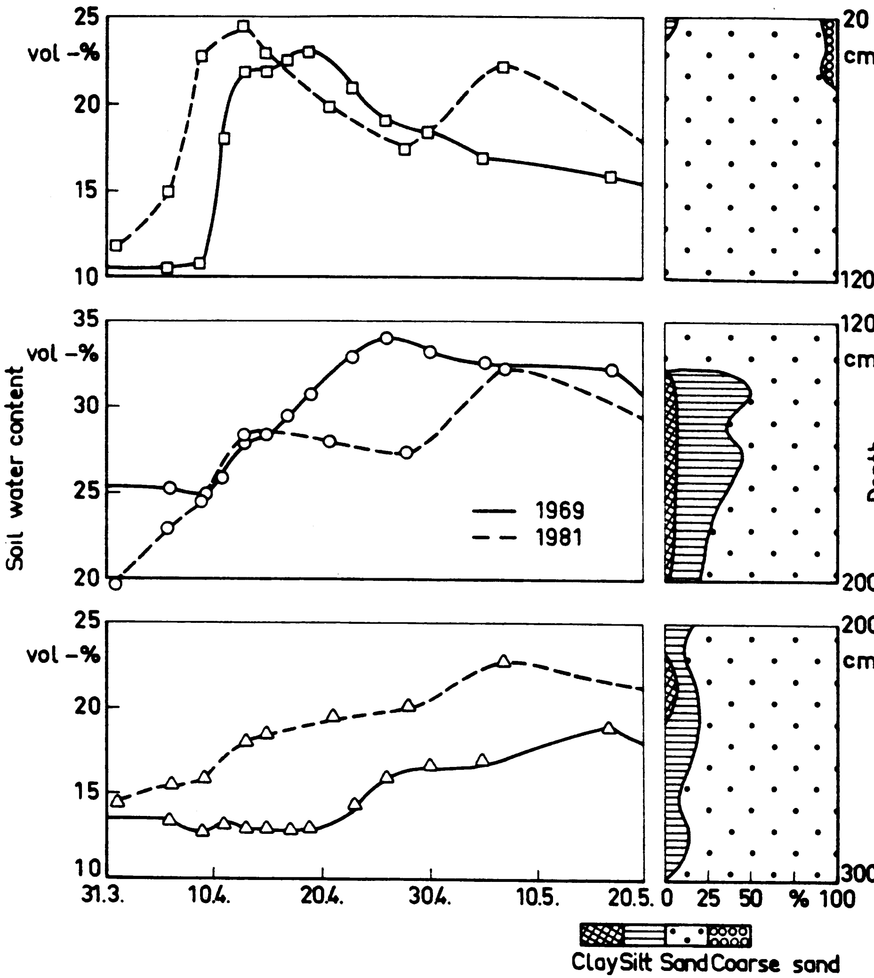

Figure 8. Variation of the soil moisture content in three consecutive layers (100 cm, 80 cm and 100 cm respectively) during the melting periods of 1969 and 1981.

6. Conclusions

The following conclusions can be drawn concerning the infiltration and variation of soil moisture in a sandy area in southern Finland in 1968-1984.

(1) The soil moisture storage varied with an average annual amplitude of 185 mm. The largest increases of this storage occured in springtime, but considerably high values were also reached in November-January.

(2) The infiltration capacity was considerably smaller in the frozen period than in unfrozen period. However, even in the frozen period in springtime, it exceeded the rate of snowmelt by more than two orders of magnitude. A considerable amount of meltwater infiltrated even before the soil frost started to melt from the surface.

(3) The cumulative percolation in springtime was on the average 182 mm; its duration was typically one month. On an average more than 60 per cent of the annual uncorrected precipitation percolated through the one meter layer into the soil. Fifty per cent of this value percolated during the melting period.

(4) A typical velocity of the wetting front movement in the springtime was 10 cm/day. The layered structure of the soil disturbed the wetting front movement to some extent.

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TIMELINESS OF A SILTY CLAY AS RELATED TO DRAINAGE INTENSITY

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Summary.

A higher intensity of artificial draining in a silty clay increaces the sub-surface drain run-off rate in proportion to the length of drain pr. square area drained. A closer grid of drains also makes the soil ready for drilling at an earlier date. The land is prepared for drilling with a higher moisture-content in the surface layer of soil, which is beneficial for agriculture. Better drainage results in better crops and the fields are easier workable. Years with complete crop failure will also become more scarce.

Scope of experiment.

The aim of the experiment is to study the influence of the drainage intensity on the value of agricultural land when used arable for spring-drilles grain production.

Farmers have for years been aware of the importance of an early spring. An old saying is that a day in the spring is equal to a week in the autumn. This may be an exaggeration, but still how important is an eary drilling ?

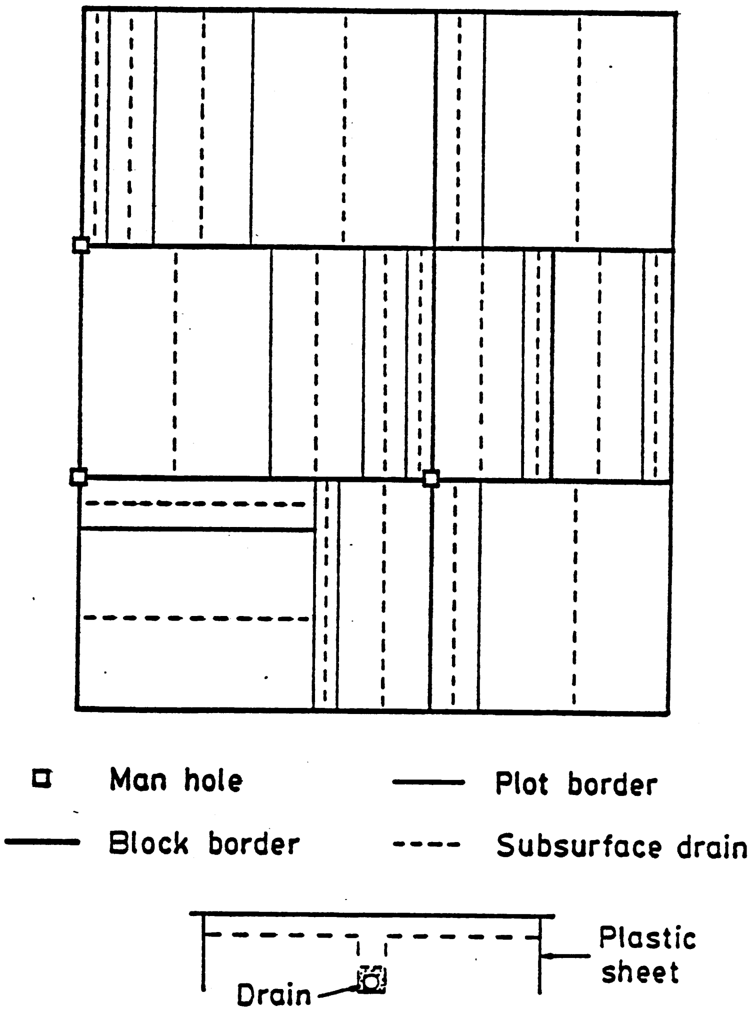

Figure 1. Experimental lay out. End view of drainage plot.

Sub-surface drains with short intervals will drain away surplus water at a rate depending on the drain distance. In the experiment at Støkken described here, drain distances at 4, 8, 16 and 32 m were tested.

A water devision consisted of á 0,4 mm thick vertical plastic sheet down to approx. 1,0 m (fig. 1), which also shows the experiment´s layout.

The silty clay at the experimental site is overlaid with 10-20 cm of organic soil. The upper 30-50 cm of the profile has got a fairly high permeability. Further down the permeability is extremely slow (fig. 2).

The run-off from the drains indicate that the permability of the backfill keeps on i high level down to the drainline, which is enveloped in gravel (fig. 1).

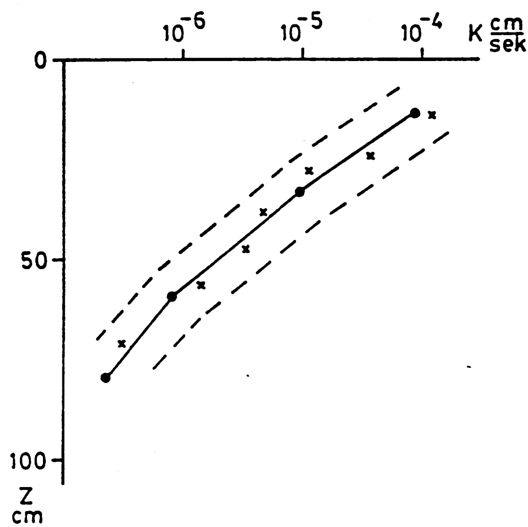

Figure 2. Permeability of various depth´s at the Støkken exp. in 1975 before experiment started. x shows measurement taken in 1982.

The runoff time diagram (durability curve) for the drain system is shown in fig. 3.

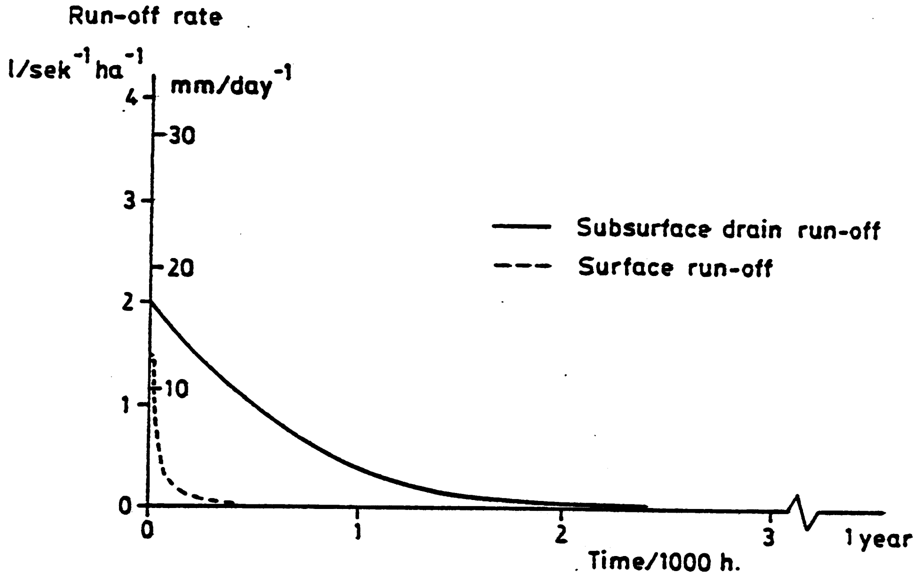

Figure 3. Intensity of run-off time diagraph for subsurface drain run-off water (solid line) and surface water run-off (dotted line) (1983).

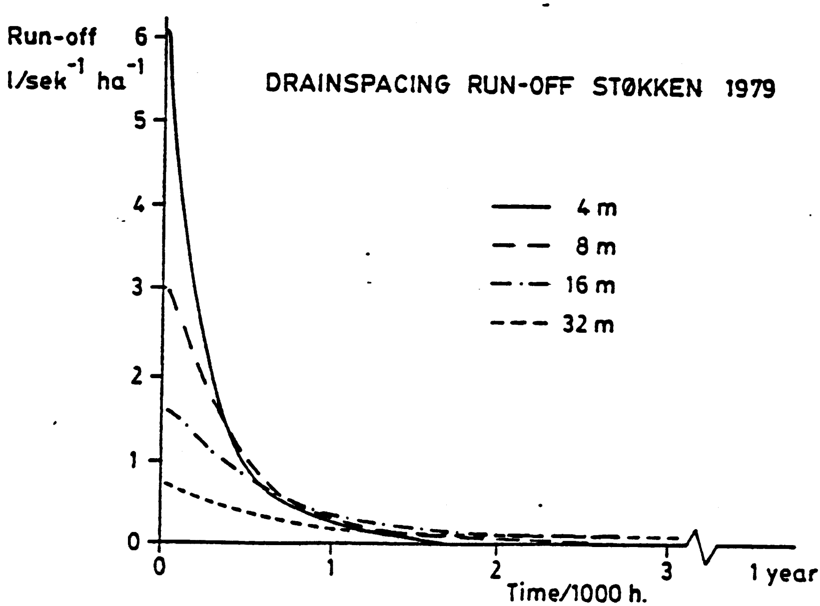

Figure 4. Intensity of run-off - time diagraph for sub surface drains at various distances.

Fig. 4 shows durability curves for various drain spacings. One sees that the max. rainfall is approx. proportional to the length of drain per square area.

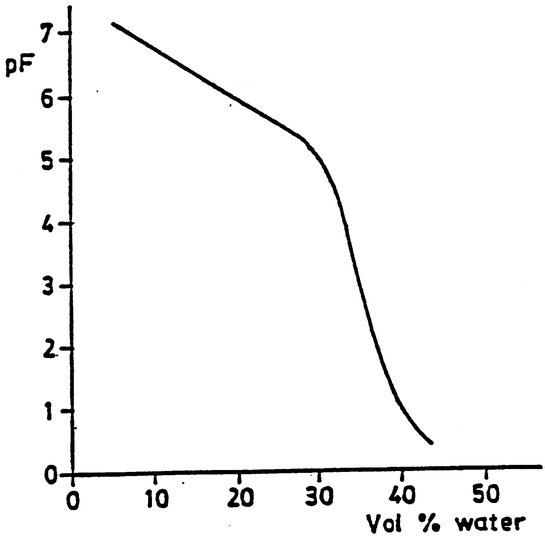

Figure 5. Retention curves for the soil at Støkken.

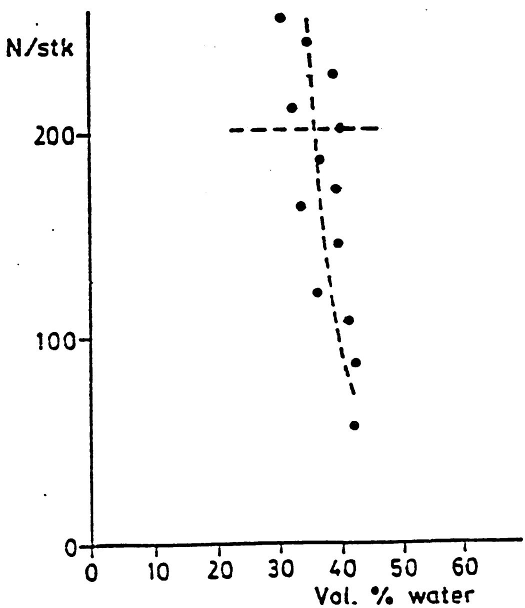

Figure 6. Relation soil strength as measured with a prector meter and water content.

Water regime - strength of soil.

Coulomb's law states that the strength of a material is porportional to the normal stresses miltiplied by the angle of friction. In the upper layer of the soil the normal stresses mainly depend on the water suction in the soil pores. A volume % moisture - suction diagram of the Støkken soil is given in fig. 5, and in fig. 6 a volume % moisture strength relation diagram is given.

The drying up process.

After snow melting the soil is usually saturated with water. The time it takes to reduce the water content to a level indicated in fig. 7 determines the drilling date.

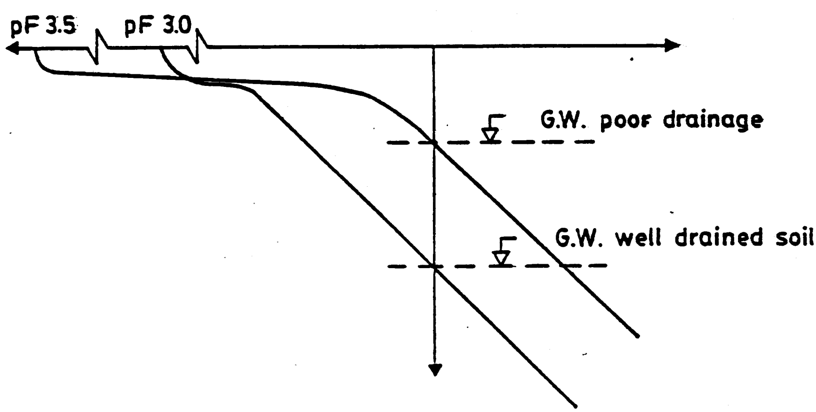

Figure 7. Tension at various depth of soil on well drained and poorly drained ground when the soil is ready for drilling.

Water can be removed by evaporation or by simple gravity drainage. It takes 2,25 MJ/kg to evaporate water (100 °C, at lower temperatures this figure is reduced). The soil at Støkken requires a removal of around 50 mm or 500 m^3/ha. That is the equivalent of over 300 000 kWh of energy, to a large extent can be saved by letting the water drain away under the force of gravity. But that requires a rather close grid of drains in such low permeability soil.

Why the drilling date is of such importance.

The dry matter producing potential of a field depends on how efficient the photosynthesis can function. This again depens on the rate of radiation, and the length of time it goes on, and that there is no bottleneck in the process like draught or lack of fertilization etc. Fig. 8 shows the incoming radiation at Ås,and it is obvious that good timing is essential, especially under these local climatic conditions.

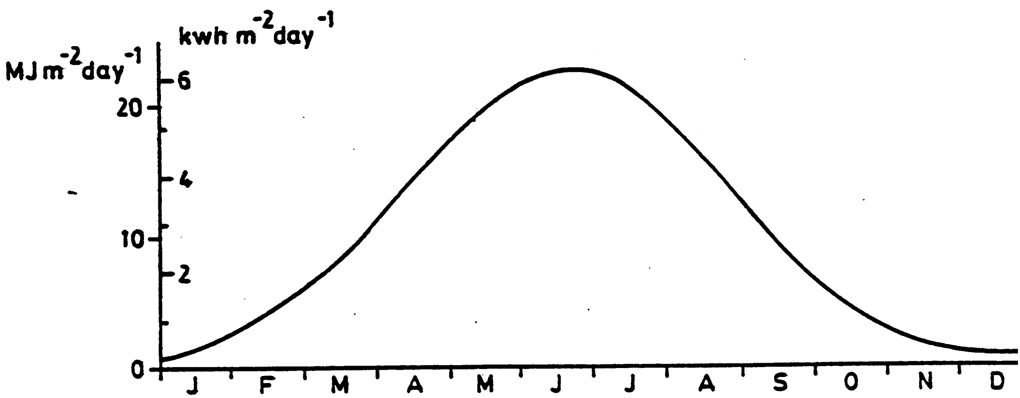

Figure 8. Incomming radiation at Ås through the year.

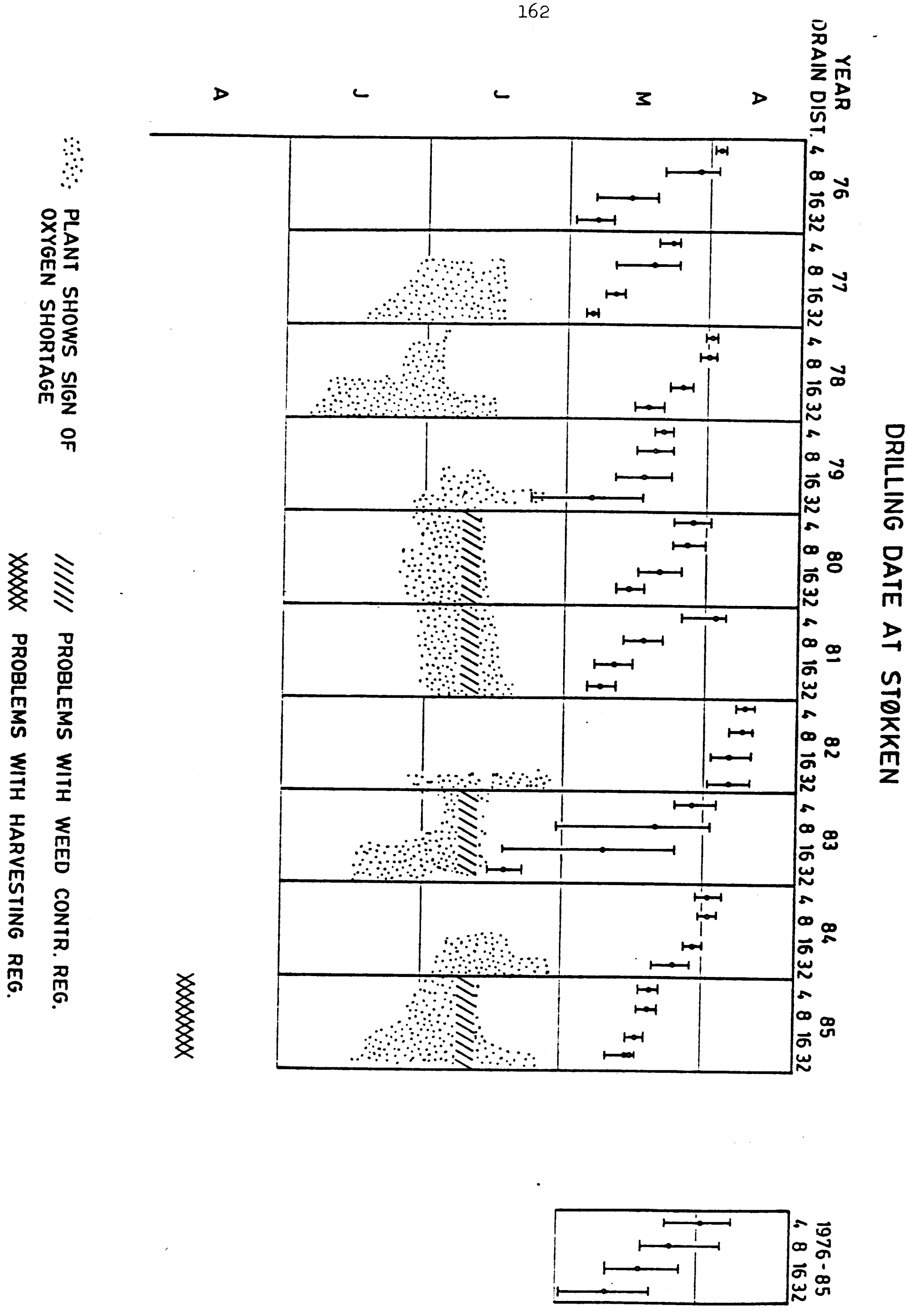

Figure 9. Drilling date on various drain distances through the years 1976-85. Vertical line indicate variation for a drain distance (5 replications).

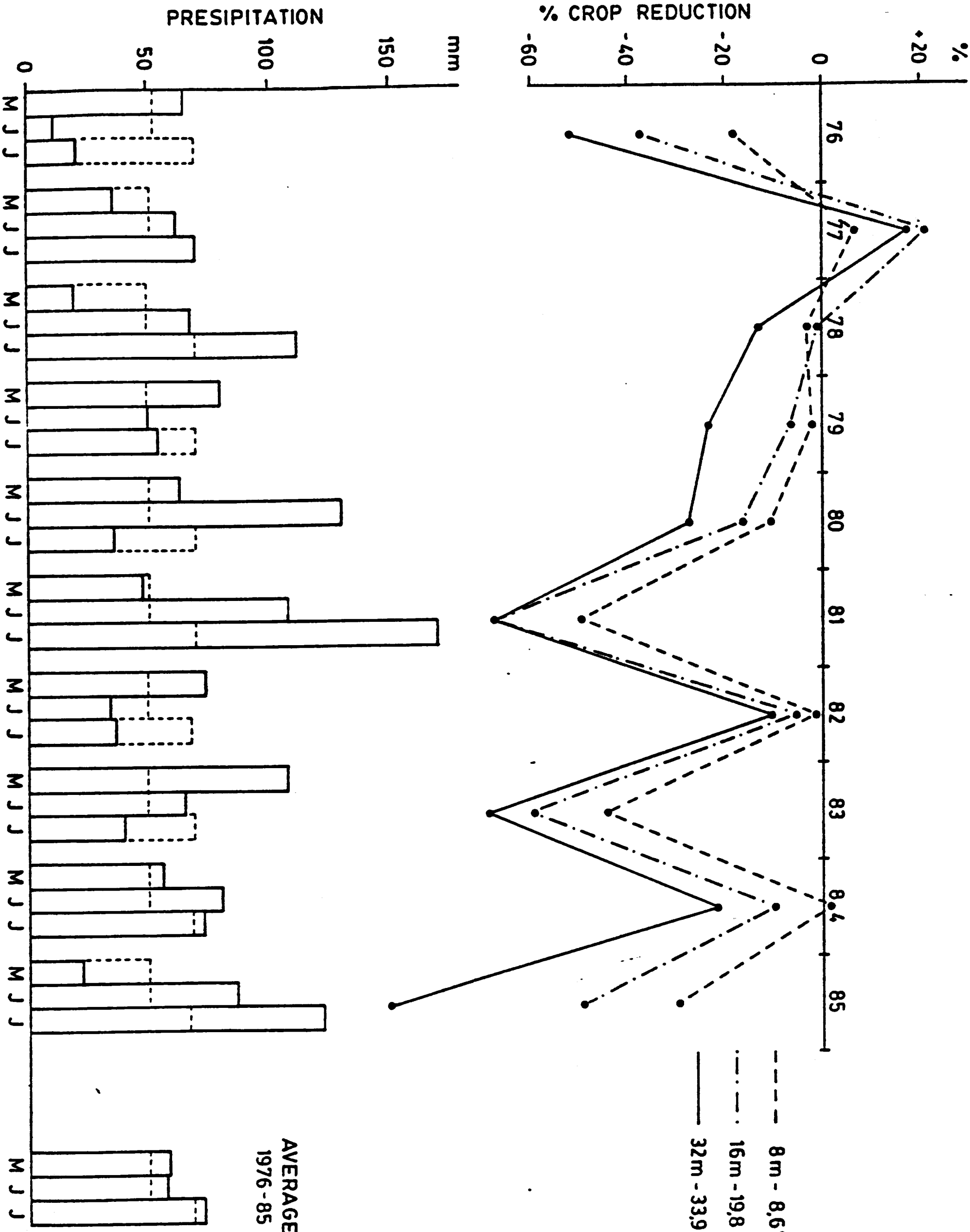

Figure 10. Crop harvested on plots with 8, 16 and 32 m drain spacing measured with the crop on plots with 4 m drain distance. Vertical column shows precipitation in May, Juni and July. Dotted line shows normal precipitation.

