USE OF SOIL SURVEY DATA TO IMPROVE SIMULATION OF WATER MOVEMENT IN SOILS

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"Science is uncertain; the moment that you make a proposition about a region of experience that you have not directly seen then you must be uncertain. But we always must make statements about the regions that we have not seen, or the whole business is no use".

(R.P. Feynman, The Character of Physical Law, Cornell, 1964)

Aan Liesbeth, Lidwien, Hedwig en Lilian

STELLINGEN

1. Differences among soil hydraulic functions should preferably be determined on the basis of functional properties to be calculated with these functions as input data, rather than on a statistical comparison of the functions themselves.
   Dit proefschrift.

2. When characterizing soil water regimes the challenge is to identify a modelling approach that provides a quantitative estimation of system behaviour using a relatively simple and cheap data set, and to provide an indication of the reliability of predictions.
   Dit proefschrift.

3. Physical characterization of horizons distinguished in soil survey and functional testing of their differences results in a substantial reduction of the number of layers with different soil physical behaviour.
   Dit proefschrift.

4. Once a sufficient large data base of measured soil water retention and hydraulic conductivity functions is established, it is attractive to use this information to derive class and continuous pedotransfer functions relating these hydraulic functions with soil survey data such as texture, organic matter and horizon designation.
   Symposium on Land qualities in space and time. 1988.
   Dit proefschrift.

5. De door Warrick et al. herontdekte schaleringstechniek beschrijft weliswaar op eenvoudige wijze de variabiliteit in bodemfysische karakteristieken, maar leidt tevens tot een onderschatting van de variabiliteit in modeluitkomsten.
   A.W. Warrick et al. 1977. Water Resour. Res. 13: 355-362.
   Dit proefschrift.

6. Because of spatial and temporal variability, it is not unlikely that predictive models eventually, or perhaps already, yield estimates for hydraulic functions that are sufficiently accurate for many field applications.
   D.R. Nielsen et al. 1986. Water Resour. Res., 22: 898-1088.

7. While the soil water retention and hydraulic conductivity functions are the crucial parameters for predicting unsaturated flow, their theoretical description and measurement remains a continuous and sometimes frustrating challenge for hydrologists and soil scientists.
   D.R. Nielsen et al. 1986. Water Resour. Res., 22: 898-1088.

8. Bij de bepaling van de bodemfysische karakteristieken van een veldlocatie door inverse modellering wordt ten onrechte verondersteld dat deze karakteristieken de enige onbekenden zijn.

9. Verantwoord generaliseren is één van de moeilijkste bezigheden, maar vormt een vaardigheid waar veel behoefte aan is.
   J. Bouma. 1988. Inaugurele rede.

10. Soil survey and land evaluation will have a future when three elements are emphasized in future work: (1) a focus on areas of land that occur in a geographical context; (2) emphasis on observations and measurements in undisturbed field soils either to obtain data for simulation modelling or to validate simulation results; and (3) the need to integrate knowledge from a wide range of disciplines from geology to socio-economics.
    J. Bouma. 1989. Advances in Soil Science, 9: 177-213.

11. As the use of soil survey data, especially soil series characterization, increases, more precise statements will be needed about the means, ranges and simple and multiple relationships of soil properties.
    M.E. Collins and T.E. Fenton. 1984. Soil Sci. Soc. Am. J. 48: 1107-1114.

12. De bepaling van bodemfysische karakteristieken mag niet de sluitpost van de projectbegroting zijn.

13. Indien de ontwikkelingen in Oost-Europa leiden tot een vermindering van de hulp aan ontwikkelingslanden, zegt dit meer over het eigenbelang van de eerste dan over de nood in de derde wereld.

14. Doelmatig trimmen: fietsen naar het werk.

Stellingen behorende bij het proefschrift van J.H.M. Wösten
Use of soil survey data to improve simulation of water movement in soils
Wageningen, 19 juni 1990.

WOORD VOORAF

Bij het hier beschreven onderzoek heb ik de hulp en steun gehad van velen.

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Karel Hulsteijn bedankt voor de fraaie omslag.

