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Spatial Variability of Physical Parameters and Processes in Two Field Soils

Part I: Water Flow and Solute Transport at Local Scale

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Natural field systems exhibit a large degree of soil heterogeneity which affects the movement of water and solutes and thus leads to highly varying observations of water contcnt and solute concentration. To investigate this problem comprehensive field investigation programs were carried out at two field sites in Denmark representing two different soil types, a coarse sand and a sandy loam, respectively.

The field investigations included collection of soil samples for analysis of textural composition, retention, and hydraulic conductivity, measurements of water content and suction, and measurements of radioactive tracer concentration, all carried out at a number of positions within the two field sites.

Models for one-dimensional vertical unsaturated flow and solute transport were applied to the two field sites, and the simulation results were compared to field measurements of water content, suction and solute concentration. This paper describes results from model simulations in individual soil profiles, while the variability issues at field scale are described in the two accompanying papers. The modelling approach was based on numerical solutions to Richards' equation for water flow and the convection-dispersion equation (CDE) for solute transport. The model results from the coarse sand field site compared relatively well to measurements of water content, suction, and concentration except for the upper soil layer (~ 10 cm depth) where the measured water contents appeared to be somewhat uncertain.

Due to the neglecting of hysteresis and macropore flow (by-pass) in the model the measured retention curves (drainage based) and the hydraulic conductivity functions at the sandy loam field site had to he modified empirically through the calibration procedure in order to match the measurements.

Introduction

Flow and transport processes in the unsaturated zone involve a complex interaction of climatic, plant, soil and chemical parameters. One of the major complications related to the prediction of flow and transport in field systems is the heterogeneity of the natural soil formations. Several field studies, *e.g.* Nielsen *et al.* (1973), Byers and Stephens (1983), Russo and Bresler (1981), and numerical studies, *e.g.* Mantoglou and Gelhar (1987 a,b,c), Curtis *et al.* (1987), have demonstrated the significance of the large variability of local-scale hydraulic properties (retention and hydraulic conductivity functions) and highlighted the problems associated with predicting the flow and transport conditions in large-scale natural soil formations.

The traditional approach to modelling of flow and transport is based on partial differential equations, which are solved by numerical techniques subject to the appropriate boundary and initial conditions, and soil hydraulic properties. These equations and their solutions have been subject to extensive validations under controlled laboratory conditions whereas direct field validations have been carried out to a less degree.

When applying these models, which originally have been developed for soil columns, to field scale problems a usual assumption is to interpret the field as an equivalent soil column characterized by a set of equivalent hydraulic functions which are established by averaging over a number of sampling locations. By solving the partial differential equations using equivalent hydraulic properties it is hypothesized that the model predictions represent the expectation values of the dependent variables within the field site. In this approach it is assumed that the flow is strictly vertical.

In a series of papers by Bresler and Dagan (1981), Dagan and Bresler (1983), Bresler and Dagan (1983 a), and Bresler and Dagan (1983 b) the concept of equivalent soil was investigated mainly by numerical analysis. They concluded that in general neither the average moisture profile nor the average concentration profile in a heterogeneous field could be predicted by the classical differential equations using effective soil properties.

Mantoglou and Gelhar (1987 a, b, c) carried out a theoretical analysis of large-scale unsaturated flow behaviour. By using a stochastic approach based on the assumption that the local soil properties are realizations of three-dimensional random fields, the partial differential equation representing the large-scale flow conditions can be derived. The resulting equation is of the same form as the local governing equation; however, the effective parameters entering the equation depend in a complicated manner on the statistical values of the soil properties, and the parameters exhibit both anisotropy and hysteresis.

Contrary to the results mentioned above an experimental study by Schulin *et. al* (1987) has provided field data where the field-averaged concentration distribution is in agreement with the classical locally-derived transport equation.

In general, more field evidence on field-scale behaviour of unsaturated flow and solute transport is needed in order to clarify the predictive capabilities of the various model approaches. These requirements have been the motivation for the present study which have included a comprehensive data collection program at two field sites combined with numerical model analysis of water and solute movement.

Experimental Field Study

Two field sites representing two diverse, yet typical, soil types in Denmark were selected for the experimental investigations. Field site Taastrup is located on Zealand 20 km west of Copenhagen in a glacial moraine landscape, and field site Jyndevad is located in Jutland 5 km north of the Danish-German border in an alluvial outwash plain landscape. The soil at the two sites can be classified as sandy loam and coarse sand, respectively.

Both experimental areas were uncontrolled in the sense that they were exposed to the prevailing climatic conditions, and both were covered by short-cut grass. The experiments were carried out during the three-year period 1985-87. The average precipitation at the two sites during the investigation period was 640 mm/year at the Taastrup field site and 860 mm/year at the Jyndevad field site. Potential evapotranspiration averaged 555 mm/year and 520 mm/year, respectively, for the two sites during the study period.

Both field sites were designed to the same size (100 x 50 m), and a total of 53 sampling and monitoring points were established in the same systematic horizontal arrangement, Fig. 1. The sampling distances varied from 1 m up to 112 m, while the sampling levels over depth ranged from three m and up to seven m.

Textural composition was determined at all sampling locations (53 in total), whereas retention characteristics, saturated hydraulic conductivity, porosity, and dry bulk density were determined in a subset of 24 sampling points. All measurements were made on undisturbed soil cores, 5 cm in diameter and 5 cm long, with three soil replicates for each location and level.

At the same 24 locations access tubes were installed to monitor the water content using a neutron probe. At 12 of the tubes a tensiometer system was installed to measure the suction. The depth to the water table was measured across the field in 10 piezometers at Jyndevad and 14 piezometers at Taastrup. The results of these field investigations and the geostatistical analysis of the data have been reported by Hansen *et al.* (1986) and Hansen and Jensen (1988).

At Jyndevad the 12 access tubes, arranged in four triangular setups as indicated in Fig. 1, were used for tracer experiments, while only 6 access tubes distributed over the experimental area were used at Taastrup. A solution of $^{60}CO(CN_6)^{3-}$, a gamma-emitting radioactive tracer, was applied to the soil surface around each of the tubes. As the tracer moved down through the soil, the transport and dispersion characteristics were determined by logging the access tube using a scintillation

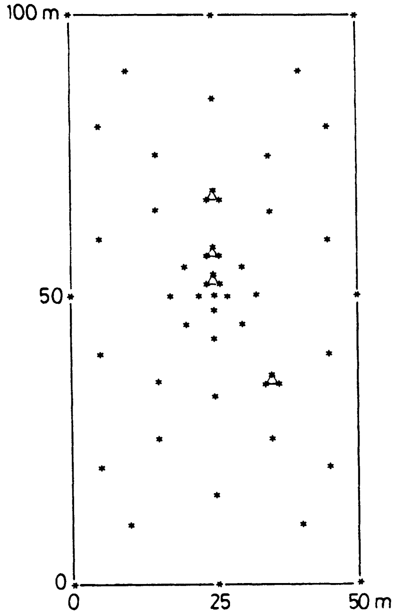


Fig. 1. Layout of sampling points (*).

counter. This technique enabled a continuous. *in-situ* logging without disturbing the sites. However, since the access tubes used for soil water determinations were too narrow to provide for a shielded radiation probe, the signal represented the radiation level over a spherical volume 30 cm or more in radius. Hence, a deconvolution procedure was required in order to obtain the actual tracer distribution in the soil profile. The experimental tracer studies have been reported by Sevel *et al.* (1988).

At both sites rainfall was recorded on a daily basis, while climatic data were obtained from a nearby meteorological station. Crop characteristics in the form of leaf area index and root density were measured on a scattered basis during the investigation period, and these measurements were used to establish functions of the seasonal variations of leaf area index and root distribution which were applied in the evapotranspiration description.

Models of Unsaturated Flow and Transport

The field data have been analyzed within a one-dimensional framework by assuming that the entire field is composed of a number of soil columns where no exchange of water or solute between the individual columns takes place. This approximation follows the modelling efforts by Dagan and Bresler (1983), Bresler and Dagan (1981, 1983a, 1983b), Schulin *et al.* (1987), Butters and Jury (1989), and Destouni and Cvetkovic (1991).

Consequently this assumption allows for adopting the general differential equation for vertical unsaturated flow (Richards' equation) in order to describe the water movement on the local scale (the individual soil columns)

$$C \frac{\partial \psi}{\partial t} = \frac{\partial}{\partial z} \left(K \frac{\partial \psi}{\partial z} \right) - \frac{\partial K}{\partial z} - S \tag{1}$$

where

Ψ – capillary pressure
C – water capacity
K – hydraulic conductivity
S – sink term representing uptake by roots
z – vertical coordinate, positive downwards
t – time

The hydraulic functions needed for solving Eq. (1) are: a) retention function (relationship between water content θ and capillary pressure Ψ) and b) hydraulic conductivity function (relationship between hydraulic conductivity K and water content θ or capillary pressure Ψ). The hydraulic functions for the individual soil profiles are determined on the basis of the experimental data. However, the available data are incomplete for an exact definition of these relationships, which renders it necessary to establish the functions partly by calibration, as described in more details below. A major simplification is introduced by neglecting hysteresis in these functions.

In any attempt to simulate the moisture conditions in the field, the water loss by evapotranspiration is an important element. In the present model approach evapotranspiration is predicted on the basis of daily values of potential evapotranspiration. These values are divided into two fractions applicable for transpiration (root uptake) and soil evaporation, respectively. To predict the actual values of these components semi-empirical functional relationships are defined mainly by relating the reduction below the potential rates to the actual water content. The predicted evapotranspiration is entered into the flow equation through the sink term S. More details on the computational procedure can be found in Jensen (1983).

The prediction of solute transport in the individual soil profiles (local scale) is based on the one-dimensional convection-dispersion equation

$$\theta \frac{\partial C}{\partial t} = \frac{\partial}{\partial z} \left(\theta D \frac{\partial C}{\partial z} \right) - q \frac{\partial C}{\partial z} - S \tag{2}$$

where

C – concentration
θ – moisture content
q – water flux
D – dispersion coefficient
S – sink term

It is assumed that the dispersion coefficient can be predicted by the following traditional expression (Hillel 1980)

$$D = \alpha V \tag{3}$$

where

α – longitudinal dispersivity
V – seepage velocity obtained as q/θ

More details on the transport model is given in Jensen *et al.* (1984).

