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Simulation of Surface Runoff and Pipe Discharge from an Agricultural Soil in Northern Sweden

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In order to test the ability of a physically based water and heat model to predict surface runoff and pipe discharge, adaptations were made to an agricultural field in the north of Sweden. A five-year period was selected, including observations of meteorological data, frost in the soil and discharge. Basic model requirements on soil properties, *i.e.,* the water retention curve and the saturated hydraulic conductivity, were available from a previous investigation. Unsaturated conductivity was estimated from the water retention curve and by assuming a substantial influence of macro pores in the subsoil. Snow properties and thermal soil properties were adjusted to obtain a reasonable agreement with observed frost depths for areas with barley and with grass leys. Surface runoff was the dominating part of the total runoff, especially during conditions of frozen soil. The simulated discharge agreed well with the general partitioning between surface runoff and pipe discharge but discrepancies occurred in their temporal patterns. A probable explanation of these discrepancies was that the model did not account for the enhanced spatial heterogeneity in water flow through snow and in partially frozen soil.

Introduction

The importance of surface runoff as a source for the formation of stream runoff has long been an area of conflicting opinions among hydrologists. Recent studies in Sweden by Rodhe (1984 and 1985) have clearly demonstrated the importance of ground water flow in forested catchments during the snow melt period. The ^{18}O

found in streamwater after snow melt indicated that the major portion of the snow melt infiltrated into the soil profile and it was assumed that only a minor portion, in the saturated region close to the stream, formed surface runoff. In contrast, Gustafson *et al.* (1984) reported that 82% of the spring flood was surface runoff from an agricultural field during a five-year period in the north of Sweden.

Occurrence of frost in the soils may be one of the key-factors behind the differences reported in the magnitude of surface runoff. The forested catchments investigated by Rodhe (1984 and 1985) were all heterogeneous with respect to an organic soil cover which prevented the frost from penetrating deeply and homogeneously. In the agricultural soil investigated by Gustafson *et al.* (1984) the frost could penetrate deeper because the soil was bare after the end of the growing season and because of the lack of an organic soil layer.

Modelling of water movements in partially frozen soils was first presented by Harlan (1973) and was followed later by a number of similar studies, *e.g.*, Jansson and Halldin (1979). Most of the models have been tested only for a limited data set, commonly covering a short time period. This restricted our hope to be able to accurately predict water movements in frozen field soils. In the study of Halldin *et al.* (1980) a considerable delay of the percolating water flow was found during a year with deep frost. During another year, with a shallow frost layer, the infiltrating water penetrated the frost without any delay. Unfortunately, these differences could not be quantitatively tested because of the lack of water flow measurements.

The purpose of this paper is to present our efforts to test the soil water and heat model (Jansson and Halldin 1979) for the agricultural soil investigated by Gustafson *et al.* (1984). Emphasis was placed on the partitioning between surface runoff and drainage through pipes and the importance of a grass cover on the development of frost in the soil. The present work could be considered as a first step towards modelling of nitrogen losses from soils strongly influenced by frost.

Materials and Methods

Site Description

The field is situated at Röbäcksdalen, immediately to the south of the town of Umeå in northern Sweden. The field was systematically tile-drained in 1952 with pipes at 1 m depth (Hallgren and Rietz 1963) and simultaneously a measuring station was constructed where the tile-drainage and the surface runoff could be measured separately in two triangular weirs. To make the collection of surface runoff possible the field was enclosed by a soil bank on three sides (Fig. 1) and at the lower short side of the field a small ditch was established from which the water was led to the measuring station through a surface water collector.