ABSTRACT

Wösten, J.H.M. 1990. Use of soil survey data to improve simulation of water movement in soils. Doctoral thesis. Wageningen Agricultural University, Wageningen, The Netherlands, (IX) + 103 pp.

Soil water retention and hydraulic conductivity functions are crucial input data for models that simulate water movement in soils. When these functions have to be generated for areas of land, the intended application of the results of the modelling determines the level of generalization at which the problem should be addressed. In the research described in this thesis a soil map at a scale corresponding to the identified level of generalization, is used as the basic document from which soil hydraulic functions are derived for an area of land. The hydraulic functions of the major pedological horizons distinguished during the soil survey are measured. Pedological differences do not necessarily correspond with soil hydraulic differences. This results in a limited number of hydraulic "building blocks" with a significantly different soil physical behaviour. Transforming the major pedological horizons into "building blocks" provides the information to transform the soil map into a map of soil physical units.

Direct measurement of the hydraulic functions is cumbersome and costly. As an alternative, the existing data base of measured hydraulic functions is analysed and pedotransfer functions are developed. The use of pedotransfer functions is found to be a cost-effective method of translating the basic soil data recorded during soil survey into hydraulic functions.

A concept of functional criteria is introduced for evaluating differences in the hydraulic functions measured in the major pedological horizons distinguished and for evaluating different methods of generating these functions. Functional criteria are practical aspects of soil behaviour calculated with the hydraulic functions as input. Hydraulic functions are not an aim in themselves, but serve as input data for simulation models. Therefore, in this study the evaluation of differences in these functions is based on the evaluation of differences in calculated functional criteria and not on a statistical comparison of the functions themselves.

Finally, aspects of spatial and temporal variability are investigated. The scaling technique is successfully applied to quantify the complex spatial variability in measured hydraulic functions in a distribution function of scale factors. However, using the results of scaling to calculate the variability in model output results in a conservative estimate of this variability. Model output is also affected by temporal variability in meteorological and water table data. Meteorological and water table data from a 30-year period are used to calculate moisture deficits and trafficability. The influence of temporal variability is reflected by presenting graphs that show the probability of the occurrence of moisture deficits and adequate trafficability.

Additional index words: soil map, map scale, water retention, hydraulic conductivity, pressure head, hydraulic function, simulation model, spatial variability, pedotransfer function, scaling

CONTENTS

1. GENERAL INTRODUCTION

Regrettably, agricultural and industrial activities are increasingly causing the quality of our soils and waters to deteriorate. The excessive use of fertilizers, pesticides and inorganic and organic chemicals has already caused considerable environmental damage.

In order to partly control and ultimately rectify this damage, scientists have developed increasingly complex computer models that simulate water and solute movement in the unsaturated zone of the earth's crust. These models have now become indispensable in research aiming to quantify and integrate the most important physical, chemical and biological processes active in the unsaturated zone of agricultural soils.

Simultaneously, models ranging from very simple to highly complex are being used to evaluate the effects of alternative management practices on crop yield and groundwater quality. The use of models for research and management has shown that very many input data have to be quantified in order to be able to make reliable predictions on a field scale. In hydrology, it is not the models themselves but the lack of accurate functions (particularly on soil hydraulics) that is considered to be the major obstacle to progress. The most important hydraulic functions in this context are water retention and hydraulic conductivity.

It could be postulated that it is acceptable to use available data as long as the uncertainty of the resulting predictions is borne in mind. The use of "perfect" data, even if they exist, is not always necessary, because many research and management problems do not require exact solutions. For every problem the challenge is to identify a modelling approach that provides a quantitative estimate of system behaviour from a relatively simple and cheap data set while providing an indication of the reliability of predictions.

The need to make realistic predictions and the fact that it is neither possible nor necessary to postpone these predictions until all unknown parameters are known, creates an interesting subject for study. Obviously, in order to deal effectively with water movement in soil on a field scale stochastical approaches that deal with soil heterogeneity are also important. However, in this thesis, the emphasis is on a deterministic approach. Eventually, both approaches must be integrated.

