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Distribution of soil moisture and groundwater levels at patch and catchment scales

S. Beldring[a,*], L. Gottschalk[b], J. Seibert[c], L.M. Tallaksen[b]

[a]*Department of Hydrology, Norwegian Water Resources and Energy Administration, P.O. Box 5091 Majorstua, N-0301 Oslo, Norway*
[b]*Department of Geophysics, University of Oslo, P.O. Box 1022 Blindern, N-0315 Oslo, Norway*
[c]*Department of Earth Sciences, Uppsala University, Norbyvägen 18B, S-752 36 Uppsala, Sweden*

Abstract

This study is a contribution to the northern hemisphere climate processes land surface experiment (NOPEX). Its purpose is to investigate the spatial variability of groundwater levels and soil moisture content at different scales in a landscape dominated by boreal forest and till soils, which is characteristic of the Nordic countries. The analysis of data from the NOPEX area are based on a review of previous studies on the spatial distribution of these state variables and their significance for runoff formation. Soil moisture content in the unsaturated zone and depth to the groundwater table show characteristic patterns which are related to the landscape elements (patches) of the drainage basins. Similar behaviour is observed in different parts of the NOPEX region. The variability of average values between areas decreases to a minimum for catchments with size larger than 1 km^2. It can therefore be concluded that the main part of the spatial variability of soil moisture content and depth to the groundwater level in the till soils of the NOPEX area is found within small drainage basins. Based on a physical description of the soil, distribution functions of soil moisture content conditioned on the depth to the groundwater table have been developed, both for the patch scale and the catchment scale.

Keywords: Till; Groundwater; Unsaturated zone; Spatial distribution

1. Introduction

This work is a contribution to the northern hemisphere climate processes land surface experiment (NOPEX). The purpose of NOPEX is to study land–surface–atmosphere interactions at local and regional scales in a Northern European landscape dominated by boreal forest and crystalline rocks and affected by a number of glaciations. Since the dynamics and water balance of land surface hydrology have profound influence on these interactions, one of the objectives of NOPEX is to investigate the relationship between hydrological state variables, in particular soil moisture and groundwater storage, and landscape characteristics at different spatial scales (Halldin et al., 1999).

The hydrological response of a drainage basin is determined by small- and mesoscale variations in its geology, topography, vegetation and other landscape characteristics. The independent input variable precipitation is transformed in the hydrological system

* Corresponding author. Fax: +47-22-95-92-01.
E-mail address: stein.beldring@nve.no (S. Beldring).

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into the dependent output variables evapotranspiration and runoff and changes in the soil moisture and ground water storage of the system. In order to develop physically based models of hydrological processes the large variability in system properties and their influence on the water balance components must be considered. This requires a subdivision of the landscape into hydrological units or patches with characteristic time and space scales with regard to runoff formation, i.e. different water pathways and travel times (O'Loughlin, 1981). A patch is any area of the landscape that has broadly similar hydrological responses in terms of the quantities of interest, in this instance, runoff production and evapotranspiration (Beven, 1995). The spatial variation of these patches or source areas of runoff and their impact on moisture conditions and runoff in catchments at different scales is a major problem in hydrological research.

The main surface deposits in the Nordic countries are till soils. Till is a non-sorted sediment deposited by glacier ice and composed by a variety of clasts of a wide range of sizes (Haldorsen and Krüger, 1990). A characteristic feature of the boreal landscape dominated by till soils is a thin soil cover with a saturated hydraulic conductivity which decreases with depth (Lind and Lundin, 1990). As a consequence, the groundwater table largely follows topography, with a depth of a few metres below ground surface in elevated areas. In the lower parts of certain hillslopes and in some low areas, the groundwater table is at or above the ground surface (Lundin, 1982; Rodhe, 1987). The depth to the groundwater table controls the process of runoff formation in these areas, both during stormflow and recession. When the depth is small, the large hydraulic conductivity of the surface near soil layers compared to the deeper layers leads to larger streamflow contribution from hillslopes (Grip and Rodhe, 1994). Several studies on runoff generation dynamics in humid temperate environments have emphasized the domination of groundwater in runoff generated by storms or snowmelt (Sklash and Farvolden, 1979; Rodhe, 1981, 1987; Espeby, 1990; Sklash, 1990; Bonell, 1993). (Nyberg, 1995, 1996) found that the runoff from a small headwater catchment increased drastically when the depth to the groundwater table diminished.

Soil moisture conditions are closely related to groundwater conditions. Between precipitation events, the soil moisture content will have a maximum value at a state of hydrostatic equilibrium which is determined by the physical characteristics of the soil and the depth to the groundwater table. The soil moisture deficit which develops due to evapotranspiration will have to be filled up before the groundwater table rises. In those parts of a catchment which are located close to the stream, parts of the storm runoff may be generated as saturation excess overland flow due to a combined effect of small soil moisture deficit prior to the storm and a rise in the groundwater table during the storm (Anderson and Burt, 1990).

In summary, the various delivery mechanisms of catchment runoff in a landscape dominated by till soils depend on soil moisture conditions and the depth to the groundwater table. They are the main hydrological variables that can be used to distinguish between different hydrological response units, i.e. patches in the landscape mosaic which produce 'similar' hydrological response on snowmelt or precipitation. Flügel (1995) defines hydrological response units as distributed, heterogeneously structured entities having a common climate, land use and underlying pedological, topological and geological associations controlling their hydrological dynamics. The variation of hydrological process dynamics within one type of hydrological response unit is small compared to the dynamics in a different type of unit. Topography has a significant effect on subsurface flow and moisture conditions in a catchment and it can therefore be used to distinguish between hydrological response units (Beven and Kirkby, 1979; Beven and Wood, 1983; O'Loughlin, 1986; Bonell, 1993; Thompson and Moore, 1996). In a study which showed that spatial response units in a landscape dominated by till soils can be identified with the help of geomorphology, Krasovskaia (1985) used a classification of catchment topography based on Hack and Goodlett (1960) to differentiate between three types of units in a catchment: (1) nose, the driest part, including the ridge crest and the nearby slopes where the contours are convex outward; (2) hollow, the central part of the basin along the stream with favourable moisture conditions, an area in which the contours are concave outward; (3) slope, the zone between nose and hollow with transitional moisture conditions where the contours are straight or nearly so.

The scaling problem (aggregation/disaggregation problem) is one of the key issues for the NOPEX project. A critical question is whether it is possible to generalize information about hydrological behaviour from small catchments to large catchments (regions) or not. Large scale hydrological models with calibrated effective parameters are frequently used, although non-linearities and structural heterogeneity in natural catchments make it unlikely that the equations of hydrological theories developed at small space and time scales can be applied to large scale problems (Beven, 1995). An alternative approach is the use of spatial distribution functions which assume that the frequency of occurrence of state variables or parameters are more important than the actual pattern of the parts within the catchment. A critical problem in the application of these models is to determine at what scale the description of actual patterns of heterogeneity in catchment properties can be replaced by their distribution functions. An attempt to define this scale is the representative elementary area (REA) concept introduced by (Wood et al., 1988, 1990). It implies that (1) for a range of spatial scales the pattern of landscape elements is no longer significant for the formation of runoff, only their relative proportion as given by a spatial distribution function matters; (2) the similarity in catchment behaviour within a region due to similar climate, soil, topography and other factors allows the same parameterization of distribution functions to be used for different catchments within this range of spatial scales.

Several studies of the REA concept have focused on runoff, since it represents the integrated effect of hydrological processes within a catchment. (Wood et al., 1988, 1990) studied the effects of spatially variable topography, precipitation and soil characteristics using synthetic data and a hydrological model. They found that the variability in runoff between catchments of similar size decreased with increasing area and stabilized at a minimum at above 1 km^2. Their conclusion was that the runoff generation population appears to be stationary and that the size of the REA is determined by topography, while the spatial variability of rainfall and soil characteristics only have a secondary role. In a similar study, Blöschl et al. (1995) concluded that the size of the REA is governed by storm duration and correlation length of precipitation. The main difference between the studies is that (Wood et al., 1988, 1990) compared catchments of similar size irrespective of their relative position, whereas Blöschl et al. (1995) considered the runoff from nested catchments of variable size around a fixed point. Woods et al. (1995) used distributed measurements of runoff in river basins to study the effect of scale. They found that for areas smaller than 1 km^2, the variance of specific discharge between catchments of similar size decreased faster with increasing area than would be expected for a random sample. At larger scales the variance decreased in a manner consistent with sampling from a stationary random field. These results were taken as an evidence of organization in catchment behaviour, supporting the idea of a REA. In an investigation of the scaling behaviour of soil moisture, Wood (1995) concluded that the size of the REA may be as large as 5–10 km^2.