Experimental results.

Naturally the effect of drain spacing on the sowing date and the crop varies from year to year. In fig. 9 the sowing date is shown for the various drain distances. On an average over ten years a doubling of the drain distande delays the drilling operations by 6 days. In years with no rain in the drilling season this delay is shorter, approximately 2 days, but in wet years it has been up to three weeks.

Economical importance of drainage.

As fig. 9 indicates, and as shown by the photographs, lack of oxygen may also hamper the crop production. In fig. 9 it is indicated in which years this happened (1977- ....).

Literature.

P. Hove   Sluttrapport   ISBN 82-7290-076-9.

SPATIAL VARIABILITY OF SOIL MOISTURE IN URBAN AREAS

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ABSTRACT

A number of Swedish soils with varying clay contents have been studied with respect to the variability of two soil physical properties: air-filled porosity at 10 kPa suction and saturated hydraulic conductivity. A comparison is made with a Swedish sports turf.
Only soils with very low clay and high sand contents show both high mean values and relatively low variability.
For all soils, except those with a high sand content, the range of values includes figures which are very low for both of the named soil properties. A thorough examination of the causes of these low values is indicated in order to provide a basis for adequate advice on soil amelioration measures and mixtures for man-made soils in urban areas.

INTRODUCTION

"Green" urban areas show a great variety in land use: from on the one hand, intensively-used sports turf or highly frequented parks and gardens and on the other hand unfrequented bush and tree plantations. Both types of areas are often damaged by severe compaction during construction.

This paper is primarily concerned with these vegetated areas. The variability of two important physical properties, air-filled porosity at 10 kPa suction and saturated hydraulic conductivity, is discussed. These soil properties have been chosen to illustrate different patterns of variation on different soil types. The figures presented indicate probable variability of Swedish urban soils.

In contrast to agricultural, horticultural and forestry applications of soil physics, crop production is of secondary importance in urban areas. However, proper evaluation of soil physical data may provide valuable information in a wide spectrum of problems, such as surface water drainage management, soil amelioration measures on intensively used areas, land use planning etc.

The soils selected for comparison of variation have not been sampled systematically. They are examples of types of soils which have been sampled under very different climatic conditions and soil moisture contents. Due to the lack of data, only one true urban soil is presented but in most cases urban soils are either of forest or agricultural origin.

In urban soils, layers with high humus contents and severely compacted layers can be found at varying depths in the profile. Therefore, within each soil type no distinctions have been made between topsoils and subsoils, or between samples which are more or less compacted. It is assumed that the data from forested or agricultural land are useful when assessing criteria for urban soils.

CHOICE OF SOIL DATA

The data on Swedish soils which are presented in this paper come from several sources:

The forest (till) soils (from Lundin, 1982) are from 6 profiles on a slope in Masby, southern Dalarna, in the central inland of Sweden. The soil was sampled at 10 cm intervals down to 1 m depth. Soil porosity data was corrected for stone and boulder contents. Peat layers have been excluded.

The agricultural soils (from Andersson & Wiklert, 1972) are from 51 profiles of sites throughout Sweden. Most of these were sampled at 10 cm intervals down to 1 m depth. In this paper, these agricultural soils are classified according to dominant soil fraction and clay content. The soil physical data presented are from Andersson & Wiklert (1972) (air-filled porosity) and Andersson & Wiklert (1977a,b,c) and Wiklert et al. (1983 a-g)(hydraulic conductivity data). Peat soils have been excluded.

For soil moisture characteristics, as well as for measurement of hydraulic conductivity, cylinders of 50 mm height (forest soils) and 100 mm height (agricultural soils) were used. Sample diameter is 72 mm in all cases.

For each layer, the soil moisture retention value is a mean of measurements on 4 undisturbed cores and the saturated conductivity value is a mean from 2 cores. Organic matter content varies but most samples are subsoils with very low or zero values of organic matter.

Finally, the soil data of the sports turf "Studenternas" in Uppsala city (80 km north of Stockholm) were collected by the author. The sampling cylinders used had dimensions 50 mm (height) x 72 mm (diameter). Samples were taken at 24 different plots distributed over the 65 x 105 m area. 6 replicates were taken from each of the two horizons distinguished at each plot. 3 of these were used for soil moisture characteristic determination and 3 for hydraulic conductivity measurements. All in all, a total of 12 soil cores were sampled at each plot. Maximum soil depth varied from 11 to 23 cm.

CHOICE OF PHYSICAL PARAMETERS

The air-filled porosity at 10 kPa suction can be used to separate micropores from meso- and macropores (Bouma, 1981). It has been shown that larger pores, which hold water at suctions less than 10 kPa, are more vulnerable to compaction than smaller ones (Eriksson, 1982b). Thus the air volume at 10 kPa suction is a rough estimate of the actual structure of the soil. It is also often assumed to represent the situation at the "field capacity" moisture content in the upper horizons of the soil. A minimum of roughly 10 % air by volume is commonly accepted as adequate to meet the oxygen requirement of most common plants (Eriksson, 1982b; Edling, 1973; Ward, 1983).

Hydraulic conductivity values are of great importance in all hydrological calculations and can effectively indicate the potential of urban soils for use as sports pitches and recreational areas.

For intensively used areas, it is often useful to compare infiltration and permeability characteristics of the soil with relevant climatic data. A suggested dimensioning figure for sports turf is that saturated hydraulic conductivity should equal the highest local precipitation intensity of at least 1/2 hour duration and with a return period of one year (Adams, 1981). For much of Great Britain, this figure is about 25 mm/h (see Table 3, "Ideal pitch"). In Sweden, the corresponding figure would be 11 mm/h in Stockholm (Eriksson, pers.comm.), and 20 mm/h in Gothenburg (Bergström, 1976). This lower limit for hydraulic conductivity can be extended to most vegetated urban areas, since surface water is very troublesome in densely populated residential quarters.

METHODS

Suction apparatus and constant head apparatus for hydraulic conductivity measurements in the laboratory have been described by Andersson (1955) and (1953), respectively.

A single-ring infiltrometer cylinder of 1 meter diameter was used for field measurements of hydraulic conductivity at "Studenternas". The ring was driven down into the soil to a depth of approximately 10 cm. Initial water height above soil surface was 20 cm. Water was allowed to infiltrate into the soil for at least one hour before measurement to ensure a field-saturated situation. 6 readings were taken at 10 minutes intervals. The method is further described by Eriksson (1982a).

RESULTS AND DISCUSSION

Variability of air-filled porosity values

The data on air-filled porosity at 10 kPa suction (Table 1) show increasing variability with increasing silt- and clay content. It can also be seen that mean values are highest for soils with a high sand content.

A comparison between the normal and log-normal distributions in Table 1 shows that the mean values are considerably lower when a log-normal distribution is assumed for all soils except for soils with high sand contents. This indicates that one may not be allowed to assume that the values are normally distributed, a fact which makes it difficult to evaluate the variability of this parameter.

The range of data has interesting implications. Except for pure sands, there is a considerable risk that, under certain conditions, very low air contents are created in most soils. Such conditions are, for example, severe compaction or high silt content. It would be interesting to examine the factors causing these minimum values, especially in relation to appropriate recommendations for amelioration of compacted soils and for the building of "man-made" soils in urban areas.

Fig. 1. shows the frequency (%) of samples from different soil types with air contents less than the critical 10 % by volume.

Table 1. Air-filled porosity at 10 kPa suction of common soil types in urban areas

Dominating particle size	%clay	n	Porosity (cm^3/cm^3 x 100) Norm. distr. mean	stand. dev.	Log-norm. distr. mean	stand. dev.	range
1.Silt(till)	< 5	50	12.9	9.3	6.1	8.1	0- 52
2.Sand	< 5	29	27.0	8.6	25.3	1.5	9- 42
3.Silt	< 5	6	5.7	4.6	4.4	2.1	2- 12
4.Sand	5-15	20	15.8	7.1	14.1	1.7	4- 27
5.Silt	5-15	23	7.7	5.2	6.2	2.1	1- 19
6.Sand&silt	15-25	28	7.5	3.7	6.5	1.7	2- 15
7.Sand&Silt	25-40	58	7.4	4.5	4.7	5.7	0- 23
8.Clay	40-60	73	5.5	4.6	2.0	13.6	0- 20
9."Studenternas"	0-34	144	9.1	5.8	6.6	3.6	0- 24
Recommended ideal pitch*	< 5						10- 25

*(Ward, 1983)

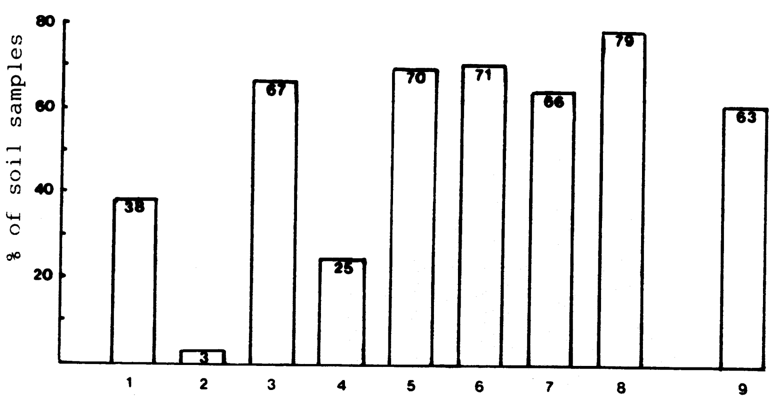

Fig. 1. Percentage of samples with lower air-filled porosity values (at 10 kPa suction) than 10 cm^3/cm^3 x 100

Variability of saturated hydraulic conductivity

In statistical analysis assuming a normal distribution saturated hydraulic conductivity values show much greater variability than air-filled porosity (table 2.) In fact, there are good reasons to assume that the distribution is log-normal. In either case, the range of data shows that measurements of this property are highly variable.

In Fig. 2. the values of hydraulic conductivity for different soil types are compared to the critical value of 11 mm/hour in the Stockholm area mentioned earlier.

Table 2. Hydraulic conductivity (K_{sat}) of common soil types in urban areas

Dominating particle size	%clay	n	K_{sat}, (mm/h) Norm. distr. mean	stand. dev.	Log-norm. distr. mean	stand. dev.	range
1.Silt(till)	< 5	50	101	144	34	7.0	0- 850
2.Sand	< 5	29	236	315	109	3.6	10- 1080
3.Silt	< 5	6	40	43	22	3.3	6- 110
4.Sand	5-15	20	55	64	25	4.9	1- 290
5.Silt	5-15	23	31	50	2	49.8	0- 170
6.Sand&silt	15-25	28	79	216	3	28.9	0- 1010
7.Sand&silt	25-40	58	259	472	16	50.1	0- 2500
8.Clay	40-60	73	666	2580	16	45.2	0-21190
9."Studenternas"	0-34	138	16	36	3	6	0- 222
Recommended ideal pitch*	< 5						50- 150

*(Ward, 1983)

Table 3. Field data on hydraulic conductivity ($K_{field-sat}$)

Soil	%clay	n	$k_{field-sat}$ (mm/h) Norm. distr. mean	stand. dev.	Log-norm. distr. mean	stand. dev.	range
10."Studen-ternas"	0-34	72	7	9.0	3.1	2.8	0- 40
Recommen-ded ideal pitch*	< 5						> 25

*(Ward, 1983)

Fig. 2. Percentage of samples with lower saturated hydraulic conductivity values values than 11 mm/hour

In a current study of laboratory and field methods for hydraulic conductivity determination on Swedish soils, the variability of laboratory measurements has been shown to increase sharply with high clay contents (Messing, pers. comm.) Measuring in the field can reduce by half the range shown in Table 2, see the results of field measurements on "Studenternas" in Table 3.

Within-pedon variability

The variation patterns described in Tables 1-3 are probably somewhat extreme, since the soils originate from all over Sweden. However, within-profile variation is also high. For example, the coefficient of variation within some of the 24 sampling sites of "Studenternas" (each sampling site being only about 0.1 m^2) was about 80 % (air content at 10 kPa suction) and 150 % (saturated hydraulic conductivity).

In natural soils, the greatest compaction often occurs in the deepest layers of the soil profile. In urban areas, however, a common practice on construction sites is to remove and store the topsoil during building. Consequently, soil 0.3-0.5 m below the final soil surface may be considerably more compacted than the layers above and below. Thus, the variation with depth may be very high, although a natural regeneration will begin as soon as construction is completed.

It has been observed in many places that compaction damage has caused the establishment of bushes and trees to be delayed by several years, with very high plant replacement costs. These problems indicate a quite slow regeneration process and underline the necessity of finding the "bottlenecks" within the profile, rather than trying to produce a mean value for the entire soil profile.

General remarks

The values of both air-filled porosity and hydraulic conductivity at "Studenternas" are too low for both international standards and with regard to the local climate. A sports turf is continually damaged by compaction due to trampling, and a comparison of values for different soils (Tables 2 and 3) supports international recommendations that sports turf should be established on a well-defined sand. These low values are most probably due to the intensive use of a soil with too high clay and silt contents (a high susceptibility to compaction).

The variability of different soil physical properties has been discussed in a number of recent works (Warrick & Nielsen, 1980; Webster, 1985). These authors stress that the future approach to soil surveying and to field work in general has to be much more systematic than in previous years. Generally, variation in the values of most soil properties is very high but the patterns of variation patterns are also different for different properties, different areas, land uses, etc. Statistical methods (i.e. geostatistics) have been developed to improve sampling and interpretation of data, and to minimize the variability problem. Future Nordic soil research must also give high priority to this line of investigation.

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INTRODUCTION TO DISCUSSION 2, BY BENGT ROGNERUD

The session has given a good overview of work done on soil moisture content, soil moisture distribution and soil water retention characteristics on forested and cultivated lands in Scandinavia. The discussion must concern different aspects. We have gone through detailed studies of water movements, for instance by using micro-fabric studies and photogrammetric methods. Larger scales were presented in the first session but have also been considered in this section, e.g., by the presentation of results from infiltration tests. The sampling density and the importance of different soil parameters on soil water contents have been discussed. It shows us the complex nature of the soil water variability problems. Even in local studies great variabilities have been reported. Till is an extremely difficult soil type to study. What do we gain when we study the variability within small till samples? How representative are the data on a larger scale? For most practical purposes, we must have a more regional view and must know how to integrate the local variability into a more regional scale. It is important to focus on water resource planning, but even then it is important to know how to sample and which parameters to study.

The papers presented today show that these aspects must be carefully considered before a field work programme is carried out.

SUMMARY OF DISCUSSION 2

During the meeting it was not possible to point out what sort of problems to concentrate on. It was necessary to make a distinction between small-scale studies and regional out-looks. The variable may be of a continuous or mosaic nature. We must decide which processes are important to study on a small scale and which processes must be studied on a larger scale.

Problems concerned with studies of tills are, for example, the following:

What types of tills are we studying? A till is probably the most variable soil from a geological point of view. There are very compact and clay-rich tills in Germany, Denmark and Scania, sandy-silty tills in central Scandinavia, homogeneous tills and tills with bands and lenses of sorted material. And even within one single till type, the variability is considerable. It is for instance shown how important it is that the method used for measuring the variability is appropriate for the variability we want to describe. Bouma has shown that by using the wrong size of a tensiometer body one may obtain an unrealistically high variability. Results from methods which have used different scales are very difficult to compare and difficult to combine in soil water variability data sets.

Over many years various studies of water in tills have been carried out. It is important to summarize our present knowledge about about spatial soil water variability, and what to focus on in the future. It should also be decided which methods are suitable for the different scales.

In till, for instance, it is not satisfactory to use only mean values. You have to know the variability because it is so great. You need to use statistics and a systematic approach, and it is necessary to relate it to the scale.

SIMULATION OF SOIL MOISTURE DYNAMICS IN TIME AND SPACE

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ABSTRACT

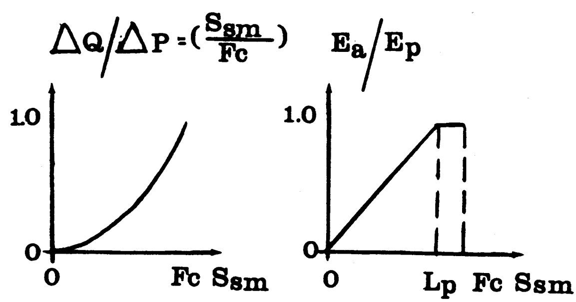
In order to map soil moisture and runoff formation dynamics over time (in a 100-years perspective) and space in Motala Ström River Basin (14 600 km^2), a soil moisture model, based on the soil moisture routine in the conceptual runoff model HBV (SMHI) has been developed. The model calculates daily values of soil water content in the first upper metre of the soil. Water content in the snowpack, actual evapotranspiration and percolation to depths below the first upper metre are also calculated. Attempts have been made to calculate the occurence of ground frost. The need for input data to the model is limited to daily mean values of air temperature and daily precipitation plus monthly mean estimates of potential evapotranspiration. The parameters are optimized with the help of pF-analyses. In this paper, simulations are presented from two sites: Grassland on clay (Källtorp) and coniferous forest on till (Velen).

INTRODUCTION

The high number of dry summers in Sweden since 1969 has caused an increased interest for a systematic study of the occurrence and effects of droughts in normally humid areas. The study described below is part of the project "Water and Regional Development in Motala Ström River Basin". The river basin is situated in the South East of Sweden. The climate in this region is characterized by relatively dry early parts of the summer. The aim of the study is to map variations in soil moisture deficits and in runoff formation in Motala Ström River Basin. The mapping will be done both in time (using up to a 100 years old climatological records) and space (different soil, vegetation and drainage types, climatological data from several climatological stations). This will be followed by studies of the effects of soil water dynamics on crop yield, forest growth and leaching of nitrogen. The historical perspective will probably make it necessary to take human impacts of drainage patterns into consideration. To be able to study soil moisture dynamics for periods and sites for which no soil moisture measurements are available, there is a need for a model that can estimate probable patterns from available data. The use of long time series makes it necessary to limit the need for input data to such information that has been collected for a long time at most meteorological stations. The conceptual runoff model "HBV" (Bergström, 1976) needs very few input data. In this papar, the testing of the soil moisture subroutine in the HBV-model and further development of the soil moisture routine will be discussed.

SOIL MOISTURE SIMULATION WITH THE HBV-MODEL

The conceptual runoff model, HBV (Bergström, 1976) which has been developed to simulate daily runoff values, includes a soil moisture accounting routine, which controls the runoff formation. This routine is based on three parameters: β, L_p and FC (Fig 1). The need for input data to the model is limited to daily values of air temperature and precipitation, plus monthly mean estimates of potential evapotranspiration. Usually, 30-years averages of potential evapotranspiration, calculated by Penmans formula (Penman, 1948) are used. Calder et al (1983) have shown that more sophisticated calculations of potential evapotranspiration do not necessary result in improved soil moisture predictions.



Ssm = computed soil moisture storage
ΔP = contribution from rainfall or snowmelt
ΔQ = contribution to the response function/runoff
Fc = maximum soil moisture storage
β = empirical coefficient
Ea = computed actual evapotranspiration
Ep = potential evapotranspiration
Lp = limit for potential evapotranspiration

Figure 1. The structure of the soil moisture routine in the HBV-model (from Bergström, 1976)

A threshold temperature decides whether precipitation will be considered rain or snow. Snowmelt is calculated by a degree-day method. The model has a simple interception routine. The output from the soil moisture routine in the HBV-model was not tested against measurements before the start of this project. When attempting to make a comparison between measured and calculated soil water contents, one has to face the problem of giving the soil moisture index (the content in a "soil moisture box") a physical interpretation. "Full box" can be defined as the maximal water content that is possible after that the macropores have drained (pF about 0.7) and "empty box" when the water content is at the permanent wilting point (pF about 4.2). Whith these definitions, the box represents the "active water" in the soil.

FIRST DEVELOPMENT STEP - FURTHER DEVELOPMENT OF EVAPOTRANSPIRATION AND PERCOLATION ROUTINES (USING SOIL MOISTURE DATA FROM KÄLLTORP)

At SMHI:s research station in Källtorp, Norrköping, soil moisture measurements have been carried out between 1976-1981 (Milanov, 1982). The measurements were made with neutron probes in three soil tubes. The area has a mixed grass vegetation. The soil profile consists of a clayey top soil down to 0.5m depth. underlain by a heavy post glacial clay. The slope is gentle and the spatial variability of the soil is small. The relatively homogeneous conditions made it possible to use means of the measurements form the three soil profiles to estimate the typical soil water content. The "typical soil water content" is assumed to represent the average soil water deficit (or surplus) for a certain area. The model has been tested against data of the integrated water content of the first upper metre of the soil. Very small changes in the soil water content take place below this depth. Some results from soil analyses are presented in Fig 2. The data on typical soil water content in the upper metre of the soil in the Källtorp area where used for further development of the model and for optimizing the parameters. In order to simulate the soil water dynamics in an acceptable way, the soil moisture routine was extended with the following routines:

* By definition (Penman, 1948), potential evapotranspiration can only occur when the vegetation is fully developed. Before the start of the vegetation period, the evapotranspiration is limited to evaporation from the soil. The soil evaporation is close to the potential when

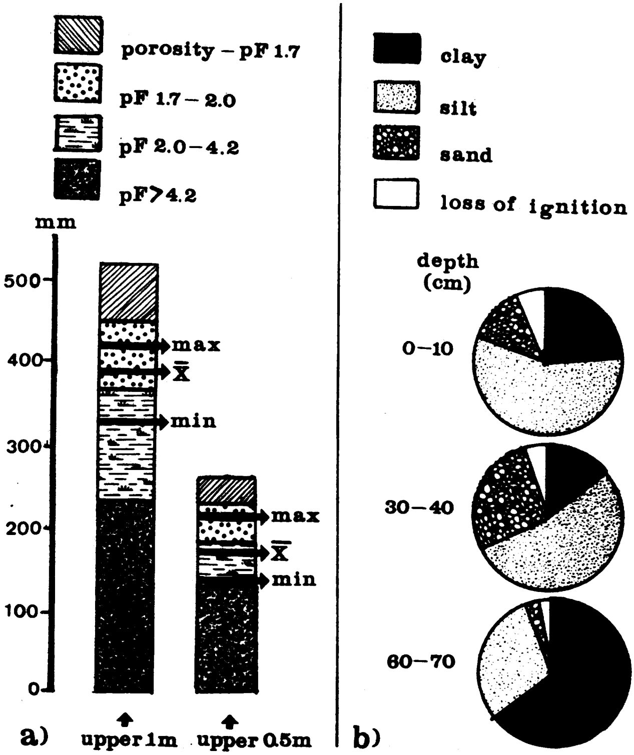


Figure 2. Soil characteristics, Källtorp (L Andersson, unpublished)

a) Porosity and soil water contents at different moisture tensions for the first upper metre and the first upper half metre of the soil profile. The mean, maximum and minimum water contents during the measurement period (1976-1981) are shown.

b) Grain size distribution and loss of ignition at different depths in the soil profile.

the soil surface is wet. However, the rate of evaporation declines to an insignificant value after 5 to 10 mm of water have been removed. It was clear from the soil moisture measurements in Källtorp, that the actual evapotranspiration rates during spring, were smaller than calculated by the original routine. Therefore, a vegetation index that decides the stage of growth was introduced. Averages of five days air temperature were used as a criterion for growth period. The onset of the growth period was considered to be when the average temperature exeeded +5 °C. In the new routine, transpiration is a function of soil water content, potential evapotranspiration and the vegetation index. The soil evaporation is a function of the water content in an "evaporation zone" (Fig 3), the potential evapotranspiration and the inverse of the vegetation index (the soil evaporation is reduced when the vegetation protects the ground from wind and radiation). The "evaporation zone" is emptied by the calculated soil evaporation and by percolation. The percolation out of the zone, is a function of the total water content in the soil.

* At Källtorp, the neutron probe measurements indicated that evapotranspiration decreased faster at reduced water contents in the soil than a linear function of the type used in the HBV-model predicts. This is probably due to the heterogeneous water distribution. Sometimes, the upper part of the soil can be close to the wilting point, when the water content of the sub soil is still close to "field capacity". To simulate a faster decrease in actual transpiration at reduced water contents for shallow rooted vegetation, a parameter "α" was introduced.

$$\text{TRANSPIRATION} = (SM/Fc)^{\alpha} \tag{1}$$

SM = actual water content
Fc = water content at pF 0.7

* When large amounts of water is added to the soil and the initial water content is low, the supplied water is not distributed homogeneously. This can create situations when the evapotranspiration is equal to or close to, the potential, although the integrated water content in the first upper metre of the soil profile is low. To compensate for this, an "upper transpiration zone" was added to the model (Fig 3). It is fed by the amount of rain which is not kept in the soil "evaporation zone". If there is water in the "upper transpiration zone", the transpiration will occur at a rate equal to a total soil water content "TMAX" (parameter). The content in the zone is maximized by:

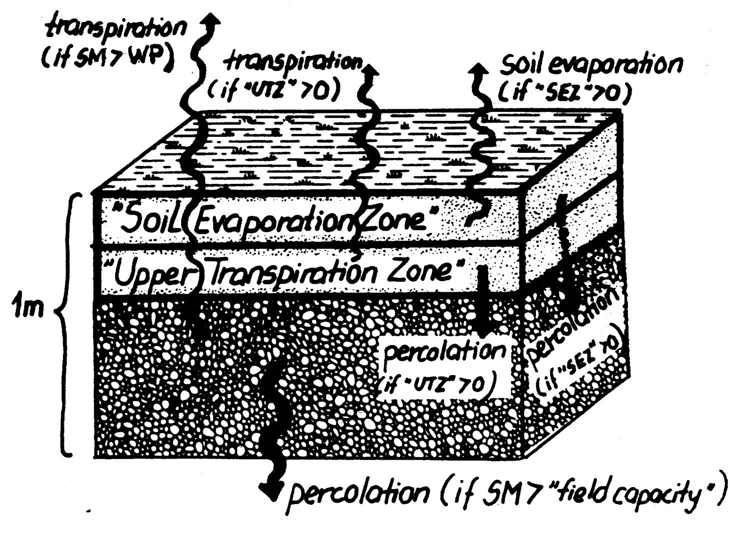
$$UTZmax = TMAX - SM \tag{2}$$

UTZmax = maximal content in the "upper transpiration zone"
TMAX = parameter

The "upper transpiration zone" is emptied by the calculated transpiration and by percolation. The daily percolation from the zone is assumed to be 10% of the content in this zone. If the total water content in the soil becomes lower than it was before the initial water contribution to the "upper transpiration zone", it is considered empty.

