Nordic Hydrology, 19, 1988, 1-18

Hydrological Analysis of Basin Behaviour from Soil Moisture Data

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Soil moisture dynamics in the Velen drainage basin (Sweden) were analyzed in order to assess the degree of and the reasons for spatial variation in basin behaviour. The main tool was a modified version of the soil moisture accounting routine in the conceptual runoff model HBV, optimized against neutron probe field data. Simulated soil moisture dynamics, interception and percolation rates agreed well with measurements and other calculations. Integration of simulated evapotranspiration from sites with different characteristics agreed well with water balance computations for the area. It was shown that unsaturated flow through macropores probably occurred after heavy rainstorms. During spring, evapotranspiration was limited to values below the potential (Penmans equation) even at times when no soil moisture deficit existed. Soil moisture differences between forest and grassland (including a deforested site) were, during summer, mainly attributed to differences in the root distribution with depth. The effect of interception on the total evapotranspiration rates was only significant during periods when transpiration demands were low. Soil moisture differences between forest sites were mainly attributed to topography but variations in soil characteristics and root distribution had to be considered, especially during dry periods.

Introduction

Soil moisture conditions within a basin are heterogeneous. Hence, the problem to describe spatial variations is well known. For large areas it is not possible to achieve a network of measuring stations of sufficient density to cover the heterogeneous structure of soils, vegetation and topography (Kutilek 1971). Therefore, it is necessary to estimate to which degree various basin characteristics will effect soil moisture dynamics. Attempts have been made to use statistical estimates where each soil moisture tube will represent moisture content for a particular proportion of the basin. For example, Moor and Clarke (1982) used this technique in basin soil moisture simulation, giving each tube an equal weight since they found it difficult to assess the particular proportion that each tube should represent.

This paper presents an analysis of the soil moisture conditions registered during the International Hydrological Decade (IHD) in the Velen research basin (Fig. 1). The Velen basin was considered representative for coniferous forest areas on moraine soils in south central Sweden. The basin was to 65% covered by forest. The trees were dominantly coniferous having a fairly high canopy density. Lakes occupy nearly 10% of the area. The remaining part consists of swamps and small areas with cultivated land. The soil is mainly sandy till, underlain by gneisses and granites.

The objective was to assess how, to what degree and during which weather conditions that various factors (vegetation, topography, soil) influenced spatial variations in the basin behavior. This knowledge can be used together with information of the spatial distribution of various factors to assess the spatial variations of hydrological behavior.

Method

The Soil Moisture Model

The main tool used in the analyses was a soil moisture (SM) model, developed from the SM-accounting routine in the HBV runoff model (Bergström 1976). The formulation of the equations in the SM-accounting routine in the HBV model is given in Appendix I.

Climatological data requirements are restricted to daily values of precipitation and mean air temperature, plus estimates of monthly long-time means of potential evapotranspiration. The SM-accounting routine was further developed into a SM-model by considering information achieved from field data from the Velen basin. The objective was to find an optimal combination of physical relevance and simplicity. The introduced equations are fully described in Appendix II.

In the traditional model, the SM-content is assumed never to exceed field capacity (Eq. (1) in Appendix I). In the SM-model, water contents can however reach values above field capacity (Eq. (1) in Appendix II).

The further development of the model also included equations which take macropore flow into consideration (Eqs. (2)-(7) in Appendix II), a temperature index that reduces transpiration rates when soil and air temperatures act as limiting factors (Eqs. (8)-(11) in Appendix II and rainfall interception routines (Eqs. (12)-(13) in Appendix II).