The purpose of this work is to investigate the spatial variation of soil moisture in the unsaturated zone and the depth to the groundwater table at different scales in the boreal forest landscape in the NOPEX region. In order to achieve a realistic description of land-surface hydrological processes, it is necessary to take into account small- and mesoscale variations in land surface properties important for runoff formation. This can be achieved by a regional scale hydrological model which is based on integrating the contributions from several sub-catchments or small scale elements. An investigation of the spatial distribution of the state variables which determine the hydrological response of the landscape is an initial step in the development of such a model.

2. Study area

The NOPEX region is an area of $\approx 80 \times 100$ km^2 in the southern part of the boreal forest zone, north of Stockholm, the capital of Sweden. The landscape is dominated topographically and morphologically by the very flat sub-Cambrian peneplane. Altitude differences are small, with the main part of the area confined between 30 and 70 m above the sea level, and extreme values at 1 and 131 m. The geology is characterized by granite, sedimentary gneiss and leptite. Lakes and bogs of different sizes are numerous. Tills are the dominating soil type in the research area, however, fine-grained clay soils together with areas of sandy

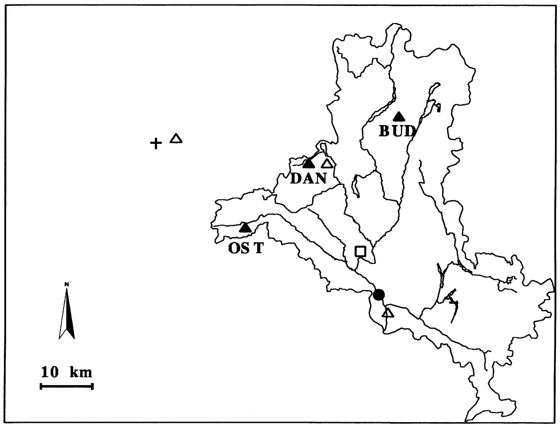

Fig. 1. The NOPEX region with the drainage basin of River Fyrisån (1982 km^2). The triangles show the location of the experimental sites in areas with till. The three experimental catchments Buddby (BUD), Dansarhällarna (DAN) and Östfora (OST) are marked with filled triangles. The esker site (sand and gravel) is represented by a cross and the agricultural site (clay) by a square. The climate station Uppsala is shown by a filled circle.

and silty material are common in the southern parts. Areas with till soils and bogs occupy more and more space towards the northern part. The area is covered by coniferous forest (spruce and pine) with a small fraction of deciduous trees. Towards the south, the forest is gradually substituted by agricultural fields (Lundin and Halldin, 1994).

The corrected mean annual precipitation at the station Uppsala in the southern part of the NOPEX area (Fig. 1) is 636 mm. The precipitation increases towards the north, with corrected mean annual values above 700 mm. Maximum precipitation occurs in August and minimum in February and March. Precipitation in the form of snow constitutes 20–30% of the total annual precipitation. Mean annual temperature for Uppsala is +6°C with maximum monthly mean +17°C in July and minimum −5°C in February (Seibert, 1994). The river flow regime of the region is characterized by spring snowmelt and autumn rain high flows and winter and summer low flows (Gottschalk et al., 1979). The flow regime is, however, unstable and the seasonal pattern can differ from this average pattern during individual years (Krasovskaia, 1995).

The data used in this study were collected at eight experimental sites in different parts of the NOPEX region in the period 1994–1996. Three of these sites

Table 1
Characteristics of the experimental catchments (Lundin and Halldin, 1994; Sulebak, 1997)

Basin	Area (km^2)	Dominating land use	Elevation (m asl)
Buddby	0.5	Forest (27 % swampy forest)	33–45
Dansarhällarna	0.9	Forest (10 % bog)	43–53
Östfora	0.45	Forest (30 % bog)	75–103

are small experimental catchments where detailed observations of soil moisture conditions and groundwater levels were performed along hillslopes. Tables 1 and 2 summarize some of the characteristics of these three catchments. The Östfora catchment is not included in Table 2 since a digital elevation model was not available. At the other five sites, soil moisture and groundwater data were collected at a limited number of locations. The bedrock of the experimental sites consists of massive or folded granitoids of precambrian age which are generally unweathered. The regional hydraulic conductivity of the bedrock is of order 10^{-7} m/s, which is considerably lower than in the soils above, where the bulk volume of groundwater flow takes place (Geological Survey of Sweden, 1982a, b, 1983a, b).

The three experimental catchments and three of the other sites are located in areas covered mainly by sandy till with low or medium boulder frequency. The vegetation is dominated by coniferous forest, mostly spruce. The soil thickness in the investigated catchments varies from several metres in the valley bottoms to nearly zero on the top of ridges where exposed bedrock is often found. The two remaining sites are located on an esker (glaciofluvial deposit of sand and gravel) and in an agricultural area with postglacial clay. These two experimental sites differ from the other sites with respect to soil type, soil depth, topography and vegetation (Geological Survey of Sweden,

Table 2
Distribution of topographic units in the experimental catchments Buddby and Dansarhällarna based on a digital elevation model with horizontal resolution 5 m (Sulebak, 1997)

Basin	Nose (%)	Slope (%)	Hollow (%)
Buddby	31.7	18.8	49.5
Dansarhällarna	38.5	11.5	50.0

1982c, 1989, 1991, 1993). Fig. 1 shows soil type and location of the experimental sites within the NOPEX region.

3. Data collection

In each of the three experimental catchments, ground water and soil moisture data were collected along transect lines extending from the nose profile in the upper parts of a slope to the hollow profile at the bottom of the slope, with the exception of transect No. 2 in the Dansarhällarna catchment where only depths to the groundwater table were measured. The number of transects were two in the Buddby and Dansarhällarna catchments and one in the Östfora catchment. Table 3 gives a description of the data collection in each experimental catchment. In order to have a soil depth sufficient to perform the measurements, the transect lines did not cover the entire hillslope from the stream to the ridge, nevertheless all three topographic units; nose, slope and hollow were represented in each transect.

Groundwater table depths in the three experimental catchments were observed in piezometers with diameter 2.5 cm located at irregular intervals ranging between 1 m and 30 m along the transect lines. The piezometers were open at the base and slotted at intervals of 5 cm along the lower 60–80 cm. They were installed by manual drilling to the maximum depth possible in the compact till soil, varying between 0.6 and 1.6 m with average depth 1 m.

Soil moisture was measured using the principle of time-domain-reflectometry (TDR), which is based on a unique relation between the volumetric water content and the dielectric constant for mineral soils. Calibration curves for soil moisture measurements with the TDR-method which are not restricted to specific soil conditions have been established by different investigators, e.g. Topp et al. (1980) and Roth et al. (1990). However, the TDR-method suffers from uncertainties due to variations in dielectric properties for soils with changing content of organic matter (Gottschalk et al., 1995). Tallaksen and Erichsen (1995) investigated the influence of organic matter on calibration curves for the till soils of the experimental basins in this study by comparison with water content determined from soil samples by the gravimetric method. No systematic variations were found. The soil moisture data used in this study have been calculated from TDR-measurements with calibration curves given by Tallaksen and Erichsen (1995) which agrees with the results of Topp et al. (1980).

The effect on the TDR-measurements of boulders and stones in the till has not been investigated, but since they are bedrock fragments just like the mineral grains, they represent parts of the soil volume with volumetric water content equal to zero. Unrepresentative sampling will occur if boulders or stones prevent the TDR-electrodes from being inserted at the correct location according to the data collection program. However, the frequency of boulders and stones within the surface layers of soil in the experimental catchments was so low that this problem was considered negligible.

The TDR-measurements were performed using 15 cm long electrodes installed perpendicular to the soil surface. This means that the soil moisture was determined in a layer below the soil surface with the same depth. Two different instruments were used: Tektronix 1502 C MTDR with two electrodes and Trime TDR-system with three electrodes.