The soil immediately beneath the plough layer consists mainly of fine grained

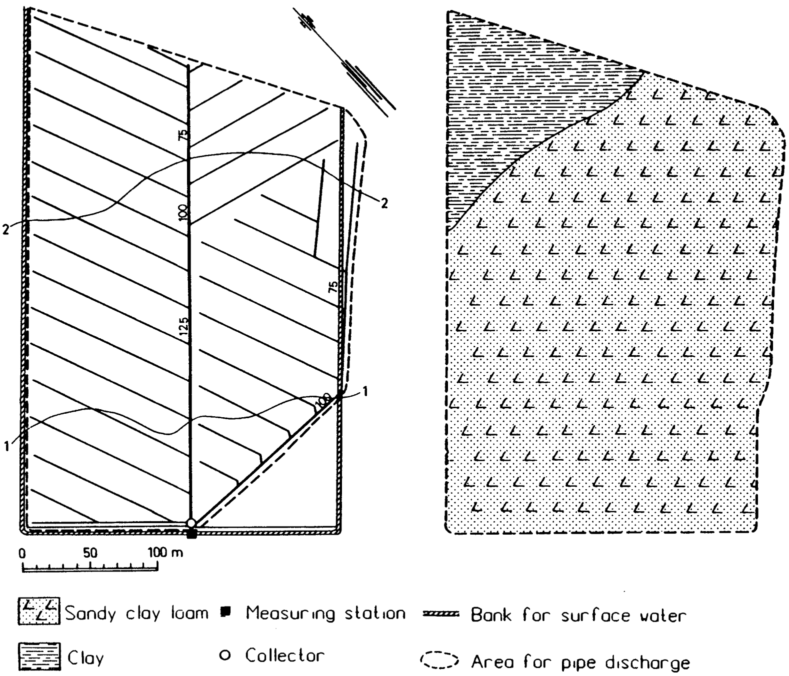

Fig. 1. Map of the experimental field at Röbäcksdalen showing both the drainage system and the geology of the area.

sediments dominated by the silt fraction. The contribution of silty clay and clay is considerable in some places. The sediments were deposited in the sea which covered the area around the field about a thousand years ago. The soil material originates partly from sludge which was transported by the river Umeälven and deposited in the sea at that time and partly from material which was surged out by wave actions on the more elevated terrain SW and SE of the field.

The soil below the plough layer is rich in vertical crack plains and has a stable and well-developed structure with rust deposits formed on the outside of the aggregates. In the crack plains, and to some extent around the aggregates, the crops often establish a relatively dense root-carpet.

Perennial crops as grass ley and annual crops as barley are cultivated at the field with different degrees of cover for each year. During the investigated period 1976 to 1980 annual crops, *i.e.,* mainly barley occupied 58, 81, 81, 66 and 64 per cent of the total area.

Model Description

The model was described in detail by Jansson and Halldin (1980). A brief description of structure and the major assumptions is given below.

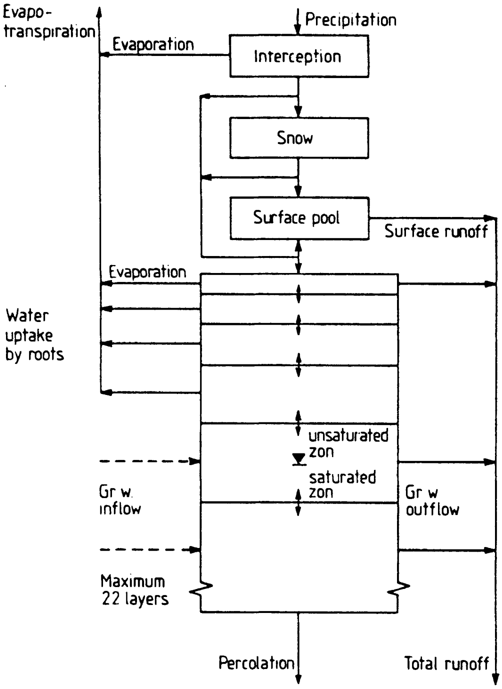

Fig. 2.
Model structure of the water and heat model. All flows in the soil profile except water uptake by roots represent both water and heat (modified from Jansson and Halldin 1979).

The water and heat model (Fig. 2) is based on two coupled differential equations describing heat and water transport (derived from Fourier's and Darcy's laws, respectively) in a one-dimensional soil profile (Jansson and Halldin 1979). Snow dynamics, frost, evapotranspiration, precipitation, groundwater flow, plant water uptake and drainage flow are included. The model predicts soil climate variables (e.g. soil temperature, water content, etc.) with a daily resolution at any level in the soil profile.

The model, originally developed for a forest soil, has been modified and tested for agricultural soils (Jansson and Thoms-Hjärpe 1986) and for application to agricultural watersheds (Lundin 1984).

Water flow during partially frozen conditions is based on a similarity between drying-wetting and freezing-thawing. The content of unfrozen water and the water tension is given as a function of the heat content of the soil (see Jansson and Halldin 1980). The flow of water is calculated by Darcy's law in the same way as for unfrozen conditions. In order to minimize unrealistically high upward flow towards the frost boundary, a new procedure of calculating the interboundary conductivity between two adjacent layers was introduced. Instead of using a weighted mean value of two conductivities, the lowest conductivity of the two adjacent layers is used. By using this procedure, empirical consideration is paid both to the numerical error caused by the extremely steep tension gradient close to the frost boundary and to the resistance for water flow caused by the development of ice lenses. The new procedure, which is restricted to upward flows, has been tested by Lundin (1987).