* In the original routine, percolation was only calculated during days when water is added to the soil. In reality, however, the percolation will continue as long as there exists drainable water. In a new routine, the percolation is proportional to the content of drainable water (water between pF 0.7 and pF 2.0). The maximum daily percolation is a parameter. It is set by parameter optimizing, but alternatively conductivity measurements could be used.

The new soil moisture routine is described in Fig 3.



WP = permanent wilting point
SM = actual water content
UTZ = "upper transpiration zone"
SEZ = "soil evaporation zone"

Figure 3. The structure of a new, extended soil moisture routine

The series of soil moisture measurements from Källtorp was divided into two parts: Data from 1976-1978 were used for model development and for parameter optimisation. Data from 1979-1981 were used for an independent testing of the model. Simulations are shown in Fig 4.

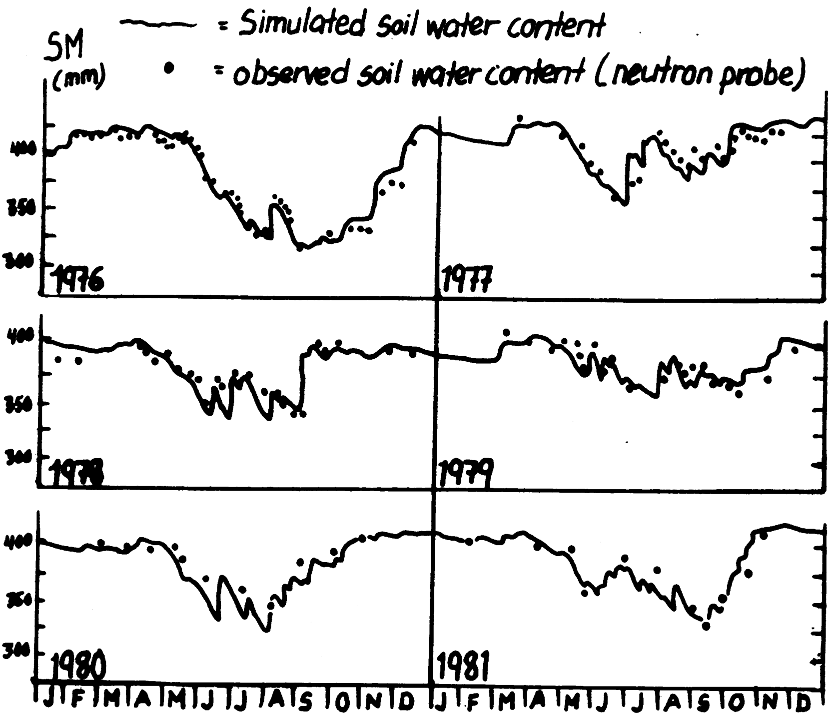


Figure 4. Soil moisture simulation (0-1m) with the extended soil moisture model, Källtorp. Soil water data from 1976 - 1978 were used for parameter optimisation. $R^2 = 0.92$ (1976 - 1981)

It was evident that the model could describe the soil water dynamics also during the independent period. This encouraged the desire to try the model against soil moisture data from other types of areas.

SECOND DEVELOPMENT STEP - TESTING AND FURTHER DEVELOPMENT OF THE MODEL IN A FORESTED TILL AREA (VELEN)

The Velen area was established as a "representative basin" during the International Hydrological Decade. It consists mainly of coniferous forest on till. Soil moisture measurements were carried out at seven soil stations 1968-1973 (Milanov, 1971). The soil moisture data from Velen were used for a study with the aim to find the variations in time of the typical water deficits. The "typical water deficit" is supposed to represent the average soil moisture deficit (or surplus) in the area. This will later be followed by studies of the spatial variability of water deficits. The first step was to estimate how typical the soil stations were for the average conditions. The number of soil stations is too small to asses which can be considered representative for the "average condition". However, the groundwater levels in Velen are relatively shallow (on average 1-2m below the soil surface). Fluctuations in the groundwater table followed the variations in soil water content without considerable time lap. Groundwater measurements were made in 35-40 wells and tubes, including all soil stations. The study showed that if one arranges the stations according to mean groundwater levels, they were in the same order as if arranged according to mean soil water content. Therefore, the classification of a station as "typical" for the average conditions was done from an inspection of the groundwater fluctuations. The stations whose groundwater oscillations always fell within one standard deviation from the average level in the whole area and were the deviations from average conditions were not remarkably changed during the measurement period were considered to be representative for the "typical" soil moisture conditions. Four out of seven stations were classified as "typical" for average conditions. The soil moisture dynamics of the other stations will be discussed in another paper (L Andersson, to be published). That paper will also include a discussion about the variations between the "typical" stations. It was clear that the variations usually increased when the water contents decreased (Fig 6). Soil analyses have been carried out for all soil moisture stations in order to get estimations of the parameters: "Maximum soil water content" (pF = 0.7), "Soil water content when the drainage ceases" (pF = 2.0) and "Wilting point" (pF = 4.2). The analyses were also used to make it possible to compare the soil water contents at the different soil stations in terms of soil water availability for percolation and evapotranspiration. The "typical soil water content" ($\overline{SW}$), was estimated as:

SWi = measured soil water content at a "typical" station

SWFi = soil water content at "field capacity" (pF 2.0) at a "typical" station

For each station the water deficit index (WDI) is calculated:

$$WDIi = SWi/SWFi \tag{3}$$

The "typical soil water content" is:

$$\overline{SW} = \overline{SWF} * \overline{WDI} \tag{4}$$

At the "typical stations", the average soil water contents were close to the contents at pF = 2.0. It was evident that the "typical stations" became more comparative when the water holding properties were taken into consideration (Fig 5)

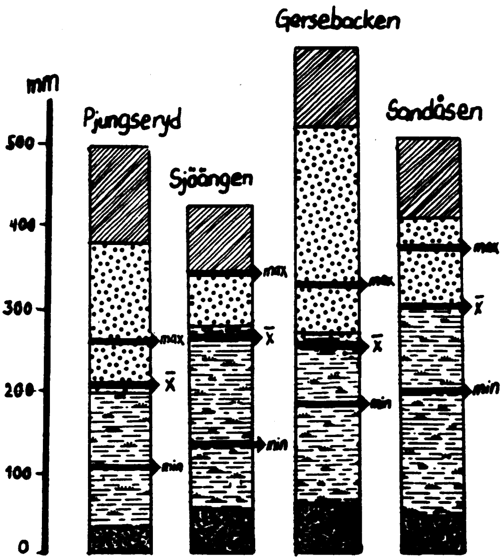


Figure 5

For explanation, see Fig 2

Porosity and soil moisture contents of the first upper metre at the "typical soil stations", Velen. The mean, maximum and minimun water contents during the measurementperiod (1968-1973) are shown.

(L Andersson, unpublished)

The estimated "typical soil moisture contents" were used for further development of the soil water model and for optimizing the parameters. The first attempt used data from 1968-1970 for estimations of parameters and left the rest of the measurement period (1971-1973) as an independent period. However, at the first run for the whole period, water contents were underestimated during spring 1971 and 1972. The winters during these years were relatively cold but with little snow. In Fig 6 simulation from Velen with a simple ground frost index added to the model is shown. The purpose of the index is to reduce percolation rates proportionally to the rate at which the water in the ground is frozen.

DISCUSSION

The results from the soil water modelling in Källtorp and Velen have shown that it is possible to estimate soil moisture dynamics with a conceptual model. The disadvantage with this sort of model is the need for soilwater measurements when estimating the values of the parameters. In this study, attempts have been made to relate some of the parameters to pF-curves. The ambition is to find typical parameter values for the main soil and vegetation types in Motala Ström River Basin. This will make it possible to map dynamics of soil water deficits and runoff formation in the river basin.

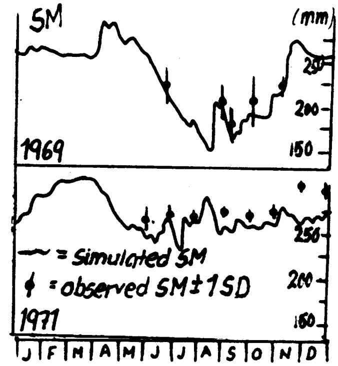


Figure 6. Simulation of "typical soil water content" (0-1m) Velen.

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SKRANINGSHYDROLOGI - AVLØPSSTUDIER I ET LITE NEDBØRFELT.

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Summary

The results from direct measurements and field investigations in a small East-Norwegian catchment with humid climate indicate that a dynamic response area model can be useful in this type of catchment.

The extension of the saturated area seems to be controlled from the different reservoirs of groundwater, soil water and surface water in the basin. This indicates that the saturated area reflects the size of the reservoirs and gives a good estimate of the moisture in the field.
In models we then are able to take into consideration the area-variation of the soil moisture, even though it in basins such as here can vary considerably within few centimeters.

Maps over saturated area at different runoff situations were composed for cases of recession ranging from minimum runoff to about half of maximum registered flood in the basin. In this work a functional relation between saturated area and the runoff was established. It also resulted that one can separate the hydrograms of "pure" saturated overlandflow (produced from rain on the saturated area) from the rest of the runoff at any moment of time.

What is said above also implies that the initial runoff can be used as a moisture indicator for a basin.
The initial moisture in the field, which is indicated by the saturated area (or the runoff), was important in explaining how the basin responded on an occurrence of precipitation. In fields where saturated overlandflow is the dominant drainage process during flood, one therefore can make an empirical model with the aid of field experience and a functional relation between saturated area and runoff. One may then among other things predict how the runoff will be after different occurrences of precipitation.

Accordingly, it seems important to take the dynamic response area into consideration both in further field investigations and in hydrological modelling.

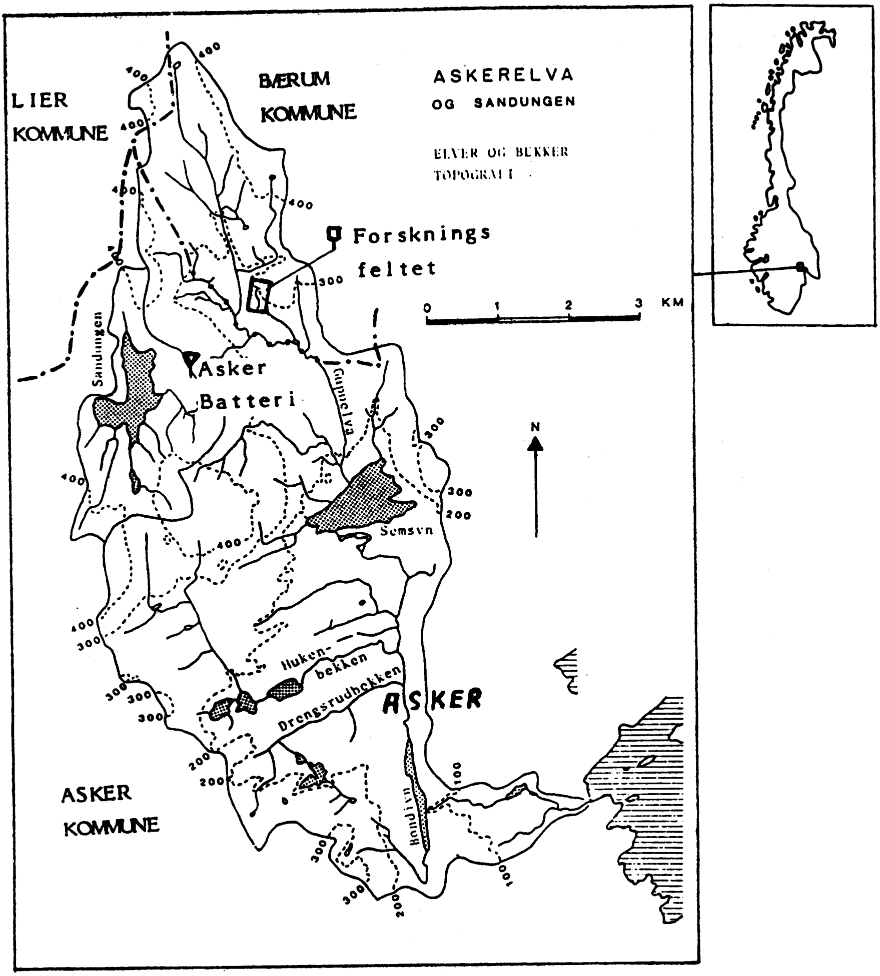

Fig. 1 Topografisk kart over Askerelvas nedbørfelt med beliggenhet av "minifelt".
Bearbeidet fra Wingård m.fl. (1981).

1 Innledning

Innlegget er en del av en hovedfagsoppgave i hydrologi. Feltarbeidet i forbindelse med denne oppgaven ble utført i øvre del av Askerelva (fig. 1) fra høsten 1981 til og med høsten 1983. For nærmere beskrivelse av de enkelte kapitlene og litteraturhenvisning, se Myrabø (1985).

1.1 Problemstilling

Et av formålene med prosjektet var å undersøke de prosessene som genererer nedbør til avløp i bekker og elver og dermed styrer vannets veier gjennom avrenningsområdet, avrenningens størrelse og vannets oppholdstid i ulike horisonter. En skulle også se på hvilke avrenningsformer som eksisterer i et lite nedbørfelt. Et viktig spørsmål var dessuten hvordan en i modeller skal ta hensyn til arealvariasjonen i markfuktighet. Dette er av stor betydning f.eks. for utvikling av matematiske avløpsmodeller og for en forståelse av de hydrokjemiske prosessene. Problemstillingen er derfor fundamental for en rekke ulike områder, som innen hydrologi, jordbruk, skogbruk, miljøvern og vannkvalitet.

En rekke avløpsmodeller er blitt utviklet gjennom tidene (se under historikk). Her har en særlig sett på respons-areal modellen. En viktig oppgave var å måle mettede områder i feltet i forhold til vannføringen, da dette også kan si en noe om arealvariasjonen av markfuktigheten i feltet.

1.2 Historikk

Omkring 1933 skapte den amerikanske ingeniøren Rober A. Horton den tradisjonelle forståelse av avløpsprosessene (se fig. 2).

Selv om begrepet lateral bevegelse av vann i markvannssonen var fraværende i Hortons modell, så er dets opprinnelse nesten like gammelt. Først i 1934, deretter mellom 1936 og 1944 kom det viktige bidrag til studier av "subsurface stormflow". Etter hvert ble det utviklet "throughflow"/ "interflow"/sigevanns- modeller (se fig. 3).

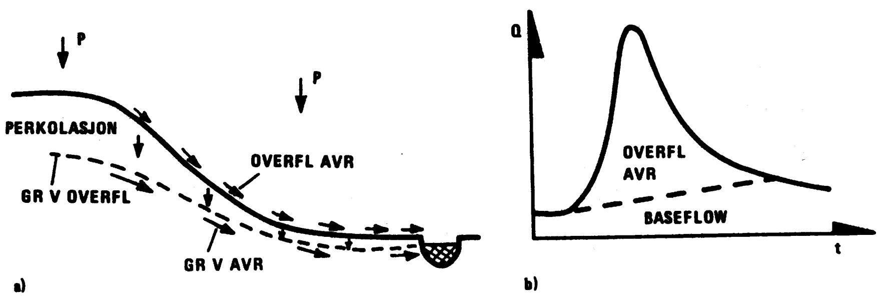

Fig. 2 Horton avløpsmodell.
a) Skråningsprofil med avrenningsformer.
b) Typisk hydrogram med mulig separasjonslinje (——)

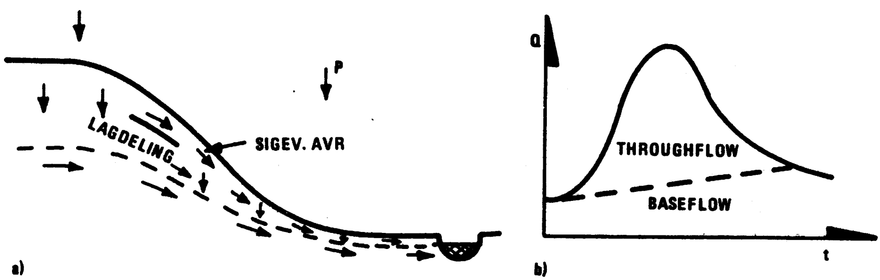

Fig. 3 Sigevanns modellen. a) og b) som i fig. 2.

Etter hvert som undersøkelsene av sigevann ble intensivert de siste 20 åra, oppstod også begrepet "partial area". Begrepet "partial contributing area"/"dynamic source area"/ respons-areal ble utviklet tidlig på 1960-tallet i USA (se fig. 4).

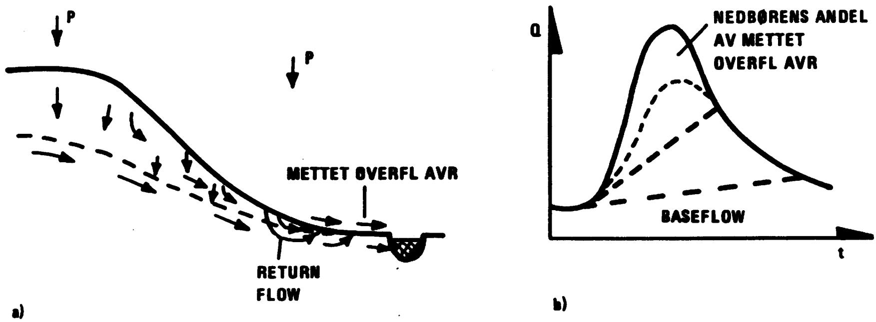

Fig. 4 Respons areal modellen. a) som i fig. 2.
b) Typisk hydrogram med mulige separasjonslinjer.

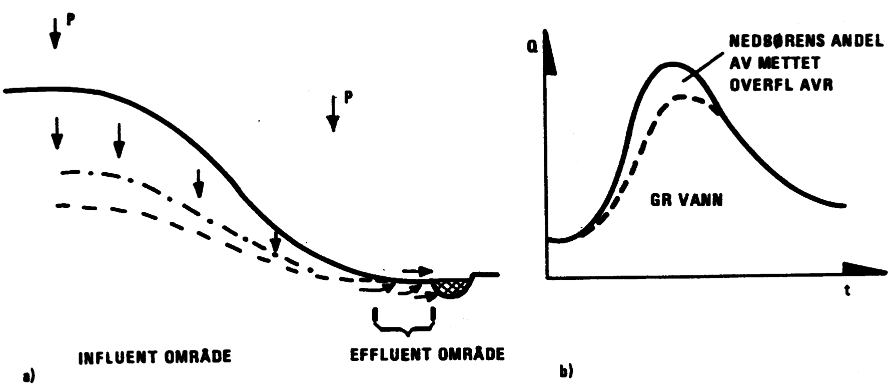

Fig. 5 Stempelstrøm modellen. a) og b) som i fig. 2.
—·—Hevningen av gr.v. speilet i løpet av et regnvær.

Det har i andre modeller blitt tatt hensyn til andre avrenningsformer som "piping" i f.eks. rot- og dyrekanaler, eller utpressing av gammelt regnvann ved hjelp av stempel-effekt fra det nye regnfallet (se fig. 5). Videre har andre forsøkt å dele nedbørfeltet opp i homogene områder, for å lage en modell for hvert typeområde.

I de fleste studier av avløpsdrenering i fuktige områder som har blitt publisert de siste åra er det i følge Kirkby (1980) anvendt en eller annen type av respons-areal modell. Den dynamiske rollen som denne type modeller spiller i skrånings/felt-hydrologi under spesielle jords og skråningsforhold, former derfor nå basis for spekteret av modeller som ligger mellom ekstremene representert ved Hortons modell og den rene sigevanns-modellen.

1.3 Feltbeskrivelse

Forskningsfeltet ligger i Bærum kommune (fig. 1). Arealet er på 0,085 km² og er en del av Askerelvas nedbørfelt på totalt 37,2 km² ,et typisk øst-norsk skogsvassdrag. Tilnærmet hele "minifeltet" er dekket med skog, mest barskog, men også noe lauvskog. Feltet inneholder en del myrer, og er for det meste dekket av løsmasser med innslag av bart fjell spredt rundt omkring. Bergartene i området er permisk rombeporfyr av Kolsås-type. Løsmaterialet består for det meste av et tynt usammenhengende morenedekke og en del glasifluvialt materiale, særlig i myrområdene og langs nedre del av bekken.

Klimaet er sterkt variabelt. Selv i vinterhalvåret vil en betydelig del av nedbøren falle som regn. Laveste og høyeste temperatur i måleperioden ble registrert til h.h.v. −19 og 31 °C. Bekken som drenerer feltet varierer fra å være helt uttørket midt på sommeren til å ha en flomvannføring i utløpet av feltet på over 800 l/s·km². Andre karakteristiske data for feltet sees av tabell 1.

maksimum høyde :	329 m.o.h.
minimum høyde :	285 m.o.h.
maksimal høydeforskjell (H) :	44 m
median høyde :	307 m.o.h.
areal :	0,085 km²
maksimal lengde (L) :	500 m
maksimal bredde :	180 m
relieff forhold (H/L) :	9/100

Tab. 1 Karakteristike felt data.

2 Målemetoder

For å teste de forannevnte modellene, ble det for utskillelse av ulike avrenningsformer bl.a. benyttet hydrogramseparasjon ved hjelp av spesifikk ledningsevne og kartlegging av mettet areal i forhold til vannføringen. En foretok også direkte målinger av de ulike prosessene nedbør, snøsmelting, fordampning og avløp, som alle er medvirkende variable i vannomsetningen i et nedbørfelt.
For nærmere beskrivelse av målemetodene, se Myrabø (1985a).

3 Presentasjon og diskusjon av måleresultatene

Utstrekningen av det mettede arealet varierte med tiden. Stikkprøver viste også at fuktigheten i feltet kunne være svært forskjellig innenfor bare få centimeter.

F.eks. kunne markoverflata være nesten helt tørr ett sted, mens den like ved siden av var mettet. I et så innhomogent felt med store fuktighetsvariasjoner over små avstander nytter det ikke å få en oversikt over arealvariasjonen i markfuktigheten uten å forbinde den med en lettere målbar parameter. Nedenfor antydes at mettet areal og dets dynamiske karakter gir et estimat for denne arealvariasjonen.

Det synes som det bare er på bestemte steder at vannet samles og at det derfra renner bort under eller på overflata. I følge Dunne, Moore og Taylor (1975) oppstår mettet areal der transportmekanismer under jordoverflata ikke klarer å transportere bort vannet. Den akkumulerte vannmengden som lagres i jorda vil etter hvert heve "vannspeilet" til jordoverflata. I følge Dunne og Black (1970) kan soner av metning eksistere helt fra fjellgrunn eller de kan bygges opp over et relativt tett lag i jorda.

De områdene som er disponert for å produsere mettet overflateavrenning er derfor bestemt av topografi og jordkarakteristikker, samt de lokale fuktighetsforhold. Størst sansynlighet er det at de mettede arealene dannes i nederste del av feltet med et betydelig dreneringsareal ovenfor, som kan forsyne de nedenforliggende områdene.

Det mettede arealet varierte både over sesongen, i løpet av og i mellom nedbørtilfeller. Det viste videre store variasjoner ved forskjellige vannføringer. For alle tilfeller synes utbredelsen å samvariere godt med vannføringen. Under flom ekspanderte de mettede områdene utover til sidene og oppover skråninger til en utstrekning hovedsaklig bestemt av nedbørtilfellet (intensitet og varighet) og jordkarakteristikkene, samt initialfuktigheten. I tillegg oppstod også nye mettede områder. Gjentatte registreringer viste at det mettede arealet var tilnærmet likt ved samme vannføring for resesjonstilfeller. Det ble derfor utarbeidet kart over arealet fra minimum vannføring opp til ca. halvparten av maksimum målt flom i feltet.

Målingene viser at det er en nær funksjonssammenheng mellom vannføringen og utbredelsen av det mettede arealet. Dette gjelder antagelig tilnærmet for alle årstider, så lenge bekken ikke blir tørr. Etter en tørkeperiode i feltet må imidlertid de magasinene som opprettholder det mettede arealet mellom nedbørtilfellene fylles opp. Ved nedbør skjer det da først en oppfylling av magasinene nederst i feltet, p.g.a. at disse områdene i tillegg til nedbøren også får tilførsel av vann fra høyereliggende områder.

I tørre perioder er det sigevann og grunnvann som forsyner de mettede arealene og bekken med vann, slik at mellom nedbørtilfellene så minker de mettede arealene i takt med de forskjellige magasinene i feltet. Utbredelsen av det mettede arealet ser derfor ut til å styres av de forskjellige grunnvanns-, markvanns- og overflatemagasinene i feltet. Med overflatemagasin menes her de mettede areal som bidrar med vann til mettede areal lenger ned i feltet. Dette indikerer at mettet areal avspeiler magasinenes størrelse og gir et godt estimat av fuktigheten i feltet. Hvis det mettede arealet er godt kartlagt i forhold til vannføringen tyder dette også på at vannføringen kan brukes som fuktighetsindikator for et felt.

Funksjonssammenhengen mellom mettet areal og vannføringen (Q) er sansynligvis komplisert. Det var derfor vanskelig å finne et analytisk uttrykk for å beskrive sammenhengen.

Den beste ligningen som ble funnet ved kurvetilpasning (minste kvadraters metode) var :

$$A = -1,95 + 3,48 \cdot Q^{0,3} \quad \text{for} \quad Q > 0,145 \text{ l/s}$$

hvor A er andel mettet areal i forhold til antatt maksimalt mettet areal (maksimalt mettet areal utgjør i dette tilfellet ca. 50% av hele feltarealet). Av figur 6 ser en at ligningen beskriver godt sammenhengen mellom vannføringen og mettet areal i området fra små vannføringer til normale flomforhold i feltet. Avvikene mellom beregnede og målte verdier er mindre enn usikkerheten ved selve målemetoden.