As a first step in generating hydraulic functions for field scale application it is necessary to identify the level of generalization at which potential users need to have their information. These levels of generalization differ for different users, as is illustrated in Chapter 2. In that chapter the changes in grass yield that result when water tables fall because water is extracted to supply drinking water are calculated. The individual farmers affected by this phenomenon are interested in a detailed study in order to obtain a just settlement of their financial losses. However, a water extraction company might be satisfied with information on a much more general scale, providing it can calculate the total amount of money needed to settle yield reductions in the water extraction area as a whole. Identification of the level of generalization can avoid money and energy being spent to provide answers that do not correspond with the questions being asked.

Once an appropriate level of generalization has been identified for a specific problem it is attractive to use the soil map on a scale that corresponds with the level of generalization, as a point of departure to derive soil hydraulic functions for areas of land. In Chapter 3 this approach is described and tested. An area of 650 ha was surveyed at a scale of 1 : 5 000. Because the soil survey data were to be used for hydrological interpretations, the description of the soil properties could not be restricted to the surface horizons; instead the properties of all horizons were recorded. As a result, a set of major pedological horizons was identified. Multiple measurements of the water retention and hydraulic conductivity curves of each of these horizons were made. Next, those major pedological horizons whose hydraulic functions did not differ significantly were combined into soil hydraulic "building blocks" which were used to transform the soil map into a simulation map that showed a characteristic sequence of these "building blocks". Because the occurrence of major pedological horizons is derived from the soil map, this map can in turn be used to extrapolate information on the physical characteristics of the soil.

In the 1980s soil water retention and hydraulic conductivity curves were measured in a large number of soils in the Netherlands in projects such as those described in Chapter 3. Large, undisturbed soil samples were measured. All the soil horizons measured in these projects were classified according to soil texture and type of horizon (the latter being either topsoil, i.e. A horizon, or subsoil, i.e. B and C horizons). This classification resulted in 20 different soil groups comprising a total of 197 individual curves. The individual curves form an unique data base covering a broad spectrum of soils in the Netherlands. The geometrically averaged curves for each of the 20 soil groups were calculated and tabulated forms of these averaged curves were presented. As a set, the curves form a "class pedotransfer function" that translates the soil survey data (soil texture and type of horizon) into average soil hydraulic functions for each texture class for top- and subsoils.

Given this data base, Chapter 4 describes how this information is used to derive relations that allow the hydraulic functions to be estimated from soil survey data, thus avoiding cumbersome and costly direct measurements of these functions. Analytical functions for the moisture retention and hydraulic conductivity curve were fitted on the 197 individual curves. A nonlinear least-squares optimization program was used to estimate the parameters of the analytical functions simultaneously from the data on soil water retention and hydraulic conductivity. After the parameters had been estimated multiple regression techniques were used to investigate to what extent the estimated model parameters are dependent on basic soil properties that are more easily measured, such as percent silt, percent clay, percent organic matter, bulk density and median sand particle size. Linear, reciprocal and exponential relationships of these basic soil properties were used in the regression analysis, and possible interactions among the soil properties themselves were also investigated. The regression model consists of various basic soil properties and their interactions, all of which contribute significantly to the optimized soil hydraulic parameters. The regression model forms a "continuous pedotransfer function" that translates basic soil survey data into individual soil hydraulic functions. Chapter 4 also provides an evaluation of the accuracy of the hydraulic functions estimated from the regression models.

Hydraulic functions obtained either by direct measurement or estimated from regression models are not an aim in themselves but serve as input data in studies that examine functional properties of soil behaviour. Therefore, differences among the hydraulic functions of different soil horizons are preferably determined on the basis of functional properties to be calculated with these functions as input data, rather than from a purely statistical comparison of the functions themselves. For this purpose, three functional properties are introduced in Chapter 5:

- calculated travel times for water to move from the soil surface to a defined water table;
- calculated water table levels which allow a defined upward-flux density to a defined level;
- calculated downward-flux densities that correspond with a defined air content in the soil.