Two replicate measurements of soil moisture were made at 5 m intervals along the transect lines in the experimental catchments. At 2–5 locations along the transects intensive soil moisture measurements were

Table 3
Description of data collection in the three experimental catchments

Basin (transect No.)	Length of transect (m)	No. of piezometers	No. of soil moisture squares in different topographic units
Buddby No. 1	130	13	nose (1), slope (1), hollow (1)
Buddby No. 2	80	4	nose (2), slope (2), hollow (2)
Dansarhällarna No. 1	110	13	nose (1), slope (1), hollow (1)
Dansarhällarna No. 2	90	6	0
Östfora	40	5	nose (1), hollow (1)

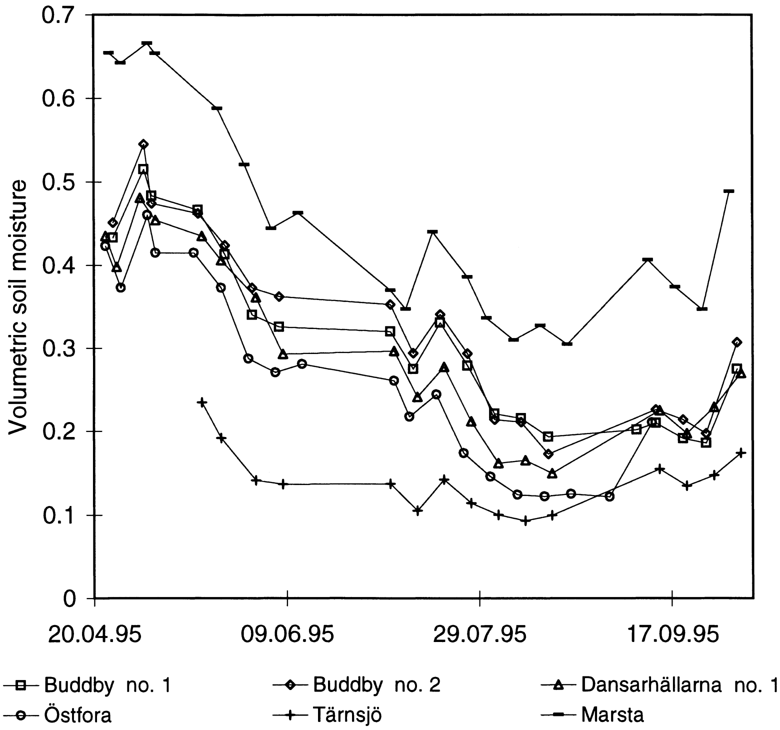

Fig. 2. Mean volumetric water content from the transect lines in the experimental catchments and from two other experimental sites in the NOPEX region. Buddby, Dansarhällarna and Östfora are located in areas with till, Tärnsjö on an esker (sand and gravel) and Marsta in an agricultural area (clay).

made in squares which define the patch scale in this paper. Mostly the squares were 8 m × 8 m with 25 electrode pairs on a 2 m × 2 m grid. The exception was transect line No. 2 in the Buddby catchment which used five squares 10 m × 10 m and nine electrode pairs at a 5 m spacing. The squares are located in areas with different topographic characteristics, either nose, slope or hollow. This means that each square represents a hydrological response unit or patch in the landscape. In general 1–3 piezometers were located within the squares, the exception was transect line No. 2 in the Buddby catchment where piezometers were located within 20 m from the squares and the Östfora catchment where a piezometer was located 10 m from the square in the nose unit, in direction of the hollow unit.

At the other five experimental sites, soil moisture data were measured in one or two squares with 25 equally spaced points in a grid net of size 2 m × 2 m, while groundwater levels were observed in at most two piezometers within a distance of 20 m from the squares. With the exception of the agricultural site, these piezometers were installed by the Geological Survey of Sweden. Their depth varies with soil type, being more than 20 m at the esker site and around 5 m at the other sites.

Fig. 2 shows mean values of volumetric water content from the transect lines in the experimental catchments and from two other experimental sites during the period from April 1995 to October 1995. Fig. 3 shows mean values of depth to the groundwater table in topographic units along the transect lines Buddby No. 1 and Dansarhällarna No. 1 during the period from April 1995 to February 1996. The data series from the nose units stop when the groundwater table is below the bottom of the piezometers.

Three groups of three piezometers and eight pairs of piezometers are located within squares with size less than 10 m × 10 m. Data from these piezometers show that the depth to the groundwater table varies slowly in space, with a maximum difference between simultaneous observations in a patch between 5 and 15 cm.

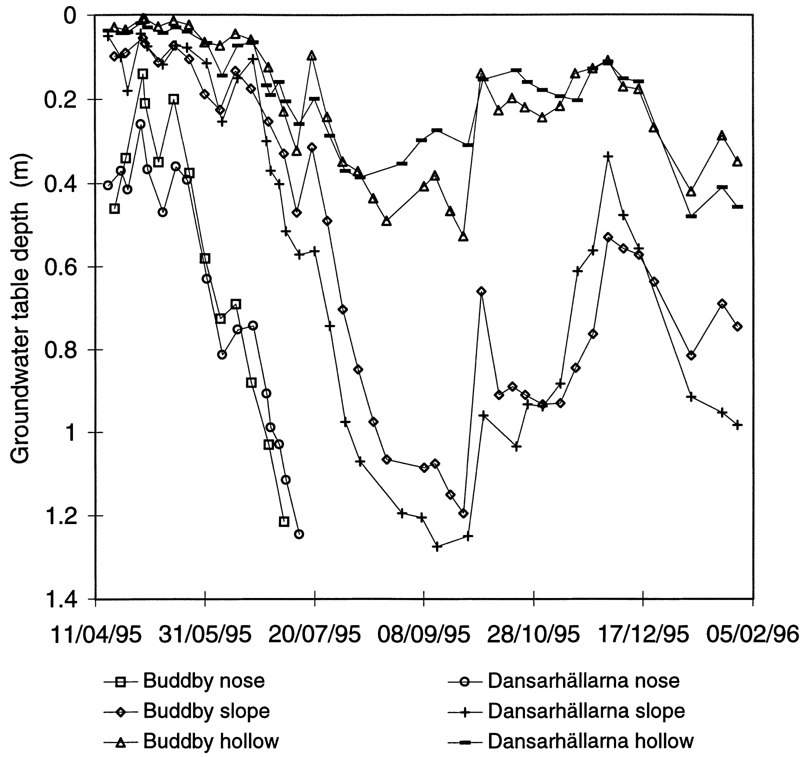

Fig. 3. Mean values of depth to the groundwater table in topographic units along the transect lines Buddby No. 1 and Dansarhällarna No. 1.

Fig. 4 shows groundwater table depths from squares with three piezometers along transect lines Buddby No. 1 and Dansarhällarna No. 1 in May and June 1996.

4. Soil moisture in the unsaturated zone

A characteristic feature of till soils is the drastic decrease in porosity and saturated hydraulic conductivity with depth. This has the consequence that the groundwater level follows the land surface at a shallow depth, generally between 0 and 2 m, which in turn leads to a close relationship between groundwater table depth and soil moisture content in the unsaturated zone (Lundin, 1982). This is due to the capillary transport of water and means that the groundwater level determines the possible range of fluctuations of the soil moisture content.

When there is no movement or uptake of water in a soil profile it reaches a state of hydrostatic equilibrium determined by the depth of the groundwater table. Soil evaporation and extraction of water by plants cause a deficit relative to the water content at equilibrium to develop in the root zone. After infiltration of water, this deficit must be filled up before the groundwater level can rise. During episodes of infiltration the water content in the unsaturated zone is above the equilibrium to allow vertical drainage, but it is quickly redistributed to equilibrium with a rise in the groundwater table.

To describe the soil moisture content in the unsaturated zone we use the relationship between volumetric water content and soil moisture tension given by Brooks and Corey (1966):

$$\frac{\theta-\theta_{\mathrm{r}}}{n-\theta_{\mathrm{r}}}=\left(\frac{\psi_{\mathrm{a}}}{\psi}\right)^{\lambda} \qquad \psi>\psi_{\mathrm{a}} \tag{1}$$

where θ is the volumetric water content, θ_{r} the residual water content, ψ the tension head of soil water, n the porosity of the soil, ψ_{a} the air entry tension head and λ a parameter that depends on the pore size distribution of the soil.