The differential equations for flow and transport are both solved numerically using finite difference techniques.

Model Application to the Jyndevad Site

The modelling procedure described above has been applied to the individual sampling profiles within the field site. The model results for a representative profile (D2) are demonstrated and discussed below. Only a part of the total investigation period is shown.

Water Flow

The input and boundary data required for predicting water flow in the soil profiles include rainfall, potential evapotranspiration, leaf area index, root zone depth, and level of water table. For the seasonal variation of the leaf area index and the root zone depth generalized functions have been defined on the basis of the available measurements, while the other data have been obtained from direct measurements.

The model parameters comprise the retention and hydraulic conductivity functions. In all profiles undisturbed soil cores were taken in triplicates at up to seven levels in order to measure these functions in the laboratory. The retention characteristics were determined by displacing water from the cores through a series of suction levels and measuring the equilibrium water content at each level. The sequence of suctions were imposed monotonically in order to produce the main drainage branch of the capillary pressure – water content relation. The saturated hydraulic conductivity was measured on all cores in a permeameter, while only scattered measurements of unsaturated hydraulic conductivity were made using the outflow method developed by Richards (1967).

As an approximation it has been assumed that the main drainage branch determined in the laboratory as described above represents the retention characteristics at field conditions. In order to facilitate the application of the data in the numerical model cubic spline functions have been fitted to the set of corresponding experimental values of water content and capillary pressure.

The hydraulic conductivity function is experimentally defined only in one point (saturation), and in order to establish the function over the whole moisture regime a power function of the following type is adopted, Mualem (1978)

$$K(\theta) = K_s \left(\frac{\theta - \theta_r}{\theta_s - \theta_r} \right)^n \tag{4}$$

where

$K(\theta)$ – hydraulic conductivity at water content θ
K_s – hydraulic conductivity at saturation
θ_r – residual water content
θ_s – water content at saturation
n – exponent

The parameter θ_r is selected as the water content at a suction of 150 m (wilting point) while the parameter n is estimated in a rather heuristic manner by utilizing the qualitative concept *field capacity* defined as the water content at which the water flow for all practical purposes has ceased. For a sandy soil type as the one at the Jyndevad site field capacity is usually defined at a suction corresponding to pF = 2 (pF defined as the logarithmic value of suction in cm), and at this suction we assume that a predetermined low value for the hydraulic conductivity applies (0.5 mm/day). On the basis of these assumptions the complete hydraulic conductivity function can be defined (Eq. (4)). The function will vary from profile to profile according to the data on the local soil properties.

Fig. 2 shows the hydraulic functions which are established for the selected soil profile D2 at the individuel depths using the procedure described above.

Fig. 3 compares the retention curve determined in the laboratory with the corresponding values of field-measured water content and suction. Note that due to the fact that the levels do not coincide where the soil cores are sampled and where the

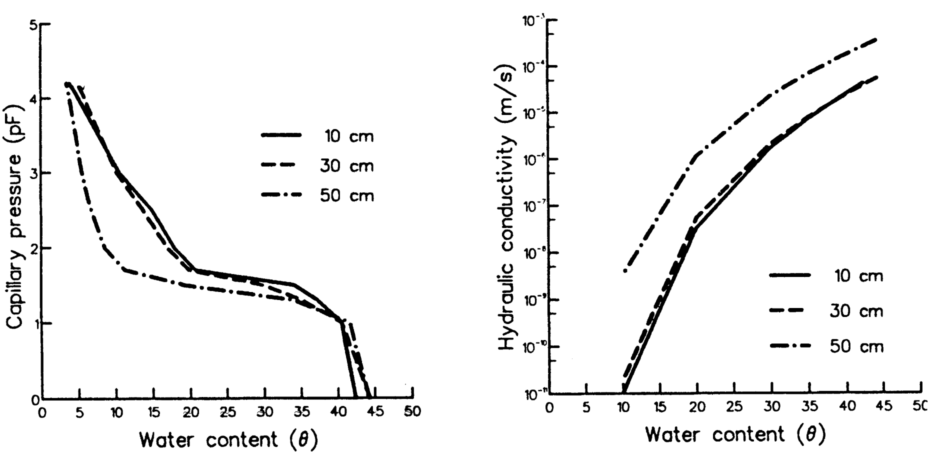
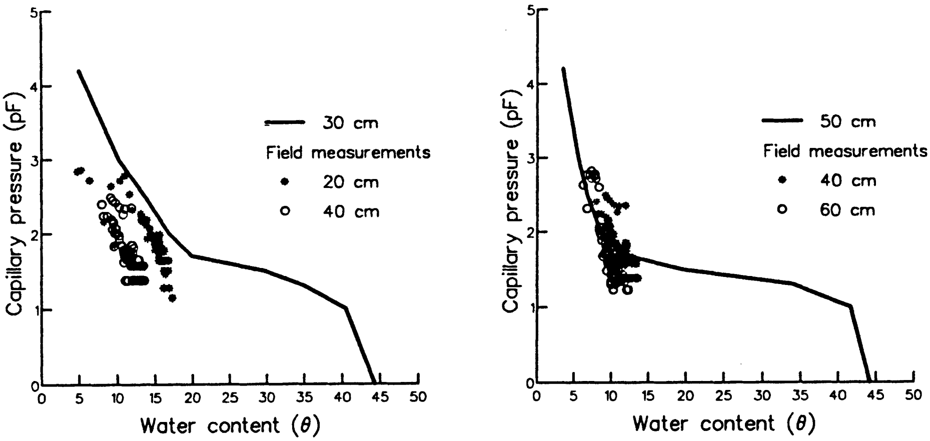


Fig. 2. Retention and hydraulic conductivity functions for soil profile D2, Jyndevad site.



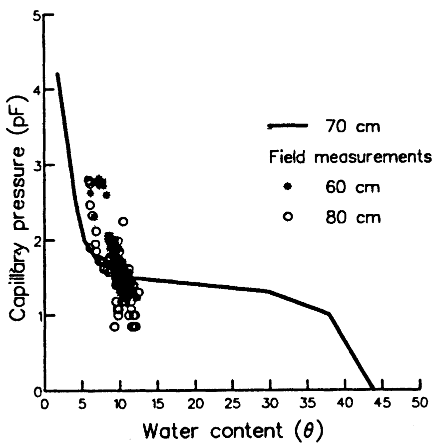


Fig. 3.
Laboratory-measured retention curve and field measurements of water content and suction at field point D2, Jyndevad site.

soil variables are measured an exact comparison is not possible. It appears that the drainage curve of the retention characteristics as determined in the laboratory tends to predict somewhat higher water content at a given suction near the soil surface in comparison with the field observation. Note, that part of this problem may be caused by inaccuracies of the neutron probe instrument as discussed below. At larger depths the drainage curve seems to provide a good approximation of the field conditions although it is evident from the figure that hysteresis influences the retention characteristics. Thus, for this field site we have decided to apply the laboratory-measured retention curves directly without any adjustments to compensate for this problem.

On the basis of the above modelling approach and procedures for defining input data and parameters, simulations of water content and suction for profile D2 have been carried out and compared to measurements as shown in Figs. 4 and 5, respectively.

In general, good comparisons are obtained. At 10 cm level major discrepancies

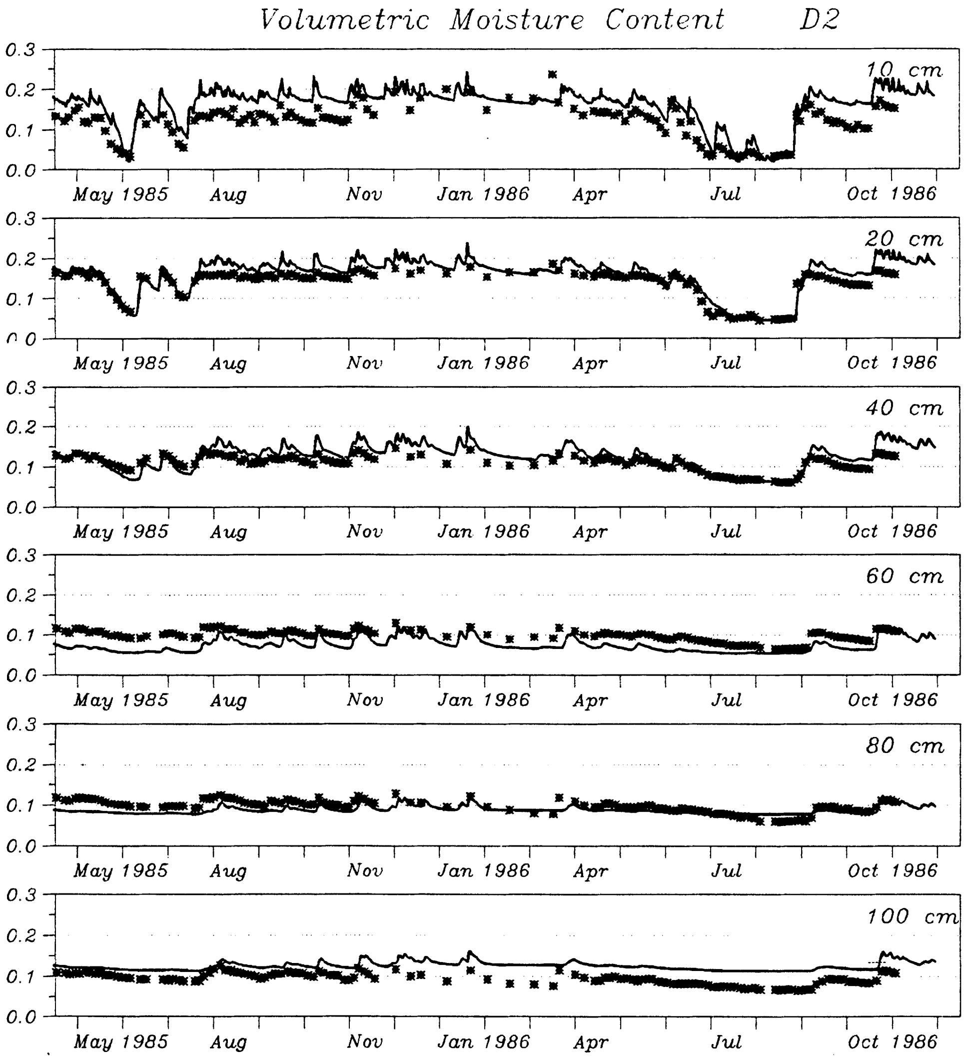


Fig. 4. Measured (*) and simulated (–) water content at field point D2, Jyndevad site.