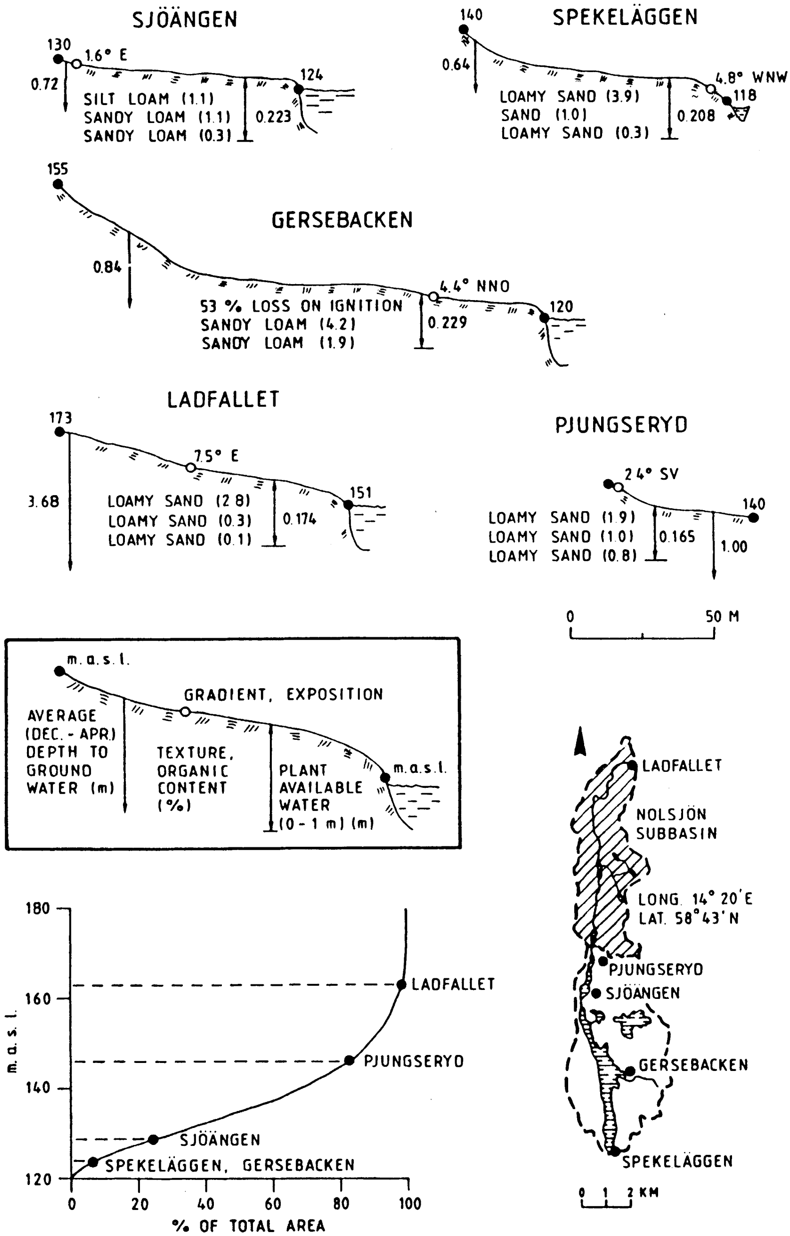

Fig. 1. Characteristics and locations of the soil stations Profiles showing position of each site (unfilled circle), soil texture and organic content at 0-10, 30-40 and 60-70 cm below soil surface, plant available SM at pF 2.0 in the top metre of the soil, highest point of slope above and lowest point of slope below soil station (filled circles), exposition, gradient, and average depth to groundwater table during December-April.

Field capacity was defined as the plant-available water in the top metre of the soil when the potential gradient is in equilibrium with the average depth to the water table (December-April). Maximum water contents (pF 0.7) and wilting points (pF 4.2) were estimated from pF-analysis made at soil samples from each site. Theoretically, it is possible to use the pF-analyses to calculate the value of field capacity. However, as soil moisture data were available for winter periods, field capacity was individually estimated for each site by calculating the average winter soil moisture content, reduced by the water content below wilting point.

Data

Precipitation and temperature data were taken from the meteorological station at Sjöängen (Fig. 1). Monthly long-term averages of potential evapotranspiration were calculated with the Penman equation, using meteorological data from a near-by station, Örebro (Wallén 1966).

Soil moisture data collected 1967-1974 at five stations dispersed over the basin (Fig. 1) were available for analysis (Forsman and Milanov 1971). The measurements were made with the BASC neutron probe and transformed into integrated soil moisture content for the top metre of the soil profiles. Intervals between measurements varied between one week and one month. A drift had occurred during winter 1970/71, which caused too high readings in terms of absolute water contents (T. Milanov, pers. comm. 1987). Data collected after the drift were transposed to absolute values by considering winter values of soil moisture.

Fits Achieved from Model Optimization and Validation

Except for Spekeläggen, only soil moisture data collected before 1971 were used for model optimization. For the optimization period (usually 1968-1970) the explained variance varied between 0.76 and 0.94 and for the validation period (1971-1973) between 0.72 and 0.85.

Analysis of Some Hydrological Phenomena

Interception of Rainfall

Bringfelt (1982) showed that the forest interception amounts achieved by subtracting rainfall in troughs below the canopy from rainfall in troughs in clearings in the Velen forest (Bringfelt and Hårsmar 1974) could be simulated by a model. Also the SM-model was well adapted to simulate interception (Eqs. (12)-(13) in Appendix II). Although Bringfelts model was more complex, both in terms of the mathematical approach and the need of input data, the difference between the outputs from the two models was small (Table 1).

Table 1 – Modelled interception rates with the SM-model and with Bringfelts forest evapotranspiration model compared to measurements

Rain collection period (troughs)	Measured (mm)	SM-model (mm)	Bringfelts model (mm)
730517-0604	12.6	14.5	11.3
0609-0613	2.8	2.6	2.3
0625-0812	32.8	33.8	34.0
0815-0904	7.4	6.2	6.2
0904-0912	1.7	0.7	1.9
0912-0915	0.5	0.4	0.5
0919-1002	16.0	17.2	15.4
740524-0611	14.2	13.6	16.0
0617-0711	14.3	19.3	18.4
0711-0730	14.5	12.6	12.0
0730-0802	0.3	1.3	0.5
0809-0820	14.1	12.8	13.0
0826-0910	14.8	14.7	12.5
0919-1010	28.8	29.7	28.8

Monthly means of interception ratio (percentage of monthly rainfall amount), calculated by the SM-model (1968-1982) were largest during spring when total rainfall amounts were small and heavy rainstorms unusual. The interception ratio was lowest in August when a large part of the rainfall is in the form of heavy convective rainstorms (Fig. 2).

Macropore Flow

In an earlier study (Andersson and Harding) it was shown that SM-models overestimated the soil moisture recharge during autumn for the forest sites in Velen as

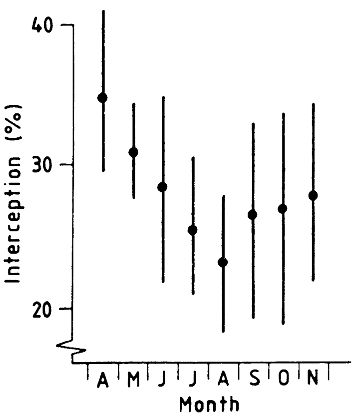

Fig. 2. Calculated interception losses (percent of monthly rainfall amounts) 1968-1982. Monthly mean ± one standard deviation are shown.

long as it was assumed that no water would percolate before field capacity was reached. We found that neither underestimation of interception or transpiration rates nor the occurrence of overland flow could be the main reason for this. It was suggested that the most probable explanation was that water under certain conditions tended to bypass the soil matrix as a result of conduction through macropores.