Because of the one-dimensional vertical structure, lateral flow is not explicitly calculated. Water flow to drainage tiles occurs when the simulated groundwater is above the level of the tiles, *i.e.,* flow occurs directly from a layer to drainage tiles when water content is at saturation. The flow rate is proportional to the hydraulic gradient (estimated from the depth of the groundwater table and the density of the drainage tiles), the thickness and the saturated hydraulic conductivity of each layer.

Surface runoff can occur because of a limited infiltration capacity or a limited conductivity in the soil profile. When the infiltration rate exceeds the infiltration capacity a surface pool of water is formed on the soil surface. Drainage of the surface pool to the stream is simply calculated as a portion of the amount of water in the pool. If water infiltrates into the uppermost soil layer and the conductivity is too low in the subsoil, the excess of water (above the porosity of the soil) will in the present version of the model be directed directly to the stream without any delay.

The water and heat model uses standard daily meteorological input data, *i.e.,* air temperature, humidity, precipitation, global radiation (or duration of sunshine or cloudiness) and wind speed. If necessary, precipitation, air temperature and a simple estimate of potential evapotranspiration can suffice as input data. Parameter values for hydraulic and thermal soil properties can be estimated from standard soil physical characteristics, or independent measurements can be used.

Most important requirements of model parameters concern soil and plant properties. Soil properties are defined by the water retention curve and the hydraulic conductivity as a function of water content or water tension. Plant properties are those controlling water uptake and transpiration. Transpiration is calculated with a combination formula (Monteith 1965), accounting for surface resistance. The reduction of water uptake because of limiting soil water availability is calculated from an empirical formula (Jansson and Halldin 1979).

Adaptation of the Soil Water and Heat Model

The meteorological data required as driving variables to the model were selected from two meteorological stations located close to the investigated field. Daily mean values of air temperature, air humidity, wind speed and cloudiness were taken from the synoptic station at Umeå (within the network of the Swedish Meteorological and Hydrological Institute), 2 km from the field. Precipitation was measured at the field.

The precipitation was not corrected for the aerodynamic error. Normally, snow precipitation is underestimated by measurements but the snow depth observations did not indicate this in the present case. One reason may be that snow drift occurred over the open field investigated. The daily mean air temperature was increased by 1°C for three days in December 1979 in order to simulate a reasonable

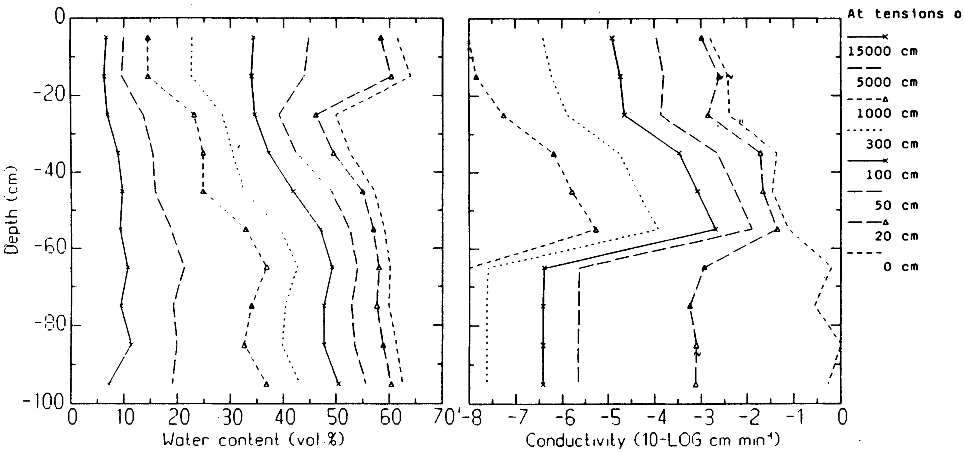

Fig. 3. Soil moisture characteristics (left) and unsaturated conductivity (right) assumed in the simulation.

distribution between rain and snow. During these days a mass of warm air entered the region with considerable amounts of mixed rain and snow precipitation. The original daily mean temperature resulted in an overestimation of snow, because it represented also the colder periods prior to and after the period with the most intensive precipitation.