are present between simulated and measured water content, particularly outside the growing seasons, but there is an excellent agreement between simulated and measured suctions at this level. This suggests, as above, that the experimental data are not consistent: either the water contents are erroneous or the measured retention curve is not representative for the field conditions. Application of the neutron method for soil water measurements implies that the neutron counts must be converted by calibration curves. Due to the interference with the free air it is always difficult to establish an accurate conversion procedure close to the soil

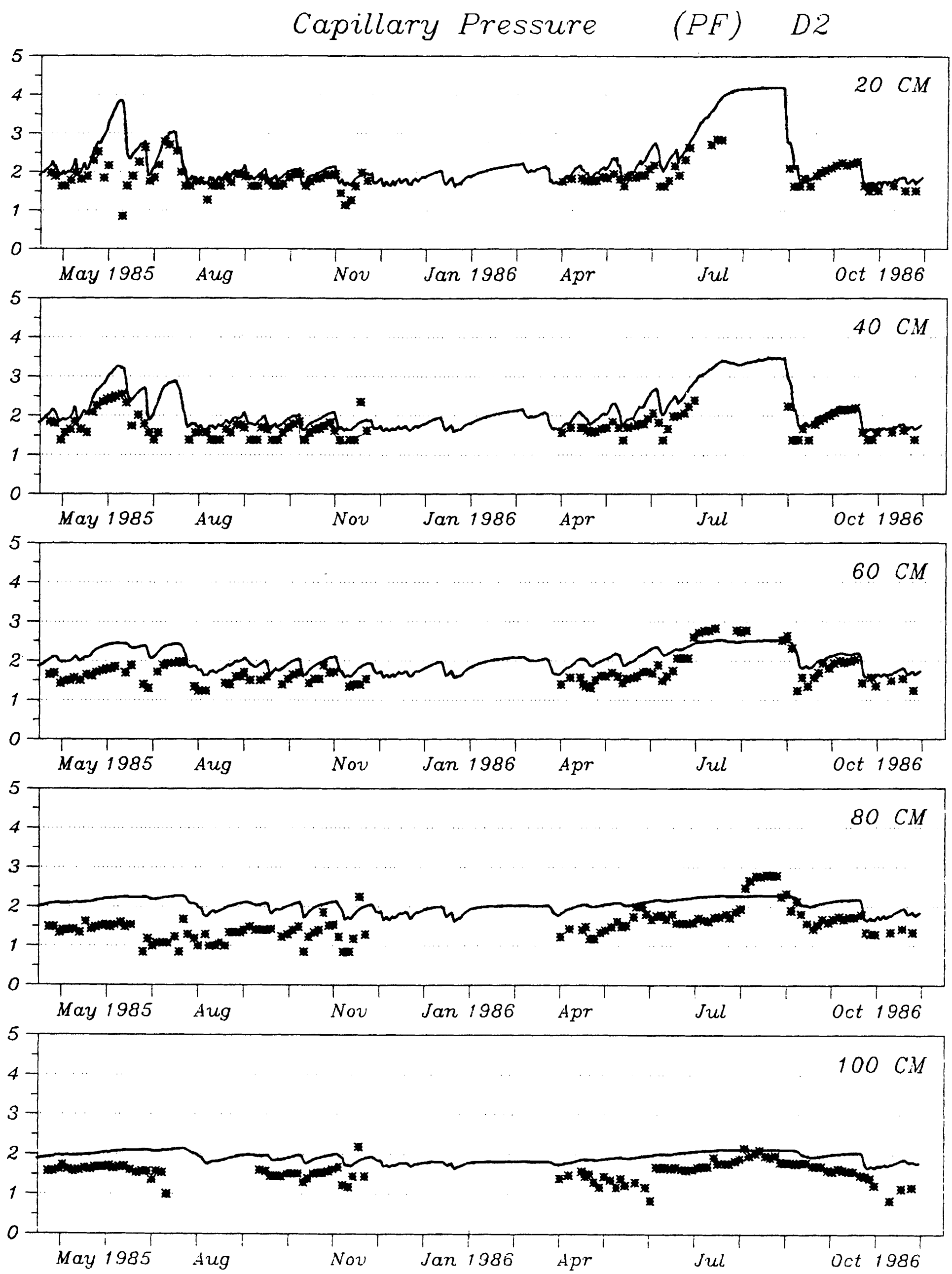


Fig. 5. Measured (*) and simulated (–) suctions (logarithmic values) at field point D2, Jyndevad site.

surface, and in addition site-specific problems have been encountered in establishing the calibration curves. Hence, it is likely that the derived figures for field water content are inaccurate, particularly those at shallow depths, and that the drainage retention curve after all may represent field conditions better than the impression given by the figures. Inclusion of hysteresis would certainly tend to decrease the predicted water content, because the main drying curve represents the highest possible value for a given suction level. However, it is not likely that a sufficient decrease can be obtained by including hysteresis, because this effect tends to diminish at suctions around 70-100 cm, and this is the range where the simulation problems are most pronounced.

Another possible source of error is the unsaturated hydraulic conductivity function which has not been measured. If a higher unsaturated hydraulic conductivity at pF = 2 was assumed for all soil types in the profile, the simulated water content during the period August to May would become lower at all depth levels and hence too low at 20-100 cm depth. Furthermore, the simulated suction would become higher at all depths resulting in an overall worse comparison with the water content and suction data. An isolated increase of unsaturated conductivities at 10 cm depth would have virtually no effect because the suction and hence the water content at 10 cm depth is to a large degree determined by the suctions at 20-100 cm depth. Consequently, it is concluded that uncertainties in the unsaturated hydraulic conductivity function cannot alone explain the deviations between measured and simulated water content at 10 cm depth.

Solute Transport

A radioactive tracer was applied to the soil surface around each of the access tubes in the triangular set-ups shown in Fig. 1. 100 μCi of ^{60}Co-complex was injected over an area of 0.785 m^2 on December 7, 1984, followed by freshwater application to create a pulse of tracer solution at the soil surface. The movement of the tracer in the individual profiles (12 in total) was measured onwards using a radiation probe.

The processed measurements for triangle D2 are shown in Fig. 6 as profiles at selected times and in Fig. 7 as time series at selected depths together with the model simulations. Three sets of measurements are shown representing the measurements from each of the tubes, whereas the simulations are based on the average soil properties within the narrow area confined by the triangle.

In the period after injection the tracer distributions in all three access tubes were similar in shape, and the deviations between the tubes were small in a relative sense. At later stages these short-range differences in tracer concentrations tended to grow as the solute moved downwards, and during the last part of the measurement period the relative differences were significant.

Part of the observed differences in transport characteristics between the tubes can probably be explained by uncertainties involved in the gamma logging technique, particularly the deconvolution procedure. These issues are discussed in detail

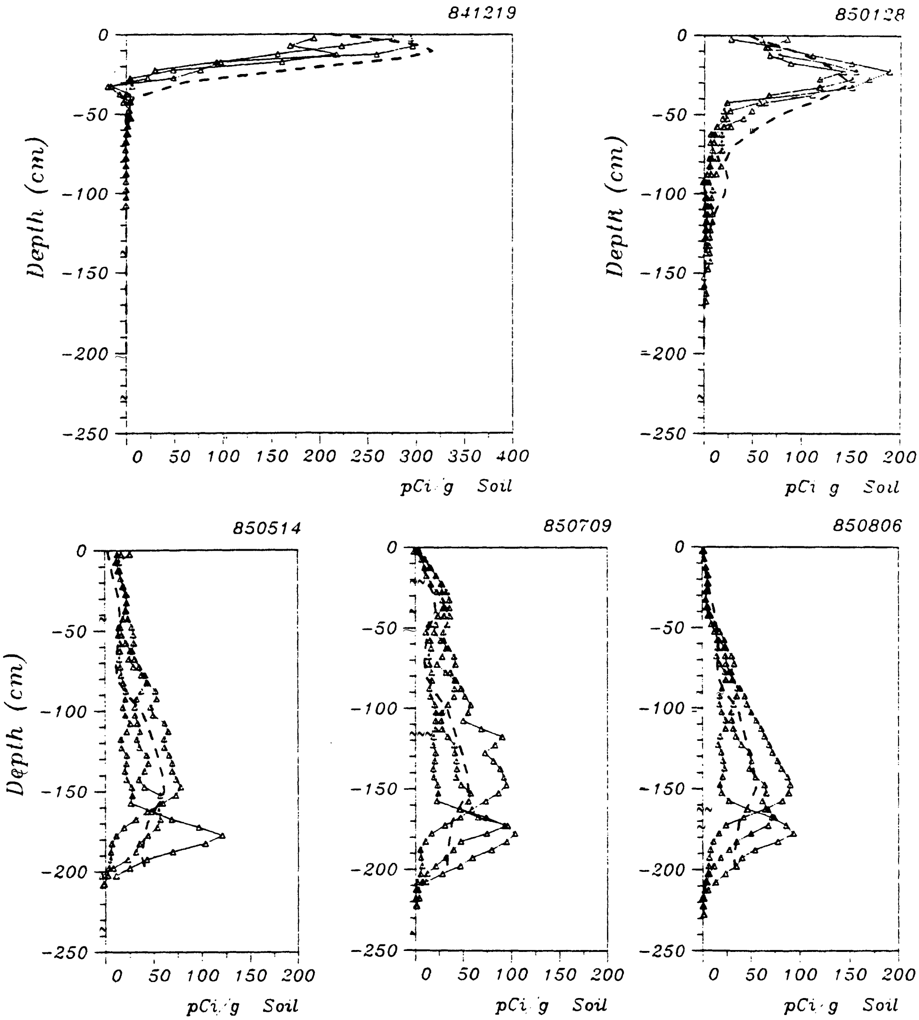


Fig. 6. Measured (Δ) and simulated (---) concentration profiles at field point D2, Jyndevad site. Measurements from three access tubes.

by Sevel *et al.* (1988). However, the observations demonstrate that the water flow exhihit a significant spatial variability even at this small scale.