There are ample evidences that macropores may conduct water rapidly through unsaturated soils (*e.g.* Beven and Germann 1982). These voids can consist of pores, formed by soil fauna or plant roots or by cracks caused by drought or ground frost.

Forest soils tend to have a high degree of anisotropy. This has been attributed to the presence of channels left behind in the soil by decayed roots (De Vries and Chow 1978). Such macropores may be very effective in transporting water through the soil, even when the surrounding soil matrix is unsaturated (Aubertin 1971; Beasly 1976; Mosley 1979,1982).

According to Richards (1950) water will enter a root channel only if it is in direct hydraulic contract with a source of free water. Initiation and maintenance of flows in the macropore system requires a supply of water exceeding all losses to the matrix. Thus, macropore flow will only be generated at certain rainfall intensities, depending on antecedent rainfall conditions. These general observations were taken into consideration when developing a macropore routine.

The macropore routine which was included into the SM-model is a simplified accounting procedure (Eqs. (2)-(7) in Appendix II). The accumulated weighted sum of rainfall for the latest three days determines if macroflow is initiated. To consider the effect that dry spells may have on the extension of the macropore system, the accumulated sum of soil moisture deficits is incorporated into the calculations.

It turned out that the inclusion of a macropore routine significantly improved the fits (Fig. 3). However, the degree of improvement and the optimized values of the two parameters in the routine varied between different sites. This was however expected as there exists a large variability in the occurrence of macropores.

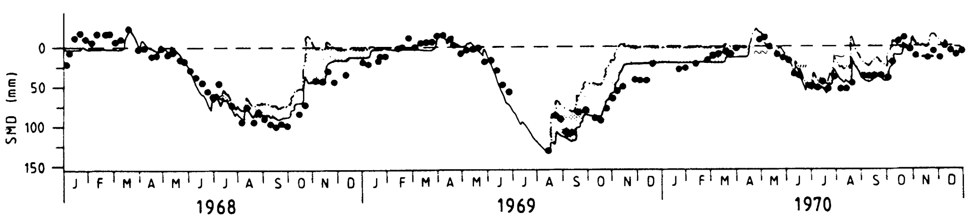

Fig. 3. Measured (points) and simulated (optimization period) soil moisture deficits for Sjöängen 7 (1968-1970). Dash-dotted curve shows predictions when macropore routine was not included.

Spatial Variations of Soil Moisture in the Basin

Effects of Vegetation Cover

The differences in soil moisture dynamics between grassland and forest were analyzed by comparing measured and simulated soil moisture from two forest and one grassland tube at Sjöängen. The forest tubes were situated just a few metres apart. The grassland tube was situated in a clearing, about 50 metres away from the forest tubes. Topographic characteristics and soil textures were similar. Gradients were below 3°. In spite of these similarities there occurred considerable differences in the soil moisture dynamics (Fig. 4).

Simulated evapotranspiration and percolation rates are shown in Fig. 5. In spring, total evapotranspirations was largest from the forest where photosyntetic activity was initiated earlier and where interception occurred. During summer, the interception losses were partly compensated by the well known negative feed-back which reduces transpiration rates when water is intercepted (Eq. (12) in Appendix II) At one of the forest sites (Sjöängen 8), the soil moisture depletion had only a

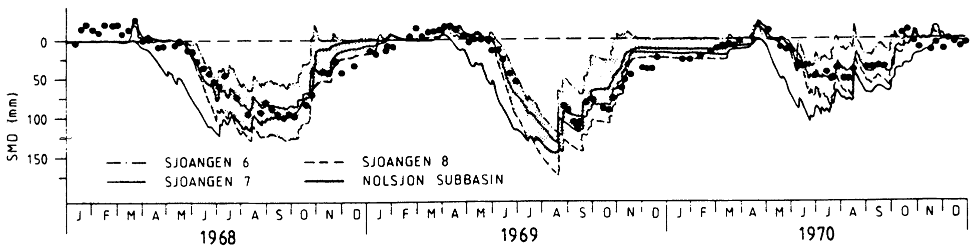

Fig. 4. Measured and simulated soil moisture deficits. Simulated with the SM-model for the forest stations 7 and 8 at Sjöängen and with the PULSE-runoff model for the Nolsjön subbasin as a whole. Measurements (points) refer to station 7 (where the most frequent measurements took place). The PULSE model simulation is discussed in the section »Groundwater Recharge Regime and Runoff Formation Process«.

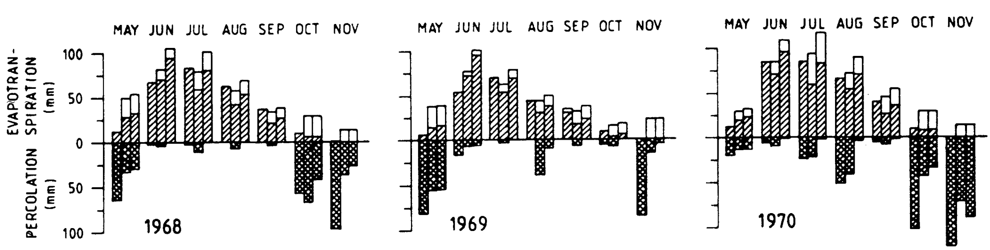

Fig. 5. Simulated interception (unfilled) and transpiration (dashed mm/month). From left to right: Sjöängen 6 (grassland), Sjöängen 7 and 8 (forest). Chequered columns show percolation (mm/month).

minor effect on transpiration rates except after prolonged dry spells. At the other forest site (Sjöängen 7), and at the grassland site (Sjöängen 6), transpiration rates were significantly reduced in times of soil moisture deficits. The large soil moisture depletion at Sjöängen 8 was probably caused by the existence of a deeply penetrating root system in the surroundings of this tube. Deep rooted trees can maintain sufficient water supply in spite of high water stress in the upper parts of the soil profile. During autumn, transpiration demands are low but interception losses can be considerable (Fig. 5).