Laboratory analyses of texture, water retention and saturated conductivity were available from three different pits, 1 m deep, in the field (Andersson and Wiklert 1977). Based on these data, functions for the water retention curve and for the unsaturated conductivity were estimated to fulfil model requirements (Fig. 3). The soil profile represented in the model was 3.5 m deep and divided into ten layers with thicknesses from 10 to 75 cm, with the thickest layers in the bottom of the profile. Soil properties for the layers below 1 m were extrapolated from the deepest layer in the investigated pits.

The major uncertainty in soil physical properties concerned the unsaturated conductivity. In sandy soils we have previously used the equation by Mualem (1976) to estimate the unsaturated conductivity from the water retention curve only. A straightforward application of this procedure would have resulted in misleading results because of the macropores in the present soil. The macropores are well developed and quite stable, especially in the deeper parts of the soil profile below 50 cm depth, giving very high saturated conductivities. In the same way as Jansson and Thoms-Hjärpe (1986), we now made a separate account for the macropores by assuming a very rapid decrease in the unsaturated conductivity when the largest pore sizes were drained. The equation by Mualem was thereby restricted to the capillary pores and a new matching value to this equation was estimated to give the unsaturated conductivity presented in Fig. 3.

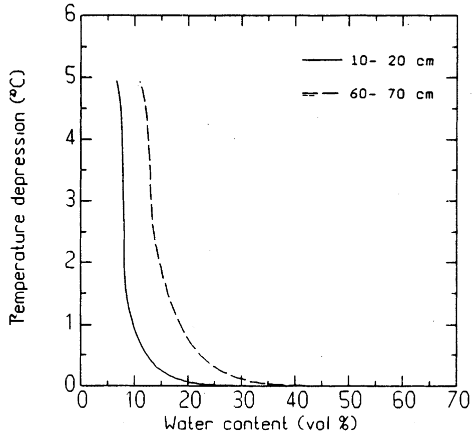

Fig. 4. The assumed content of unfrozen water at temperatures below 0°C.

Thermal soil properties were estimated from the dry bulk density as calculated from the porosity (Fig. 3) and a particle density of 2.65 g cm^{-3}. Thermal conductivity was estimated by using Kersten's equations (1949). Differences in soil surface boundary conditions between the ley and the barley were considered by reducing the thermal conductivity of the uppermost 10 cm of the grass-covered soil to a value that corresponds to humus. The humus thermal conductivity was chosen according to de Vries (1975). The heat capacity as well as the hydrological properties of the soil were kept unchanged. The soil surface temperature was assumed to be similar to the daily mean temperature except in situations with snow (see Jansson and Halldin 1980).

The freezing point depression of the soil (Fig. 4) was based on the water retention curve by using the pore size distribution index in the expression by Brook and Corey (1964) and the water content at the wilting point. Since no independent measurements of the freezing point were available, the estimation of the freezing point was based on information from Beskow (1935) and by adjusting the freezing point to achieve an agreement between simulated and measured frost depths.

Snow melting coefficients accounting for the effects of air temperature and of global radiation were adjusted until reasonable agreement was achieved with snow depth observations. The temperature coefficient was put to 3 kg m^{-2} day^{-1} $°C^{-1}$ and the global radiation influence was varied from 0.1 to 0.35 kg MJ^{-1} depending on snow age. No information on water equivalent in the snow was available which made the simulation of snow density and the corresponding thermal conductivity of snow uncertain. The density of new snow, free from rain, was put to 50 kg m^{-3}. Depending on wetness, the density of old snow increased by up to 100 kg m^{-3} and, depending on the total mass of snow, it was assumed that there was a further increase of 1.5 kg m^{-3} for each 1 kg m^{-2} increase in the total mass of snow. The thermal conductivity of snow was calculated using a function derived from Snow Hydrology (1956).

No efforts were made to adjust the crop properties after comparison with the measurements of drainage and surface runoff from the field. The adaptation of the model was made different for the two dominating crops barley and grass ley. The annual course of their respective surface resistances and leaf area indexes were based on experiences gained when the model was adapted to similar crops at Kjettslinge in Central Sweden (Jansson and Thoms-Hjärpe 1986; Jansson *et al.* 1987). Unfortunately, the higher transpiration and the higher interception losses simulated from the grass ley could not be tested since the drainage measured did not represent the treatments separately.

The depth distribution of roots and the maximal root depth (70 cm) was assumed to be similar for the two crops at the height of the summer, with 80% of the water uptake demand in the 0-25 cm layer. The roots of the ley were only slightly varied during the year whereas the barley roots developed in the same way as assumed for the leaf area index above ground. The critical soil water tension, where the reduction of potential transpiration starts, was put to 400 cm water for both crops.