The CDE-based transport model has been applied to the site using a longitudinal dispersivity of 8 cm and assuming that the tracer behaves as a non-reactive constituent.

As shown by the figures the model-predicted solute concentrations provide a reasonable approximation to the average conditions within the investigated triangle, thus supporting the applicability of a one-dimensional Darcy and CDE-based model approach at this scale.

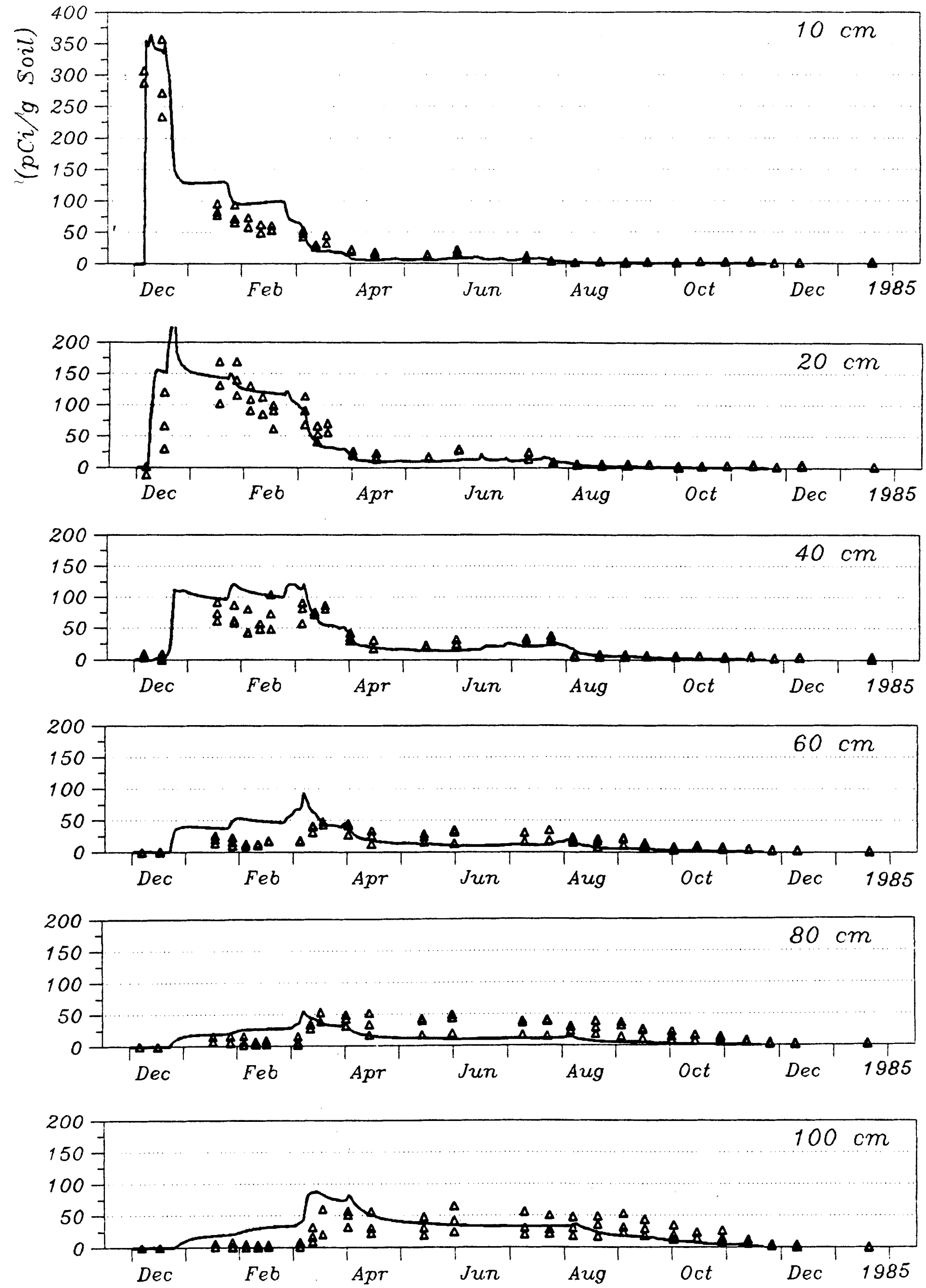


Fig. 7. Measured (Δ) and simulated (–) time series of concentration at field point D2, Jyndevad site. Measurements from three access tubes.

Model Application to the Taastrup Site

Basically, the same modelling procedure has been applied to the two field sites. However, due to differences in experimental conditions and soil characteristics, site-specific adjustments have heen necessary. The modifications required for the Taastrup site will be discussed with reference to the specific profile, C4, which is the only profile where supplementary measurements of unsaturated hydraulic conductivity are available.

Water Flow

From preliminary model simulations it appeared that the approach used for the Jyndevad site for establishing the retention and hydraulic conductivity functions was not directly applicable for the Taastrup site.

Also for this site the soil water retention characteristics were measured in the laboratory in the form of desaturation curves from field saturation, Hansen *et al.* (1986). Saturated hydraulic conductivities were measured as part of the present project, while Blem and Clausen (1986) made supplementary measurements of unsaturated hydraulic conductivity on selected soil samples using the outflow method developed by Richards (1967).

By comparing the field data on water content and suction and the measured soil water retention curves it is evident that due to hysteresis the measured drainage part of the retention curves cannot be applied directly in the model simulations. As an example, the measured drainage curves and some field data are shown in Fig. 8 for four depths at the field point C4.

The figure shows that the field data for 1985 and 1986 are generally significantly below the main drainage curve indicating that either the main wetting curve or some scanning curve here would be more representative. The data from 1987 are on the other hand in better agreement with the main drainage curve.

The effective porosity (natural saturation) can be assessed from the water content data in periods when the water tables are above the lowest measurement level in the individual profiles. The effective porosities in 1985-1986 are found to be about 8 vol. % lower than the total porosities measured in the laboratory (drainage curves). In 1987 the effective porosities are found to be somewhat higher.

To some extent these observations may he explained by the hysteresis phenomenon. The moisture content-capillary pressure relationship as derived from the field data was most of the time below the main drainage curve, and in 1985-1986 the residual air content at saturation was about 8 vol. %. In order to move to the main drainage curve, a total redistribution of air and water was necessary. Such a situation occurred apparently during the very cold winter January – March, 1987 where the soil was frozen down to more than 50 cm depth. The jump in the retention characteristics resulting in higher water content can be observed in the measured water content as shown in Fig. 14 p. 296. In January 1987 the increase in water

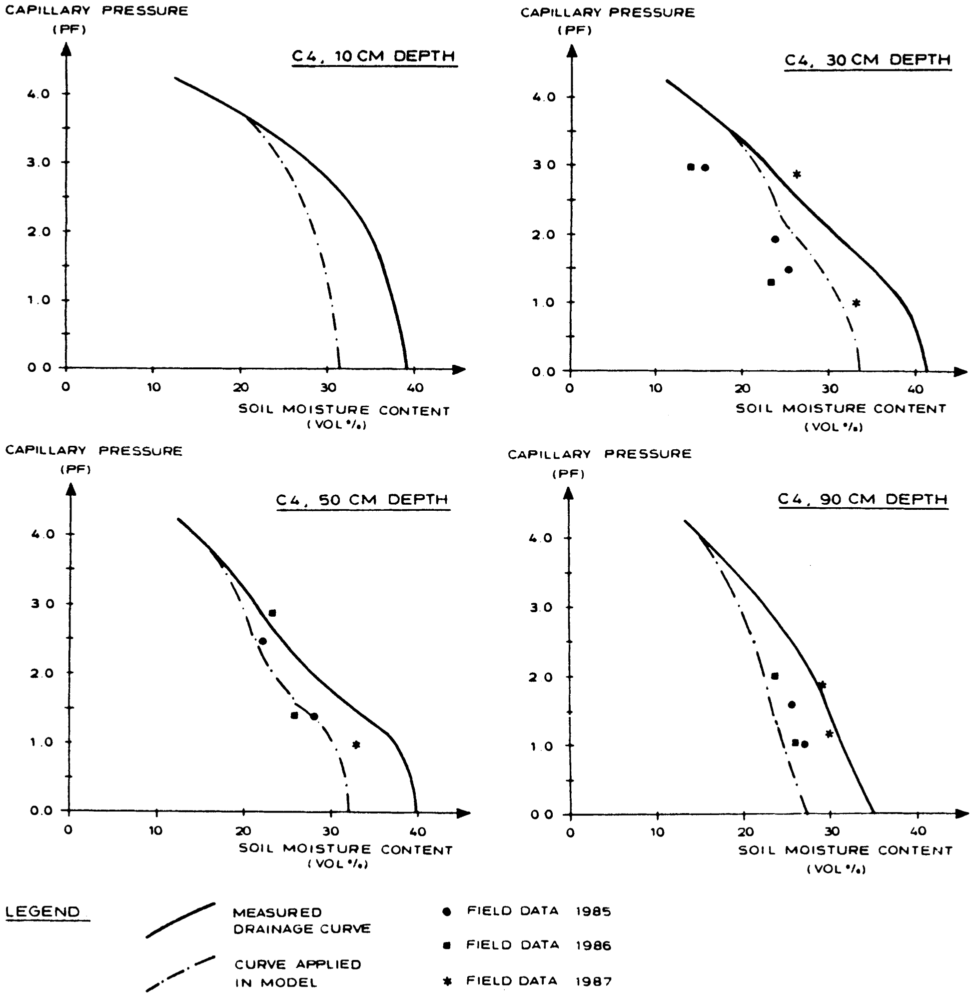


Fig. 8. Laboratory-measured retention curves, and field measurements of water content and suction at field point C4, Taastrup site.

content started at 10 cm depth and moved downwards with the penetration of the freezing front.

The flow model is based on the assumption that single-valued functions for the hydraulic properties apply and that hysteresis cannot be accounted for explicitly. Since the laboratory-measured drainage curves apparently are not appropriate overall representations of the field conditions, a general correction procedure illustrated in Fig. 8 is applied to all field points throughout the profile.