The largest difference of soil moisture deficits between the grassland site and the forest site Sjöängen 7 occurred during autumns when heavy rainfalls initiated macropore flow to a larger extent at the forest soil and total evapotranspiration rates were larger from the forest due to interception.

Recharge to field capacity was reached simultaneously at the forest stations, in spite of the larger soil moisture deficits that occurred after dry periods at Sjöängen 8. This was probably due to a larger degree of macroporosity around the tube Sjöängen 7.

In summary, the distribution of living roots, determined the relationship between actual and potential transpiration. The distribution of decayed roots and other macropores determined the degree of recharge following a heavy rainfall. The variations in intercepted amounts had affected the total evapotranspiration rates when transpiration demands were low.

Effect of Deforestation

At the soil station Spekeläggen (Fig. 1) a clear cut took place during the winter 1971/1972. The vegetation in the clearing consisted of a mixture of grasses, mosses and small shrubs.

Groundwater measurements were made in 30-40 wells and tubes in the Velen basin. At each soil station, measurements of groundwater levels were made. By comparing groundwater levels at Spekeläggen with averages from the Velen basin, it was concluded that the water table was increased after the clear cut. The increase was most significant in periods when the average groundwater level in the area was low (Fig. 6).

The effects on soil moisture dynamics were studied by comparing the observations after the clear cut with simulations made with the model, optimized against field data collected before the clear cut (1968-1970, Fig. 7). Compensation was made for the earlier mentioned drift which had caused too high absolute values of the neutron probe measurements.

The comparison showed that the measured soil moisture was significantly higher than the simulated values. Three hypotheses that might explain the wetter conditions in the soil after the clear cut can be suggested:

1) Reduced interception losses after the clear-cut.

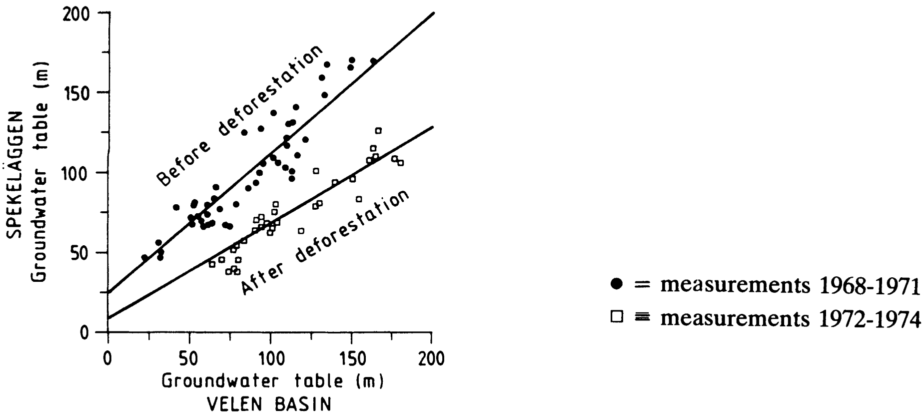

Fig. 6. Relation between groundwater levels (below soil surface) at Spekeläggen before (1968-1971) and after deforestation (1972-1974), achieved by regressing the time series against average groundwater levels in the Velen basin as a whole.

2) A faster decrease of actual transpiration rates when soil moisture deficits are developed and thus less transpiration losses.
3) A time delay in the initiation of transpiration during spring after the deforestation.

The first hypothesis was tested by running the model without the interception routine but otherwise with the same set of parameters that were used before the deforestation. The effects on summer soil moisture deficits were negligible.

The next step in the analysis was to reoptimize the SM-model against measurements made after the deforestation (1972-1973). The optimization showed that the parameter LP which decides at which soil moisture deficit level that transpiration is reduced, (Eqs. (2)-(3) in Appendix I) was increased from 0.40 to 0.83. This was probably an effect of the removal of deep rooted vegetation. There also seems to

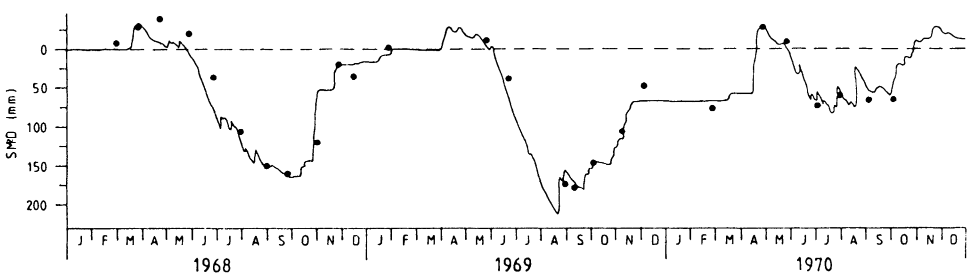

Fig. 7. Measured (points) and simulated soil moisture deficits at Spekeläggen before deforestation (1968-1970). All available soil moisture data where used for the parameter optimization.