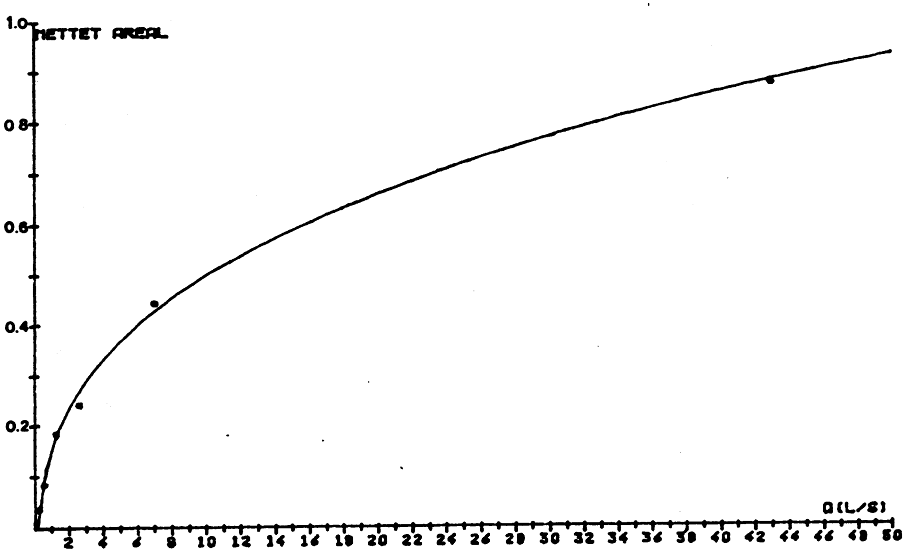

Fig. 6 Kurve for funksjonssammenhengen mellom mettet areal (angitt i andel av maksimalt mettet areal) og vannføringen (Q angitt i l/s).
■ : de målte verdiene.

Ligningen er altså god nok til å beregne mettet areal til enhver tid under og etter et nedbørtilfelle på grunnlag av vannføringen. En kan videre beregne nedbøren som faller ned på det mettede arealet og som i følge Dunne og Black (1970) antas å gå direkte til overflateavrenning. Vannføringen i starten av hvert nedbørtilfelle blir da brukt til å estimere initielt mettet areal. Da det mettede arealets dynamiske karakter må ansees å være svært viktig, må en korrigere for at det mettede arealet varierer med vannføringen i løpet av flommen. For å estimere størrelsen av det mettede arealet ved et gitt tidspunkt setter en derfor vannføringsverdien for tidspunktet inn i ligningen. For samme tidspunkt finner en netto nedbør som faller på det mettede arealet ved å korrigere for magasinering og fordampningstap. For hvert tidspunkt kan deretter den utledete størrelsen for mettet areal (gitt i andeler av feltarealet) multipliseres med netto nedbørmengde. Produktet gir et estimat av den hurtige flomavrenningen produsert som "ren" overflateavrenning (ikke medregnet "return flow"). En må også ta i betraktning en forsinkelsestid, bl.a. p.g.a. magasineringseffekter. Nedbøren fordeles derfor over et visst tidsrom før den transformeres til utløpet av feltet. Forsinkelsestiden i feltet er her ca. en time. Regnet må derfor fordeles slik at det største bidraget kommer omkring en time etter at regnet har falt. Hvis dette taes i betraktning kan en for hvert tidspunkt trekke de estimerte verdiene for ren overflateavrenning fra den målte vannføringen i feltet. Denne metoden for hydrogramseparasjon skiller således ren overflateavrenning fra resten av avløpet.

Eksempler på en enkel bruk av denne metoden sees i figur 7. Her er forsinkelsestiden av nedbøren på det mettede arealet fordelt med en antatt prosentverdi innenfor et tidsrom fra en til femten timer. Hvis en her for hvert tidspunkt beregner netto nedbør på de mettede områdene og korrigerer for tap ved intersepsjonsmagasinering, magasinering på det mettede arealet og fordampning i perioden,

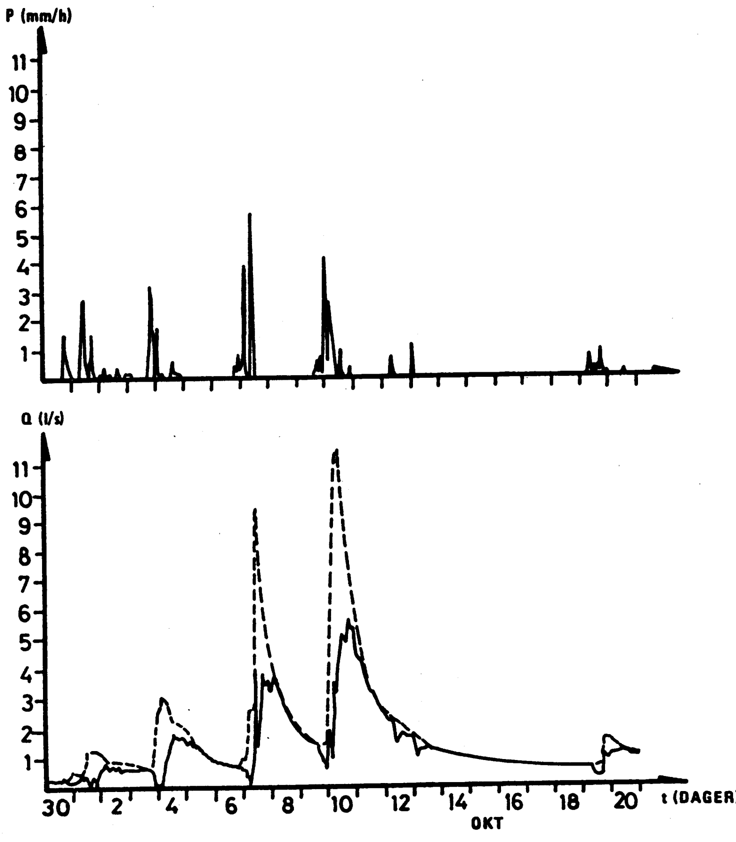

Fig. 7 Forsøk på å skille ut "ren" overflateavrenning (arealet mellom (——) og (– –)) ved hjelp av ligningen, nedbøren og avløpsdata høsten 1983.
Øverst (——) vises nedbøren.

vil en få jevnere kurver og et bedre resultat. Enda bedre blir resultatet om en også fordeler netto regnvann bedre innenfor tidsrommet til det når måledammen som avløp.

Det eksisterer selvfølgelig en del usikkerheter ved målingene og beregningene omtalt i dette kapittelet. Trolig av størst betydning er den valgte definisjonen for mettet areal (vannspeilet ned til 5 cm under markoverflata). Dette fører til at størrelsen på det mettede arealet som estimeres å bidra direkte til den rene overflateavrenningen er en del for stort.

En bør også tilføye at det sansynligvis eksisterer hysterese-virkninger av mettet areal i forhold til vannføringen. Alle målingene av mettet areal i forbindelse med undersøkelsen foregikk imidlertid ved resesjonstilfeller, da simultane målinger ved stigende vannføring var umulig for et så stort område. Både vannføringen og det mettede arealet ville ha gjennomgått store forandringer før en rakk å gå gjennom hele feltet. Dette fører kanskje til at ligningen ikke er representativ ved stigende vannføring, og at en da muligens burde legge inn et tidsforsinkelsesledd på f.eks. ett eller to tidsskritt. Resultatet i figur 7 ville da sansynligvis forbedre seg ytterligere.

Samme område kan altså gi uli respons på en viss nedbørmengde alt etter de rådende fuktighetsforhold. Initial-fuktigheten i feltet , som her indikeres av mettet areal eller vannføringen, er derfor viktig for hvordan feltet vil reagere på et nedbørtilfelle. Hvis en som Taylor (1982) konkluderer med at mettet overflateavrenning er den dominer-ende avløpsprosessen under flom, så er dermed initialvannføringen sammen med ligningen avgjørende f.eks. for hvor stor flommen skal bli, andel av regnet involvert i flommen og hvordan hydrogrammet skal se ut. Hvis samme type nedbørtilfelle gir likt avløp når initialvannføringen er den samme og ellers like tilstander i feltet (f.eks. fordampning og årstid), så åpner dette nye muligheter. Ved å lage en empirisk modell ved hjelp av erfaringer i felt, kan en f.eks. forutsi hvordan avløpet blir etter forskjellige nedbørtilfeller og hvor mye nedbør som må til (over en viss tid) ved forskjellige vannføringer for å produsere et bestemt flomavløp.

4 Konklusjon og videre arbeid

Resultatene av undersøkelsene understøtter teorien om at det i denne type skogsfelt kan anvendes en respons areal modell, og mettet areal antydes å gi et bra estimat av fuktigheten og dens arealvariasjon i feltet.
Horton overflateavrenning ble f.eks. aldri observert i felt. En så bare overflateavrenning fra de registrerte mettede områdene. Stempelstrøm og/eller sigevannsavrenning er trolig heller ikke dominerende i flomavrenningen, p.g.a. den raske responsen og relativt stor forskjell i forsinkelsestid for snøfritt felt i forhold til under snøsmeltingen.

Mulige prosesser som genererer nedbør til avløp er bl.a. mettet overflateavrenning, som består av ren overflateavrenning og "return flow", stempelstrøm, "pipeflow", og sigevanns- og grunnvannsavrenning direkte i bekken. Den relative betydning av de forskjellige avløpsprosessene varierer fra felt til felt, med topografi og med jordkarakteristikker. Innen et bestemt felt avhenger det inbyrdes forhold også av nedbørtilfellet (intensitet og varighet) og initialfuktigheten i feltet.

Når en her konkluderer at variasjonen i flomavløpet fra flom til flom i hovedsak kontrolleres av nedbørtilfellet og mettet areal (korrigert for magasinerings- og fordampningstap) så er det viktig at en i denne type felt får en oversikt over mettede områder.
Feltkartlegging er i følge Dunne, Moore og Taylor (1975) den beste metoden for å lokalisere og vurdere størrelse og variasjon av mettet areal. Kartlegging i felt er imidlertid tungvindt i lengden, og kan bare skje leilighetsvis. For å beskrive utbredelsen av det mettede arealet som funksjon av tida er det derfor nødvendig med en hydrologisk teknikk som involverer en eller flere parametere som kan måles rutinemessig. Feltkartlegging må derfor korreleres med andre lett målbare feltkarakteristika. Her antydes at vannføringen er en god indikator innen et enkelt felt.

I dette arbeidet har en påvist et funksjonsforhold mellom mettet areal og vannføringen. Det fører også til at en ved hydrogramseparasjon kan skille ren overflateavrenning (produsert av netto regn på det mettede arealet) fra resten av avløpet.

Det mettede arealet varierte i størrelse med tida. Initialfuktigheten i feltet som indikeres av mettet areal var viktig for hvordan feltet reagerte på et nedbørtilfelle. I felt der mettet overflateavrenning er den dominerende avløpsprosessen under flom, kan en derfor lage en empirisk modell ved hjelp av felterfaringer og et funksjonsforhold mellom mettet areal og vannføringen. En kan da bl.a. forutsi hvordan avløpet blir etter forskjellige nedbørtilfeller.

Det mettede arealets dynamiske karakter er i alle tilfelle viktig i denne type felt som her er omtalt, og må taes hensyn til i fremtidige modeller.

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RELATION BETWEEN CHEMISTRY OF SOIL SOLUTION AND RUNOFF IN TWO CONTRASTING WATERSHEDS: LANGE BRAMKE (WEST GERMANY) AND RISDALSHEIA (NORWAY)

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1 Introduction

The regional pattern of surface water acidification can be largely explained by the combination of acid deposition and sensitivity factors (WRIGHT, 1983). Site sensitivity towards acid deposition is highly dependent on the pathway of seepage water through the terrestrial ecosystem. Thus the influence of geology is linked to catchment hydrology. In this paper I examine the influence of hydrology on the acid-base status of runoff water by a comparison of two contrasting watersheds. Lange Bramke (West Germany) is situated in the Harz mountains at 600 m and Risdalsheia (Norway) lies in southern Norway at about 300 m above sea level.

At both catchments the pathways of water through the soils are known for different runoff conditions. The chemistry of soil solution extracted by tension lysimeters along these pathways will be compared to runoff chemistry.

2 The sites

The soils at Lange Bramke are relatively old and deeply weathered whereas soil development on the mountain ridge at Risdalsheia started after deglaciation about 10 000 years ago. Loss of water to deep seepage is minimal at both sites. At Lange Bramke a layer of coarse stoney material rests on top of the bedrock. Thus saturated hydraulic conductivity increases with depth towards the bedrock (HAUHS, 1985a). Table 1 shows a comparison of the two sites. At EGIL catchment (Risdalsheia) runoff stems from a thin, patchy and poorly developed soil (LOTSE and OTABBONG, 1985). Figure 1 gives a typical example of the runoff response at Lange Bramke.

	Lange Bramke	EGIL catchment
Precipitation (mm)	1350	1313
Size (ha)	76	0.04
Relief (m)	160	2.5
Soil cover (%)	100	50
Soil depth in the flow region (m)	2.5	0.25
saturated conduc. (m/sec)	> $4*10^{-4}$	> $1*10^{-2}$
Bedrock	Sandstone	Granite
Vegetation	Norway spruce (30-38 yrs old)	Heather, Grasses Pine, Birch
Sulfur Depos. (Kg S/ha/yr)	43	23
pH runoff	6.2	4.0

Table 1: Comparison of Lange Bramke and EGIL-catchment (Risdalsheia)

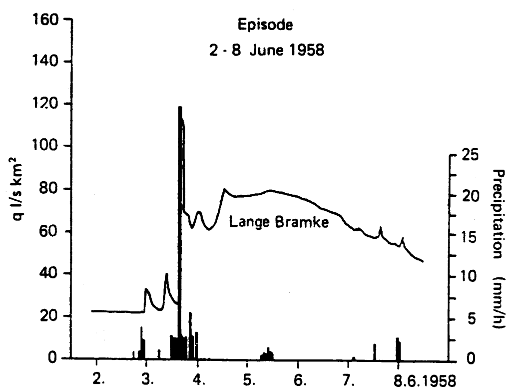


Figure 1: A stormflow episode at Lange Bramke shortly after afforestation (after Balazs et al.1974)

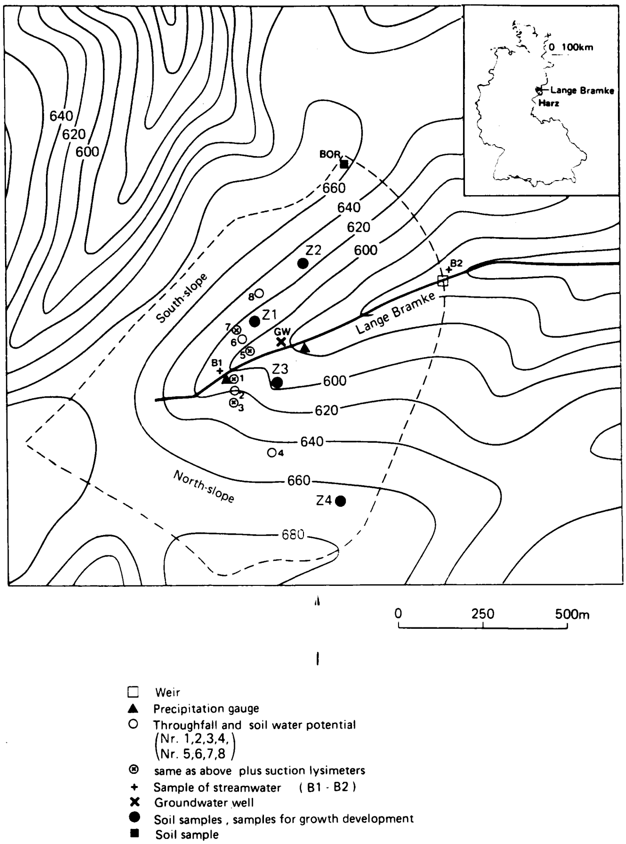


Figure 2: Lange Bramke

3 Methods

At both sites pF-curves and the unsaturated hydraulic conductivity were determined in undisturbed soil samples. At Lange Bramke precipitation, throughfall and runoff are recorded continously. Soil water potential down to 105 cm was measured from 1977 until 1980 at eight sites along a cross section through the watershed. After 1980 these measurements were reduced to two representative sites.

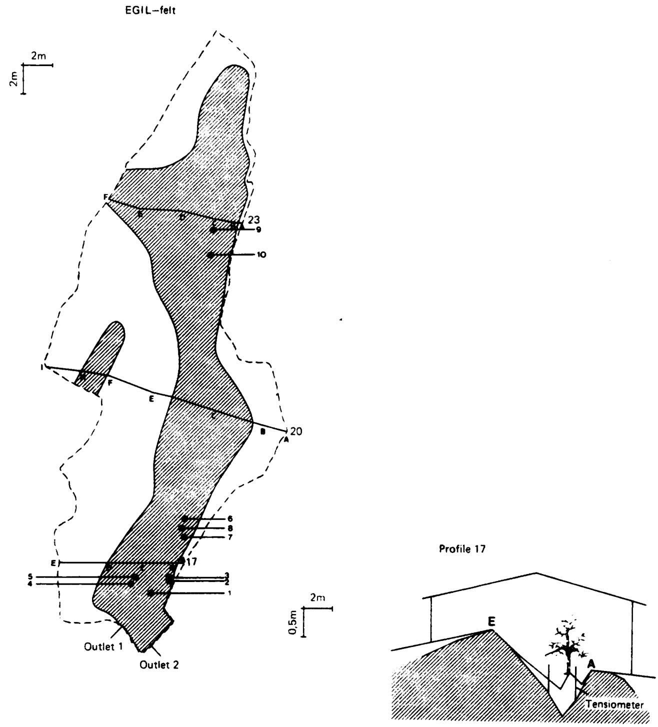


Figure 3: EGIL catchment, parts covered by soil are hatched. Numbers indicate positions of tensiometers. The cross section 17 is shown in the small sktech (vertical exageration 4x).

Soil solution is extracted twice a month by 24 suction lysimeters from a depth of 80 cm. These lysimeters operate at a suction of 150 cm. Groundwater is sampled at a well in the middle of the catchment (figure 2). Samples for the chemical composition of precipitation, throughfall and streamwater (at the headwater and at the weir) are taken weekly. A detailed description of the methods is given by HAUHS (1985a,b).

The EGIL catchment at Risdalsheia is covered by a roof. This catchment is part of the RAIN project (Reversing Acidification In Norway) and serves as a control for the treatment at the adjacent KIM catchment (WRIGHT,1985). Rain is collected on the roof and recycled by a sprinkler system underneath the roof. The experimental setup gives control over the rate and volume of rainfall and runoff. Samples from runoff are taken automatically at each 0.94 mm of outflow. In the soil of the EGIL catchment 10 pressure transducer tensiometers were installed in three groups (figure 3). They provide a continuous measurement of soil water potential. Within the group of tensiometers 1-5 suction lysimeters were installed at four depths below the groundwater table (figure 4). Soil solution was extracted at a suction of 300 cm during a runoff event on 5 November 1985.

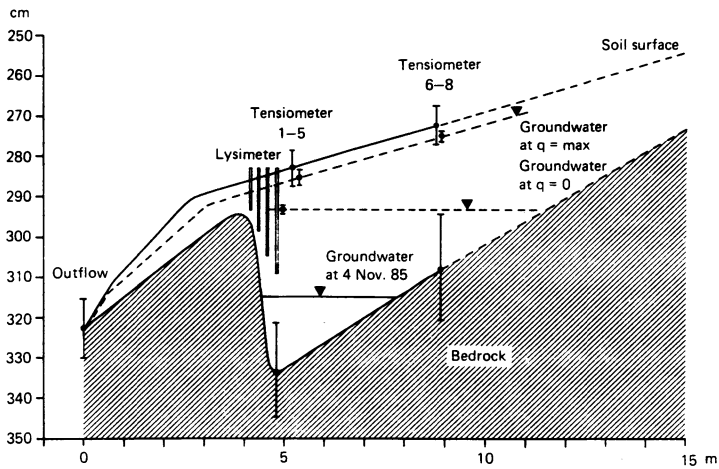


Figure 4: Cross section through the lower part of the EGIL catchment (vertical exageration 10x) and the position of the suction lysimeters.

4 The Model

The hydrology at Lange Bramke was evaluated with a physical/mathematical model of transient saturated/unsaturated water flow through the cross section of the catchment. The model is described by HAUHS (1985a,b). It calculates daily amounts of runoff and soil water potential from daily infiltration sums and daily unstressed transpiration. A comparison with the measured runoff and soil water potential for the years 1979-81 showed that water movement within the first meter of the soil at the slopes of the watershed followed a vertical direction (HAUHS,1985a).

The highest runoff in the last 25 years at Lange Bramke occured during snowmelt in March 1981. The measurements of soil water potential during this episode showed that the soil stayed unsaturated but approached a suction of about 10 cm at all depths. The model calculated the streamlines for this event. Figure 5 shows the result for the situation at peakflow in the cross section of the south-facing slope.

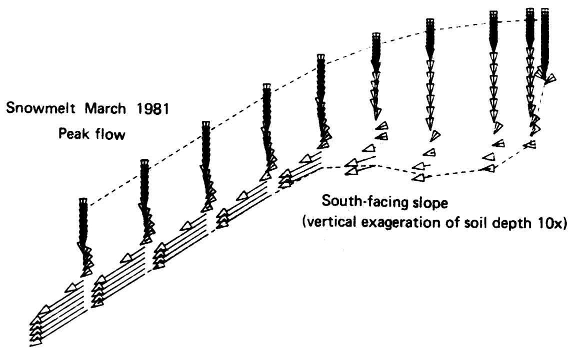


Figure 5: Flow field in the cross section of the south slope of Lange Bramke.

5 Results - Lange Bramke

This section gives a summary of the results for Lange Bramke (HAUHS, 1985a).

- On the investigated cross section saturation was not observed within the first meter of soil. The seepage of water across the depth at which the lysimeters were installed was vertical for all episodes in the period from 1979-81. This includes the exceptional highflow in March 1981.
- Streamwater at B1 (figure 2) stems from a small subcatchment where deeper seepage is prevented. Most of the runoff at the weir comes from the groundwater in the bottom of the valley or the transient groundwater that builds up in the deeper soil on the slopes during highflow.
- Runoff chemistry is independent of flow (exception: HCO_3^+, H^+).
- Runoff chemistry responded to the flow conditions only during the highflow in March 1981 (table 2). Streamflow chemistry at peakflow corresponded to a mixture of groundwater with soil solution from

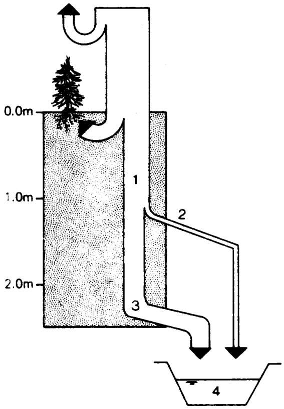
the upper layers of soil (figure 6,table 2). About 25% of the strong acid anions were not accompanied by base cations at peakflow. Therefore pH was as low as 4.3 compared to the volume weighted average of 6.0 (excluding this event). An evaluation of O_{18} in the same event suggests that 25 % of runoff was event water (HERRMANN pers. comn.).

- The sketch of hydrology at Lange Bramke that is presented in figure 6 is confirmed by the chemistry along these pathways. Headwater corresponds to the soil solution chemistry at 80 cm depth, and runoff is similar to the groundwater from the well in middle of the catchment.

(µeq/l)	Runoff peak 12 March 1981	Volume weighted avg.⁺ for 1981
H	53	1
(pH)	(4.28)	(6.04)
Na	70	75
K	26	21
Ca	170	190
Mg	125	154
Mn	11	< 2
Al*	67	< 2
Cl	96	90
SO_4	320	257
NO_3	114	45

⁺ Runoff from 12 March 1981 excluded.
* All Al regarded as Al^{3+}

Table 2: Runoff chemistry at the weir (B2) of Lange Bramke.



	Concentration ($\mu eq/l$)			
	1 = Soil solution	2 = Headwater	3 = Groundwater	4 = Runoff
H	40	34	2	1
(pH)	(4.40)	(4.47)	(5.70)	(6.12)
Na	77	65	73	72
K	11	17	25	16
Ca	47	105	179	176
Mg	32	73	159	140
Mn	15	12	2	2
Al	244	111	<1	<1
SO_4	347	271	215	228
NO_3	42	65	41	46
Cl	90	89	88	87

Figure 6: A Model of ion transport through the soils at Lange Bramke (Means for period 1 Aug. 1983 - 30 Sep. 1984).

6 Results - EGIL catchment

- Despite a high saturated hydraulic conductivity (on the order of 10^{-2} m/sec) all measurements of soil water potential show that the water flux follows a well-defined gradient. There are no signs of flow components bypassing the soil matrix.
- At highflow streamlines converge to a 5-cm interval about three meters upstream from the outlet (figure 4).
- The absolute change in water content between both the onset of runoff and the end of runoff and the maximum runoff (50 mm/day) is about 5 mm.
- Runoff chemistry at peakflow on 5 November 1985 was similar to precipitation chemistry from the preceeding day (figure 7, table 3). Compared to the composition of rainfall the runoff was enriched in Na and Cl, probably due to the washout of seasalts from dry deposition. A 27-day dry period preceeded the event shown in figure 7.
- Before the onset of rain all suction lysimeters were in the unsaturated zone (figure 4). Soil solution collected by these lysimeters during 5 November 1985 showed a chemical composition that is completly different form rainfall or runoff (table 3). Nitrate in soil solution was by a factor of 50 lower than the peak in runoff and precipitation (figure 7, table 3).