These three functional properties are not unique; other properties that enable differences in hydraulic functions to be assessed can also be identified. In Chapter 5 the physical behaviour of different soil horizons is compared by comparing the calculated functional properties for these horizons. Using the results, pedological horizons can be combined into soil hydraulic "building blocks" that have significantly different soil hydraulic functions.

The need to evaluate soil physical behaviour on the basis of functional properties in combination with the fact that methods to estimate hydraulic functions range from complicated and costly to simple and cost-effective inspired the studies described in Chapters 6 and 7. Four different methods were used to generate hydraulic functions:

- direct measurement of the hydraulic functions of the horizons of the soil profiles at three sites;
- hydraulic functions obtained by averaging the measured hydraulic functions of major soil horizons sampled on a regional scale;
- the same as above, but now sampled on a national scale;
- the hydraulic functions sampled on a national scale were fitted to analytical expressions and the model parameters were related to more easily measured soil properties.

In Chapter 6 the water storage in the upper 0.5 m of the soil profile is used as a functional property to compare the hydraulic functions generated with four different methods for three sites in a catchment. Independent measurements of water storage with a neutron probe, for a period of seven years, serve as a reference for the evaluation of the performance of the different methods to generate soil hydraulic functions. The results of the comparison of calculated and measured water storage in the upper 0.5 m, in combination with an indication of the costs of using each method provide information that enables the best way of spending limited available resources to be chosen.

In Chapter 7 the same four methods to generate soil hydraulic functions are used to calculate two functional properties that are relevant for agricultural and environmental use:

-evapotranspiration deficit;

-flux through a plane at 0.3 m below soil surface.

In contrast to the water storage in the upper 0.5 m, used in Chapter 6 as a functional property, no independent measurements of deficits and fluxes were available. Meteorological data strongly affect the calculated deficits and fluxes. These meteorological influences are eliminated by using rainfall deficit as a covariable. The remaining variability was used to evaluate differences between the four methods of generating soil hydraulic functions.

Model output is influenced considerably by the soil hydraulic functions that are used as input. Therefore, in an increasing number of studies it is necessary to quantify the variability in hydraulic functions, so that model output can be presented in the form of a mean and variance. Using the measured individual curves of the data base developed for the Netherlands, in Chapter 8 scaling techniques are used to quantify the variability in the hydraulic functions of three large soil groups; coarse-textured, medium-textured and fine-textured. It is investigated if scaling is an attractive technique to simplify the description of the complex hydraulic heterogeneity of soils and whether the technique provides a good estimation of the variability in hydraulic functions.

In addition to spatial variability, temporal variability is also present. It is dealt with in Chapter 9, where meteorological and water table data are used to calculate moisture deficits and trafficability over a period of 30 years. The results of this study are condensed and described in terms of the probability of the occurrence of moisture deficits and of adequate trafficability; this provides the user with a choice of possible management strategies.

2. LAND EVALUATION AT DIFFERENT SCALES: YOU PAY FOR WHAT YOU GET!

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LAND EVALUATION AT DIFFERENT SCALES: YOU PAY FOR WHAT YOU GET!

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ABSTRACT

In a survey area of 1435 ha, water tables are being lowered due to water extraction for drinking-water supply. The resulting change in yield of grass in this area was simulated by comparing the former hydrological situation without extraction to the present hydrological situation with extraction. A detailed soil survey of the area on scale 1 : 10 000 was used to derive additional soil maps on scales 1 : 25 000, 1 : 50 000 and 1 : 250 000. The 'representative profiles' of the mapping units of the four soil maps were physically interpreted and were used thereafter to simulate changes in yield.

Working on a certain scale has implications for the detail of information to be provided and for costs incurred. The average change in yield for the area as a whole could, with relatively low cost, be calculated on scale 1 : 50 000. However, changes in yield for a specific farmer's field could most accurately be calculated on scale 1 : 10 000 at relatively high cost. Taking these values as a reference, increasing deviations were found when using smaller map scales. Ideally, a user should first select the degree of detail by which a given answer has to be known. Then, a corresponding mapping scale can be selected. This procedure assures an optimal use of resources.