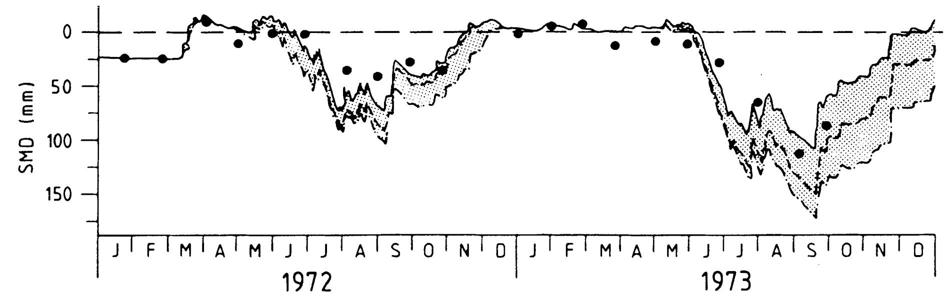

Fig. 8. Measured and simulated soil moisture deficits at Spekeläggen after deforestation (1972-1973). Dash dotted curve shows simulation with the forest optimization, dashed curve with forest optimization but without interception. The solid curve represents simulation with the deforestation optimization.

be a time delay in the initiating of the transpiration in spring. This was also compensated for in the model optimization (Eq. (10) in Appendix II).

In Fig. 8, it is demonstrated that reduced interception did only explain a minor part of the increased wetness that occurred after the deforestation. When the changes in the relationship between soil moisture deficit and transpiration rate and the changes in the time of initiating of transpiration were taken into consideration, the results were significantly improved.

Differences in Topography

Observed soil moisture dynamics (1968-1970) from five forest soil stations (Fig. 1) were compared. Field capacities varied between the stations, due to differences in soil characteristics and due to varied depths to the water table. To make the soil moisture measurements comparative in order to analyze the importance of differences in topography, the measurements were normalized on basis of the relative amounts of plant available water in the top metre of the soil (Fig. 9).

$$\left(\frac{SM}{AW}\right)_i = \frac{SM_i - SM_{wp}}{SM_{fc} - SM_{wp}} \tag{1}$$

where

$(SM/AW)_i$ – relative amount of plant available water on day i (mm)
SM_i – SM on day i (mm)
SM_{wp} – SM at pF = 4.2 (mm)
SM_{fc} – SM at pF = 2.0 (mm)

During times with large amounts of snowmelt and after prolonged rainy periods, soil moisture contents above field capacity (SM/AW = 1.0) existed at all stations except Ladfallet, situated in the highest part of the basin. At Spekeläggen, close to the outflow from the basin, saturated conditions could be kept for prolonged times.

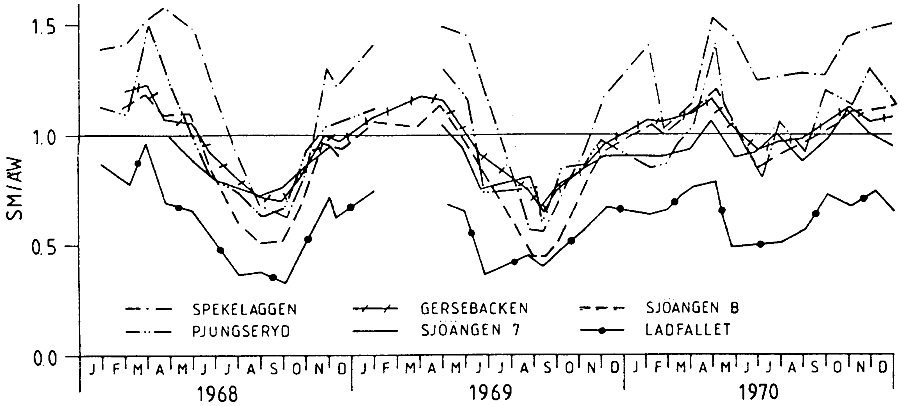

Fig. 9. Ratios between monthly measurements of plant available soil moisture and available soil moisture at pF 2.0 (AW) at six forest stations (1968-1970).

A high degree of saturation was achieved during wet spells at Pjungseryd, situated on the top of a hill in the upper part of the basin. This was probably caused by a low water storage capacity in the soil moisture zone and a low effective porosity, which lead to large fluctuations of the groundwater table. However, the saturated conditions were not kept for long times.

The sandy soil at Spekeläggen, which was part of the contributing area during wet periods, lost large amounts of water during dry periods (before the clear-cut) and achieved SM/AW ratios comparative to higher situated sites (Fig. 9). This demonstrates that sites which are outflow areas during wet periods are not necessarily among the wettest after prolonged dry spells. During summer, the type of root distribution and the water holding capacity seemed to be the most important factors determining the degree of dryness.

Ladfallet was the only site for which the inclusion of the temperature effect routine into the SM-model did not increase the explained variance (Andersson and Harding). This can have been due to the faster warming up during spring that occurs in a well drained soil.

Simulation as a Tool in Estimating Overall Basin Behavior

In the previous section it has been argued that relatively simple SM-models can describe local soil moisture fluctuations with good accuracy. The next question is: Is it possible to use the calculated percolation rates, which represents a few points in the area, in an estimation of groundwater recharge and runoff formation on a river basin scale? And does the simulated evapotranspiration agree with the actual evapotranspiration losses from the basin?