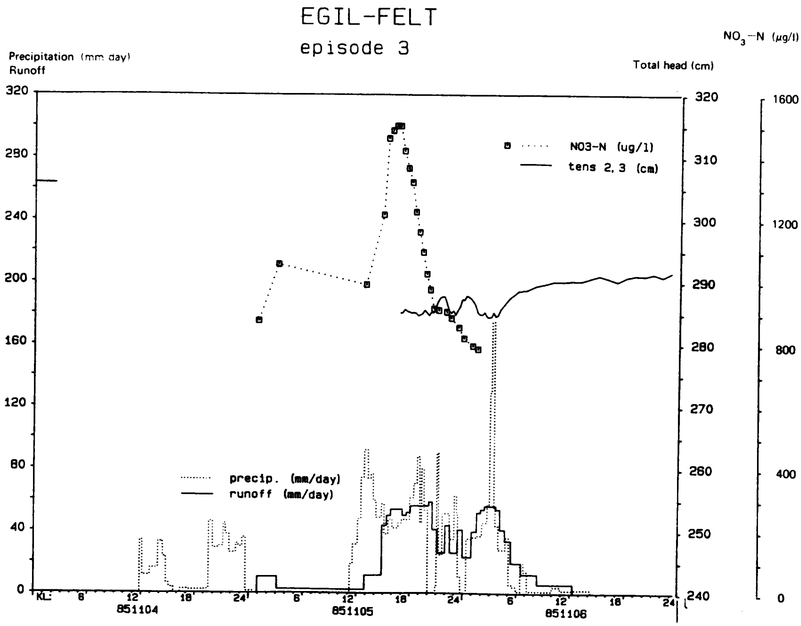


Figure 7: Precipitation, runoff, soil water potential and nitrate-concentration in runoff during the episode from 4 Nov. - 6 Nov. 1985. The line for the tensiometers gives the position of the groundwater table on the same scale as shown in figure 4.

7 Discussion

At Lange Bramke seepage in the top soil occurs at a suction of 20 - 50 cm. Despite an extreme pore-size distribution (HAUHS, 1985a) the seepage water quality can be determined from soil solution collected in suction lysimeters. At exceptional highflow events, however, water movement occurs partly at pore velocities that are probably to high to sustain chemical equilibrium with the rest of the soil water. This water passes through the soil without interaction with "old" water.

	Precip. 4 Nov.85	Runoff peak conc.	Precip. 5 Nov.85	Runoff 6 Nov.85	Soil Solut. Lys. 1-4
(mm)	8.6	5.7-6.6⁺	36.0	21.5-22.4⁺	-
(µeq/l)					
H	158	141	51	93	35
(pH)	(3.80)	(3.85)	(4.29)	(4.03)	(4.45)
Na	30	148	8	92	-
K	13	21	4	14	-
Ca	2	34	1	18	-
Mg	69	59	17	27	-
Al**	-	32	-	13	-
NH_4	63	67	13	41	-
NO_3	116	107	25	57	2 (0.9)*
Cl	31	251	9	129	117 (41)
SO_4	150	169	50	131	54 (12)

⁺ the sample was taken from this runoff-interval since the onset of runoff
* st. deviation
** labile Al

Table 3: Chemical composition of rainfall, soil solution and runoff during the episode shown in figure 7.

This case which is the exception at Lange Bramke is the norm at a site such as EGIL catchment. Under such conditions suction lysimeters are no longer suited to collect water that is moving through the soil. Event water infiltrates the soil but has only limited access to the bulk buffering capacity of this soil. Only that part of the soil that is in contact with the pore fraction in which the water moves under saturated conditions is able to provide base cations to buffer the acid event water.

Under highflow conditions soils such as at Risdalsheia are extremely sensitive to water acidification. The deeply weathered, permeable soils at Lange Bramke provide access to much larger buffer capacity for all flow conditions. Therefore catchments such as Lange Bramke are able to maintain a pH in runoff above 6.0 despite the fact that acid deposition in this area is at least twice as high as in southern Norway.

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Introduction

Numerical models provide possibilities to solve the physical equations for soil water flow with realistic boundary conditions and soil properties. The solutions can be made in one- two- or three dimensions but most models are restricted to one dimension. The models are commonly applied to a areal scale substanially larger than the typical scale for determination of physical soil properties. Furthermore the required soil properties are seldom available for more than a few profiles. The poor information on spatial variability of soil physical properties sometimes makes the applicability of numerical models uncertain especially if areal mean moisture conditions are requested or if the actual variability within a field is of interest.

The purpose with this paper is to demonstrate how simulated soil moisture conditions are influenced by different soil and plant properties and to compare this influence with observed differences between different plots within an agricultural soil. The model used has previously been adapted to the field and a reasonable agreement between simulated and observed tensions was obtained but no efforts were made to explain the found spatial variability within the field (Jansson & Thoms-Hjärpe, 1986).

The field site and measurements

The site was the experimental field, Kjettslinge (Lat. 60 10′ N,17 38′E), in the "Ecology of Arable Land" project in Sweden. A detailed field description is given by Steen et al. (1984) and only information relevant in the present context will be repeated here.

The experimental design consisted originally of four cropping systems each with 4 replicates (=blocks). Only two of these treatments representing unfertilized and N-fertilized barley are considered here. The size of each plot was 30x12 m and within each of these plots one neutron probe tube and a profile of tensiometers were installed.Both the neutron tube and the tensiometers were used for measurements down to 60 cm depth with increments of 10 cm for the neuton probe and 15 cm for the tensiometers. Duplicates of tensiometers were used at 15 cm depth in each plot. Measurements were made during the growing season 1982, three times each week with tensiometers and two times each month with the neutron probe (for details see Jansson & Thoms-Hjärpe ,1986).

THE IMPORTANCE OF SOIL PROPERTIES WHEN SIMULATING WATER DYNAMICS FOR AN AGRICULTURAL CROP-SOIL SYSTEM

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Abstract

A physically based model was used to obtain the sensitivity for a variation of the surface resistance, the critical tension for reduction of water uptake, a subsurface source flow, the thickness of a layer of fine sand, the slope of the water retention curve and the unsaturated conductivity. The purpose was to compare the obtained sensitivities with differences observed within a field. The model was previously adapted to the field and a reasonable agreement with areal mean values was acheived.

Plant properties only slightly influenced the simulated tensions for the topsoil and a variation of plant properties could therefore hardly explain any greater part of the observed spatial variability. A subsurface source flow strongly influenced simulations but the agreement with measurements became simultanously worsen for all plots within the field. The thickness of the fine sand layer seemed to explain a great deal of observed variability but som exceptions occured. The slope of the water retention curve affected the agreement with measured tesions and water storages in two different directions which made it unreasonable to assume an important role for this property. Finally, the unsaturated conductivity caused the strongest influence on simulated tensions but the influence on simulated water storages was less pronounced.

Model description

The model is based on the one-dimensional form of the Fokker Plank-equation or generalized Richard's equation:

$$\frac{\partial \theta}{\partial t} = \frac{\partial}{\partial z}\left(k\left(\frac{\partial \psi}{\partial z} + 1\right)\right) + S(t) \qquad (1)$$

where

θ = soil water content ψ = water tension k = unsaturated conductivity
S = sink/source term

A detailed technical description of the model, also including heat transport in the soil, was given by Jansson & Halldin (1980). The previous use of the model has mainly been restricted to forest soils (Jansson, 1980), so a number of modifications were needed, especially with regard to boundary conditions and to the sink term.

Soil properties - the water characteristic curve and the unsaturated conductivity function - are adapted to modified expressions of Brooks & Corey (1964) and Mualem (1976), respectively. In a soil with a developed structure of macropores, an air-filled volume may be considered below the so-called air entry pressure (ψ_a) in the expression of Brooks & Corey, where

$$S_e = \left(\frac{\psi}{\psi_a}\right)^{-\lambda}$$

$$S_e = \frac{\theta - \theta_r}{\theta_s - \theta_r} \qquad (2)$$

and ψ is the water tension, λ is the pore size distribution index, S_e is the effective saturation, θ is the volumetric water content, θ_s and θ_r are the correspondings water contents at saturation and at high tensions respectively. Equation (2) is used when the air-filled porosity is higher than 4 volume per cent and the tension is below pF 3. At lower air-filled porosities a simple linear relation between the tension and the water content is used. The linear relation is thus valid between the tension that corresponds to an air-filled porosity of 4 volume per cent according to Equation (2) and the null tension that corresponds to complete saturation. At higher tension than pF 3 a log-linear relation is used between the wilting point pF 4.2 and pF 3.

The unsaturated conductivity is calculated using the equation given by Mualem (1976):

$$k = k_s \left(\frac{\psi_a}{\psi}\right)^{2+(2+n)\lambda} \qquad (3)$$

where k_s is the conductivity at saturation and n is a parameter accounting for the tortuosity of flow path. The saturated conductivity in this equation do not account for rapid flows in macropores which may be important in clayey soils. Too account for this, an additional contribution to the conductivity can be added to the one given by Equation (3) when the air-filled porosity is below 4 volume per cent.

To account for the drainage to pipes, a net horizontal water flow is calculated for each layer, i.e. a contribution to the sink term in Equation (1). This water flow is calculated below the depth of the mean ground water table and above the depth of the drainage pipes. The horizontal water flow from each layer is calculated by the Darcy law, i.e., by the saturated conductivity and a gradient estimated as the difference between the mean depth of the ground water table and the depth of the pipes divided by the distance between drainage pipes. Vertical redistribution of water is calculated, below the ground water table, by assuming conservation of mass and the water contents exactly at saturation.

The interception of vegetation is treated as a single storage controlled by a simple threshold formulation.

The interception storage is described by a function accounting for the development of the crop. Evaporation from water intercepted on the crop is calculated with a form of Penman combination equation as given by Monteith (1965). The surface resistance is assumed to be close to zero.

Water uptake from the soil profile and evaporation from the soil surface is calculated as a function of potential demand, a depth distribution and actual tensions in the soil. The potential demand is calculated from the combination equation with a surface resistance, treated as a temporal function. When evaporation of intercepted water occurs, a reduction of the demand from the soil profile is considered. The remaining demand from the soil, E_p, is then calculated as:

$$E_p = \left(1 - \frac{E_{aint}}{E_{pint}}\right) E_{pot} \tag{4}$$

where E_{pot} is the unreduced demand, E_{pint} is the potential demand of intercepted water and E_{aint} is the actual amount of evaporated water from the interception storage.

The potential demand of water from the soil is distributed according to the root development and the degree of bare soil. The actual amount of water uptake, from each layer, is reduced when the tension exceeds a critical value, ψ_c. The actual uptake above this tension is proportional to

$$R = \left(\frac{\psi_c}{\psi}\right)^{a \cdot E_{pot} + b} \tag{5}$$

where a and b are empirical constants.

Adaptation of the model

The adaptation of the model to the treatments unfertilized and N-fertilized barley is previously presented by Jannson & Thoms-Hjärpe ,1986. Only the most important plant and soil properties will be discussed in this paper.

Driving variables to the model consist of daily mean values of air temperature, humidity, wind speed, and daily sums of precipitation and global or net radiation. These variables were normally available at the experimental field with exceptions for some periods when data from a nearby meteorological station had to be selected. This was due to problems with the used equipement for data collection.

To analyse the role of certain plant and soil properties a number of model parameters were selected to be altered in different simulations. Two of these selected parameters, representing plant properties, the minimum surface resistance for transpiration (rs) and the critical tension for reduction of water uptake by roots (ψ_c, eq. 5) had

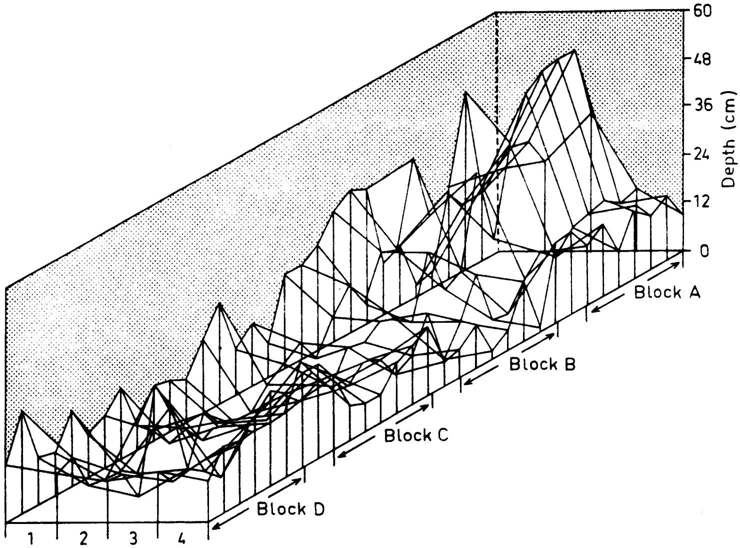

Figure 1. Thickness of the fine sand layer in the the Kjettslinge field (from Steen et al, 1984)

Table 1. Chosen values on a selected set of parameters used in the sensitity test.

Parameter	Symbol	unit	Run 1	2	3	4	Previously used values unferilized Barley	fertilized Barley
Minimum surface resistance	r_s	(s/m)	30	40	50	60	50	30
Critical water tension	Ψ_c	(cm water)	200	400	600	800	400	800
Subsurface water flow	q_s	(mm/day)	0.	0.5	1.0	1.5	0	0
Thickness of fine sand layer	Δz_s	(cm)	0	10	20	30	19	9
Pore size distribution index, topsoil	λ	(-)	0.15	0.13	0.11	0.09	0.15	0.15
Conductivity eq. 3,topsoil	k_s	(10-3cm/min)	1.6	0.5	0.16	0.05	1.6	1.6

previously been identified as important to partly explain the observed differences in soil moisture conditions between unfertilized and fertilized barley (Tab. 1).

Measurements of ground water pressure revealed an upward movement of water from subsoil and extremely high drainage rates were observed from some field lysimeters (Lars Bergström, pers com.). To account for this a subsurface source flow (q) was introduced in the model considered with a constant rate during the whole season (Tab. 1).

Soil properties were determined from soil cores and from a detailed survey of the depths of the different textural horizons in the field (Steen et al., 1984). The top soil is rather homogeneous and approximately of similar depth throughout the field. Below this horizon, the variability is substantial, with 94 per cent of the field covered by a layer of fine sand with a thickness of up to 50 cm. The mean thicknesses were 9 and 19 cm for the fertilized and unfertilized barley, respectively. The loamy fine sand is underlain by a loamy coarse sand in 20 per cent of the field with a thickness never exceeding 5 cm. A clay soil with increasing clay content with depth is found below the sand all over the field. Especially the thickness of the fine sand (z) varies substantially within the field (Fig. 1 and Tab. 2) which made it interesting to analyse how this layer affected simulated moisture conditions. Also a variablity in the pore size distribution index (,eq 2) was introduced (Tab. 1) to obtain a variation of water retention for the topsoil (Fig 2).

The unsaturated conductivity function given by eq. (3) and an estimate of the saturated conductivity (excluding marcopores) was not compared to any independent measurements of unsaturated conductivity in the previous study (Jansson & Thoms-Hjärpe, 1986). A possible variation in this property was therefore introduced (Tab. 1 & Fig. 2)

A variation in the selected properties resulted in a change in the simulated moisture dynamics and thereby the agreement between simulated and measured variables was affected. To quantify the degree of agreement the root mean square (RMS) and the arithmetic mean (M) of the residuals were calculated.

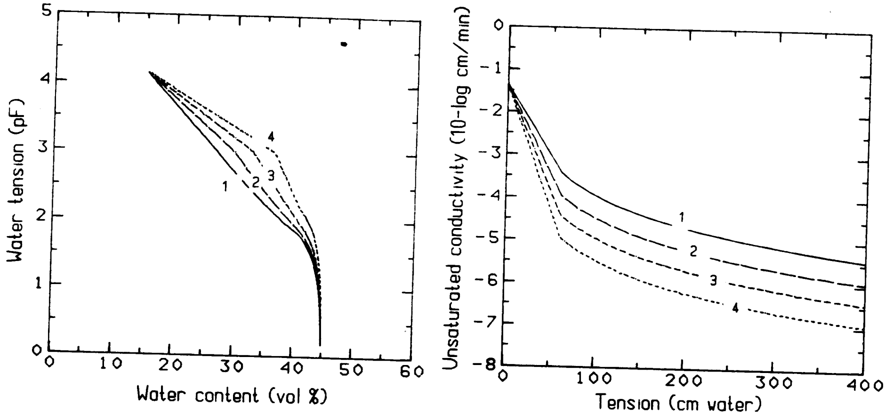

Figure 2. Seleted variation in the water retention curve and in the unsaturated conductivity function for the topsoil.

Table 2. Areal mean values of the thickness of the fine sand layer at different plots within the field (from Steen et al, 1984).

	Treatment			
	Unfertilized Barley		fertilized Barley	
Block	Degree of cover (-)	Thickness (cm)	Degree of cover (-)	Thickness (cm)
A	1.0	34	1.0	14
B	1.0	21	0.91	10
C	1.0	12	0.71	5
D	1.0	12	0.97	13

Results

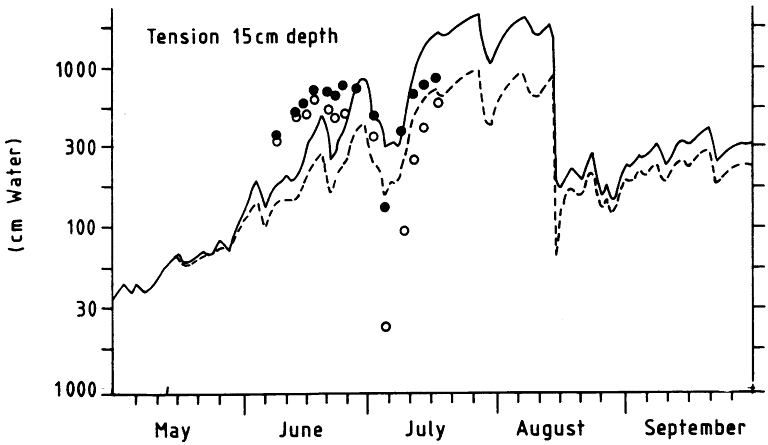

Figure 3. Simulated and measured tensions at 15 cm depth.
Solid line represents simulations of fertilized barley and dashed line unfertilized barley. Filled cirlces represent measurements from fertilized barley and open cirlces measurements from unfertilized barley. The measurements represent areal mean values of 4 blocks within the field (from Jansson & Thoms-Hjärpe, 1986)

A number of different sensitivity pattern were obtained depending on: point of measurement in the field, N-fertilization rate of barley, compared variable and changed property . The result (Fig. 3) from Jansson & Thoms-Hjärpe (1986) will here be used as a reference, and it should be kept in mind that they also considered another growing season and they did not used numerical techniques to estimate model parameters.

A decrease of the minimum surface resistance (r_s) caused an improved fit for unfertilized barley (Fig. 4) whereas an increase similarily worsen the fit for N-fertilized barley (Fig. 5). However, a change of the surface resistance could hardly explain the differences between different blocks in the field, since the fit changed similarily for the different blocks and the difference between blocks was much greater than what could be caused by a change in the surface resistance.

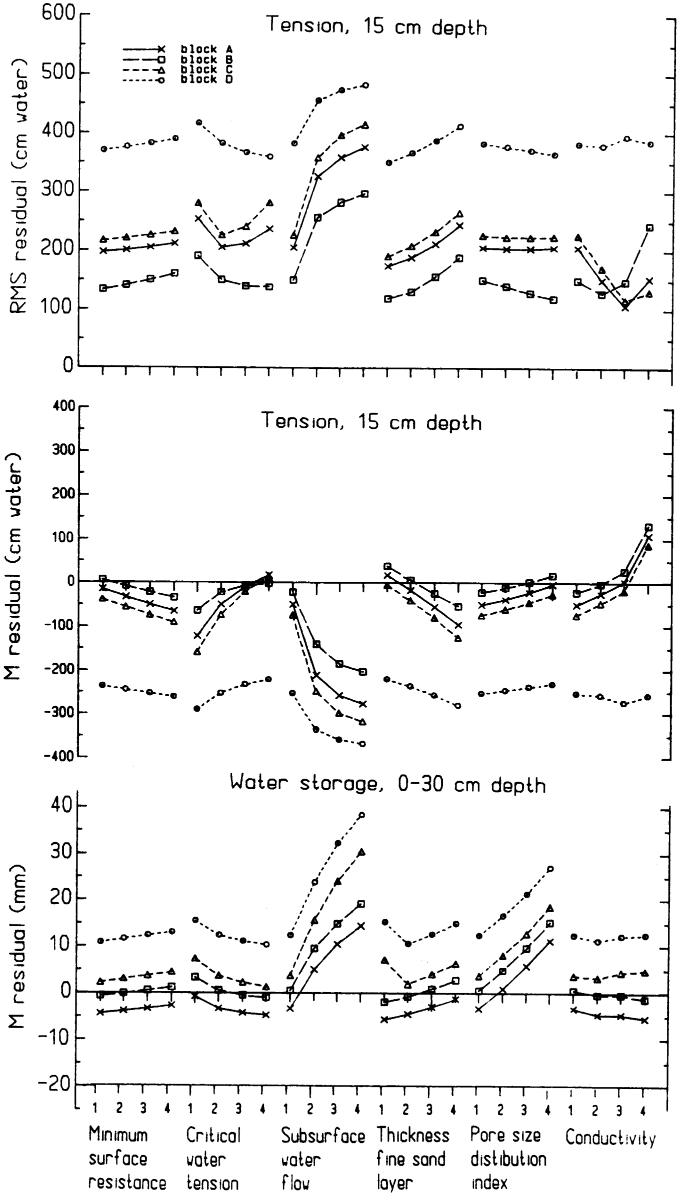

Figure 4. Residual between simulated and measured variables for unfertilized barley, 1982.

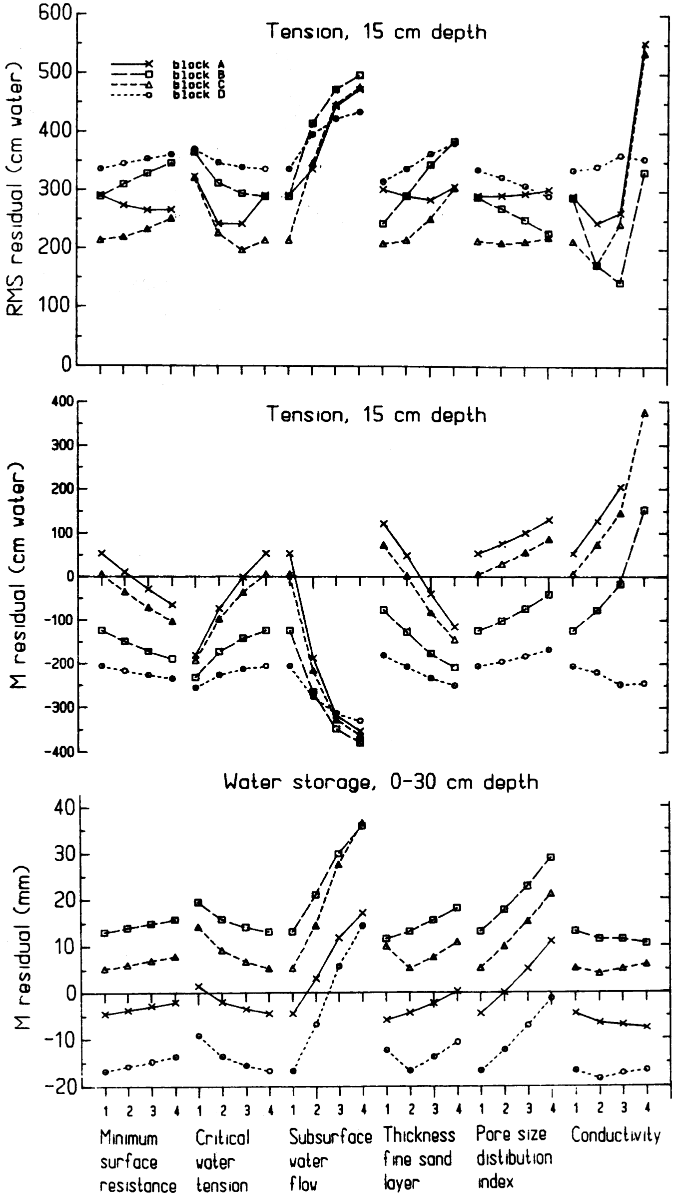

Figure 5. Residuals between simulated and measured variables for fertilized barley, 1982.

A change of the critical tension (ψ_c) for reduction of water uptake showed a quite different pattern compared to that obtained when testing the sensitivity for a change of the surface resistance. A minimum of RMS (Fig. 4 & 5) were obtained for some of the blocks both for unfertilized and for N-fertilized barley. Different values on the critical tension (similar as the reference runs) resulted in the best agreement for the two treatments. Similar as for the surface resitance it was concluded that a variation of the critical tension hardly could explain the differences between observations in the individual plots.

The introduction of a subsurface source flow totally failed to improve the fit with measurements (Fig. 4 & 5). This reject the hypothesis that a subsurface flow would have been important to explain observed spatial variablitity in this case.

The sensitivity for the thickness of the fine sand layer was of special interest because of the available information on this property (Fig. 1 & Tab. 2). The mean residuals for the tension at 15 cm depth (Fig. 4 & 5) showed differences between blocks which partly could be explained by differences in the thickness of this layer according to model behaviour. The sensitivity for a variation of the fine sand layer was especially strong for the N-fertilized barley. A thicker sand layer resulted in lower tensions in the topsoil, probably because of a more efficient capilary rise in sand compared to a clay horizon. The two most moist blocks of unfertilized barley were also the ones with the thickest fine sand layer. The same was true for N-fertilized barley with respect to block A, with the thickest fine sand layer, but not with respect to the other blocks. Especially block C was too moist when considering its thin layer of fine sand. Obviously this property could not entirely explain the observed differences between the different blocks.

When analysing the sensitivity of a change in the pore size distribution factor it was not enough to consider the influence on the agreement with measured tensions. The sensitivity pattern for tensions and water contents were very similar, but opposite in direction, with respect to tested evapotranspiration properties and to tested thickness in the fine sand layer (Figs 4 & 5). In the case of changing the pore size distribution factor, tensions became higher simultanously as the water storages also became higher. This seemed less realistic because the agreement with observed water storages got more worse than the agreement with tensions improved. Consequently a variability in the water retention curve did not seem to explain any part of the observed variability within the field.

The strongest sensitivity on the agreement with measured tensions was caused by the unsaturated conductivity. A decreased conductivity with one power of ten resulted in a substantial change of agreement, with different patterns for different blocks (Figs. 4 & 5). Generally, better fits than the reference runs were obtained, especially when considering RMS. Providing that a spatial variability exist in unsaturated conductivity, a great deal of the observed differences in tensions between different blocks could be explained. However, the observed differences in water storages could not at all be explained by differences in the unsaturated conductivity. The agreements with the water storage, 0-30 cm, were not affected at all by the unsaturated conductivity.