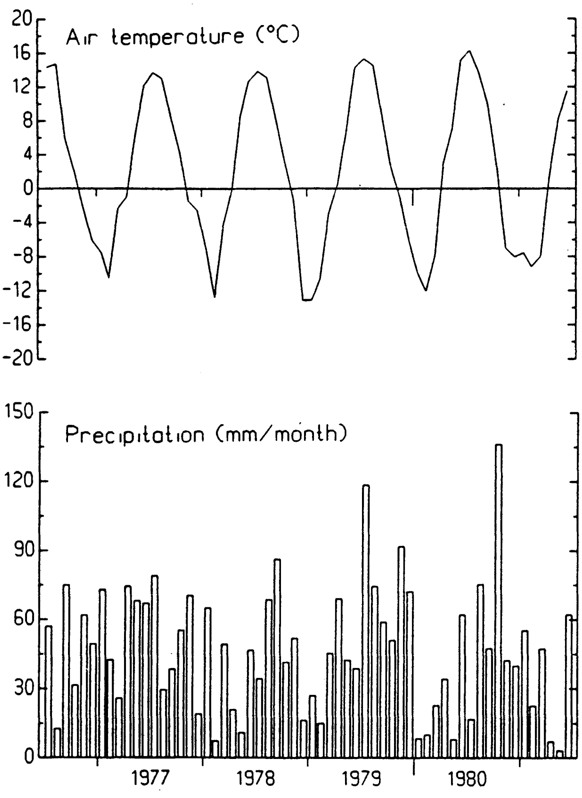

Fig. 5. Monthly mean values of air temperature and monthly totals of precipitation.

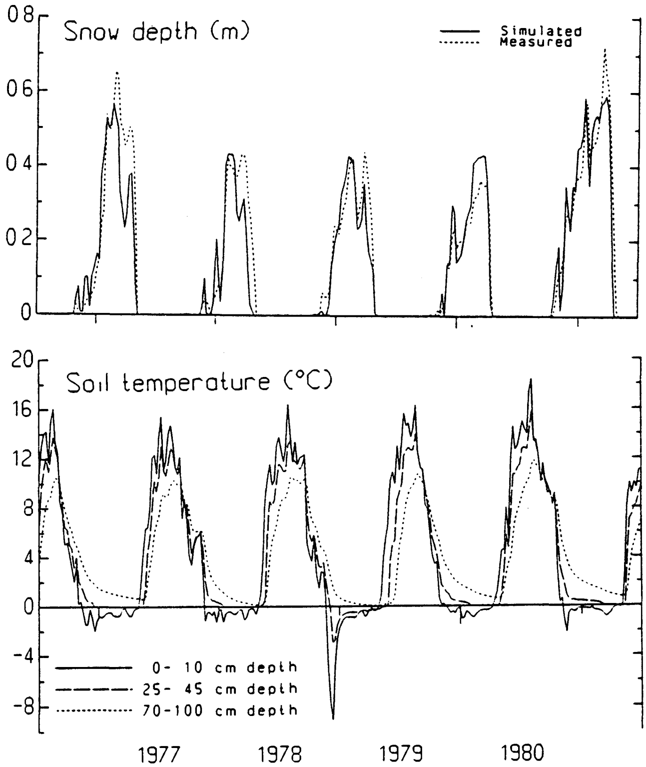

Fig. 6. Simulated and measured snow depth and simulated soil temperatures.

Results and Discussion

The simulation period, covering five years from July 1st 1976 – June 30th 1981, represented a wide range of shifting temperature and moisture conditions (Fig. 5). A tendency for summers to be warmer from 1977 to 1980 occurred but no similar tendency was seen for the winter temperatures. The coldest winter occurred in 1978/79 and the winter with the longest duration was 1980/81, the shortest being in 1977/78. The lowest monthly precipitation totals normally occurred during winters and the corresponding highest totals occurred during summers and autumns. The summer of 1979 was the wettest and it was followed by an equally wet autumn which was followed in early 1980 by five months which together received less than 90 mm of precipitation.