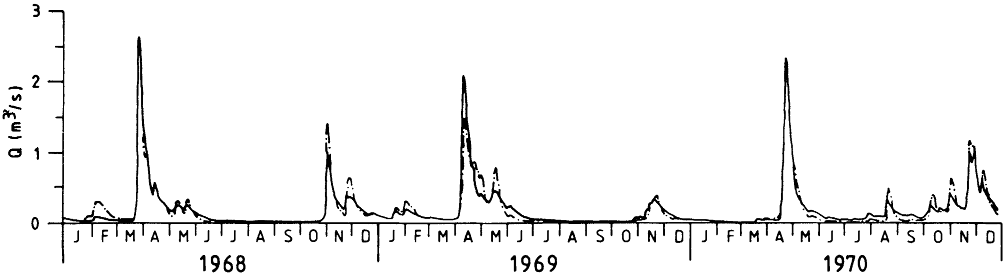

Fig. 10. Measured (dash dotted curve) and simulated (solid curve) runoff (PULSE-model) for Nolsjön 1968-1970. Model optimization from Bergström *et al.* (1985).

In order to assess this, simulated evapotranspiration and percolation were compared to measurements and calculations of various water balance components.

Groundwater Recharge Regime and Runoff Formation Process

Basin wide groundwater recharge can be calculated from a subroutine in the PULSE-runoff model. Bergström *et al.* (1985) optimized that model for Nolsjön which is a subbasin of the Velen drainage basin. For the independent period 1968-1970, the model gave an explained variance of 0.88. Observed and simulated runoff are shown in Fig. 10.

The sub-basin covers 18.1 km and has a low lake percentage (1.5 %), from which follows that responses on percolation are relatively fast. Seventy percent of the Nolsjön basin was covered by forest.

Measured (relative to the soil surface) groundwater levels at Sjöängen were similar to average groundwater levels in the Velen basin. This implicates that the wetness at Sjöängen can be considered representative for »average conditions« in the basin as a whole.

Monthly percolation rates, calculated with the PULSE-model (optimized against runoff data) and with the SM-model (optimized against soil moisture data) were compared.

Basin percolation was simulated with the SM-model, considering the allocation of the land use in the Nolsjön basin (Eq. (2)).

$$\mathrm{PERC}_{SJ} = 0.30\ \mathrm{PERC}_{G6} + 0.35\ \mathrm{PERC}_{F7} + 0.35\ \mathrm{PERC}_{F8} \tag{2}$$

where PERC_{SJ} = monthly percolation simulated with SM-model, representing average conditions in the Nolsjön basin. PERC_{G6}, PERC_{F7}, PERC_{F8} represents monthly percolation from Sjöängen 6 (grassland), Sjöängen 7 and 8 (forest) respectively.

In general, the differences between the simulated monthly percolation rates were small (Fig. 11).

Introduction of the »temperature effect« (Eqs. (8)-(11) in Appendix II) made

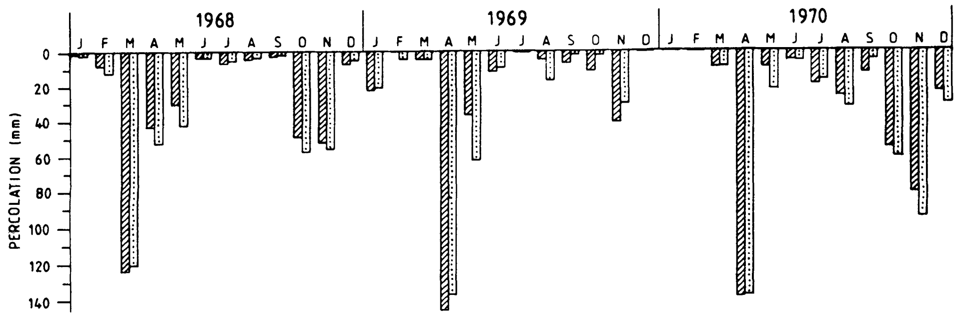

Fig. 11. Monthly percolation (mm/month) 1968-1970 from the top metre of the soil simulated for the Nolsjön basin with the PULSE runoff model (dashed columns) and for Sjöängen with the SM-model (dotted columns).

simulated soil moisture deficits develop slower in spring. The PULSE model, which does not include a »temperature effect«, underestimated soil moisture contents in spring (Fig. 4). Consequently, the runoff response to spring rains that occurred after the snowmelt 1969 was underestimated (Fig. 10).

According to simulations with the SM-model which includes a unsatuated macropore flow routine (Eqs. (2)-(7) in Appendix II), percolation occurred at several sites during two occasions with heavy rainfall in spite of large soil moisture deficits. The first event occurred in August 1969 when following a very dry summer, 50 mm of rain fell within one day. At that time, at most sites the groundwater levels were below the bottom of the access tubes. However at all those tubes which still could be used (Sjöängen 6, Gersebacken and Spekeläggen), a rise of the groundwater table took place in spite of considerable soil moisture deficits. At Sjöängen 6, the rise between 20 August and 3 September (85 mm rain) was from 1.35 m to 0.85 m below the soil surface. However, groundwater levels were still low and the percolation did therefore not result in any runoff.

The second event occurred in connection with a rainstorm of 60 mm (August 1970). At this occasion the runoff model simulated an increase from initial flow to peakflow which was only 60% of the measured. The following recession was much faster in reality than simulated (Fig. 12). Macropore flow might explain the fast increase of the runoff and the quick recession that followed.

Evapotranspiration Regime

Basin evapotranspiration, calculated from the residual in the water balance for the Velen basin (Waldenström 1977) was compared with monthly sums of simulated interception and transpiration rates by letting simulations from Sjöängen 7 and 8 represent 50% each of the forest and Sjöängen 6 represent the open land. For the lakes, potential evapotranspiration was set equal to the potential value (Penman equation for water surfaces).