INTRODUCTION

In the Netherlands, soil maps are increasingly used for environmental interpretations in the context of new laws on soil protection. Such interpretations have to be quantitative so as to allow reliable estimates of soil properties in relation to critical levels that are being defined in soil protection laws. These critical levels may, for example, relate to chemical contents of heavy metals and phosphates or to physical characteristics such as the density of compacted soil horizons and associated physical land qualities.

Environmental interpretations are needed at different levels. At a national level, broad information is needed, for example, as to the location of soils that have a low trafficability or a low capacity to adsorb phosphates. The new 1 : 250 000 soil map of The Netherlands (Steur et al, 1985) is successfully applied in this context. Interpretations at a regional or provincial level are usually more detailed, and standard 1 : 50 000 soil maps are being used here. Quite specific interpretations at farm level are made using 1 : 10 000 maps that are made for all land reallotment plans in The Netherlands.

Interpretations of soil maps of any scale are generally based on descriptions of 'representative profiles' for each mapping unit which have been made by soil surveyors who are familiar with the area. Usually, these 'representative profiles' are described in considerable detail irrespective of the scale of the maps for which they are prepared. As a consequence, interpretations are also quite detailed. Ideally, however, interpretations for small-scale maps should be less detailed than those for large-scale maps because mapping units occupy smaller areas of land in the latter case and variability within these smaller areas is generally lower than in similar mapping units of small-scale maps which occupy larger areas of land.

Modern quantitative interpretations, as discussed above, require specific statements about the variability within mapping units being considered. Rather than average values, more emphasis is placed on probability assessments (Wösten & Bouma, 1985). For this, statistical or geostatistical procedures have to be used. Unfortunately, emphasis on geostatistics for characterizing soil spatial variability has resulted in many scientific papers but not yet in practical operational procedures to be applied in the context of soil survey.

An additional aspect addressed in this study relates to benefit:cost ratios of soil surveys at different scales. Small-scale surveys are less costly than large-scale ones but the information provided is less specific, at least in principle. Ideally, a user should consider the degree of detail he needs to answer his questions. Then, in turn, the number of observations to be made during mapping and the associated map scale can be chosen. So far, this benefit:cost analysis has not been available for our soil survey users. This study was made, therefore, to compare application of soil surveys at different scales ranging from 1 : 250 000 to 1 : 10 000. The application was focused on predicting the sensitivity of soils to lowering of the watertable following water extraction for drinking-water supply.

MATERIALS AND METHODS

The survey area 'Mander' covered 1435 ha in the eastern part of The Netherlands near Tubbergen. Out of this survey area a sample area of 404 ha was considered in this study. The higher eastern part of the sample area is underlain by Tertiary clay sediments starting 150 cm below the present land surface. The area was affected by glaciers. After the glacial period, rivers transported sediments from the higher eastern part towards the lower western part forming deposits of clay and loam in the subsoil. Later, aeolian sands were deposited over the entire area.

The soil survey was made at a scale of 1 : 10 000, with an average observation density of 1.5 borings per hectare (Stoffelsen & Van Holst, 1985). Borings extended to a maximum depth of 3.2 m, to the upper surface of the Tertiary clay, or to the Mean Lowest Water table (MLW). Existing mapping units and classifications of the Dutch soil survey (Table 1) mainly reflect properties of horizons near the soil surface (De Bakker & Schelling, 1966). However, when the soil survey is to be used for hydrological interpretations, it should also include data on those subsurface soil horizons expected to be important for the soil physical behaviour of the profile. Observations were made of:

- thickness and sequence of the different soil horizons;

- organic matter content, percentage loam and the coarseness of the sand fraction of all horizons;

- presence of special horizons such as gravel and Tertiary clay;

- rooting depth of the dominant grass vegetation;

- estimated fluctuation of the watertable in the present situation in terms of the mean highest (MHW) and the mean lowest (MLW) levels;

- estimated fluctuation of the watertable in the former situation that was not influenced by extraction.

Mapping units occurring in the sample area represent major units of sandy soil in The Netherlands. Mapping units used in detailed Dutch soil surveys, are shown in Table 1.

Table 1 Soil classification according to Soil Taxonomy (Soil Survey Staff, 1975) of the mapping units occurring in the sample area. All families are sandy, siliceous and mesic.