The simulated snow depth agreed reasonably well with the observed depths after the adjustments of snow melt coefficients (Fig. 6). The exceptions, all orginating from individual meteorological events, were believed to be of minor importance

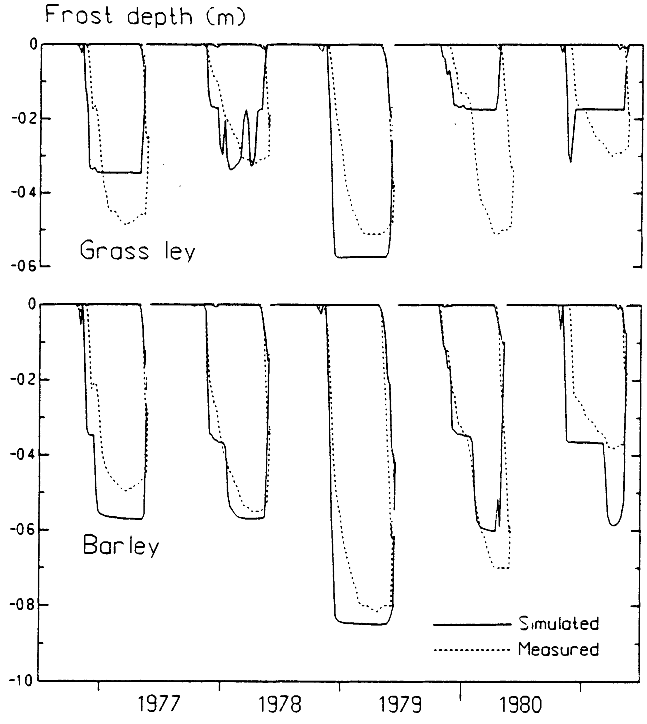

Fig. 7. Simulated and measured frost depths.

for the simulation of frost so they were accepted. During both 1977 and 1978 the model exaggerated the compaction of the snow in connection with a melt period. In 1976/77 and 1980/81 the maximal snow depth was underestimated, in contrast to the overestimation during the winter of 1979/80. We believe that a further improvement of the simulated snow depth could be achieved only by increasing the precision in the meteorological input to the model. The use of daily mean temperatures for the partitioning between rain and snow may be questioned (see the section under Adaptation of the Model).

When simulated and measured frost depths were compared (Fig. 7) the best agreement was obtained for barley, where the frost was deepest. The inclusion of a low thermal conductivity of the uppermost soil as a way of accounting for the grass cover in the ley was not successful in all years. Especially during the years 1979/80 and 1980/81, simulated frost depths were too shallow and with different temporal distributions compared with the measurements. To some extent the difference in patterns could be explained by the rough discretization of the soil profile in the model, as for 1977/78, but the differences for the ley were too large to be satisfactory. An improvement of how a grass cover influences the boundary between soil and snow or between soil and atmosphere in the model is needed. A more detailed approach based on the energy balance at the soil surface has previously been used by Alvenäs *et al.* (1986) and Jansson *et al.* (1987) to explain differences in soil temperatures between barley and grass ley during summer. Such an approach may also be used during autumn and winter. However, it was not used in the present study since the soil temperature measurements needed for a careful test were not available.

Simulated soil temperatures (Fig. 6) illustrated the need for a better quantification of soil frost than the frost depth only. The lowest temperatures at 5 and 35 cm depth occurred in connection with the growth of the deep frost during 1978/79. But at the time for maximal frost depth the same soil levels were substantially warmer with temperatures only slightly below 0°C. In the other winters the temperature at 35 cm depth never dropped below 0°C, but the duration for the 0°C temperature varied in the same way as the frost depth. The temperature rise during spring varied substantially from year to year, largely reflecting the prevailing frost. Normally, when the frost had been shallow, the temperature rise was only slightly delayed between the different soil depths, as in the spring of 1981. On the other hand, as in the spring of 1979 when the frost had been deep, a substantial delay occurred in temperature rise between the different soil depths.

An interesting phenomenon was the variation of simulated soil temperatures at 85 cm depth. The lowest simulated temperature at that depth commonly occurred in connection with the infiltration of melt water from snow. Especially during the spring of 1978 it was noticed that the simulated temperature dropped to 0°C at 85 cm depth simultaneously as a rapid temperature rise was simulated at 5 cm depth. In this case, the temperature would have been a good tracer for the water flow, which will be discussed further below.