The hydraulic conductivity function is assumed to follow a power function, Eq. (4). K_s is taken as the median of the three measurements at each site, while θ_s and

Table 1 – Determination of θ_f and θ_r, in the hydraulic conductivity function from the retention characteristics

Depth	θ_f	θ_r
10 cm	θ (pF $\equiv$ 2.5)	½ (θ (pF $\equiv$ 3.0) + θ (pF $\equiv$ 4.2))
30 cm	θ (pF $\equiv$ 2.0)	θ (pF $\equiv$ 3.0)
50 cm	θ (pF $\equiv$ 2.0)	θ (pF $\equiv$ 2.5)
70 cm	θ (pF $\equiv$ 1.5)	θ (pF $\equiv$ 2.5)
90 cm	θ (pF $\equiv$ 1.5)	θ (pF = 2.0)

θ_r are determined from the modified soil moisture retention curve following the procedure indicated in Table 1. Furthermore, it is assumed that $K = 3 \times 10^{-9}$ m/s at $\theta = \theta_f$, where θ_f is also determined from the retention curve, *cf.* Table 1. On the basis of this empirically based procedure the exponent n in Eq. (4) can be estimated.

The hydraulic conductivity curves derived hereby for field point C4 are shown in Fig. 9 together with values measured by Blem and Clausen (1986).

It should be emphasized that the hydraulic conductivity function and in particular the values given in Table 1 are strictly empirical and have been obtained by calibration against field data for water content and capillary pressure rather than from theoretical considerations. Further, the K_s values which were measured on soil cores at full saturation are assumed valid at effective saturation where the residual air content is assumed to be 8 %. This assumption is certainly theoretically questionable.

If the hydraulic conductivity curves are shifted 8 vol. % to the right in Fig. 9, they would fit reasonably well with the measurements by Blem and Clausen (1986). However, in this case the conductivity at the effective saturation would be 3-4 orders of magnitude lower than K_s with the physically unrealistic result that ponding and overland flow would be generated even with rainfall intensities lower than 1 mm/day.

In the top layers, 10-30 cm, water movement takes place even at pF values 2.5 and 3.0. This phenomenon can easily be seen from the field data, *e.g.* in June 1985, *cf.* Fig. 10 where a rainfall event is seen to penetrate down to 50 cm depth when the soil water tension is about pF 2.5. This essentially implies that the hydraulic conductivity retains some value even in this suction regime.

In the bottom layer (90 cm depth), the resulting hydraulic conductivity function has a very steep slope in order to eliminate any water movement at pF values above 2.0. If the θ_f and θ_r values used in the top soil were applied at 90 cm, the upward capillary water movement from the shallow water table (1-1.5 m depth) would be several mm/d resulting in an almost constant water content. Such a flow behaviour is not confirmed by the field measurements.

From the comparison of measured and simulated water content and suction,

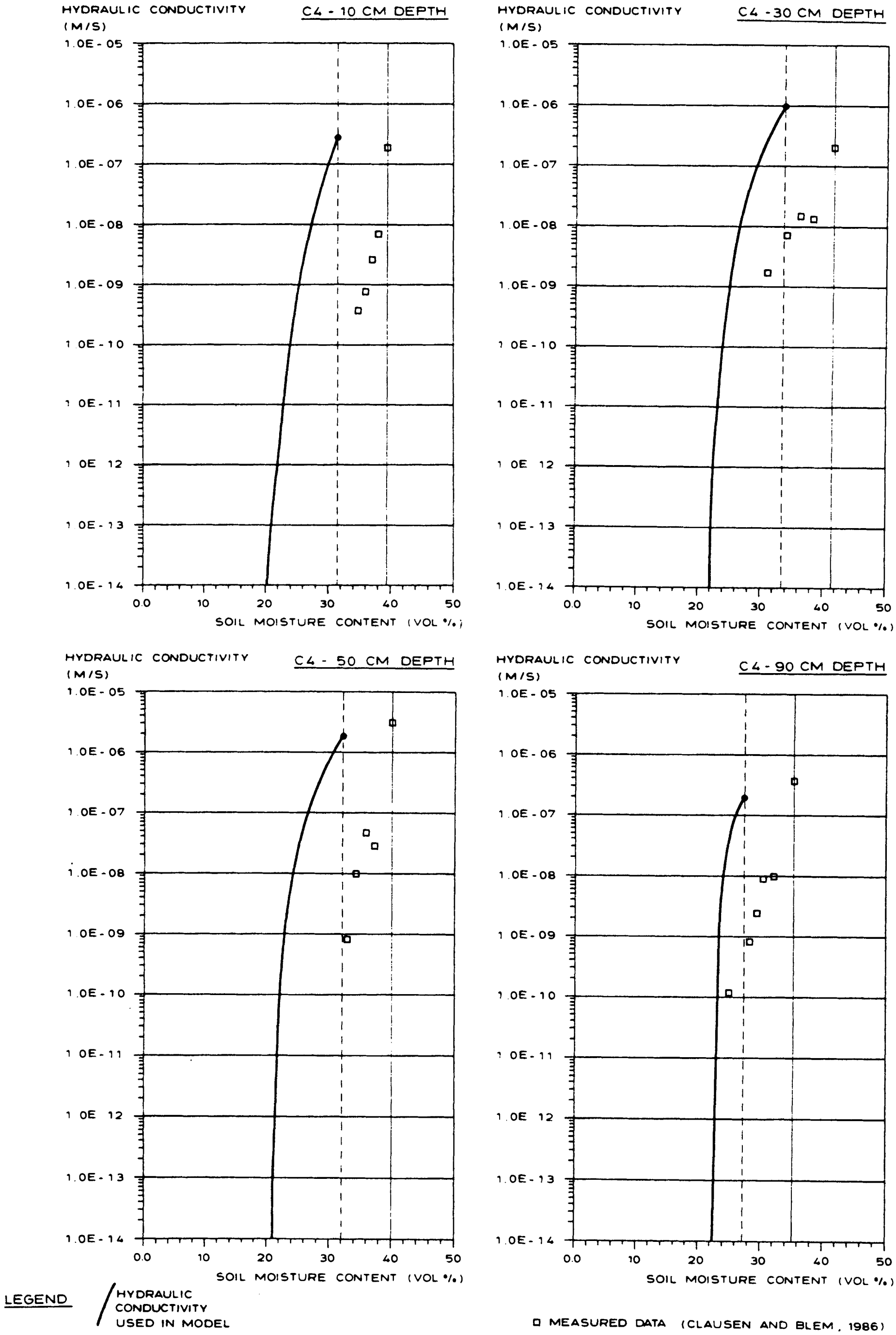


Fig. 9. Estimated hydraulic conductivity function for field point C4, Taastrup site.

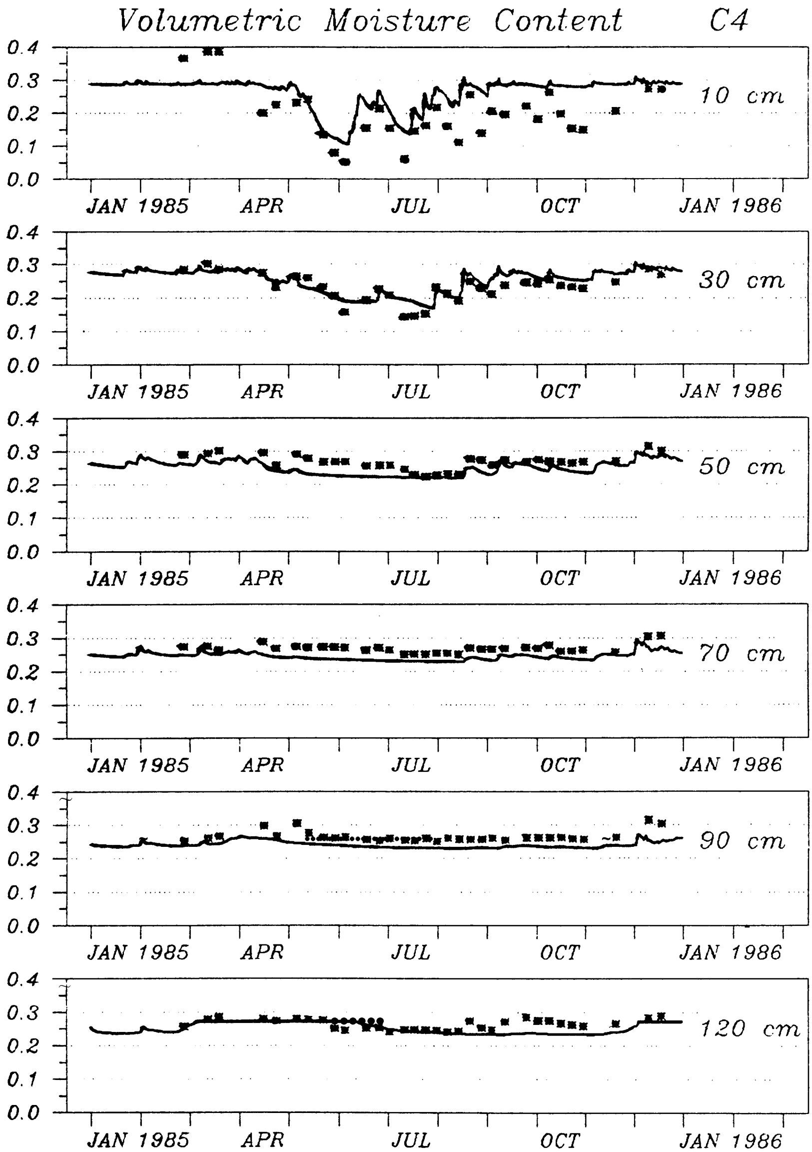


Fig. 10. Measured (*) and simulated (–) water content at field point C4, Taastrup site, 1985.