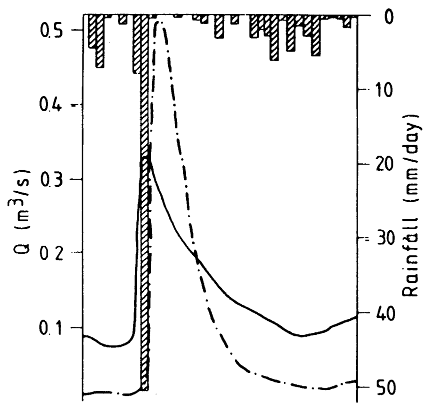

Fig. 12.
Simulated (solid curve) and observed (dash-dotted curve) runoff response to a heavy rainfall in August 1970, when soil moisture deficits still existed in the main part of the basin. The columns shows the daily rainfall amounts. Model optimiza tion from Bergström *et al.* (1985).

The soil moisture term in Waldenströms calculation was not available for August 1969 and therefore calculated as half of the soil moisture increase between July and September. In October 1968 and October 1970 large rainfall amounts were added to the soil after the monthly soil moisture measurements were made. Therefore, in order to get reasonable water balance residuals, in addition to Waldlenströms calculations, water balance residuals with the soil moisture term taken from the soil moisture simulations were calculated for these three months.

Comparison showed a good agreement between the evapotranspiration, calculated from water balance studies and evapotranspiration simulated with the SM-model (Fig. 13).

It was in other words shown that basin evapotranspiration can be estimated from the SM-model. Inclusion of the »temperature effect« and of the interception routine were probably factors that contributed to this promising result.

The model estimates of transpiration are built on the assumption that transpira-

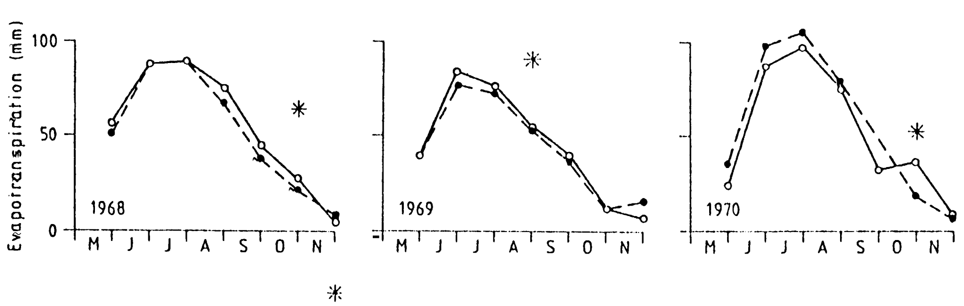

Fig. 13. Monthly evapotranspiration rates, calculated from the water balance equation for the Velen basin (solid curve) (Waldenström 1977) and as the sum of simulated transpiration and interception rates for Sjöängen (dashed curve) 1968-1970. For months without accurate estimates of the soil moisture term, Waldenström's values are shown, using asterisks.

tion will be reduced during periods when soil moisture deficits occur. This is verified by the water balance computations, which indicated the lowest evapotranspiration during the dry summer of 1969 and the highest during the wet summer of 1970.

Summary and Conclusions

The »contributing area« concept is usually confined to areas of saturation. However, this study has shown that the heterogeneous structure of forest soils may lead to flow through unsaturated soil. The dynamic source area will thus not simply be confined to areas of saturation. The slow soil moisture recharge that occurred as a response to autumn rainstorms and the rise of groundwater tables as a response to rainstorms in times when soil moisture deficits existed, imply that drainage can occur in this type of environment before the storage capacity of the soil is filled.

It was shown that the topographic locality became less important in determining spatial differences of soil moisture contents after prolonged dry spells. Therefore, it is necessary also to include other factors when mapping the extension of areas with various hydrological behavior. The most important are the water holding capacity and the type of vegetation, which not only determine the transpiration and interception losses but also influence the degree of macroporosity.

The general conclusion is that the analyzed type of SM-model is very useful to assess soil moisture dynamics and other hydrological phenomena on a local scale. Admittedly, further improvement including careful examinations of the approximations made so far, is needed before basin general simulations are reliable. However, the fit between the evapotranspiration, calculated as residual in the water balance, and the simulated evapotranspiration shows that this type of model can be useful even on a basin scale. The fact that the amount of basin evapotranspiration seldom is possible to determine from direct measurements makes this result especially interesting.

Acknowledgements

This study was supported by the Swedish Natural Science Research Council (NFR) and by the Bank of Sweden Tercentenary Foundation. The modelling part of the work was mainly carried out at the Swedish Meteorological and Hydrological Institute (SMHI) and during my stay at Institute of Hydrology, Wallingford, UK. I am especially grateful to Prof. Sten Bergström who provided me with the HBV-model and to Mr. Todor Milanov who was in charge of the soil moisture measurements in Velen. Prof. Malin Falkenmark and Prof. Ulrik Lohm gave many valuable comments on the manuscript. Figures were drawn by Lisbeth Samuelsson.

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Received: 9 September, 1982

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Appendix I – Equations used in the SM-accounting routine in the original HBV-model

Percolation

Calculation of how much percolation (PERC) that each mm of rain or snowmelt (INSOIL) will cause. SM ≡ calculated soil moisture storage above wilting point (mm), SMAX ≡ maximum water content above wilting point (mm), ß ≡ parameter.

$$\frac{\mathrm{PERC}}{\mathrm{INSOIL}} \equiv \left(\frac{SM}{S\mathrm{MAX}}\right)^{\beta} \quad \text{(mm)} \tag{1}$$

Evapotranspiration

Calculation of actual transpiration (Ea) on day i. Ep_m ≡ monthly long time average of daily potential evapotranspiration for actual month (mm day^{-1}), Lp ≡ limit for potential evapotranspiration (% of SMAX).

$$\text{when} \quad SM_i > Lp \; S\mathrm{MAX} \qquad Ea_i = Ep_m \qquad (\text{mm day}^{-1}) \tag{2}$$

$$\text{when} \quad SM_i < LP \; S\mathrm{MAX} \qquad Ea_i = Ep_m \left(\frac{SM_i}{S\mathrm{MAX}_i}\right) \qquad (\text{mm day}^{-1}) \tag{3}$$

Source: Bergström (1976).