Conclusions

The complicated sensitivity patterns obtained, sometimes similar and sometimes different for different plots in the field and for different variables, demonstrates the problem of estimating parameters with fitting techniques to a model where many parameters can be changed.

The result can be used to evalute the importance of an independent quantative knowledge on different plant and soil properties.

The unsaturated conductivity was found to be very important when simulating water tensions. When only the water balance is of interest a weak dependence of unsaturated conductivity for a sandy forest soil has previously been demonstrated (Jansson, 1984). In this application also a weak dependence was obtained on the water storage of the topsoil but not on water tensions.

Acknowledgements

I am grateful for comments on the manuscipt from Lars Bergström and Holger Johnsson. The soil moisture studies at Kjettslinge is part of the integrated research project "Ecology of Arable Land. The Role of Organisms in Nitrogen Cycling.", supported by grants from the Swedish Council for Planning and Coordination of Research, the Swedish Council for Forestry and Agricultural Research, the Swedish Natural Science Research Council and the Swedish National Environment Protection Board.

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SPATIAL VARIABILITY OF SOIL PROPERTIES

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SUMMARY

The present study includes an experimental and theoretical analysis of the spatial variability of soil properties for two 0.5 ha large fields: a sandy loam situated at Tåstrup and a coarse sand situated at Jyndevad. In situ soil samples were taken at 24 sampling sites systematically distributed in each of the fields. Saturated hydraulic conductivity, dry bulk density, porosity, the soil water retention curve, and the textural composition were determined for each sample. The textural composition of the soil was determined at additional 29 sampling sites. The experimental data were subjected to statistical treatments including ordinary statistics as well as geostatistical methods. The dry bulk density is increasing with depth at Tåstrup while it is decreasing at Jyndevad. Regarding the soil moisture characteristic especially the top layer 10 cm at Tåstrup and 10-30 cm at Jyndevad differs from the lower layers. The saturated hydraulic conductivity increases with depth at Jyndevad while the variation at Tåstrup shows no clear pattern. Spatial dependence is described by semivariogram models. The jackknifing procedure to validate the selected semivariogram models shows good results at Tåstrup, the results are not quite as satisfactory at Jyndevad. The spatial dependence shows consistency between depths at Jyndevad, but not at Tåstrup.

INTRODUCTION

In statistical analyses of soil physical properties the observations are often assumed to be spatial independent of each other, and hence, a set of observations are reduced to their mean value and a measure of its uncertainty expressed in terms of an assumed probability density distribution estimated by a set of observations without regard to their spatial positions. However, we do not expect soil physical properties to be necessarily spatially independent. We would expect measurements made close together to yield nearly equal values, and measurements made at some distance to be correlated to each other. The present paper describes statistical analyses of some important soil properties for two soils by using ordinary statistics as well as geostatistical methods.

SOIL SAMPLING

In the present experimental study two test sites were selected for soil sampling. Test site Tåstrup is situated on Zealand in a moraine landscape whereas test site Jyndevad is situated in Jutland in a moorland plain landscape. The soil at test site Tåstrup is a sandy loam whereas the soil at test site Jyndevad is a coarse sand.

At each test site a rectangular testfield of 0.5 ha was employed. In the testfield 53 sampling points were systematically located, Figure 1. Soil samples for determination of soil texture were taken at 53 sampling points whereas common soil samples for determination of soil water characteristics, saturated hydraulic conductivity, dry bulk density and porosity, respectively, were taken at 24 sampling points.At test site Tåstrup soil samples were taken in 10, 30, 50, 70 and 90 cm depth at each sampling point. At test site Jyndevad soil samples were taken in 10, 30 and 50 cm depth at each sampling point.

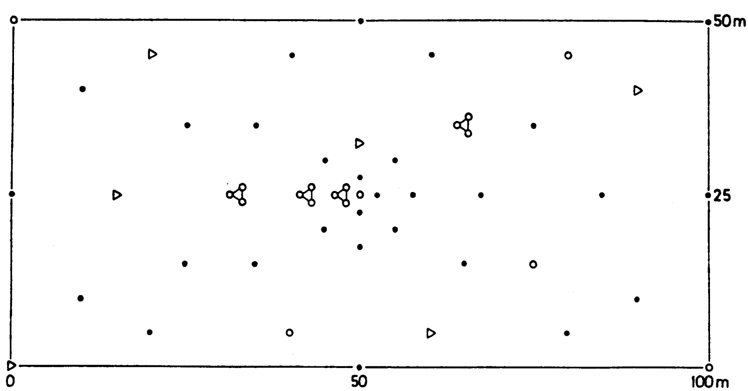

Figure 1 Test site and location of sampling points.

Regarding soil texture each value was based on analysis of one single soil sample whereas each mean value of soil water characteristics, saturated hydraulic conductivity, dry bulk density and porosity, respectively, was based on analysis of three individual soil samples.

STATISTICAL ANALYSES

VARIATION ACCORDING TO DEPTH

For a given soil physical property it is assumed that the sample originating from a given depth is normally distributed and that the observations are independent. The only exception is the saturated hydraulic conductivity where the samples are assumed to come from log-normal distributed populations.

Test site Tåstrup
The profiles of the mean clay, silt, fine sand and coarse sand content are shown in Figure 2. A 95%-confidence interval for the sample mean is given as a bar at each depth. It is apparent that there is no significant variation in the mean value of either of the clay, silt, fine sand or coarse sand content according to depth. The dry bulk density is increasing with the depth and the porosity is decreasing, Figure 3.The soil water characteristic data are illustrated in Figure 4. It is apparent that the profile

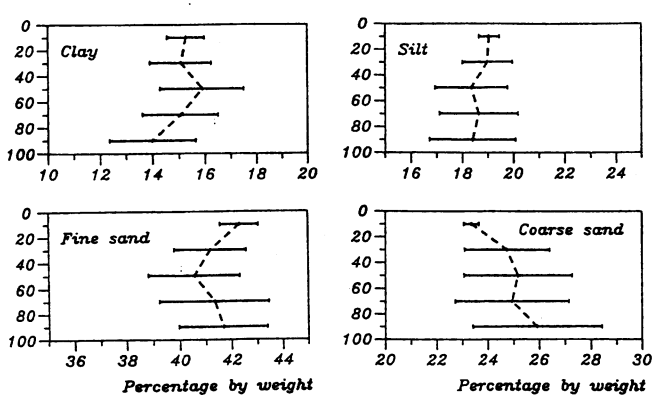

Figure 2 Profile of textural composition. Mean and 95%-confidence limits, Tåstrup

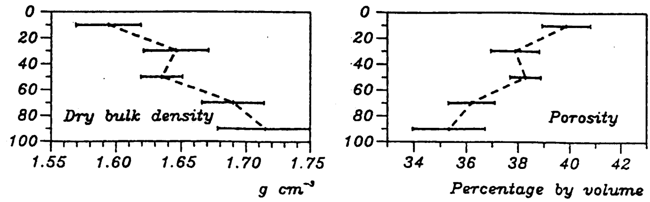

Figure 3 Profile of dry bulk density and porosity. Mean and 95%-confidence limits, Tåstrup.

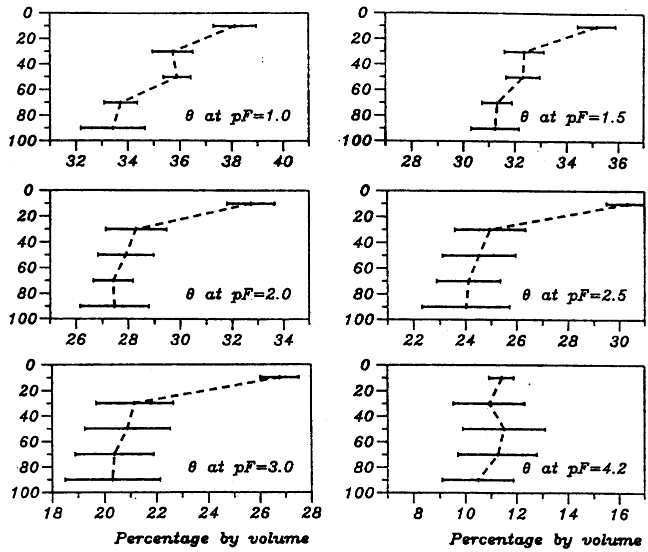

Figure 4 Profile of soil water characteristic. Mean and 95%-confidence limits, Tåstrup.

is not uniform except at the very dry end of the soil water characteristic. It is evident that especially the top layer differs from the other layers. At pF=1.5 to pF=3.0 no significant difference exists between the lower layers (30-90 cm). It is concluded that the lower layers (30-90 cm) can be treated more or less alike while the top layer must be treated separately. In Figure 5 the available water content (AWC) is illustrated. The conclusion made for the soil water characteristic is also valid in this case. The saturated hydraulic conductivity profile is not uniform, Figure 6.

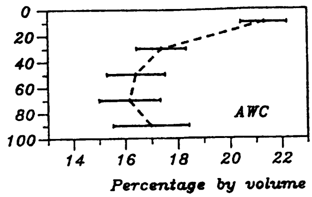

Figure 5 Profile of available water content. Mean and 95%-confidence limits, Tåstrup.

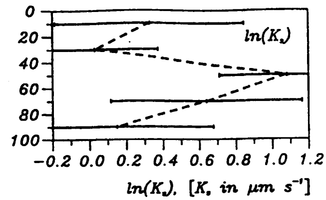

Figure 6 Profile of log-transformed saturated hydraulic conductivity. Mean and 95%-confidence limits, Tåstrup.

Test site Jyndevad

The average profiles of the humus, clay, silt, fine sand and coarse sand content are shown in Figure 7. It is evident that not any of the profiles is uniform. The mean of the dry bulk density and the mean of the porosity are shown with 95% confidence limits of the mean in Figure 8. It can be seen that the profiles are not uniform. Regarding both the dry bulk density and the porosity especially the top layer (10 cm) differs from the other layers.

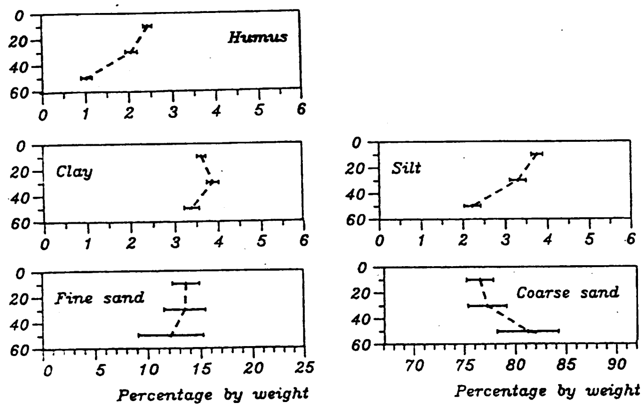

Figure 7 Profile of textural composition. Mean and 95%-confidence limits, Jyndevad

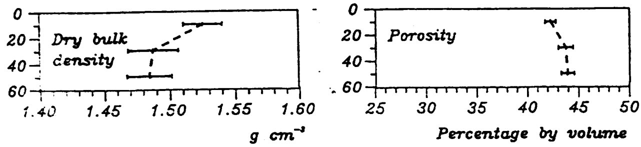

Figure 8 Profile of dry bulk density and porosity Mean and 95%-confidence limits, Jyndevad.

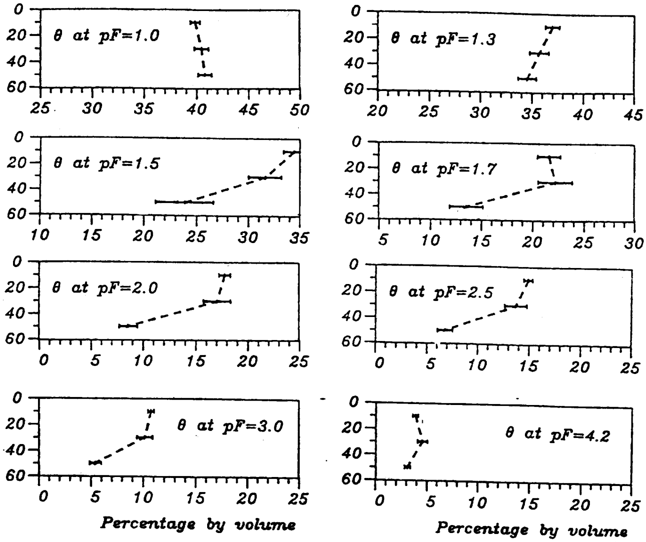

Figure 9 Profile of soil water characteristic. Mean and 95%-confidence limits, Jyndevad.

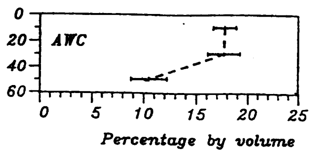

Figure 10 Profile of available water content. Mean and 95%-confidence limits, Jyndevad.

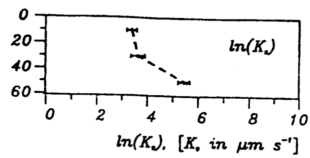

Figure 11 Profile of log-transformed saturated hydraulic conductivity. Mean and 95%-confidence limits, Jyndevad.

It is evident that the top layer must be treated separately while the other layers may be treated as one. The soil water characteristic data are illustrated in Figure 9. In the wettest end of the soil water characteristic (pF=1.0) the profile may be assumed uniform. But at pF=1.3 to pF=4.2 it is not the case. The mean available water content (AWC) is illustrated in Figure 10. It is noted that the 10 and 30 cm layers differ from the lower layer. The 10 and the 30 cm layer may be assumed to have a common mean. The variation in saturated hydraulic conductivity according to depth is illustrated in Figure 11. Apparent is that the two upper layers (10-30 cm) may be assumed to have a common mean (log-transformed observations).

SPATIAL DEPENDENCE

The spatial dependence in a given soil property can be estimated from the semivariogram γ, estimated by

$$\gamma^*(h) = \frac{1}{2N(h)} \sum_{i=1}^{i=N(h)} (Z(X_i) - Z(X_i + h))^2 \qquad (1)$$

where $N(h)$ is the number of sampling pairs of observations $(Z(X_i), Z(X_i+h))$ separated by a distance h. An experimental semivariogram displays a series of discrete points corresponding to each value of h. The spatial dependence is evaluated by fitting a continuous function - a semivariogram model - into these points. Several theoretical models may be used to describe the experimental semivariograms. Most common are the linear, spherical and exponential models. In the present study we confine ourselves to discuss a linear model which in a normalized form may be expressed as

$$\gamma^n(h) = \frac{\gamma^*(h)}{S^2} = C_0 + (C_1 - C_0)h/a \quad 0 < h < a$$

$$\gamma^n(h) = C_0 \qquad h \geq a \qquad (2)$$

where $\gamma^n(h)$ is the normalized semivariogram function (dimensionless), S^2 is the sample variance, C_0 is the normalized nugget effect (dimensionless), and C_1 is the normalized sill (dimensionless) and a is the range (meter).

The theoretical semivariogram can be used in the kriging method. This interpolation method is an optimum interpolator because it interpolates without bias and with known minimum variance. The method requires a second-order stationarity. We might expect that the estimated $Z^*(X_K)$ values are different from the measured $Z(X_K)$. The error introduced is $e(X_K)=Z(X_K)-Z^*(X_K)$. By dividing by the standard deviation of the estimate $\sigma_K(X_K)$ we obtain the reduced error $r(X_K)=e(X_K)/\sigma_K(X_K)$ which is dimensionless.

The selected theoretical semivariogram can be tested by using the jackknifing technique in which the kriging procedure is used. By suppressing each sampling point, one at a time and estimate a value $Z^*(X_i)$ in that point from the remaining N-1 data, the reduced error $r(X_i)$ can be analysed. The mean reduced error, RE,should be close to zero, which indicates no systematic overestimation or underestimation. The mean reduced variance, RV, (Var $r(X_i)$),should be close to unity. Note that this indicates consistency between the 'observed' kriging errors $e(X_i)^2$ and the estimation variance $\sigma^2(X_i)$. In the present study the linear semivariogram is the result of a least square fit of the data points. The values of RE and RV may differ from zero and one, respectively. In cases where large discrepancies are found, the theoretical semivariogram should be modified before further application.

Test site Tåstrup

Examples of the calculated semivariograms are shown in Figure 12. As mentioned earlier, each point in the semivariogram is the average value of a number of sampling pairs separated by a distance h. The value of h itself is the average distance of all pairs falling within a certain interval. The size of the interval has been specified to 5 metres. From a visual inspection the semivariograms have been grouped into three categories A, B and C. Semivariograms falling into category A show a significant spatial dependence. This is characterized by a small nugget effect and an increase in the $\gamma^*(h)$ values as h is increasing. Category B includes semivariograms which show a rather larger nugget effect and a short range. Finally, the semivariogams in category C display no spatial dependence. It appears, Table 1, that no textural fractions show sign of spatial dependence in all depths, and there is no specific depth in which the semivariograms show spatial dependence for all soil textures. There seems, however, to be more spatial dependence at small depths.

The porosity and dry bulk density data have no spatial correlation at all.The soil water characteristics in 10 cm and 90 cm depths fall mostly into category A whereas for the three other depths the dominating is 'no spatial dependence'. There is a very poor agreement between the results of the clay data and the soil water characteristics at pF=4.2. The available water content AWC displays a significant spatial dependence for all depths.

Table 1 Main features of the experimental semivariograms at test site Tåstrup.

Soil property	Depth, cm 10	30	50	70	90
Clay	A	A	B	B	C
Silt	C	B	C	B	C
Fine sand	A	A	B	C	A
Coarse sand	B	A	A	C	C
Dry bulk density	C	C	C	C	C
Porosity	C	C	C	C	C
pF = 1.0	A	C	C	C	C
pF = 1.5	A	C	C	C	A
pF = 2.0	A	B	C	C	A
pF = 2.5	A	B	C	C	A
pF = 3.0	A	B	A	C	A
pF = 4.2	C	B	A	C	A
AWC (pF = 2.0 - pF = 4.2)	A	A	A	A	A

A: Significant spatial dependency
B: Spatial dependency not significant (large nugget)
C: No spatial dependency (pure nugget)

Table 2 Parameters for the fitted normalized semivariogram, sample variance, mean reduced error R, and reduced variance S^2 at test site Tåstrup.

Soil property	Depth (cm)	Nugget	Sill	Range (m)	Sample variance	$\bar{R}$	S^2
Clay	10	0.10	1.17	27	6.64	-0.038	1.442
	30	0.34	1.24	40	17.9	0.012	0.905
	50	0.57	0.97	20	34.2	-0.004	0.908
	70	0.38	1.07	15	27.0	0.050	1.209
Fine sand	10	0.18	1.15	35	7.3	0.008	1.022
	30	0.49	1.20	39	25.4	-0.052	0.566
	50	0.54	1.07	15	40.2	-0.050	0.881
	90	0.27	1.14	21	38.6	0.006	1.641
Coarse sand	10	0.60	1.04	15	1.18	-0.074	0.806
	30	0.15	1.00	10	36.2	0.0	1.355
	50	0.0	1.06	13	58.4	-0.032	1.460
pF = 1.0	10	0.40	0.90	25	4.0	0.109	0.619
1.5	10	0.24	1.26	25	2.89	0.136	0.552
2.0	10	0.0	1.41	30	4.84	0.129	1.251
2.5	10	0.09	1.53	39	3.61	0.106	0.987
3.0	10	0.0	1.51	34	3.24	0.087	1.732
pF = 3.0	50	0.54	1.15	25	15.2	-0.012	1.033
4.2	50	0.26	1.19	26	14.4	0.018	1.322
pF = 1.5	90	0.42	1.27	30	4.84	0.011	1.042
2.0	90	0.38	1.20	26	9.61	0.067	1.056
2.5	90	0.35	1.14	25	16.0	0.066	1.322
3.0	90	0.43	1.18	30	18.5	0.047	1.285
4.2	90	0.54	1.26	40	10.9	0.044	1.151
AWC	10	0.10	1.00	32	4.41	0.118	1.124
	30	1.14	1.13	21	4.84	0.002	1.650
	50	0.46	1.10	16	6.76	-0.062	1.457
	70	0.39	1.11	16	7.84	0.005	1.065
	90	0.55	1.17	22	11.6	-0.004	1.049

A theoretical semivariogram has been fitted to the experimental variograms by using a least square criterion for a stepwise linear curve. This has been made for all category A semivariograms as well as for the category B semivariograms of the soil texture except silt, Table 2. It appears that the calculated range-values lie between 10 m and 40 m, smallest for coarse sand. The soil water characteristics show a range between 25 m and 40 m.

The jackknifing procedure to validate the fitted semivariograms has yielded good results (see Table 2). The kriging estimates are all unbiased and the computed kriging variances are consistent with the data. This is confirmed by a reduced variance S^2 value which in most cases is close to unity.

Test site Jyndevad

Examples of the calculated semivariograms for the Jyndevad site are shown in Figure 13. The general impression is that the soil properties show either spatial dependence in all three depths or no spatial dependence at all, Table 3. Spatial correlations have not been found for silt, humus, dry bulk density, porosity and the soil water characteristics at pF=1.0. Spatial dependance appears only at one depth at pF=1.3 and pF=4.2.

Clay displays correlations up to distances of approximately 20 m. For fine sand and coarse sand range values have only been found at 50 cm depth. These are approximately 45 m. In both 10 cm and 30 cm, however, the data points do not level off at a sill value within the distances investigated. It should be noted that the semivariograms for the two textural fractions are almost identical. The experimental semivariograms for 10 cm and 30 cm have been approximated by a straight line starting from origin. A better fit might be a curve of parabolic shape, indicating some sort of nonstationarity in the data. A two dimensional regression model $Z = ax + by + c$ describes the trend well.

An example of a semivariogram for the trend-free coarse sand data (residuals) is shown in Figure 13. This show a sill-value. The cross-validation or jackknifing test of the fitted semivariograms, given in Table 4, shows however that kriging on the basis

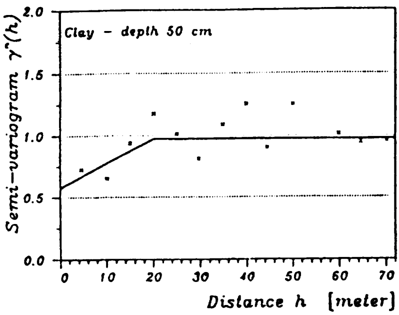

Figure 12 Examples of semivariograms, Tåstrup.

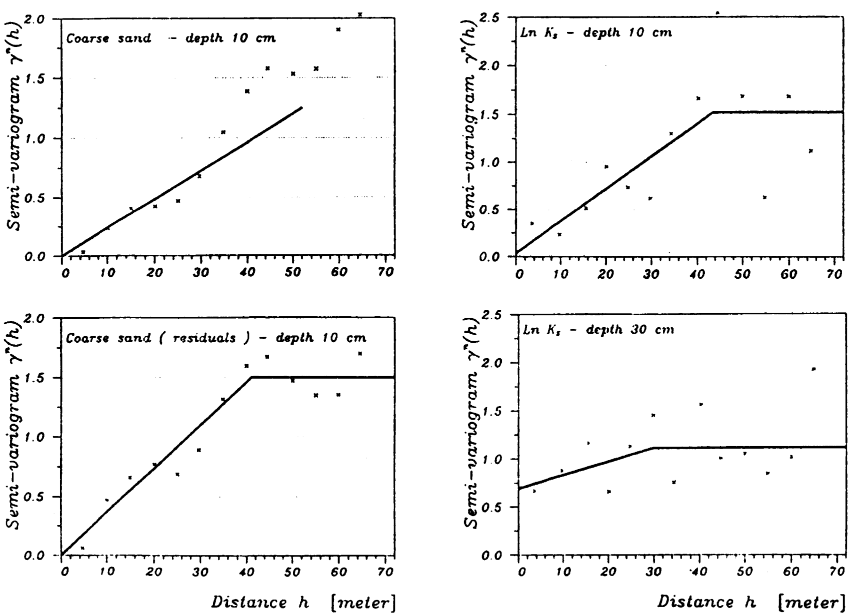

Figure 13 Examples of semivariograms, Jyndevad.

Table 3 Main features of the experimental semivariograms at test site Jyndevad.

Soil property	Depth (cm) 10	30	50	Soil property	Depth (cm) 10	30	50
Clay	A	A	B	pF = 1.0	C	C	C
Silt	C	C	C	pF = 1.3	C	A	C
Fine sand	A	A	A	pF = 1.5	C	A	A
Fine sand (residuals)	A	A	A	pF = 1.7	A	A	A
Coarse sand	A	A	A	pF = 2.0	A	A	A
Coarse sand (residuals)	A	A	A	pF = 2.5	A	A	A
Humus	C	C	C	pF = 3.0	A	A	A
				pF = 4.2	A	C	C
Dry bulk density	C	C	C	AWC (pF=1.7-pF=4.2)	A	A	A
Porosity	C	C	C	ln K_s	A	A	A

A: Significant spatial dependency
B: Spatial dependency not significant (large nugget)
C: No spatial dependency (pure nugget)

Table 4 Parameters for the fitted normalized semivariogram, sample variance, mean reduced error R, and reduced variance S² at test site Jyndevad.