The whole field which was as mean values for the whole five-year period covered to 70% of annual crops and to 30% of grass ley. Since the measurements of runoff represent the mixture of both annual crops and of grass leys it was difficult to interprete the difference obtained in simulated runoff between the two soil covers. The accumulated sum of runoff simulated for barley was about 200 mm above the measured sum of 1,180 mm and the corresponding simulated runoff for the ley was about 100 mm less than the observations (Fig. 8). A similar tendency was also observed for the simulated pipe discharge but in this case the barley simulation was just above the measured sum of 380 mm whereas the ley simulation was considerably below the measurements. The model overestimated the accumulated sum of surface runoff but, on the other hand, the best agreements with regard to the runoff patterns were obtained between the barley simulation and the measured runoff (Fig. 8).

The barley simulation overestimated the total runoff during the winters of 1977/78 and 1979/80 (Fig. 9). The reason may be that the assumed water retention capacity of snow was too low (7% of the total snow mass) or that input data on precipitation was erroneous. The simulated snow depth was reasonable when com-

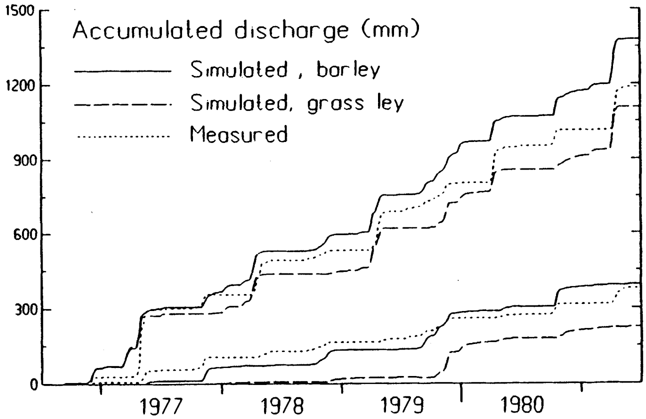

Fig. 8. Simulated and measured accumulated water flows, pipe discharge (only the lower curves) and the total runoff (the upper curves).

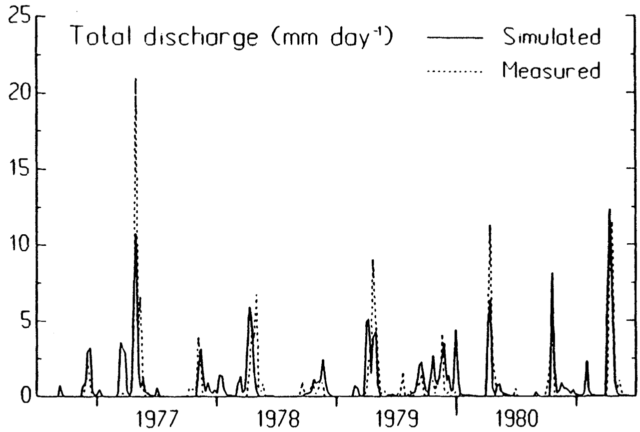

Fig. 9. Simulated and measured weekly mean values of the total runoff. Simulated values represent barley.

pared to observations but a tendency for a too sensitive compaction of the snow in connection with melt periods indicated an error in the assumed water movement in the snowpack. Experimental studies have demonstrated that water movement in snow takes place in a heterogeneous structure (Wankiewicz 1979) which makes it likely that assumptions of constant retention capacity for the whole snowpack may be unrealistic. The retention capacity may be considerably higher in an initial phase before a stable flow pattern has been developed.

The overall simulated partitioning between surface runoff and drainage through the pipes agreed with the observations, especially for the barley simulation which

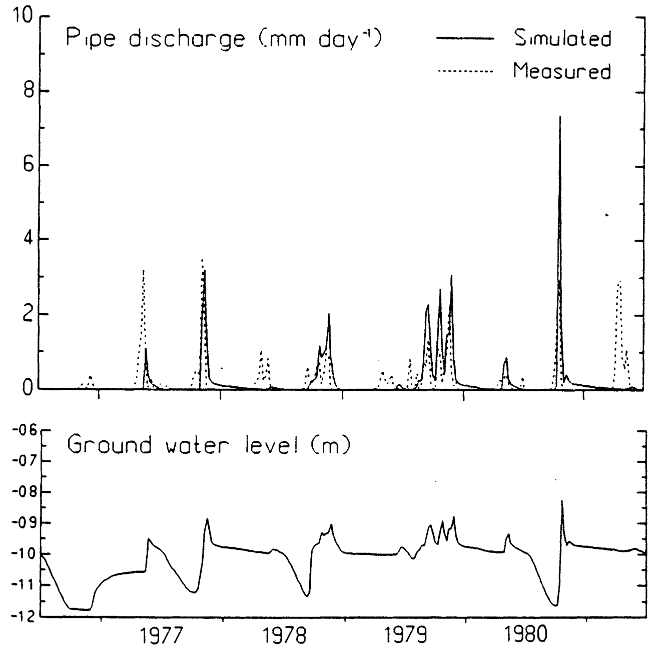

Fig. 10. Simulated and measured weekly mean values of pipe discharge and simulated ground water levels. Simulated values represent barley.