Appendix II – Used equations which are not included in the traditional HBV-model

Percolation when calculated SM-content exceeds water content at field capacity

In the traditional HBV-model, the parameter called FC is representing both field capacity and maximum water content. In the SM-model, water contents can reach values above field capacity. When SM contents exceeds field capacity (FC), drainage is calculated as the sum of PERC (Eq. (1) in Appendix I) and SPERC (Eq. (1) in Appendix II). SPERC is calculated every day when SM exceeds FC. SMAX ≡ water content when the soil is fully saturated (mm), MAXPERC ≡ parameter, representing daily percolation at soil saturation.

$$S\mathrm{PERC}_i = MAX\mathrm{PERC} \left(\frac{SM_i - FC}{S\mathrm{MAX}-FC}\right) \quad \text{(mm)} \tag{1}$$

Drainage through macropores

Drainage through macropores when the average SM-content in the soil is below FC is calculated. MACROFLOW1 is dependent on the accumulated soil moisture deficit (ASMD) and recent rainfall conditions. MACROFLOW2 is only dependent on recent rainfall conditions.

Calculation of accumulated soil moisture deficits (ASMD) on day i. ASMD is put to zero if SM exceeds FC or otherwise the first of March each year.

$$ASMD_i \equiv ASMD_{i-1} + (FC - SM_i) \quad \text{(mm)} \tag{2}$$

Calculation of an accumulated, weighted sum of rainfall (WETSUM) for the latest three days (no macropore flow is calculated when WETSUM is less than 30 mm).

$$\mathrm{WETSUM} = \mathrm{RAIN}_i + 0.5\;\mathrm{RAIN}_{i-1} + 0.25\;\mathrm{RAIN}_{i-2} \quad \text{(mm)} \tag{3}$$

Calculation of the value (0-1) of the index MACRO, showing the stage of development of cracks and fissures (WETMAX ≡ parameter, deciding at which value of ASMD that the macropore net reaches its maximum extension).

$$\text{MACRO}_i \equiv \frac{ASMD_i + 1}{\text{WETMAX}} \tag{4}$$

Calculation of the maximum capacity of the SM-zone to store INSOIL on day i, determined by antecedent rainfall conditions.

$$\text{STAY}_i = (30 \; (\frac{1}{\text{MACRO}_i})) - \text{WETSUM} \quad \text{(mm)} \tag{5}$$

Calculation of macroflow through unsaturated soil matrix on day i (KDRAIN ≡ parameter).

$$\text{if} \quad \text{WETSUM}_i > 30 \quad \text{MACROFLOW } 1_i = \text{RAIN}_i - \text{STAY}_i \quad (\text{mm day}^{-1}) \tag{6}$$

$$\text{MACROFLOW } 2_i \equiv K\text{DRAIN } (\text{WETSUM}_i - 30) \quad (\text{mm day}^{-1}) \tag{7}$$

Temperature effect

A temperature index is used to reduce actual transpiration rates when soil and air temperatures act as limiting factors on evapotranspiration.

Calculation of five days averages of air temperature (TFIVE) on day i. T_5 = average air temperature on day i (°C)

$$T\text{FIVE}_i = \sum_{i=1}^{5} (T_i/5) \quad (^{\circ}\text{C}) \tag{8}$$

Calculation of the value of the temperature index (INDEX) on day i (varying between 0.0 and 1.0). TSTART = parameter, deciding at which value of TFIVE that evapotranspiration will be initiated, DAYS ≡ parameter, deciding the number of days with TFIVE above TSTART needed until the temperature ceases to have an effect on transpiration rates.

$$\text{when} \quad T\text{FIVE}_i < 0.0 \quad \text{or} \quad i = \text{Oct 1st} \quad \text{INDEX}_i \equiv 0.0 \tag{9}$$

$$\text{when} \quad T\text{FIVE}_i > T\text{START} \quad \text{INDEX}_i = \text{INDEX}_{i-1} + (1/\text{DAYS}) \tag{10}$$

Calculation of actual transpiration (E) on day i when the temperature effect is considered. Ea is calculated from Eqs. (2)-(3) in Appendix I.

$$E_i = Ea_i \; \text{INDEX}_i \quad (\text{mm day}^{-1}) \tag{11}$$

Rainfall interception

Calculation of the fraction of day i when canopy is wet (W_i) and no transpiration is considered to take place. RAIN_i ≡ rainfall on day i, TIME ≡ the ratio between the time that the canopy is wet during and immediately following rainfall and the number of rain hours, I = mean rainfall intensity (mm/hr). Standard values of TIME (1.5) and I (1.4) were used.

$$W_i = 1 - (\text{RAIN}_i \; \text{TIME} / (I \;\; 24)) \tag{12}$$

Source: Calder and Newson (1979)

Calculation of rainfall interception (IC) on day i. γ = parameter, optimized against interception measurements.

$$IC_i \equiv (1 - e^{(-0.5 \; \text{RAIN}/\gamma)} \; W_i) \quad (\text{mm day}^{-1}) \tag{13}$$

Source: Andersson and Harding.