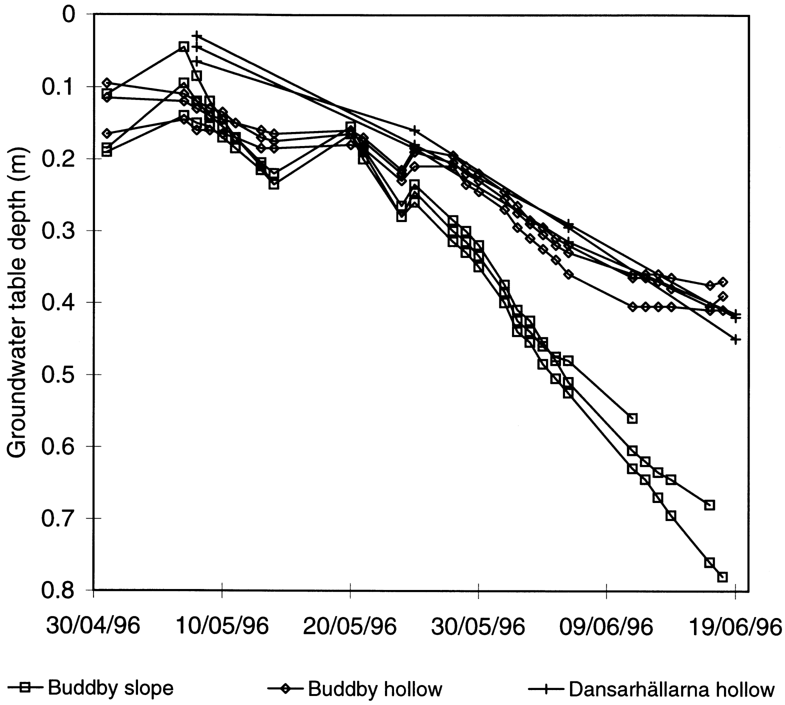

Fig. 4. Groundwater table depths from squares with three piezometers along transect lines Buddby No. 1 and Dansarhällarna No. 1 in May and June 1996.

A large range of grain sizes gives till soils a heterogeneous structure accentuating the spatial variation of soil properties (Nyberg, 1995). The porosity shows a systematic variation with depth, with the highest values near the ground surface (Johansson, 1986; Nyberg et al., 1993). Freezing and thawing, root and soil fauna activities and chemical processes change the texture and structure of the soil (Lundin, 1982). These processes are most active in the upper layers. The air entry tension and the pore size distribution do not show a systematic variation with depth (Udnæs, 1991; Stähli et al., 1995).

The residual water content is defined as the water content at which the water films in the soil become discontinuous (Dingman, 1994). Soil moisture retention curves for till soil samples from the NOPEX Central Tower Site were used to determine the residual water content as a parameter in Brooks and Corey's relation. Stähli et al. (1995) found that θ_r was of the order 10^{-3} which is much smaller than the volumetric water content and the porosity. By neglecting θ_r and assuming a state of vertical hydrostatic equilibrium, the volumetric water content in the unsaturated zone is given by:

$$\theta_e = n_z\left(\frac{\psi_a}{h-z}\right)^{\lambda} \quad h-z > \psi_a \qquad (2)$$
$$\theta_e = n_z \quad h-z \leq \psi_a$$

where the z-axis is assumed positive downwards with zero at the ground surface and h the depth to the groundwater table. The porosity, n_z, depends on the depth below the ground surface.

The equilibrium water content at the soil surface ($z=0$) is equal to:

$$\theta_{e0} = n_0\left(\frac{\psi_a}{h}\right)^{\lambda} \quad h > \psi_a \qquad (3)$$
$$\theta_{e0} = n_0 \quad h \leq \psi_a$$

where n_0 is the porosity near the ground surface.

In the analysis that follows soil water content at the surface given by Eq. (3) will be compared to measured soil water content from the upper 15 cm of the soil.

Strictly, the measured water content could be considered to apply to a soil depth of 7.5 cm, however it is assumed to be valid at the ground surface.

The equilibrium state assumes that the soil moisture tension at a given depth depends on the distance to the groundwater table. In the root zone the water content is generally below the equilibrium value due to evapotranspiration. Since sufficient data on root geometry or root pressures are not known, this soil moisture deficit must be evaluated by treating the root system as a lumped system which penetrates the upper layer of the soil uniformly (Guymon, 1994). The moisture content at the surface is then given by:

$$\theta_0 = n_0\left(\frac{\psi_a}{r_0 h}\right)^{\lambda} \quad h > \psi_a \tag{4}$$

$$\theta_0 = n_0 \quad h \leq \psi_a$$

The factor r_0 is the ratio between the actual soil moisture tension at the surface and the soil moisture tension at equilibrium. If the groundwater table is below the depth given by the air entry tension head, r_0 is generally >1. During episodes of infiltration, the soil moisture content may rise above the equilibrium value with $r_0 < 1$, followed by a rise of the groundwater table and a new equilibrium state in the unsaturated zone. When the depth to the groundwater table is less than or equal to the air entry tension head, the soil is saturated at the surface and $r_0 = 1$.

The depth to the groundwater table and the parameters of Brooks and Corey's equation; porosity, air entry tension and pore size distribution index varies in space and may be considered as sources of variability of the volumetric soil moisture at the soil surface. In addition, so will microscale topography and the factor r_0. Although the depth to the groundwater table is nearly constant at the patch scale (Fig. 4), the volumetric soil moisture in the upper 15 cm varies considerably over short distances due to variations in soil characteristics, root concentration and microscale topography. This implies that the variability of surface soil moisture conditions at the patch scale depends on the heterogeneous nature of soil characteristics. Above a certain threshold scale, the variations of soil characteristics are smoothed out and the spatial variability of soil moisture content is determined by the depth to the groundwater table.

The lack of knowledge regarding the factors controlling the soil moisture content can be expressed by treating them as random variables (Romanowicz et al., 1995). In principle, it would be correct to consider a joint distribution function of all relevant factors which determine the volumetric water content at the soil surface. However, in order to make the problem mathematically tractable, some simplifications must be made and the variability of porosity at the patch scale is assumed to account for the variability in the other soil parameters as well as the variability in root concentration, microscale topography and the factor r_0. The reason for selecting porosity as the stochastic variable which is explaining the variability in soil characteristics is that it is the parameter in Brooks and Corey's equation which resembles most what we want to model; the volumetric water content. In addition, the study by Tallaksen and Erichsen (1995) shows a variability in saturated water content at small scale which cannot be neglected. This approach implies that air entry tension head and pore size distribution index are considered as effective parameters, while r_0 is considered constant at the patch scale.

At the catchment scale, we assume that the depth to the groundwater table is the main factor determining variations in soil moisture at the surface. The frequency distribution of volumetric soil moisture near the ground surface at this scale can now be determined. However, we must take into account the following result from the experimental catchments: at a given instant of time, r_0 increases with increasing depth to the groundwater table (Table 4). This means that the soil moisture deficit relative to the equilibrium value is largest in the upper parts of slopes where the groundwater table is deep, a result which is supported by Lundin (1982). This is the recharge area where a vertically downward movement of groundwater leaves less water available for filling the deficit than in the discharge areas at the bottom of slopes where an upward directed groundwater flow supplies water. Since only limited information about this relationship is available, we assume that r_0 at a given instant of time has a linear variation over a range of values determined by the observed depths to the groundwater table:

$$r_0 = Ah + B \tag{5}$$

Table 4
Mean depth to the groundwater table within squares along transect line No. 1 in the Buddby catchment, vs. r_0, the ratio between the actual soil moisture tension at the soil surface and the soil moisture tension at equilibrium, as given by Eq. (11)