Figs. 10-15, it appears that the soil water flow dynamics is simulated reasonably well both over depth and time, and the responses to rainfall events are well described. However, as could be expected from the discussion above, the water content in 1987 is simulated too low. It should be emphasized again that the adjustments introduced to the retention curves and the procedure used to establish

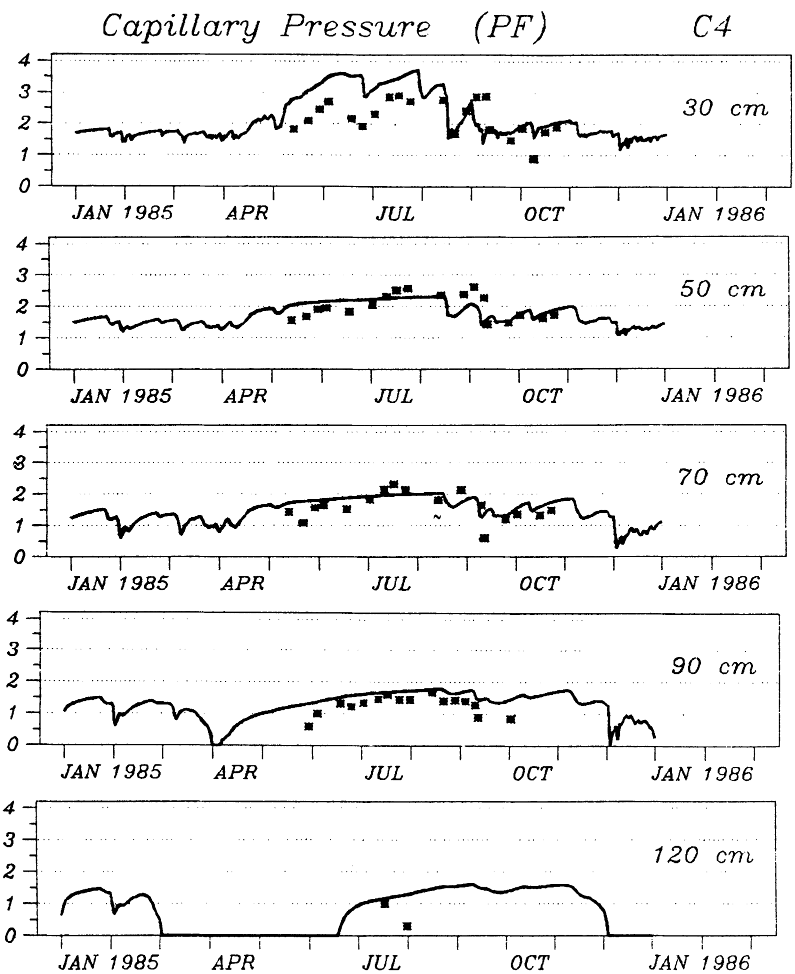


Fig. 11. Measured (*) and simulated (–) capillary pressure (logarithmic values) at field point C4, Taastrup site, 1985.

the hydraulic conductivity functions are not justified from theoretical considerations but are developed entirely by calibration. Particularly, the difference in shape between the upper horizons and 90 cm depth may at a first glance seem dubious.

This apparent inconsistency is believed to be caused by the fact that the model does not account for the following two natural processes: 1) hysteresis in the retention and the hydraulic conductivity functions and 2) bypass flow. The hysteresis process was discussed with respect to the water retention curve above; however. also the hydraulic conductivity function appears to be affected by hysteresis.

Bypass is the process by which infiltrating water enters the larger pores and

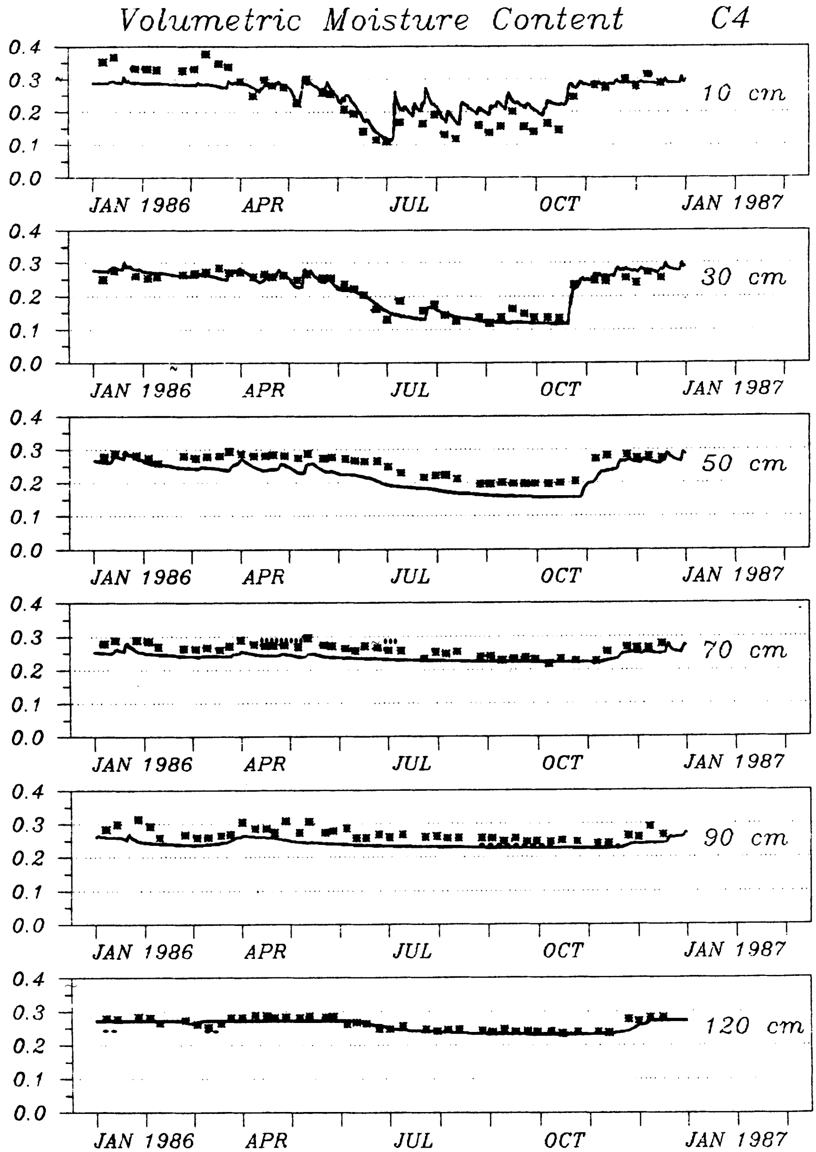


Fig. 12. Measured (*) and simulated (–) water content at field point C4,Taastrup site, 1986.

bypasses the major part of the soil even under conditions when a considerable soil moisture deficit may exist in the main soil matrix. Thus the soil may in this perspective be interpreted as a double-porosity matrix. Bypass has been reported in several studies, usually in soils with high clay contents, see *e.g.* Villholth *et al.* (1991). As a result of bypass, the effective hydraulic conductivity should be dependent on

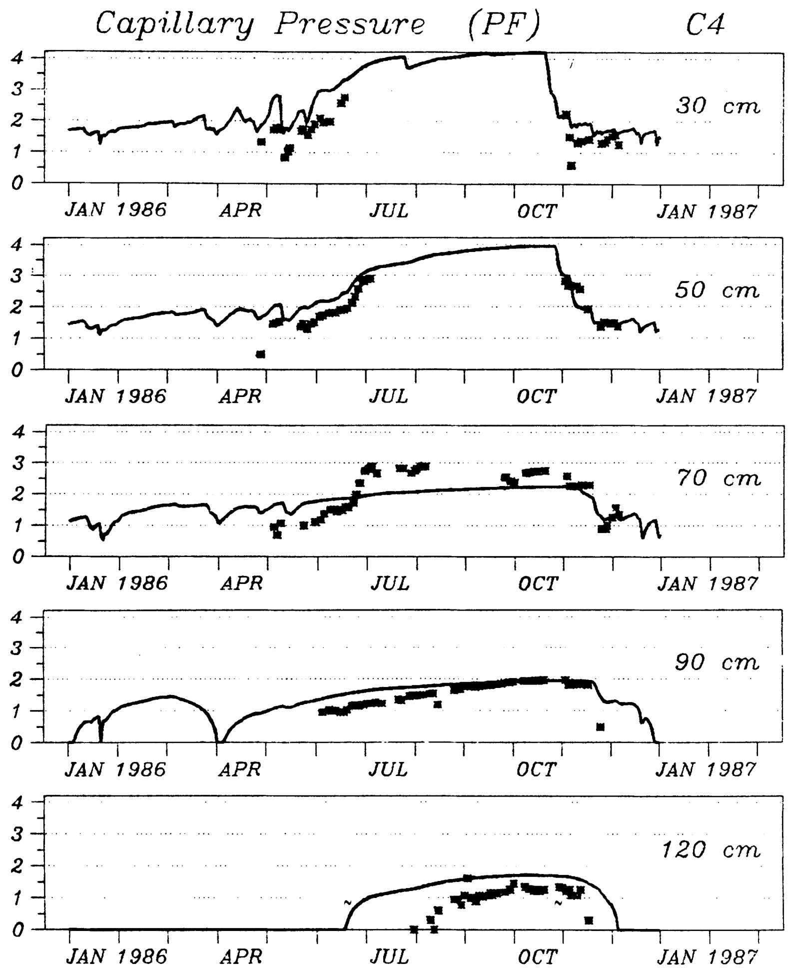


Fig. 13. Measured (*) and simulated (–) capillary pressure (logarithmic values) at field point C4, Taastrup site, 1986.

whether the wetting occurs from above (infiltration) or below (capillary rise).

The model does not in the present version account for these two important physical processes; but we have compensated for the incomplete process description in the calibration of the hydraulic conductivity function so that the dynamics of the soil water flow generally is simulated satisfactorily.