Soil property	Depth (cm)	Nugget	Sill	Range (m)	Sample variance	$\bar{R}$	S^2
Clay	10	0.48	1.08	15	0.13	0.006	0.146
	30	0.15	1.13	20	0.25	0.005	1.016
Fine sand	10	0.0	-	-	19.0	-0.066	1.316
	30	0.0	-	-	48.4	-0.029	2.686
	50	0.07	1.36	43	124.	-0.012	2.952
Fine sand (residuals)	10	0.0	1.43	40	10.11	-0.036	2.539
	30	0.0	1.41	40	31.85	-0.016	3.120
	50	0.0	1.19	26	95.75	-0.082	3.410
Coarse sand	10	0.0	-	-	18.2	0.061	1.472
	30	0.0	-	-	43.5	0.014	2.021
	50	0.13	1.34	44	120.	0.006	2.647
Coarse sand (residuals)	10	0.0	1.49	41	8.74	0.038	2.255
	30	0.02	1.41	40	27.37	0.007	3.435
	50	0.02	1.23	27	91.91	0.057	3.560
pF = 1.7	10	0.0	0.95	30	6.76	0.083	0.971
2.0	10	0.28	1.25	34	1.69	0.041	0.322
2.5	10	0.30	1.27	34	1.0	0.047	0.354
3.0	10	0.21	1.27	40	0.49	0.091	1.010
4.2	10	0.09	1.30	55	0.49	0.010	2.637
pF = 1.3	30	0.0	1.50	50	4.84	0.039	4.606
1.5	30	0.0	1.46	45	14.44	-0.015	2.649
1.7	30	0.04	1.54	50	15.2	-0.046	1.755
2.0	30	0.31	1.33	45	9.61	-0.069	0.878
2.5	30	0.25	1.41	45	6.76	-0.065	0.888
3.0	30	0.11	1.49	45	3.24	0.070	1.109
pF = 1.5	50	0.40	1.44	45	42.3	-0.024	0.988
1.7	50	0.0	1.43	45	15.2	-0.025	2.794
2.0	50	0.26	0.86	16	4.41	-0.151	1.198
2.5	50	0.21	1.04	16	3.24	-0.152	1.229
3.0	50	0.23	1.08	14	1.69	-0.157	0.998
AWC	10	0.27	1.15	45	6.76	0.044	0.917
	30	0.12	1.12	34	13.69	-0.025	1.583
	50	0.0	1.47	45	16.81	0.003	1.896
ln K_s	10	0.03	1.52	43	0.22	-0.041	2.447
	30	0.69	1.11	30	0.44	0.084	0.762
	50	0.18	1.17	25	0.30	0.0	2.116

of the original data is more reliable than on the residuals, illustrated by the S^2-values. It should be emphasized that a range value is only of importance in cases where an interpolation involves data points with mutual distances apart larger than the range-value.

The ranges of the soil water characteristics fall between 30 m and 50 m except for the driest part at 50 cm depth. In this case small range-values of approximately 15 m have been found. The reduced variance S^2 in Table 4 varies between 0.322 and 4.606. This calls for a modification of several of the fitted semivariograms before any application. In cases with large S^2-values (e.g. pF=1.3 at 30 cm) the very first $\gamma^n(h)$-point lies often above the fitted curve, and by increasing the nugget effect the S^2-value could be reduced. In cases of unsatisfatorily small S^2-values (e.g. pF=2.0 and pF=2.5 at 10 cm) a smaller nugget effect is acceptable, which would yield a larger value of S^2.

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SIMULATION OF UNSATURATED FLOW IN HETEROGENEOUS SOILS

Part I: Application to individual soil profiles

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Abstract

Water flow and solute transport in the upper soil layers are processes which are of importance in many hydrological and agricultural applications. These processes are occurring in a medium which has a complex composition, since the soil is a product of both the geological formation history and various man-induced activities like agricultural development. In addition, the upper soil is subject to cyclic variations in water content in response to the variable exposure of rainfall and outer evapotranspiration demand.

The paper describes an attempt to simulate the field conditions at different points inside a research field where comprehensive field investigations are carried out. The simulations are based on a solution to the partial differential flow equation, subject to the measured climatic conditions and the appropriate soil physical parameters representing the hydraulic properties in the individual soil columns. The simulations of the variation in water content are compared with measurements.

Introduction

The moisture conditions in the soil horizons above the water table, i.e. the variation in time and space, play an important role in many applications within hydrology and agriculture, such as water balance calculations, irrigation and drainage, crop production, migration of pollutants etc. The moisture conditions in the unsaturated zone are determined by a number of interacting factors involving soil, crop as well as atmospheric processes which are difficult to separate and formulate in quantitative terms.

The level of the moisture content is to a large degree determined by the soil hydraulic properties, i.e. the retention and hydraulic conductivity functions, whereas the seasonal fluctuations are driven by the variation in rainfall and evaporative demand. Soil properties as well as climatic variables exhibit a spatial variation although the correlation scales are quite different. The correlation scale for climatic variables is much larger than for soil properties, and the present study focuses on the processes within a homogeneous climatological unit.

The study is part of a larger experimental and theoretical study which is carried out jointly by 5 Danish hydrological and agricultural institutes. The main objectives are:

- to investigate the magnitude of spatial variation in soil physical properties and variables within two fields belonging to two different mapping units.

- to evaluate the applicability of existing theories for predicting water flow and solute transport under field conditions.

- to develop appropriate methods for including variability and uncertainty in mathematical models.

The field layout illustrated in Fig. 1. shows the locations of the sampling and monitoring points (53 in total with variable distances between the individual points) within a rectangular area of 0.5 ha covered by grass.

A comprehensive laboratory and field measurement programme has been carried out including measurements of retention characteristics, saturated hydraulic conductivity, textural composition and continuous measurements of water content and tension. Some of these results have been summarized by Hansen et al. (1986). Further, radioactive isotopes have been injected at the soil surface around 12 sampling points for in-situ monitoring of the transport and dispersion characteristics.

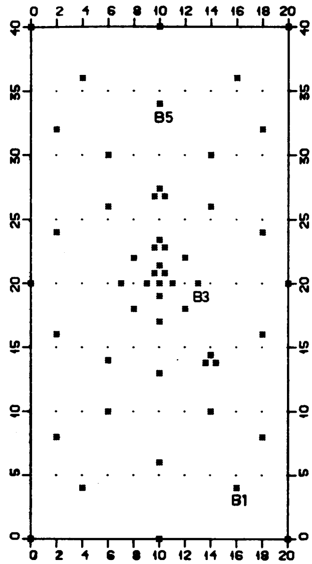


Fig. 1 Plan of field site showing mutual locations of sampling points. (Coordinates unit: 2.5 m).

The present paper discusses the potential in simulating the moisture variations in the individual soil profiles on the basis of a solution to the flow equation subject to the appropriate boundary conditions and results from the soil analyses. The mathematical modelling analysis forms the basis for the following mathematical description of unsaturated flow characteristics over the whole heterogeneous field (Part II, Jensen and Storm (1986)).

The results and analyses presented here are preliminary in the sense that they are based on the first data being processed. When larger time series become available and more data are processed, in particular the isotope measurements, this new insight may lead to other approaches and conclusions.

Model components

Simulation of water flow in the unsaturated zone involves the flow components shown in Fig. 2 below:

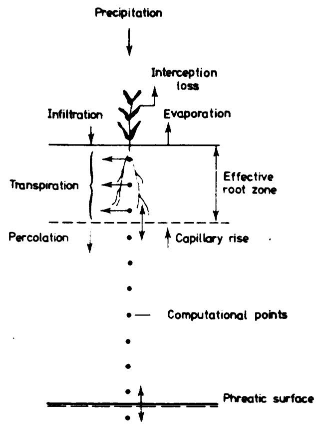


Fig. 2 Flow components in soil water flow dynamics.

The individual components are elaborated below.

Soil water flow

Assuming that water flow in the individual sampling profiles is strictly vertical the differential flow equation for unsaturated flow reads:

$$C \frac{\partial \phi}{\partial t} = \frac{\partial}{\partial z} \left(K \frac{\partial \phi}{\partial z} \right) - \frac{\partial K}{\partial z} - S \qquad (1)$$

where ϕ - capillary pressure
C - water capacity
K - hydraulic conductivity
S - sink term representing uptake by roots
z - vertical coordinate, positive downwards
t - time

This equation is derived by combining the generalized Darcy law, assuming the hydraulic conductivity to be a function of water content or capillary pressure, and the mass conservation equation.

Two functions are required for solving this equation: (a) the retention function (relationship between moisture content θ and capillary pressure ψ) and (b) the hydraulic conductivity function (relationship between hydraulic conductivity and moisture content θ or capillary pressure ψ). It is here assumed that these functions are not affected by hysteresis effects. For a further discussion of soil water flow dynamics, reference is made to Jensen (1983).

Evapotranspiration

In any attempt to simulate naturally occurring moisture conditions in the field, the evapotranspiration processes will enter as essential factors. The evapotranspiration calculations are based on a recorded value for daily potential evapotranspiration. The potential evaporative demand is divided into a fraction absorbed by the crop and a fraction reaching the soil surface according to the following relationship, Jensen (1979):

$$
\begin{aligned}
E_{ps} &= E_p \cdot \exp(-0.4\ Lai) \\
E_{pt} &= E_p \cdot (1 - \exp(-0.4\ Lai))
\end{aligned}
\qquad (2)
$$

where

E_{pt} - fraction of E_p available for evaporation of intercepted rainfall and transpiration

E_{ps} - fraction of E_p available for soil evaporation

Lai - leaf area index.

Part of the rainfall is intercepted on the crop surface from where it will evaporate directly. It will here be assumed that the interception capacity can be calculated from the relationship:

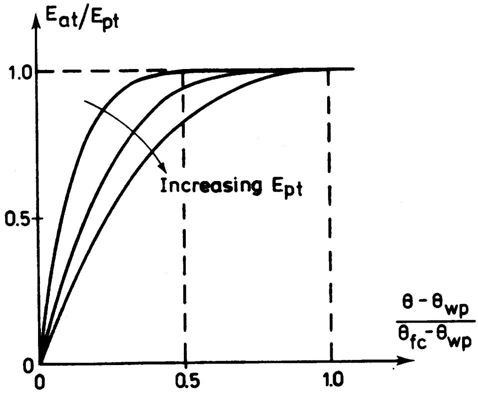
$$I_m = 0.05\ Lai \qquad (3)$$

where

I_m - interception capacity (mm)

The evaporative demand E_{pt} is first applied to the intercepted water and if it is not fulfilled the remaining part is applied for transpiration.

The external evaporative demand can only be met if the moisture supply in the root zone is adequate. For lower water contents the actual transpiration (root extraction) will be predicted according to the procedure illustrated below; (Kristensen and Jensen, 1975).



E_{at} - actual transpiration

E_{pt} - potential transpiration

θ - moisture content

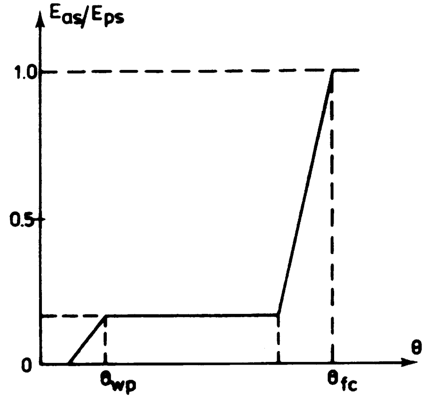
θ_{fc} - moisture content at field capacity

θ_{wp} - moisture content at wilting point

Fig. 3 Relative transpiration as a function of moisture content and potential transpiration.

The functional relationship shown above is applied to all computational points in the root zone and the derived values are subsequently multiplied by a distribution function to account for the extraction pattern of the root system. The values obtained hereby are introduced in the sink term S, Eq. 1. More details on the computational procedure can be found in Jensen (1983).

Similarly the soil evaporation demand is reduced below the potential value for moisture contents below field capacity, according to the procedure illustrated in Fig. 4.



θ_{as} - actual soil evaporation

E_{ps} - potential soil evaporation

θ - moisture content at the soil surface

θ_{fc} - moisture content at field capacity

θ_{wp} - moisture content at wilting point

Fig. 4. Relative soil evaporation as function of moisture content.

The derived value for actual soil evaporation is also introduced in the soil water flow equation through the sink term S.

Boundary conditions and computational procedure

The differential flow equation is solved by finite difference techniques on the basis of specified soil hydraulic functions and boundary conditions in the form of rainfall and the level of the water table. Meteorological data are entered on a daily basis; however the time step in the numerical procedure is generally 1 hour or less in case of high rainfall intensity.

The model predicts for each time step the vertical variation in the soil profile of water content, capillary pressure and water flow and velocity. Further, the actual values of interception loss, transpiration and soil evaporation are predicted.

Parameter requirements for model simulation

The modelling procedure described briefly above has been applied to several sampling profiles indicated in Fig. 1 with a view to establish a basis for modelling the flow in a heterogeneous field.

The results for three selected profiles are presented below and comparisons are made with field measurements.

As a first approach it is assumed that crop characteristics described by leaf area index, root zone depth and root extraction pattern, are spatially invariable. The same assumption applies to the observations of rainfall and potential evapotranspiration leaving the soil characteristics as the only parameters allowed to vary spatially.

Retention properties have been analyzed in the laboratory on undisturbed soil cores removed at five depths, (triplicate sampling). The moisture content - capillary pressure relationship for one of the profiles is shown below in Fig. 5.

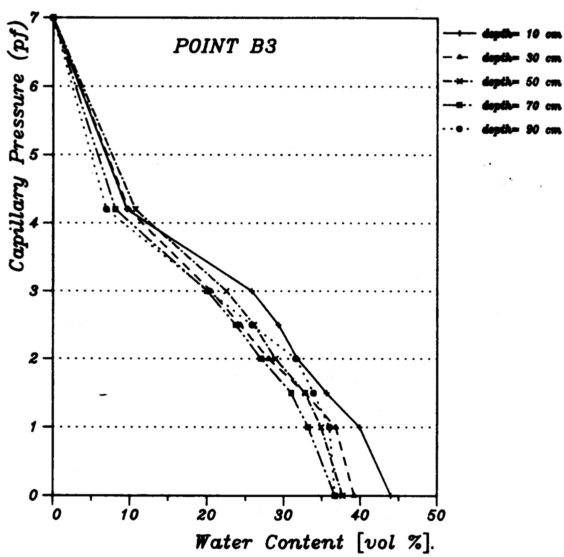


Fig. 5 Moisture content - capillary pressure relationships for B3 profile.

Saturated hydraulic conductivity has been determined on undisturbed soil cores (triplicate sampling) taken from the same levels, whereas no experimental results are available so far on the hydraulic conductivity for lower moisture contents. The measurements have shown a very large degree of variability (see Hansen et al. 1986) even within the triplicates. The variation appears to be of erratic nature, because the measurments are more than other soil physical analyses, very sensitive to variations in pore geometry and in particular to incidentally existing interconnected macropores. It appears that the measuring scale (100 cm^3 cores) is too small to accurately represent the hydraulic conductivity properties on the field scale. Added to these circumstances a possible channelling effect between soil and core ring leaves behind a limited experimental knowledge of the hydraulic conductivity function.

Hence, for establishing a hydraulic conductivity function for each profile the following procedure has been adopted : Assuming that the experimentally determined retention characteristic curve is more representative for the field conditions, the method by Kunze et al. (1968) is used for predicting a saturated hydraulic conductivity value utilizing the information embodied in the retention curve. The saturated conductivity values predicted by this procedure show less erratic variations than the measured values, although they are generally somewhat higher. However, this is a fact which is generally experienced, and a comparison analysis between the two sets of values suggests the application of a general matching factor of 0.02.

In order to establish the complete hydraulic conductivity function, the concept of field capacity is utilized. Field capacity is the moisture content which is approached in a soil after a few days of draining by gravity forces. At this state no significant water flow occurs corresponding to a very low hydraulic conductivity. This value is here assumed to be approximately 0.3 mm per day. The moisture content at field capacity is generally found at a capillary pressure of - 1.0 m.

To describe the hydraulic conductivity over the whole moisture regime a power function of the following type is adopted:

$$K = K_S \left(\frac{\theta}{\theta_S}\right)^n \tag{4}$$

where

K - hydraulic conductivity at a given moisture content

K_S - hydraulic conductivity at saturation

θ - moisture content

θ_S - moisture content at saturation.

n - exponent

The exponent n is calibrated on the basis of the assumptions described above.

In Fig. 6 below is shown, as an example, the adopted hydraulic conductivity functions for the B3 profile.

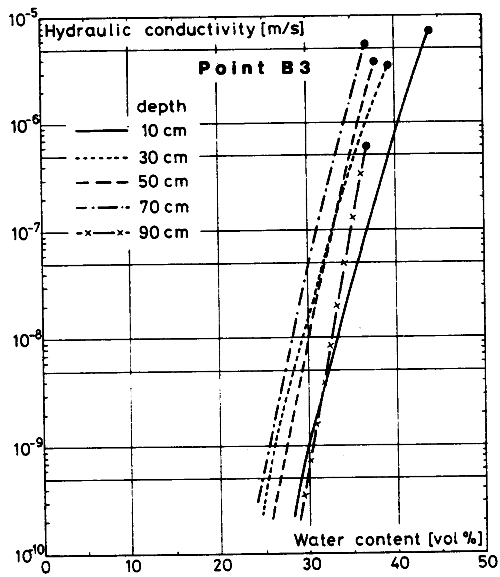


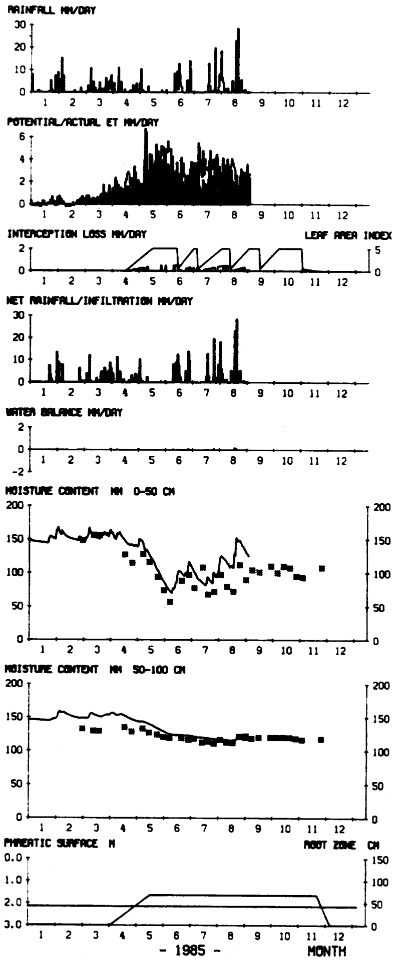
Fig. 6. Hydraulic conductivity - moisture content relationships for B3 profile.

Results and discussion

The simulation results are illustrated in Fig. 7 - 12 below. For profile B3 a detailed presentation of simulation results are given in Fig. 7 - 10. Figs. 11 and 12 illustrate the simulation results for moisture content for two other profiles, B1 and B5.

The simulation efforts presented here are based on the first data which have been processed from the research project. The field measurements of moisture content represent the only verification data available at the moment. Evaluated on the basis of this variable, the simulation results are acceptable as a first approach, although some obvious discrepancies are present for the top layer, where the simulated moisture contents are too high. On the other hand there is a tendency for the simulated values in some of the profiles to be somewhat lower than the measurements in the two next layers. However on an accummu-lated basis the simulation is rather accurate. Hence, this problem may likely be attributed to the root extraction function although the higher inaccuracy of the neutron probe measurements close to the soil surface may also contribute to the discrepancy.

When more data become available this may lead to some refinements of the modelling procedure and hence improved simulation results. In particular a refined root extraction procedure has to be developed. However, it appears that the described modelling approach can lead to an acceptable description of the water flow processes in the individual soil profiles and hence form a basis for establishing a modelling procedure for heterogeneous fields.



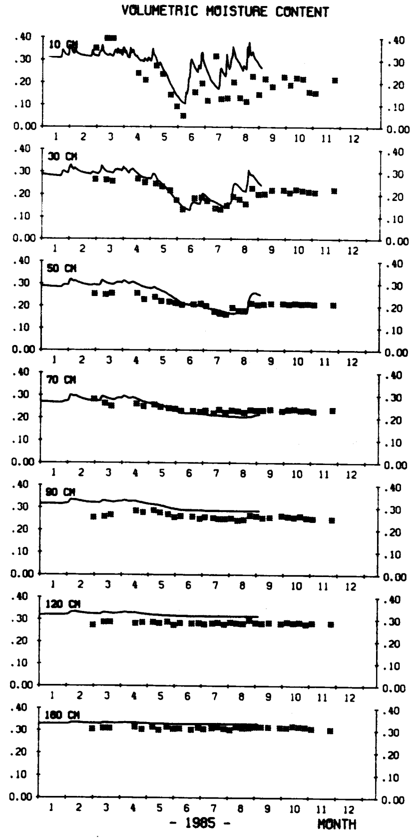


Fig. 7 Profile B3.
Input data, boundary conditions, evapotranspiration and soil moisture storage.
* Measured.
—— Simulated.

Fig. 8 Profile B3.
Measured and simulated moisture contents.

* Measured.
—— Simulated.

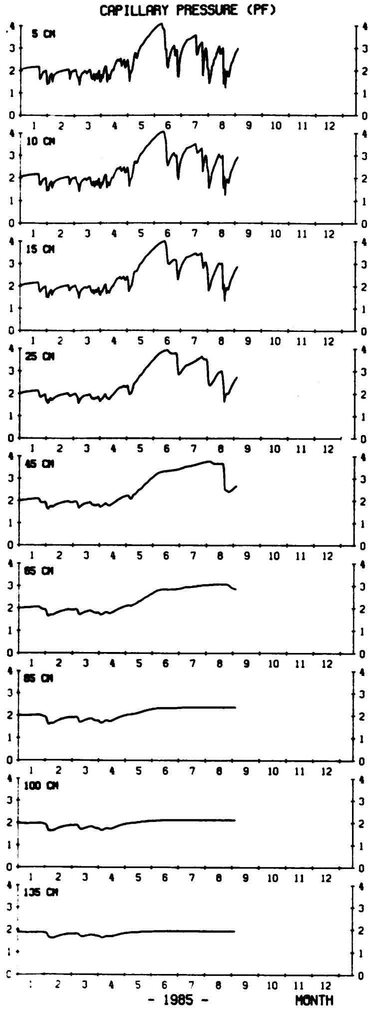


Fig. 9 Profile B3.
Simulated capillary pressure.
(logarithmic values).

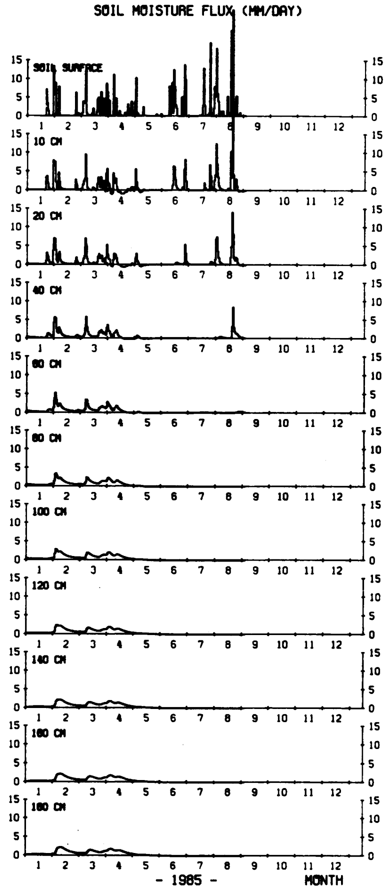
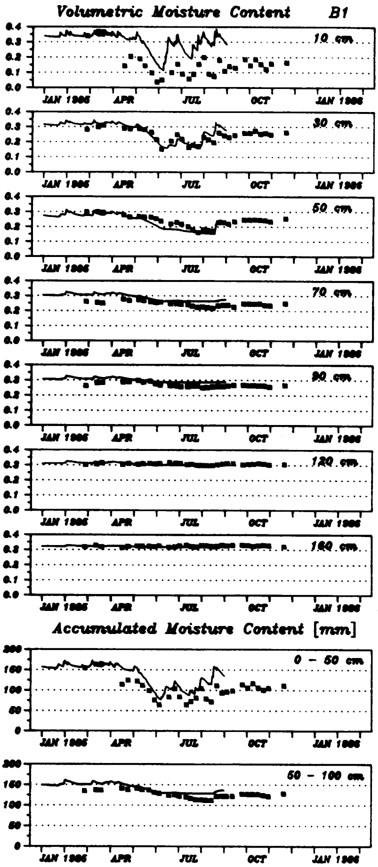


Fig. 10 Profile B3.
Simulated water flow
(Darcy).



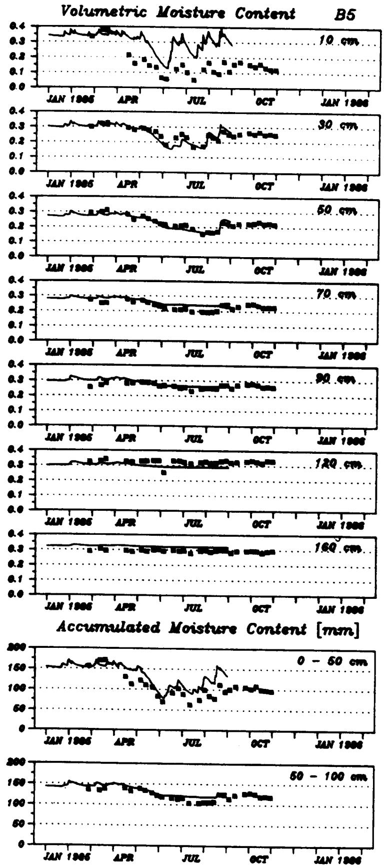


Fig. 11 Profile B1
Measured and simulated
moisture contents.
* Measured.
—— Simulated.

Fig. 12 Profile B5
Measured and simulated
moisture contents.
* Measured.
—— Simulated.

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MODELLING OF UNSATURATED FLOW IN HETEROGENEOUS SOILS

Part II: Application to a field

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Abstract

Spatial variability in moisture content in a plant covered field is partly caused by field variability in soil physical properties. The present analysis is an attempt to explain the observed variations in moisture content from the variations in retention properties. This approach possesses some evident deficiencies close to the soil surface, but provides a very accurate description of the variability deeper in the field at larger depths.

The concept of field effective hydraulic properties is discussed and a preliminary analysis has indicated that this concept may provide a reasonable approximation to the space averaged moisture content despite the highly non-linear phenomenon inherent in soil water dynamics.

Introduction

For analyzing flow and transport processes in the field the most common approach has been to formulate the physical laws in macroscopic quantities and to make the predictions by solving the differential flow equations. It is generally assumed that the hydraulic parameters are uniform over the area of interest together with the boundary and initial conditions. In other words, it is assumed that the field can be regarded as a fictitious homogeneous media characterized by some "equivalent properties" (Freeze, 1975) which are established by sampling over a few locations and by an appropriate averaging procedure.