Date	Groundwater table depth (m)	r_0	Groundwater. table depth (m)	r_0	Groundwater. table depth (m)	r_0
25/04/95	0.055	1.000	0.055	1.000	0.340	5.691
03/05/95	0.025	1.000	0.020	1.000	0.140	4.098
05/05/95	0.030	1.000	0.040	1.000	0.250	3.849
17/05/95	0.300	1.000	0.040	1.000	0.200	6.253
24/05/95	0.045	1.000	0.090	1.000	0.375	4.859
31/05/95	0.080	1.000	0.195	1.933	0.580	7.376
07/06/95	0.080	1.000	0.240	1.977	0.725	7.633
06/07/95	0.240	0.610	0.390	1.191	1.215	4.458

where

$$A = \frac{r_{\max} - r_{\min}}{h_{\max} - h_{\min}}; \quad B = r_{\min} - h_{\min}\frac{r_{\max} - r_{\min}}{h_{\max} - h_{\min}}$$

The soil physical parameters of the surface layer have been determined from available data and by reference to other investigations in areas with till soils. The porosity of samples from the upper horizons of till soils is generally in the range 0.5–0.7 (Lundin, 1982; Johansson, 1986; Nordén, 1991; Udnæs, 1991; Nyberg et al., 1993; Stähli et al., 1995). This is in agreement with maximum values of water content in samples from the experimental basins (Tallaksen and Erichsen, 1995). The air entry tension head in till soils is generally in the range 0.1–0.2 m (Grip and Rodhe, 1994). Soil moisture characteristic curves determined from till soil samples by Udnæs (1991) and Stähli et al. (1995) gave values of ψ_a below 0.1 m, while Johansson (1986) found good agreement between empirical and theoretical pF-curves using $\psi_a = 0.1$ m. The value of $\psi_a = 0.09$ m has been chosen in this study. When the depth to the ground water table is <0.09 m, the mean value of observed volumetric soil moisture near the surface is equal to 0.6 (Fig. 5), which is used as a mean value of porosity near the soil surface at the patch scale. Some values in Fig. 5 are above 0.6, however a certain range of values of saturated water content must be expected. Results from Udnæs (1991) and Stähli et al. (1995) give values of the pore size distribution index, λ, in the range 0.11–0.3. With the selected values of porosity and the air entry tension head, we find that $\lambda = 0.3$ gives a good agreement between empirical and theoretical distribution functions of soil moisture at patch and catchment scales. These values of the soil physical parameters are used at both scales. At the patch scale, the porosity of the soil near the surface varies in space, while the mean porosity, the air entry tension head and the pore size distribution index are assumed constant. This means that the spatial variation of soil characteristics is accounted for by the porosity. At the catchment scale the soil physical parameters of the surface layer are assumed constant in space.

The simplifications regarding the sources of variability of soil moisture at patch and catchment scales are appropriate since the purpose of this paper is not to give a detailed description of soil characteristics, root

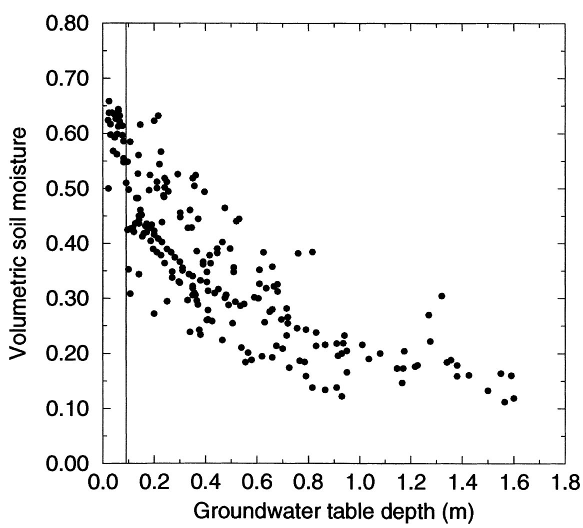

Fig. 5. Mean volumetric soil moisture vs. mean depth to the groundwater table for each set of observations in the squares of the experimental catchments.

concentration and microscale topography at the patch scale, but rather to describe the distribution of the state variables which determine the hydrological response of catchments at scales ranging from 10 to 10^4 m. A detailed description at the patch scale would require a different data collection program which in turn would prevent collection of data from a large area. We have assumed that the physical parameters of the upper layer of the soil are constant in space at the catchment scale. This assumption is justified by Nyberg (1995) who considers it possible to establish catchment representative values for porosity and water retention and Lundin (1982) who shows that the porosity values do not show large differences between subareas of a catchment.

5. Patch scale variability

If the influence of variability of the air entry tension head, ψ_a, and the pore size distribution index, λ, is neglected at the patch scale, the vertical equilibrium model states that the variations in volumetric water content in the upper soil layers is caused by variations in the stochastic variable N_0, the porosity near the ground surface. Since N_0 has a lower bound of 0 and an upper bound of 1 its variation can be described by the β-distribution which takes values in the same interval (Yevjevich, 1972). The β-distribution fits easily to empirical data since it can take a variety of shapes without restrictions regarding, e.g. symmetry. It has previously been used for approximating frequency distributions of recorded soil properties (Haskett et al., 1995). The β-frequency distribution is given by (Kendall, 1986):

$$f_{N_0}(n_0) = \frac{1}{B(p,q)} n_0^{p-1}(1-n_0)^{q-1} \quad 0 \le n_0 \le 1; \quad p > 0; \quad q > 0 \tag{6}$$

where $B(p,q)$ is the β-function and p and q are parameters.

The mean and variance of the β-distribution are given by:

$$E(N_0) = \mu_{N_0} = \frac{p}{p+q} \tag{7}$$

$$\mathrm{Var}(N_0) = \sigma^2_{N_0} = \frac{pq}{(p+q+1)(p+q)^2} \tag{8}$$

When the depth to the groundwater table in a patch is larger than the air entry tension head, the frequency distribution of the stochastic variable Θ_0, the volumetric soil moisture content near the ground surface, can be found by transforming the distribution of N_0. The frequency distribution of a monotonic function $Y(X)$ of a univariate continuous random variable X is given by (Haan, 1977):

$$f_Y(y) f_X(x) \left|\frac{\mathrm{d}x}{\mathrm{d}y}\right| \tag{9}$$

Combining Eqs. (4), (6) and (9) we find:

$$f_{\Theta_0}(\theta_0) = \left(\frac{r_0 h}{\psi_a}\right)^{\lambda} \frac{1}{B(p,q)} \left[\theta_0 \left(\frac{r_0 h}{\psi_a}\right)^{\lambda}\right]^{p-1} \times \left[1-\theta_0\left(\frac{r_0 h}{\psi_a}\right)^{\lambda}\right]^{q-1} \quad h > \psi_a; \quad \theta_0 < n_0 \tag{10}$$

The mean and variance of this distribution can be expressed in terms of those of the β-distribution of N_0:

$$E(\Theta_0) = \mu_{\Theta_0} = \mu_{N_0}\left(\frac{\psi_a}{r_0 h}\right)^{\lambda} \tag{11}$$

$$\mathrm{Var}(\Theta_0) = \sigma^2_{\Theta_0} = \sigma^2_{N_0}\left(\frac{\psi_a}{r_0 h}\right)^{2\lambda} \tag{12}$$

For each set of data from a patch, r_0 can be calculated from Eq. (11) using the sample mean of volumetric soil moisture and depth to the groundwater table within the square.

Method of moments estimators of the parameters p and q are given by:

$$\hat{p} = \frac{m_{\Theta_0}}{s^2_{\Theta_0}}\left[m_{\Theta_0} - m^2_{\Theta_0}\left(\frac{r_0 h}{\psi_a}\right)^{\lambda} - s^2_{\Theta_0}\left(\frac{r_0 h}{\psi_a}\right)^{\lambda}\right] \tag{13}$$

$$\hat{q} = \frac{(\psi_a/r_0 h)^{\lambda} - m_{\Theta_0}}{s^2_{\Theta_0}}\left[m_{\Theta_0} - m^2_{\Theta_0}\left(\frac{r_0 h}{\psi_a}\right)^{\lambda} - s^2_{\Theta_0}\left(\frac{r_0 h}{\psi_a}\right)^{\lambda}\right] \tag{14}$$

where m_{Θ_0} and $s^2_{\Theta_0}$ is the sample mean and variance of Θ_0.