From the comparisons of simulated and measured values of water content and capillary pressure shown in Figs. 10-15 it appears that there are some general disagreements between measured and simulated water content at 10 cm depth. The measured values are fluctuating significantly more than the simulated ones, and very high values appear at some occasions during the winter seasons. Furthermore,

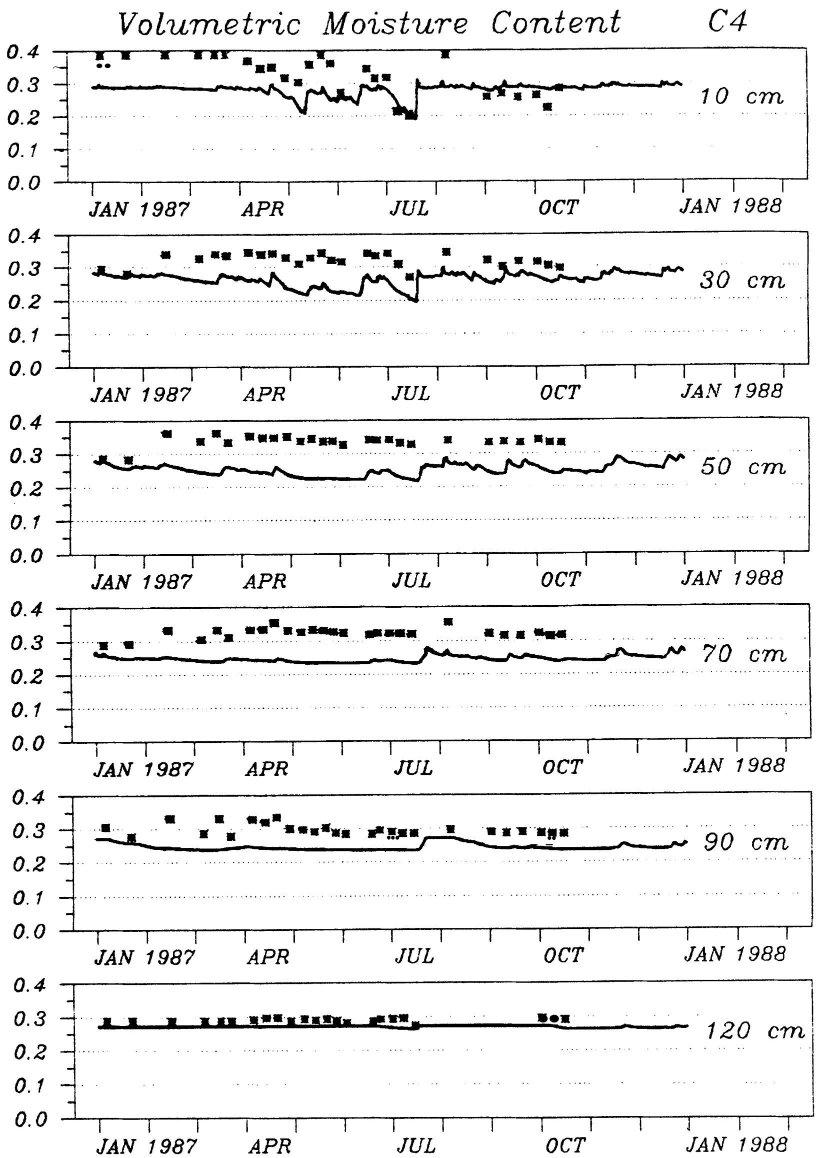


Fig. 14. Measured (*) and simulated (–) water content at field point C4, Taastrup site, 1987.

the simulated values during the autumn seasons (August-October) are generally too high. These deviations may to some extent be caused by neglecting the hysteresis and macropore flow processes.

The meteorological regime in 1987 differed significantly from those of 1985 and 1986 in two respects. Firstly, January-March 1987 was extremely cold with frozen

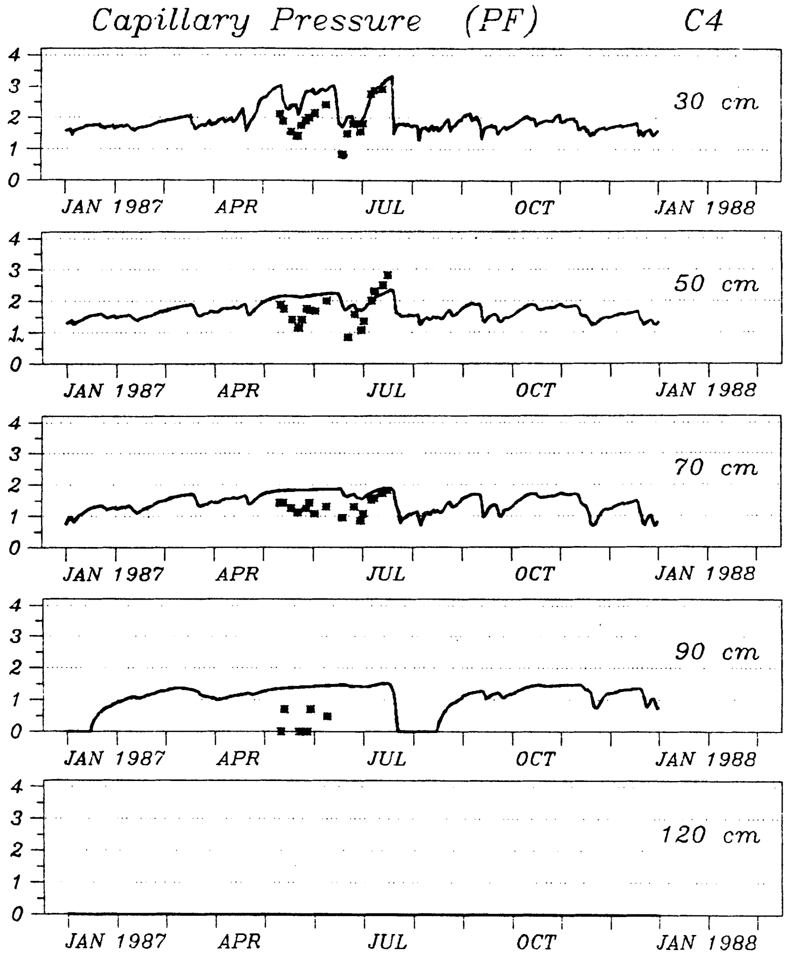


Fig. 15. Measured (*) and simulated (–) capillary pressure (logarithmic values) at field point C4, Taastrup site, 1987.

soil down to at least 50 cm. As discussed above, the freezing of the soil is believed to have caused a water-air redistribution in the soil pores and hence resulted in higher retention properties. This may be the explanation for the apparent inconsistency that the measured water content is significantly higher than simulated, while the agreement in capillary pressure is reasonably good. Secondly, the summer season was extremely wet with almost no build-up of soil moisture deficits. The nearly constant water content throughout the year is reflected both in the measured and the simulated values, although the general levels deviate.

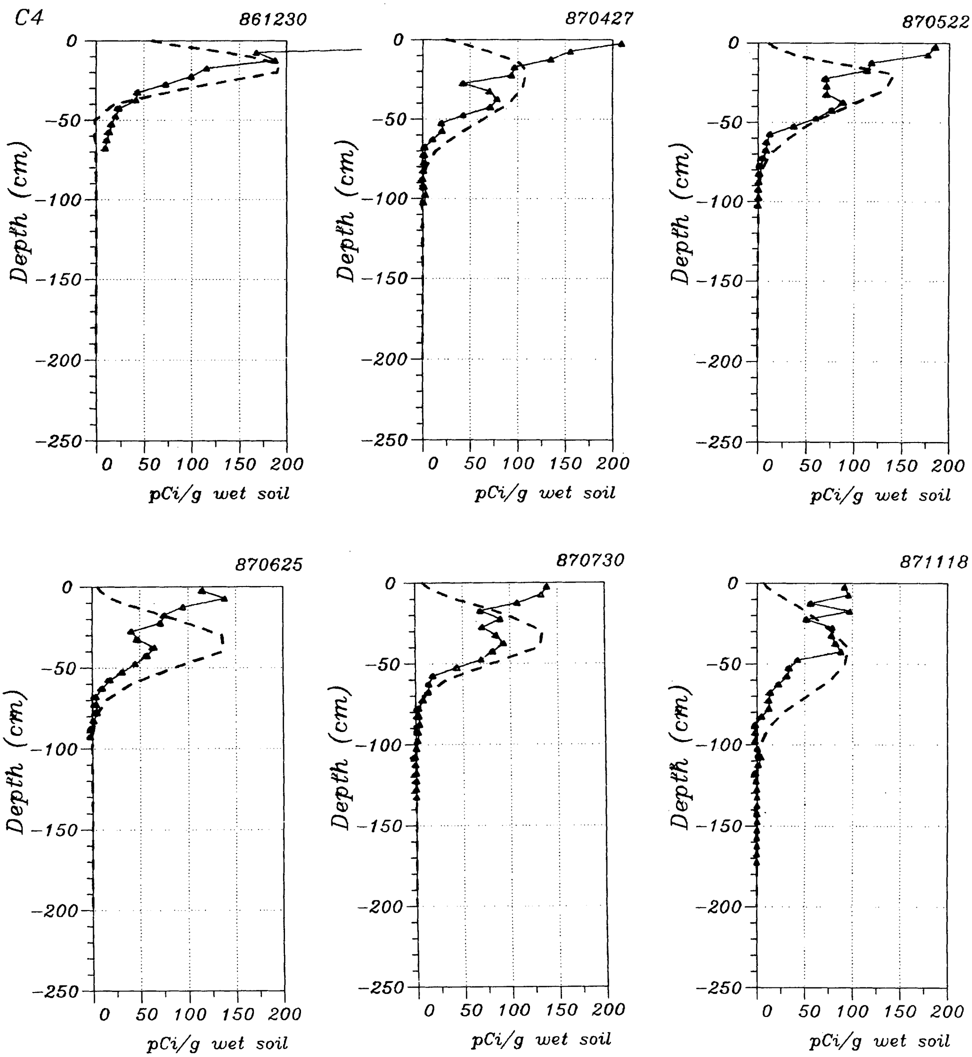


Fig. 16. Measured (Δ) and simulated (–) concentration – depth profiles for field point C4, Taastrup site.

Solute Transport

Tracer experiments were carried out at the six field locations Cl-C6 by injecting 171 μCi of ^{60}Co-complex at each point over an area of 1.767 m^2. The injections were made on December 11, 1986, and 4 litres of freshwater were applied shortly afterwards at the soil surface. The movement of the tracer was measured by using the gamma logging method until November, 1987, when the field investigations were terminated.