also represented the largest area of the field. However, substantial differences between simulated and observed water flows occasionally occurred. Generally, the simulation underestimated the pipe discharge in connection with snow melt and overestimated it during autumns (Fig. 10). Only small water flows were simulated through the frost layer which could be seen in the rise of the water table each spring (Fig. 10). The simulated pipe discharge was also delayed compared to the measurements, especially during spring but also in the autumn. The delay of the pipe discharge could either be an effect of how the frost influenced the water flow in the model or an effect of an overestimation of the soil water deficit at the time when water infiltrates into the soil. These effects were influenced in opposite directions when the simulations were modified from grass ley to barley. The deeper frost delayed the response between infiltration and pipe discharge but the moister soil conditions resulted in a faster response. A simulation using hydrological properties according to barley but soil thermal properties according to grass ley, demonstrated the crucial role of these effects separately (Fig. 11). Especially during the snow melt period in 1977, the shallower frost resulted in a substantial change in simulated discharge, improving the agreement with the measured pipe discharge.

We believe that the main reason behind the inability of the model to correctly depict the pattern of the observed drainage flow was the spatial variability in the

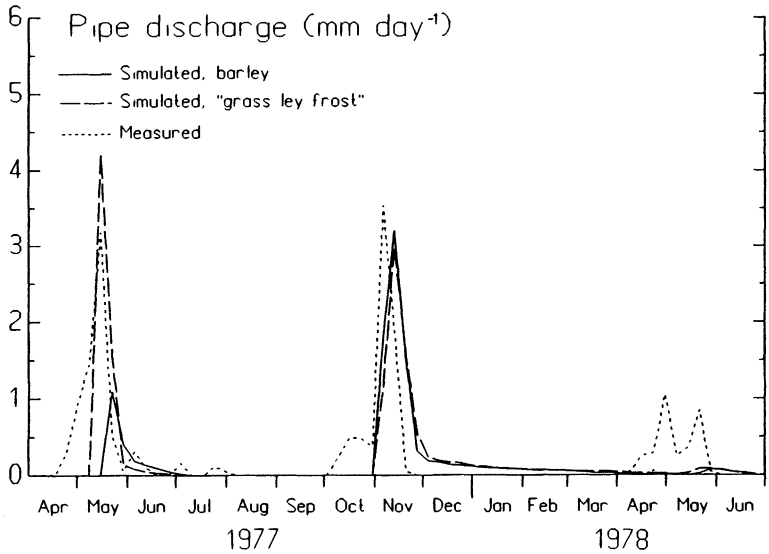

Fig. 11. Simulated and measured weekly mean values of pipe discharge during a selected period. Simulated values represent barley (the same as in Fig. 10) and one additional simulation with frost depths according to grass ley but with hydrological properties according to barley.

field. Important effects of spatial variability could occur both on a small and a large scale. On the small scale it may be reasonable that water flows may occur in macropores even if the soil is partially unsaturated, and on the large scale it is similarily reasonable to assume that low located wet areas will respond to changes in infiltration well before dry areas.

Concluding Remarks

The model application demonstrated the importance of correct quantification of soil frost if drainage from an agricultural soil is to be estimated. Frost penetration in the soil can be accurately simulated providing that detailed information on meteorological conditions and soil physical properties are available.

The partitioning between surface runoff and pipe discharge was, on the whole, correctly simulated but discrepancies between simulation and measurements concerning the temporal distribution occurred. The main reason behind these discrepancies was believed to be the spatial structure of the field, not accounted for in the one-dimensional model. The spatial heterogeneity was probably enhanced by the occurrence of snow and frost, especially during melting periods.

Further research on small-scale plots, where detailed measurements could be made, is of vital importance to improve the understanding of frost phenomena and to improve the applicability of the model for winter conditions. If the requirements are restricted to seasonal estimates of the water balance components from agricultural fields, the present stage of knowledge seems satisfactory. However, a number of problems remain if the simulated results should be used for calculation of solute transports, *e.g.,* on how the solutes are influenced by freezing and how to consider the substantial heterogeneity of water flow-paths.

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