When the depth to the groundwater table in a patch is less than the air entry tension head, the vertical equilibrium model states that the entire soil profile is

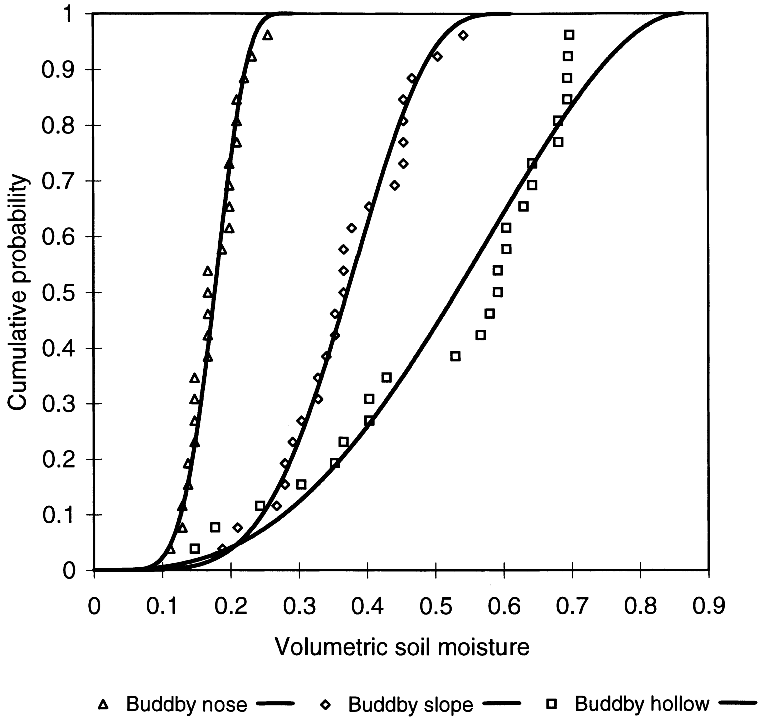

Fig. 6. Cumulative probability distributions of soil moisture within squares of size 8 m × 8 m along transect line No. 1 in the Buddby catchment on 6 July 1995 based on observed data (points) and the theoretical expression in Eq. (10) (lines).

saturated. This means that the frequency distribution of Θ_0 is equal to the frequency distribution of N_0.

The theoretical distribution function of volumetric soil moisture near the surface at the patch scale is given by Eq. (10). For each set of simultaneously observed data in a square, the parameters of this distribution were estimated using Eqs. (13) and (14). Fig. 6 shows cumulative distribution functions of soil moisture at the patch scale in Buddby on 6 July 1995. The observed soil moisture varies considerably within the patches. This is in agreement with results from Myrabø (1986) and Nyberg (1996) who observed large variations in soil moisture content over short distances in catchments with till soils in south–eastern Norway and south–western Sweden. Fig. 6 also illustrates the variability of soil moisture conditions along a hillslope.

6. Catchment scale variability

We select the two parameter γ-distribution distribution to describe the stochastic variable H, depth to the groundwater table, at the catchment scale at a given instant of time. The γ-distribution takes values in the range from 0 to ∞ just like the variable we want to model and it has the flexibility in shape which is necessary to fit empirical data. It has been widely used in hydrology (Haan, 1977). The γ-frequency distribution is given by (Kendall, 1986):

$$f_H(h) = \frac{\alpha^{\gamma}}{\Gamma(\gamma)} h^{\gamma-1} \mathrm{e}^{-\alpha h} \quad h \geq 0; \quad \alpha > 0; \quad \gamma > 0 \tag{15}$$

where $\Gamma(\gamma)$ is the gamma function and α and γ the parameters.

The mean and variance of the γ-distribution are given by:

$$E(H) = \mu_H = \frac{\gamma}{\alpha} \tag{16}$$

$$\mathrm{Var}(H) = \sigma_H^2 = \frac{\gamma}{\alpha^2} \tag{17}$$

Method of moments estimators of the parameters α and γ are given by:

$$\hat{\alpha} = \frac{m_H}{s_H^2} \quad (18)$$

$$\hat{\gamma} = \frac{m_H^2}{s_H^2} \quad (19)$$

where m_H and s_H^2 is the sample mean and variance of H.

The frequency distribution of volumetric soil moisture near the ground surface consists of two parts; one continuous part for the case when the depth to the groundwater table is larger than the air entry tension head and one discrete probability for saturation near the soil surface. The continuous part is found by combining Eqs. (4), (5), (9) and (15):

$$f_{\Theta_0}(\theta_0) = \frac{\alpha^{\gamma}\psi_a n_0^{1/\lambda}}{\Gamma(\gamma)\lambda\theta_0^{1+1/\lambda}} \frac{1}{C}\left(\frac{C-B}{2A}\right)^{\gamma-1} \exp\left[-\frac{\alpha(C-B)}{2A}\right]$$
$$h > \psi_a; \quad \theta_0 < n_0 \quad (20)$$

where

$$C = \left[B^2 + 4A\psi_a\left(\frac{n_0}{\theta_0}\right)^{1/\lambda}\right]^{1/2};$$

A and B are defined in Eq. (5).

The probability of saturation at the soil surface is equal to the probability that the depth to the groundwater table is less than or equal to the air entry tension head. This probability is given by the incomplete γ-function:

$$f_{\Theta_0}(n_0) = P(\gamma, \alpha\psi_a) = \frac{1}{\Gamma(\gamma)}\int_0^{\alpha\psi_a} e^{-t}t^{\gamma-1}dt \quad (21)$$

Data from the hillslopes describe the different moisture conditions in the experimental catchments and can therefore be considered representative for conditions at the catchment scale. The distribution function of volumetric soil moisture at this scale is given by Eqs. (20) and (21). For a given situation, the parameters of this distribution can be found from Eqs. (18) and (19) using the sample mean and variance of depth to the groundwater table along a slope. A, B and C in Eqs. (5) and (20) are found by inserting minimum and maximum values of r_0 from the squares along the transects as given by Eq. (11). Figs. 7–9 show cumulative distribution functions of soil moisture along hillslopes in the Buddby, Dansarhällarna and Östfora catchments on two occasions with different moisture conditions.

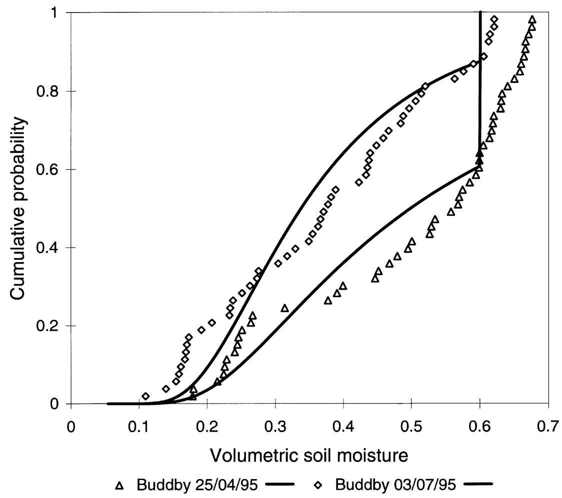

Fig. 7. Cumulative probability distributions of soil moisture along transect line No. 1 in the Buddby catchment based on observed data (points) and the theoretical expression in Eqs. (20) and (21) (lines).

These and similar data on soil moisture variations along the slopes confirm that the groundwater level controls the soil moisture conditions in the unsaturated zone in the experimental catchments. The step in the distribution function gives the proportion of the catchment area where the depth to the groundwater table is less than the air entry tension head. These are the discharge areas with a shallow groundwater table and

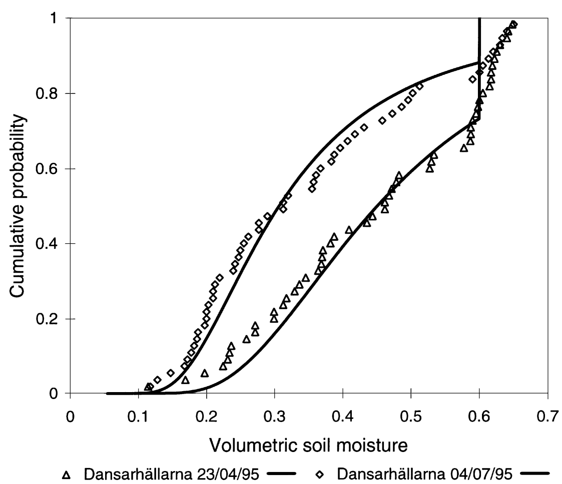

Fig. 8. Cumulative probability distributions of soil moisture along transect line No. 1 in the Dansarhällarna catchment based on observed data (points) and the theoretical expression in Eqs. (20) and (21) (lines).