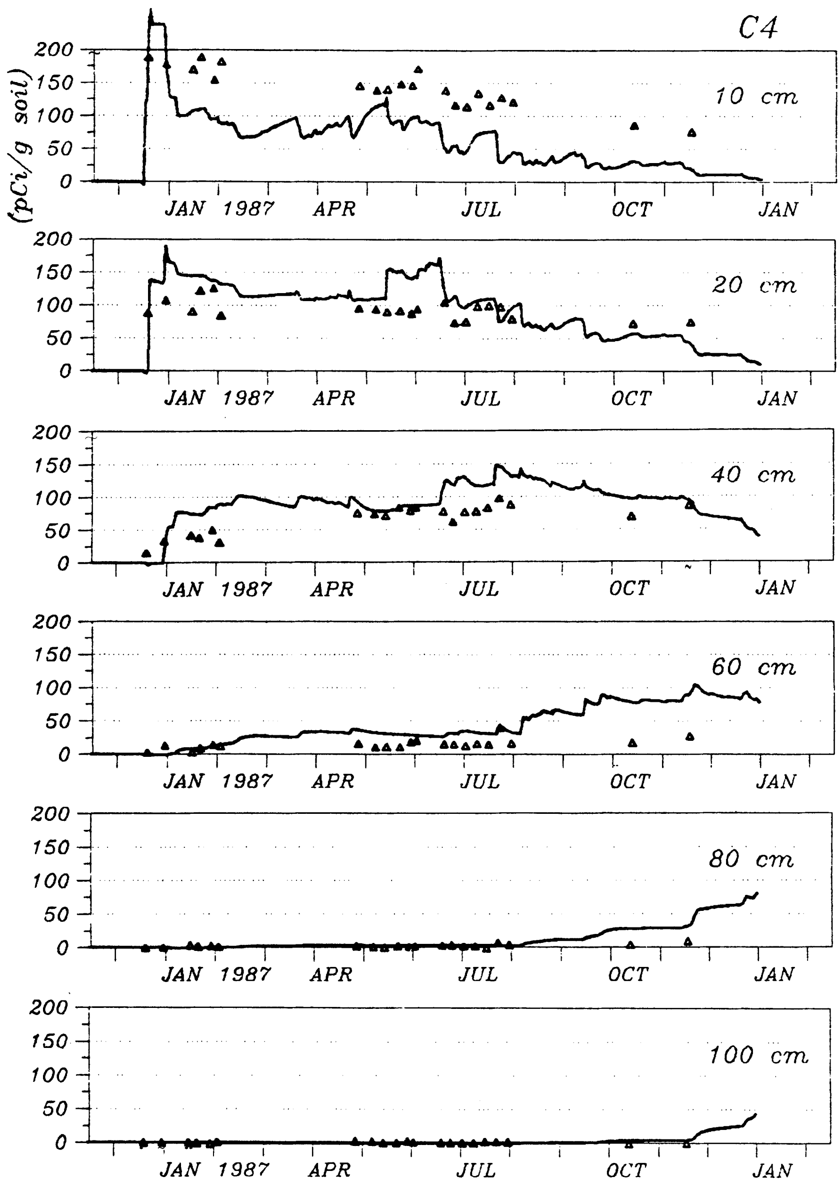


Fig. 17. Measured (Δ) and simulated (–) time series of concentration at selected depths for field point C4, Taastrup site.

The deconvoluted measurements at field location C4 converted to concentrations are shown in Fig. 16 as concentrations *versus* depth profiles at selected times and in Fig. 17 as time series at selected depths. In the same figures the model simulations are shown assuming vertical transport and dispersion, which is justified in view of the relatively large tracer application area. The longitudinal dispersivity is estimated to 5 cm on the basis of calibration, and the tracer is assumed to behave as a non-reactive solute.

Some deviations are observed between measured and simulated concentrations, particularly in the upper 20 cm where the measured values are generally higher. However, due to the experimental procedure involving gamma logging and deconvolution of directly measured data, the processed data are more uncertain close to the surface than in the deeper layers. Altogether, the simulated concentrations provide a reasonable approximation to the measured values.

Generally, the model results indicate that the transport of a conservative solute can be simulated reasonably well. However, the travel distance of the injected tracer was just about 50 cm due to an observation period of only 11 months.

On the basis of the available data, the model results are encouraging although a model verification for a travel distance beyond 50 cm would be more interesting from a practical point of view. Under such conditions the effect of spatial variability may be more significant.

Conclusions

Mathematical models for unsteady vertical water flow and solute transport in the unsaturated zone were applied to soil profiles within two research fields representing a coarse sand and a sandy loam, respectively. Model simulations were compared to field measurements of water content, suction, and solute concentration. The parameter assessment of the retention and hydraulic conductivity functions was based on laboratory measurements of the main drying retention curve and of the saturated hydraulic conductivity. For the coarse sandy soil the laboratory data were used directly together with a power function for unsaturated hydraulic conductivity. The parameters in this function were determined from the retention characteristics in combination with an assumption of a very low hydraulic conductivity prevailing at field capacity.

For the sandy loam this rather simple parameter identification technique was not applicable due to the influence of hysteresis and hypass flow. It was necessary to introduce empirical corrections to both the retention and the hydraulic conductivity functions in order to obtain reasonable comparisons with the observations of water content and suction.

With these adjustments an overall good agreement was obtained between simulated and observed flow and transport variables at both field sites.

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References

Blem, B., and Clausen, T. (1986) Hydraulic conductivity and dispersion in unsaturated soil. Institute of Hydrodynamics and Hydraulic Engineering (ISVA), Technical University of Denmark. M.Sc. Thesis (in Danish).

Bresler, E., and Dagan, G. (1981) Convective and pore scale dispersive solute transport in unsaturated heterogeneous fields, *Water Resour. Res., Vol. 17 (6)*, pp. 1683-1693.

Bresler, E., and Dagan, G. (1983 a) Unsaturated flow in spatially variable fields, 2. Application of water flow models to various fields, *Water Resour. Res., Vol. 19 (2)*, pp. 421-428.

Bresler, E., and Dagan, G. (1983 b) Unsaturated flow in spatially variable fields, 3. Solute transport models and their application to two fields, *Water Resour. Res., Vol. 19 (2)*, pp. 429-435.

Butters, G. L., and Jury, W. A. (1989) Field scale transport of bromide in an unsaturated soil 2. Dispersion modelling, *Water Resour. Res., Vol. 25*, pp. 1583-1589.

Byers, E., and Stephens, D. B. (1983) Statistical and stochastic analyses of hydraulic conductivity and particle size in fluvial sand, *Soil Sci. Am. J., Vol. 47*, pp. 1072-1087.

Curtis, A. A., Watson, K. K., and Jones, M. J. (1987) The numerical analysis of water and solute movement in scale heterogeneous profiles, *Transport in Porous Media, Vol. 2*, pp. 479-496.

Dagan, G., and Bresler, E. (1983) Unsaturated flow in spatially variable fields, 1. Derivation of models of infiltration and redistribution, *Water Resour. Res., Vol. 19 (2)*, pp. 413-420.

Destouni, G., and Cvetkovic, V. (1991) Field scale mass arrival of sorptive solute into the groundwater, *Water Resour. Res., Vol. 27 (6)*, pp. 1315-1325.

Hansen, S., and Jensen, H. E. (1988) Spatial variability of soil physical properties. Theoretical and experimental analyses, II. Soil water variables – Data acquisition, processing and basic statistics. Royal Veterinary and Agricultural University, Denmark (internal report).

Hansen, S., Storm, B., and Jensen, H. E. (1986) Spatial variability of soil physical properties. Theoretical and experimental analyses, I. Soil sampling, experimental analysis and basic statistics of soil physical properties. Royal Veterinary and Agricultural University, Denmark (internal report).

Hillel, D. (1980) *Fundamentals of soil physics*. Academic Press, New York.

Jensen, K. Høgh (1983) Simulation of water flow in the unsaturated zone including the root zone. Institute of Hydrodynamics and Hydraulic Engineering. Technical University of Denmark, Series Paper No. 33, 259 pp.

Jensen, K. Høgh, Ammentorp, H. C., and Sevel, T. (1984) Modelling of solute transport in the unsaturated zone, *Nordic Hydrology, Vol. 15*, pp. 223-234.

Mantoglou, A., and Gelhar, L. W. (1987 a) Stochastic modelling of large-scale transient unsaturated flow systems, *Water Resour. Res., Vol. 23 (1)*, pp. 37-46.

Mantoglou, A., and Gelhar, L. W. (1987 b) Capillary tension head variance, mean soil moisture content, and effective specific soil moisture capacity of transient unsaturated flow in stratified soils, *Water Resour. Res., Vol. 23 (1)*, pp. 47-56.

Mantoglou, A., and Gelhar, L. W. (1987 c) Effective hydraulic conductivities of transient unsaturated flow in stratified soils, *Water Resour. Res., Vol. 23 (1)*, pp. 57-67.

Mualem, Y. (1978) Hydraulic conductivity of unsaturated porous media. Generalized macroscopic approach, *Water Resour. Res., Vol. 14 (2)*, pp. 325-334.

Nielsen, D. R., Biggar, J. W., and Erh, K. T. (1973) Spacial variability of field-measured soil.water properties, *Hilgardia, Vol. 42 (7)*, pp. 215-260.

Richards, B. G. (1967) A review of methods for the determination of the moisture flow properties of unsaturated soils, Technical Memorandum No. 5, CSIR0, Melbourne.

Russo, D., and Bresler, E. (1981) Soil hydraulic properties as stochastic processes 1) Analysis of field spatial variability, *Soil Sci. Soc. Amer. J., Vol. 45*, pp. 682-687.

Sevel, T., Butts, M. B., and Genders, S. (1988) Spatial variability of soil physical properties. Theoretical and experimental analysis. Radioisotope tracer studies, field measurements and analysis. Danish Isotope Center (internal report).

Schulin, R., van Genuchten, M. T., Flühler, H., and Ferlin, P. (1987) An experimental study of solute transport in a stony field soil, *Water Resour. Res., Vol. 23 (9)*, pp. 1785-1794.

Villholth, K., Jensen, K. Høgh, and Fredericia, J. (1991) Field investigations of preferential flow behaviour. Hydrological Interactions between Atmosphere, Soil, and Vegetation (Proceedings of the Vienna Symposium, August 1991). IAHS Publication No. 204, pp. 245-261.

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