Several investigations have shown that fields which generally are considered uniform relative to most cultural practices have a significant spatial variation in the hydraulic properties (Nielsen et al., 1973; Russo and Bresler, 1981). Since the flow processes are highly non-linear, the validity of the concept of "equivalent porous medium" is indeed questionable. Bresler and Dagan (1983) have analyzed this approach and concluded that effective properties are meaningful only under very restricted and special flow conditions and cannot in general be justified in spatially variable fields.

A possible framework for approaching some of these problems related to field variability is offered by stochastic modelling. In this approach the soil properties and flow variables are interpreted as stochastic variables from a reasoning that the deterministic variation cannot be known in all details and therefore is subject to uncertainty. The variables are consequently defined in terms of their statistical moments and the actual field is interpreted as a realization of the ensemble of fields which all have the same properties at the sampling points as the given field.

Various techniques have been used for the stochastic analysis of the heterogeneity problem ranging from spectral methods (Yeh et al., 1985) and numerical methods (Smith and Schwartz, 1980) to a more simple statistical averaging procedure (Dagan and Bresler, 1983; Bresler and Dagan, 1983).

The approach adopted here is a statistical averaging procedure, where the statistical moments of the flow variables are predicted on the basis of the moments of the soil properties for given boundary conditions. The data are obtained from the research field described by Hansen et al. (1986).

Statistical analysis of soil properties

The two basic relationships which determine unsaturated flow processes are the retention function and the hydraulic conductivity function. As argued in Part I the measurements of the retention properties are considered to be more representative for the field conditions than the measurements of saturated hydraulic conductivity. Hence, in the following analysis of field variability, the effect of soil heterogeneity is evaluated on the basis of the retention function alone, and the hydraulic conductivity properties will be derived from this function following the method described in Part I of this paper.

The retention properties are related to the pore size distribution, which is correlated partially to the textural composition of the soil. In Figs. 1 and 2 below, the spatial variation within the field site of these properties are shown as frequency distribution. Two horizons of different characteristics have been recognized and analyzed separately.

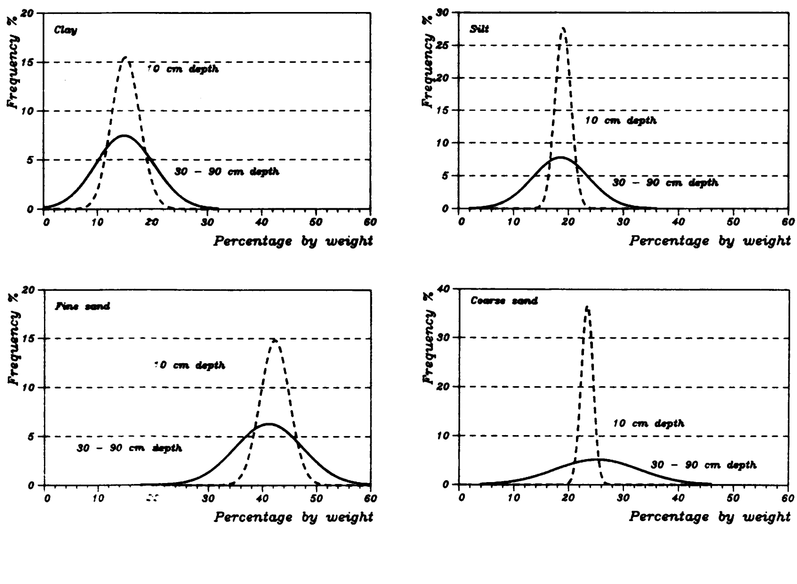
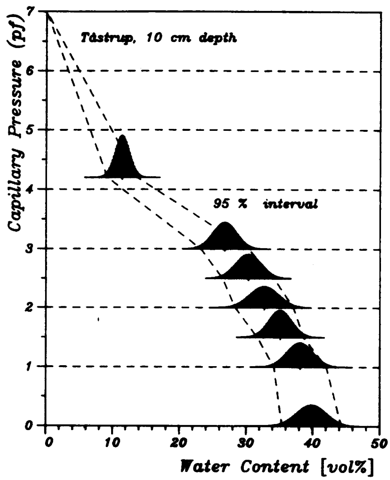


Fig. 1 Spatial variability of the textural composition of the soil within the field site.



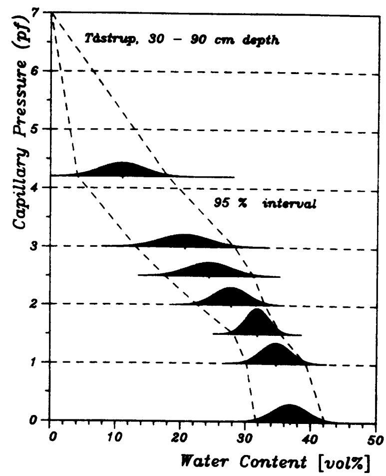


Fig. 2 Spatial variability of moisture retention properties within the field site.

It is evident from the figures that the soil properties are more homogeneous in the upper soil horizon than in the lower. It should be emphasized that the larger variability is a result of a larger horizontal variability and not caused by the larger depth interval.

Statistical analysis of moisture content

The moisture content has been measured regularly in the field in order to analyze the spatial variability. Fig. 3 below illustrates the horizontal variability with time for various depths. The full line represents the spatial average and the vertical bars represent twice the standard deviation. For a normal distribution of the moisture contents these intervals represent 95% of the probability mass. As indicated by the figure the horizontal variability is largest in the top layer, somewhat smaller at 30 cm depth after which the range of variation is more or less of the same magnitude. The figure also indicates that there is a minor seasonal variation in the variability pattern. For the upper layer, the variability is less pronounced in late winter/ early spring, but afterwards it increases and remains more or less of the same magnitude for the remaining part of the investigated period. In general the variation is larger here than at the deeper levels. At 30 cm the variability has diminished somewhat and from 50 cm and downwards no marked differences in the range of variation are observed. Also the seasonal variation seems less pronounced here.

Model analysis of the spatial variability in moisture content

The following analysis is a first-order approximation to predicting the statistical moments of the soil moisture contents on the basis of the moments of the soil properties. In many applications the average and the standard deviation in moisture content, reflecting the distribution in the horizontals, constitute the required information rather than a detailed description of the variable in space and time.

The present analysis will be based on a number of assumptions, which is discussed below. In stochastic modelling of soil water flow it is generally assumed that statistical stationarity and the ergodic hypothesis apply which means that ensemble averages and space averages can be interchanged. Hence, the expectation of a variable represents the average over the field and similarly will the standard variation represent the variability over the field.

Another basic simplification concerns the flow direction, which here is assumed strictly vertical. This further leads to a simplified representation of the soil geometry where the medium is represented by an ensemble of vertical soil columns with statistically independent hydraulic properties. Each of the columns represents possible field conditions giving no considerations to the spatial structure.

The randomness of the soil properties is introduced through the stochastic nature of the retention curves, Fig. 2. The cumulative probability functions of the water content for the individual suctions (pF = 0, 1.0, 1.5, 2.0, 2.5, 3.0, 4.2) are divided into N equal classes. As a first-order approach it is assumed that full correlation exists between water contents at the different suctions which can justify a linkage between similar probability fractiles. This procedure leads to N retention curves, each representing a certain soil

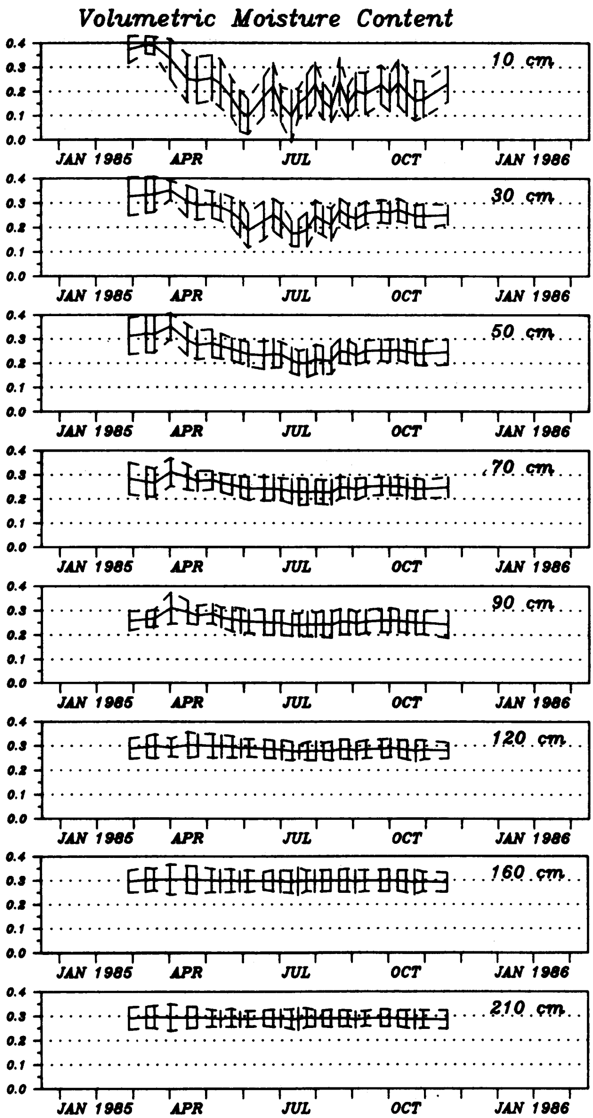


Fig. 3 Horizontal variability in measured moisture content with time for various depths.
Full line: Spatial average
Dashed lines: ± 2 × standard deviation.

class of the field with hydraulic characteristics as embodied in the retention curve. For a given retention curve the hydraulic conductivity function is derived as described in Part I.

The statistical analysis of the spatial variability in water retention characteristics has identified two horizons with different properties, cf. Fig. 2. Hence, the procedure described above will be carried out for both horizons. In defining the N different soil columns a retention curve has to be affiliated to each soil horizon. Preliminary analyses have shown that there is a rather weak correlation between soil properties of the upper and lower horizons implying that high retention properties in the upper horizon may well coexist together with low retention properties in the lower horizon and vice versa. However, as a first and preliminary approach it is nevertheless assumed that the soil properties are having the same nature throughout the profile.

Having established N statistically independent soil profiles each representing possible field conditions, the model described in Part I will be used for simulating the flow conditions in each of the columns.

Results and discussion

The simulation results are presented in Fig. 4 in terms of space average and plus/minus twice the standard deviation for each of the layers. In the same figure are shown the corresponding measured values.

The simulation of the space averaged moisture contents shows some of the same discrepancies in the root zone as experienced for the individual sampling profiles. Some of these discrepancies may be explained by an insufficient distribution of the root extraction. Assessed from the measured moisture contents, the extraction is too small from the upper layer in the mid growing season, whereas the withdrawal appears somewhat too high from the next two layers. However, some of the discrepancies for the upper layer may also be referred to an inaccurate calibration of the neutron probe close to the surface. The simulations of the spatially averaged moisture conditions in the lower depths are very accurate.

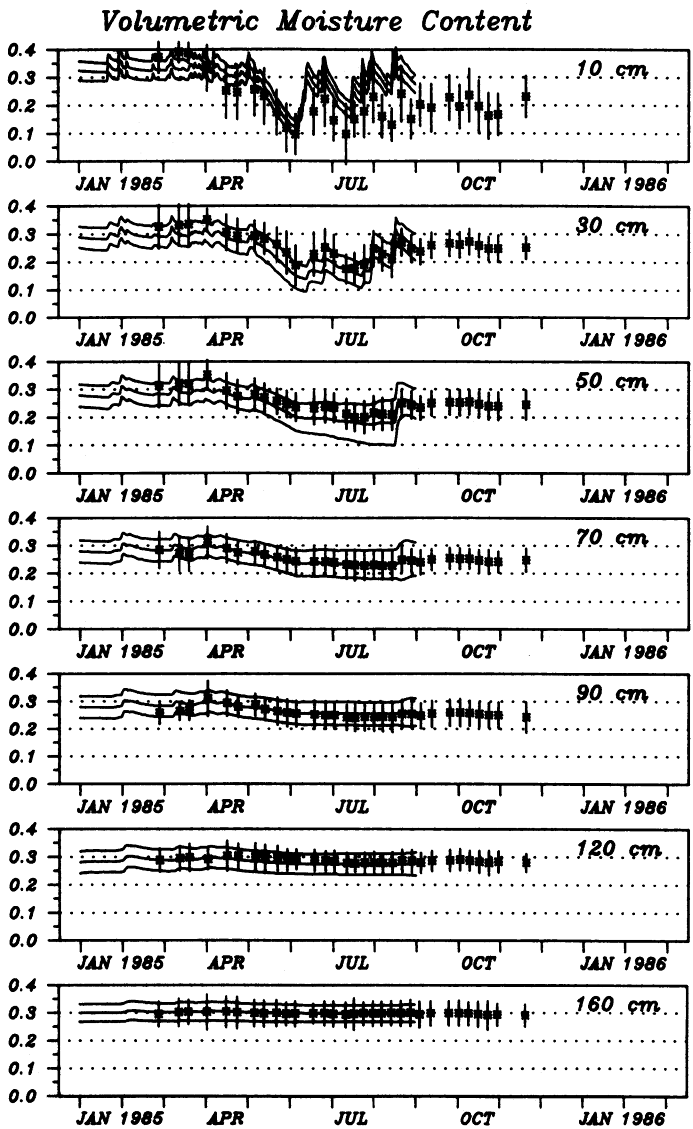


Fig. 4 Measured and simulated variability in moisture contents.
ǂ Measured range of variation.
≡ Simulated range of variation.

With respect to the range of variation an obvious deficiency exists for the upper layer, where the simulated range is much too small in comparison with the measurements. The narrow simulated range is obviously a result of the small variation in retention properties, cf. Fig. 2. Hence, the spatial variation in moisture content cannot alone be explained by the variation in soil physical properties, using the retention curves as the basis. Other factors are apparently affecting the variability in this horizon. This may well be a soil physical property which is not described in details by the retention curve such as the hydraulic conductivity property; however, in addition, plant related variables and their variability like the density of the ground cover, the green active material, albedo conditions and root growth will certainly also contribute to the observed variations.

The possible influence from plant related variables is supported by the reasonable accordance between observed and simulated ranges of variation for 30 and 50 cm depths and the excellent agreement for the deeper layers, where the plant effect is gradually diminished.

A further analysis of the simulation results also showed that the simulation based on the average retention curve corresponds very closely with the spatially averaged values. This is surprising, in view of the highly non-linear phenomena inherent in soil water dynamics and in view of the investigations reported by e.g. Bresler and Dagan (1983), which oppose the existence of the concept of "effective soil properties". A reason for this apparent disagreement may be referred to the very different flow conditions being considered. Bresler and Dagan (1983) consider short-term, rather abrupt flow events, whereas the present investigation analyses naturally occurring series of unsaturated flow. Under such conditions the non-linear effects may tend to balance out implying that "effective soil properties" may constitute a reasonable concept.

Conclusions

The study has shown that soil variability has a large influence on water and flow conditions in the field, and it is consequently of importance to account for this phenomenon in distributed catchment modelling.

The present analysis is a first approach to the problem of simulating the flow conditions in heterogeneous fields. The approach adopted, i.e. predicting the spatial variability basically on the basis of the variability in the retention properties, can explain nearly all the variation outside the sphere of root influence. Closer to the surface, however, only part of the variation can be explained indicating that other variables such as plant related properties have some effect. Such factors are obvious to include in future more refined analyses.

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SUMMARY OF MODELLING SESSION BY LARS GOTTSCHALK

During the seminar we have got an excellent exposé over all types of modells and how they can utilize and/or represent spatial variability in soil physics and soil moisture content. We started the first day of the seminar with two presentations from approaches used in Great Britain for the calculation of integrated basin response to precipitation using statistical and simple conceptual models. It was demonstrated that rather coarse classification of soil types in a regional scale can be useful in this case. The sizes of basins can, however, not be too small so that local factors can give dominant influence for instance in the form of springs.

This morning we got an illustration of the difficulties in modelling soil moisture variations with a conceptual box model, that does not take into consideration spatial variability in soil properties. We can thus not expect that a box representing basin soil moisture will behave as individual soil moisture observations. The large variability in such observation does not allow us to define representative observation sites. Another problem that was touched upon was the fact that good agreement with soil moisture variations at a certain site was only accounted for when these observations were explicitely used for calibration. Good fit to soil moisture variation gave a worse fit to runoff observation and vice versa if the runoff observations were used for calibration.

The problem with present conceptual box models that each box is a general concept for a large part of the basin that is not easily characterized. New concepts are developed that are directly identifiable in the basin. The second presentation this morning gave a good illustration of the complex variation patterns of soil moisture within a small catchment but also promising approach, the source area concept.

We then got an extensive presentation of the application of a deterministic physical model of hillslope flow to conditions in West Germany and in Norway. The behaviour at the model in the first case was excellent due to very homogeneous conditions. In the second case the heterogeneity of the Norwegian morain terrain caused problems, mainly numerical. The work in Norway has just started and will continue.

In cases of large heterogeneity deterministic approaches thus show limitations. The following presentations during the morning session more explicitely took into consideration spatial variability. In one case by presenting the possible range of variations in soil moisture properties caused by variations in soil physics and in the other two by applying the theory of stochastic dynamic modelling. In the latter two presentations the type of statistics needed to describe variation patterns of stochastic variables in space was thoroughly treated. In these approaches the range of variability of soil moisture, water flow etc. can be derived analytically or by Monte Carlo methods from the statistical description of soil physical parameters. We need of course, very large statistical samples to determine these properly.

The presentation thus included statistical models, conceptual models, deterministic physical models as well as probabilistic - dynamic physical models. We got at the same time good examples of the problems and limitations of the different types of models.

We can formulate our problem as to parametrize physical phenomena to a proper scale. In the deterministic approach we only use one representative value (usually the mean) to characterize a more or less complex variation pattern. In case of probabilistic - dynamic models the variation patterns are more properly described.

Our knowledge of physical processes are basically founded in a microscale. This fact has not stopped us to apply the same theoretical models to meso - and macroscale problems, with at least in most cases acceptable results. The physical interpretation of parameters is however, partly lost when we without modifications apply models to large scale phenomena. With the stochastic - dynamic approach it is possible to in a formally correct way integrate the models in space coordinates. It leads to very difficult theoretical problems, which scientific journals to day are filled up with.

It is not necessary so, that the answer to our problems is to find a characteristic macroscale hydraulic-conductivity or a representative pF-curve. May be there are some macroscale parameters that in a better way characterize the phenomena in the actual scale.

To talk about models in general without giving the purpose is rather meaningless. A model is a tool to solve a certain problem not an aim as such. We can identify the following four purposes for model applications.

- Integrated basin response for hydrologic design and forecasting.
- Determination of available water for vegetation.
- Determination of flow paths of water through a catchment or parts of it.
- Scientific analysis and discription.

As long as we deal with water quantity there is already a vast experience that statistical models as well as more or less complex conceptual box models are very useful for the first mentioned purpose. The models can of course be further improved mainly by finding the proper macroscale characteristics that describe the response, to make the concepts identifiable in the field.

In determining available water for vegetation the spatial variation patterns become more important. It is not enough to characterize the physical properties of soils within a field by some representative values. The whole range of possible values is needed.

Flow paths are even more sensitive to heterogeneities in soil physical properties. We need thorough descriptions of spatial variation patterns to determine structures in the flow path network. Knowing the flow paths we can then turn to the problem of hydrochemical modelling.

What the scientific analysis and description concerns it must be a fundamental task to map and describe the spatial scale of hydrologic processes of water movement in relation to the scale of variations in soil properties and landscape formation.

SUMMARY AND CONCLUSIONS

Sylvi Haldorsen and Einar Berntsen

SOIL MAPS AND INTEGRATED HYDROLOGICAL MODELS

Soil mapping is a way to present large scale spatial variability of the soil itself. Some data which are important for soil water conditions are usually included in soil mapping, or they may be derived from these maps. Most commonly, the soil maps give information about permeability, pore-size distribution and drainability, and may also include information about the groundwater level. The application of such data in integrated basin modelling was demonstrated in two lectures given by representatives from the Soil Survey of England & Wales and Institute of Hydrology, Wallingford, England. England & Wales are covered by soil maps in scale 1:250 000. The maps include data on permeability and depth to impermeable layer. Based on this soil classification, "winter rain acceptance potential maps" have been worked out for predicting flood events, and the data have also been used for predicting low flow events. The scale and classification system seem to work well for the modelling. The method is also applied in several other countries in the EEC.

Denmark is covered by regional soil maps in scale 1:50 000. Data on soil texture, content of organic material, and calcium carbonate contents are stored in data banks. Calculation of root zone capacities is based on these data. The data have been applied to urban land use planning, water planning, nitrate-runoff evaluation, acidification and erosion risk evaluations and drainage requirements.

In Norway a few soil maps in scale 1:5000 and 1:10 000 have been worked out in agricultural areas. The aim is to cover the arable lands (approximately 10 000 km^2) with soil maps during the next 20-25 years. It is not planned to cover forested areas with soil maps. The maps are based on geological genesis, depth to bedrock, texture, drainage and decomposition of organic components. It is too early to tell to what extent the maps may be applied for hydrological modelling. The maps are, however, detailed enough to be applicable for rather detailed estimates of spatial variability.

In Sweden there are, so far no systematic soil map series. No information was presented about soil mapping in Finland at the seminar. In Norway and Sweden maps of the Quaternary geology in scale 1:50 000 are far more abundant than soil maps. Some data do exist on the variability of soil-water relevant parameters in soil types of different geological genesis. Such data might be combined with geological maps for use in large-scale hydrological models.

One important factor which limits the applicability of soil maps in Scandinavia is that the spring flood here is so closely connected with the snow melt and is rather independent of the soil properties. For low flow estimates during the summer, on the other hand, the soil properties and soil variability are the controlling factors and soil maps may be of importance. For more detailed hydrological models where water quality is involved, the present soil maps are probably not detailed enough to satisfy the need for data on spatial variability, and more detailed observational programmes must be carried out.

SOIL PROFILE STUDIES

Colour, content of humus and type and degree of podsolization may be important parameters for estimating spatial soil water variability. These are information collected in connection with most soil studies. Combined with measurements of soil water conditions by neutronmeter gauges, pF-measurements and density measurements the variability in space and time might be well described. So far, hydrologists have paid little attention to soil profile data. A further advance is dependent on cooperation between soil scientists and hydrologists.

DETAILED STUDIES OF SOIL PARAMETERS

Forested areas - till

Till is probably the most variable sediment type in Scandinavia. The combination of micro- and macro-pore drainage, the high content of cobbles and boulders and the great spatial variability makes it extremely difficult to develop models that can handle the soil water movement in tills in detail. There are data for the variability of, for instance, the saturated hydraulic conductivity and infiltration capacity, while data for unsaturated hydraulic conductivity are almost lacking. Some studies of micro- and macro-pore drainage were presented at the seminar, but we still lack a general description of the unsaturated water movement in tills. Some detailed descriptions of the soil water variability in vertical profiles were given, but we do not know if these are representative for all till types.

In addition to the presentations at the seminar, there are several previous studies of the hydrological properties of tills, although most of them consider saturated conditions only. As tills play such a dominant role in Norden, it was proposed that a separate workshop be arranged on soil water variability in tills during the last part of 1987. It was recommended not to start any new comprehensive studies on tills before we have tried to systematize existing data.

Cultivated land

The studies presented from cultivated land dealt both with the variability due to different soil types and the variability affected by packing and artificial drainage. It was shows that the content of organic material is the factor of greatest importance for the plant available water capacity. Easily available water increases with increasing content of silt and decreases with increasing clay content. It was also concluded that sampling along lines with a 1000 m distance between plots may give about the same variability

for the root zone capacity as sampling with 200 m distance. More studies similar to those of Riley and Samuelsen are needed to make sampling and laboratory studies as efficient as possible.
It was clearly demon strated how the cultivation methods may affect soil water capacity and movement patterns, hence affecting the spatial soil water variability. The effect of cultivation may, in some cases, be more important than the original variability of the soil.

HYDROLOGICAL MODELLING

Several presentations at the seminar illustrated the difficulties in modelling soil moisture variations with a conceptual box model that does not take into consideration the spatial variability of soil properties (Andersson 1986). A box representing basin soil moisture will not behave like individual soil observations. A source area concept, on the other hand, gave a good illustration of the complex variation patterns of soil moisture within a small catchment. An extensive deterministic physical model of hillslope flow behaved reasonably satisfactorily in very homogeneous areas, while serious problems were connected with the same model in a heterogeneous area.

The last part of the seminar was focused explicitly on spatial variability and how to model it. The application of a physically based model to study the sensitivity for different physical soil factors was demonstrated. Sensitivity tests of this kind are very important, and it is hoped that future work will include sensitivity studies. Types of statistics most important for describing spatial variability. were discussed. Semivariograms and kriging in soil water studies will probably prove to be important tools in many detailed models. Stochastic models otherwise seem to have very limited application in soil water studies in Norden today. It is hoped that the seminar will stimulate further development of such models.

CONCLUSIONS

Hydrological models should serve specific purposes and are not an aim in themselves. One of the main problems regarding spatial variability of soil water in models is to define the proper scale. For rough integrated basin analyses, the spatial variability within a distance of 100 m is without importance. For more detailed models this variability may be quite critical. Even in cases where we may define the scale, we may lack methods which would enable us to measure the critical parameters in the field. That is clearly demonstrated for till. The choice of parameters is also important. In a great number of studies the macroscale variability of the hydraulic conductivity is most important, but very difficult to measure. Often, other parameters can better describe the actual soil water variability. These questions will have to be focused on in future work.

Soil scientists probably know more about the soil water variability at macro- and microscales than most hydrologists. On the other hand, the hydrologists know more about the sensitivity of models and what is realistic to model. We believe that cooperation between soil scientists and hydrologists will show that for many purposes our present knowledge about spatial variability is already at a level where new, national data series are not needed. In cases where new data is needed, co-operation may help to work out a realistic field programme. It is our belief that co-operation across professional boundaries will be of great importance for successful modelling of spatial variability, and we hope that the seminar has been an inspiration in that respect.

Appendix 1

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