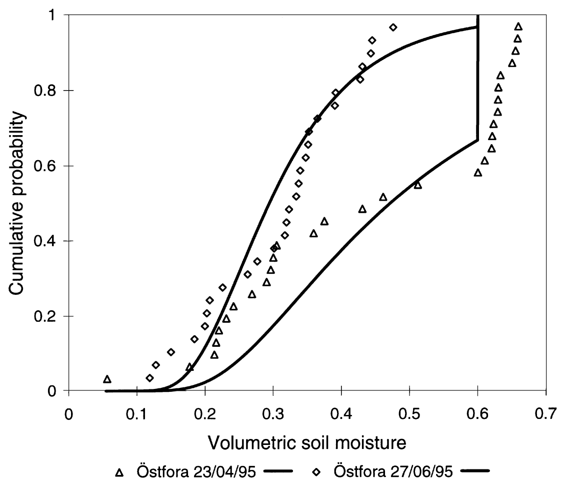

Fig. 9. Cumulative probability distributions of soil moisture along the transect line in the Östfora catchment based on observed data (points) and the theoretical expression in Eqs. (20) and (21) (lines).

saturated conditions near the soil surface where saturation excess overland flow will occur during precipitation or snowmelt (Anderson and Burt, 1990). This catchment proportion varies with time as a result of varying precipitation, evapotranspiration and downslope flow through the soil. The total area which can be expected to contribute actively to runoff during storms is larger, since the high proportion of groundwater must be caused by water which has infiltrated in parts of the recharge area (Rodhe, 1987).

7. Regional scale variability

Soil moisture and groundwater data from catchments in till soils in different parts of the NOPEX region show more or less identical fluctuations (Figs. 2 and 3), indicating similar hydrological behaviour. This similarity suggests that above a certain threshold scale a catchment might contain a sufficient sample of the geology, topography, vegetation and other landscape characteristics of a region to decrease the variability of catchment average fluxes to a minimum. This is a prerequisite of the representative elementary area concept, which assumes that the distribution of characteristics within a catchment may be important in determining these fluxes, but the pattern of those characteristics is no longer important (Wood et al., 1988, 1990). The REA can be interpreted as a scale at which the local patterns in runoff production are sufficiently well integrated to produce a similarity in response, before non-stationarity in catchment characteristics or hydrological processes starts to increase the variances at larger scales. Such non-stationarity may be due to changes in geology, physiography or land-use, changes in the scale of rainfall variability or changes in the nature of hydrological processes (Beven, 1995). The existence of a REA is equivalent to a minimum in the power spectrum of catchment properties. This refers to a process consisting of a small scale component superimposed on a much larger component. The REA relates to a preferred element size for distributed catchment modelling of large regions or drainage basins, since this requires deterministic representation of the large scale variability as different values in different model elements and parameterization of small scale processes within the elements. Provided a minimum in the power spectrum of the processes which governs catchment response exists, the same parameterization can be used for a range of spatial scales for catchments with similar distributions of catchment characteristics and precipitation (Blöschl et al., 1995).

The state variables which are at the focus of this work determine how the climatic input to a catchment is transformed to fluxes in the form of evapotranspiration and runoff. According to Beven (1995) it is the distribution of hydrological responses in the landscape which must be achieved to provide realistic predictions of discharge and evapotranspiration fluxes within heterogeneous terrain. A criterion for the existence of a scale at which the average hydrological response varies only slowly with increasing catchment area is that the variability of the responses between different areas falls to acceptably low values for a sufficiently large area, for larger areas this variability may rise again (Wood et al., 1988, 1990). This effect must also be manifest in the mechanisms which control the responses of the catchments, i.e. the state variables soil moisture content and groundwater level.

Mean values of volumetric water content and depth to the groundwater table over areas with increasing size have been calculated using all available observations from experimental sites in forest areas with till soils at specific instants of time. Figs. 10 and 11 show that the large variability between areas of similar size in till soils decreases as the averaging area increases

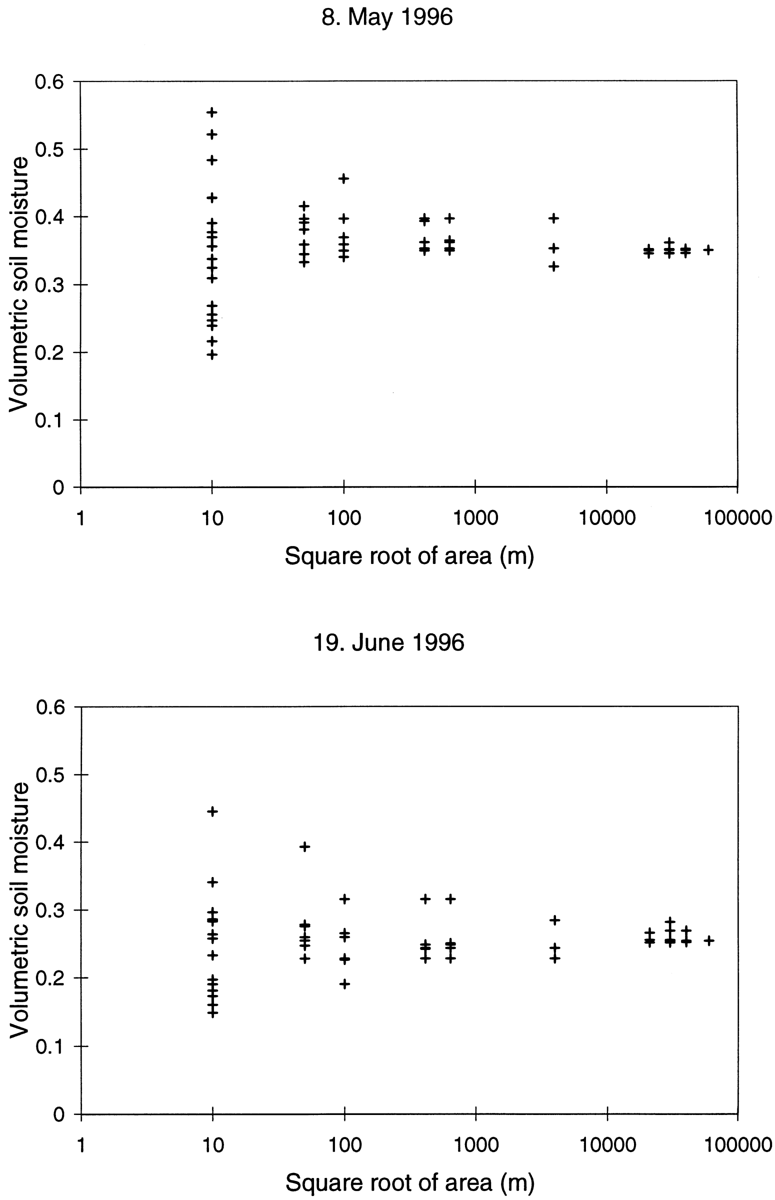

Fig. 10. Mean volumetric water content in the upper 15 cm of the soil versus size of the averaging area. Data from experimental sites in areas with till on 8 May and 19 June 1996.

from the patch scale (100 m^2) to the size of the small experimental catchments (1 km^2). By comparison, averaging data from experimental sites in different landscape and soil types does not yield the same decrease in variability, as shown for soil moisture data in Fig. 12. Groundwater data from different landscape and soil types vary even more, as the depth to the groundwater table in eskers is of the order of 10–20 m.

8. Conclusions

This study has focused on the spatial variation of soil moisture and groundwater levels in small catch-

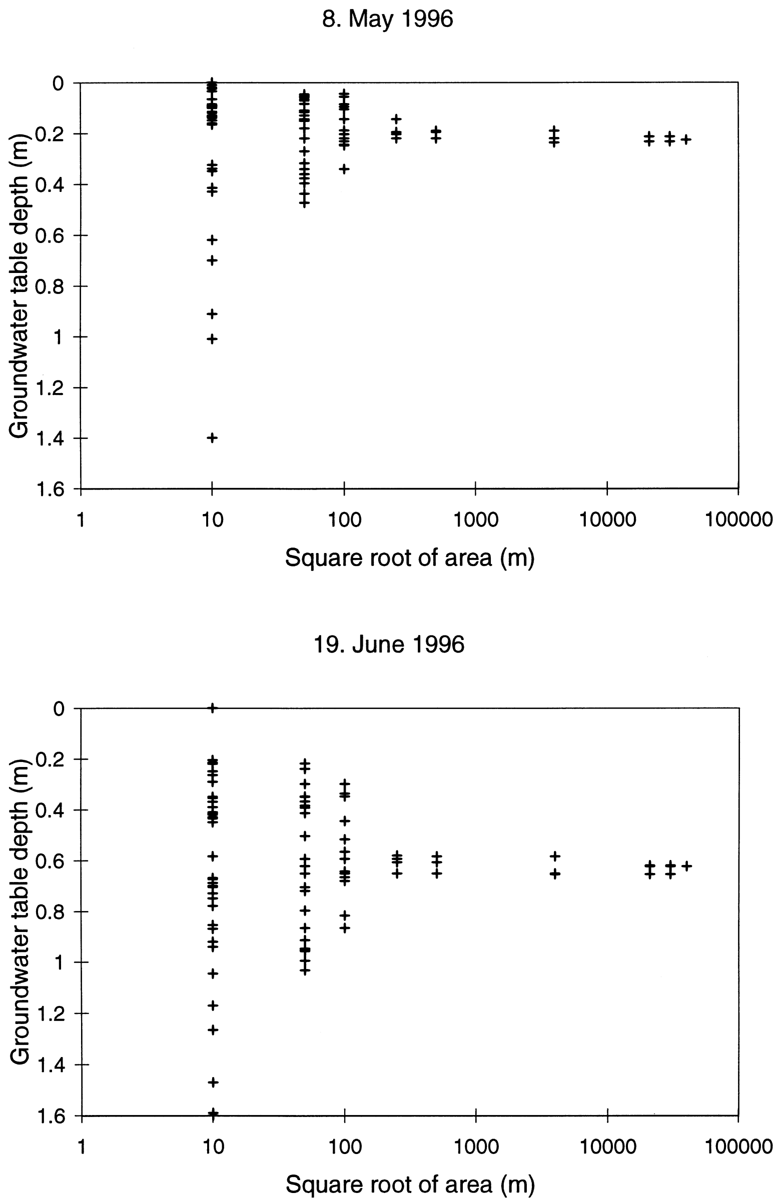

Fig. 11. Mean depth to the groundwater table versus size of the averaging area. Data from experimental sites in areas with till on 8 May and 19 June 1996.

ments in areas dominated by boreal forest and till soils. These state variables are the main factors controlling the response of a drainage basin on climatic input in this landscape type. The spatial variability of surface soil moisture conditions within hydrological response units at the patch scale depends on the heterogeneous nature of soil characteristics. Above a certain threshold scale, the variations of soil characteristics are smoothed out and the spatial variability of soil moisture content is determined by the depth to the groundwater table. Statistical distribution functions of soil moisture content conditioned on the depth

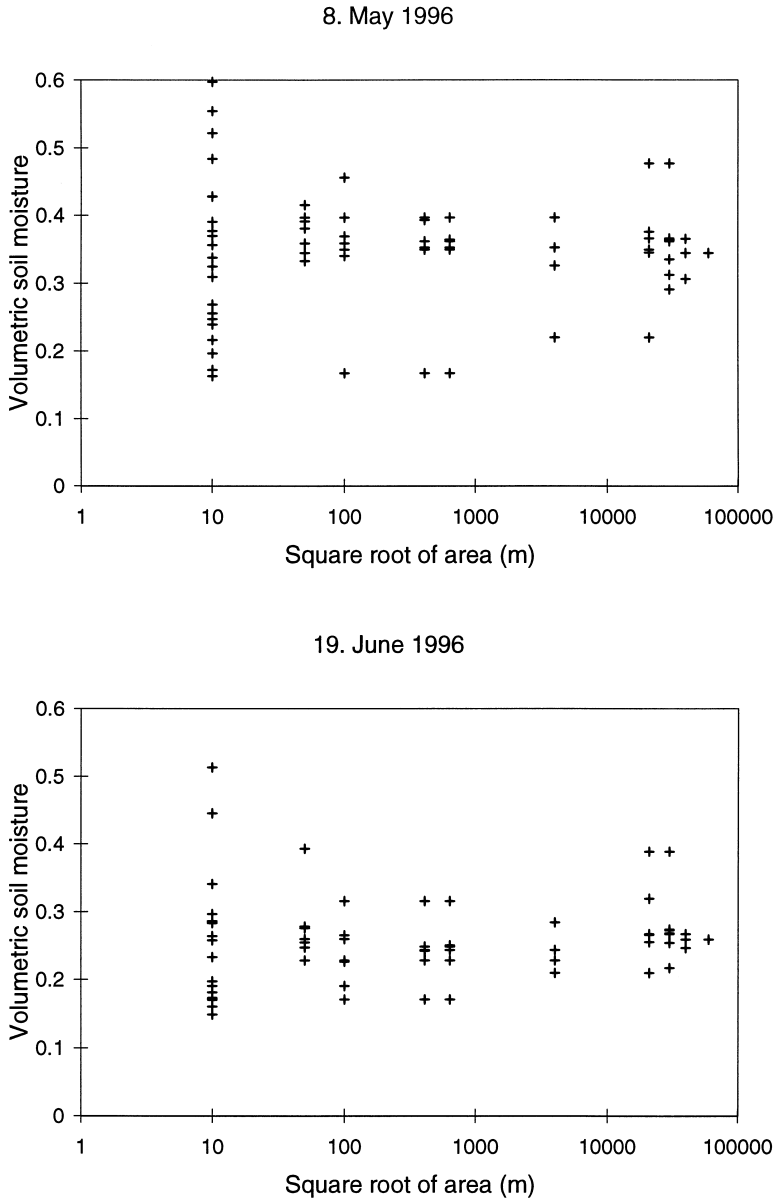

Fig. 12. Mean volumetric water content in the upper 15 cm of the soil versus size of the averaging area. Data from experimental sites in areas with till, on an esker (sand and gravel) and in an agricultural area (clay) on 8 May and 19 June 1996.

to the ground water table are well suited to the description of the observed patterns of soil moisture conditions at patch and catchment scales. The area which is capable of saturation excess runoff generation can be identified from the distribution function of soil moisture at the catchment scale.

Since the soil moisture content is conditioned on the depth to the groundwater table, which in its turn is closely related to topography, the variations of soil moisture content in a catchment are also controlled by topography. The observed data show that moisture conditions vary along a hillslope, with high soil moisture content and shallow groundwater table at the bottom of hillslopes and low soil moisture content and deep groundwater table at the top of hillslopes. The spatial variation of soil moisture deficit expressed by the ratio between the actual soil moisture tension in the root zone and the soil moisture tension at equilibrium confirms the significance of topography for moisture conditions. In many cases the measurements of hydrological state variables are too widely spaced and the natural variability too large for reliable estimation based on observations. When this occurs the results of this study indicate that topography can be used as a covariate in order to estimate the spatial distribution of wetness states in a catchment.

The small variability in hydrological behaviour between catchments located in till soils in different parts of the NOPEX region is due to more or less identical climatic input and catchment geology, soil type, topography and vegetation. However, the variability may also be attributed to difficulties with obtaining a representative sample of moisture conditions for all the different geomorphological units in a catchment, which would require a more extensive data collection program. Nevertheless, the results from this study confirm that the dynamics of hydrological response units with similar physiographic conditions are identical in different parts of the NOPEX region. This is in agreement with results from Krasovskaia (1985) who showed that groundwater and soil moisture conditions differed significantly between hydrological response units within a catchment, but not between basins with similar physiographic conditions or between the same type of units in different basins with similar climatic input. The conclusion is that moisture conditions in the till soils of the NOPEX region are dominated by variations in topography, vegetation and soil characteristics at scales $<1–2\ km^2$. Although a separation of scales or equivalently a minimum in variability between small and large scale processes could not be detected, the similarity in groundwater and soil moisture conditions suggests that the variability of catchment responses in the form of evapotranspiration and runoff decreases to a minimum at this spatial scale. This means that the same parameterization can be used for modelling hydrological processes in catchments or grid elements at scales of ca. $1\ km^2$ in till soils in different parts of the NOPEX region.

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