Mapping unit	Thickness A1 or Ap horizon (cm)	% <50 um in surface soil	Classification (*Soil Taxonomy*)
Hn51	30	0 - 10	Typic Haplaquods
Hn53	30	10 - 17.5	Typic Haplaquods
cHn53	30 - 50	10 - 17.5	Plaggeptic Haplaquods
zEZ53	50 - 100	10 - 17.5	Plaggept
dzEZ53	100	10 - 17.5	Plaggept
tZg53	15 - 30	10 - 17.5	Typic Humaquepts
tZg55	15 - 30	17.5-32.5	Typic Humaquepts
tZn51	15 - 30	0 - 10	Typic Humaquepts
tZn53	15 - 30	10 - 17.5	Typic Humaquepts
tZn55	15 - 30	17.5-32.5	Typic Humaquepts
cZn53	30 - 50	10 - 17.5	Plaggeptic Humaquepts
zWz	30	-	Histic Humaquepts

Water table fluctuations in mapping units are described in Table 2 according to Van Heesen (1970) and Van der Sluijs & De Gruijter (1985).

Table 2. Description of water table classes.

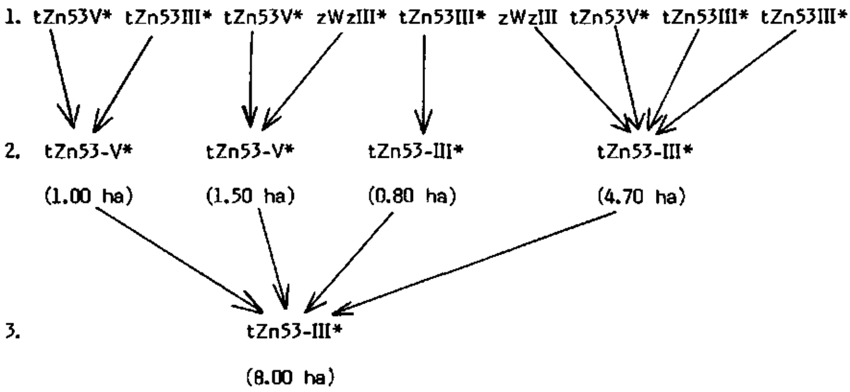
Water table class (Gt)	II	III	V	VI	VII
MHW	40	40	40	40-80	80
MLW	50-80	80-120	120	120	120

MHW = mean highest water level and MLW = mean lowest water level in cm below soil surface.
A * behind the code Gt III and V refers to a 'drier part', i.e. a MHW deeper than 25 cm below soil surface.
A * behind the code Gt VII refers to a 'very dry part', i.e. a MHW deeper than 140 cm below soil surface.

For every mapping unit of the 1 : 10 000 soil map, the soil surveyor defined a 'representative profile' on the basis of the borings in that unit. This included the water table class. Based on the 1 : 10 000 soil map, a soil map on scale 1 : 25 000 was derived, by the same soil surveyor, through generalisation without new fieldwork. The 'representative profile' for every mapping unit of the 1 : 25 000 soil map was assumed to be identical to the 'representative profile' of the mapping unit occupying the largest area on the 1 : 10 000 map, within the particular 1 : 25 000 mapping unit being considered. In the same way a soil map on scale 1 : 50 000 was derived based on the soil map on scale 1 : 25 000 and a soil map on scale 1 : 250 000 was derived based on the soil map on scale 1 : 50 000. Only one experienced soil surveyor was involved in making delineations, thus excluding errors due to different operators. The number of borings corresponded to the 1 : 10 000 map. When deriving the smaller-scale maps, the surveyor used his field experience, assuming availability of fewer borings as map scales decreased. The number of mapping units was much reduced as the map scale decreased.

In Table 3 the approach is illustrated for the mapping unit tZn53-III* (sandy, siliceous, mesic Typic Haplaquepts) on the soil map scale 1 : 50 000. Table 3 shows that this mapping unit was created by combining 4 mapping units of the soil map on scale 1 : 25 000 and by 9 mapping units of the soil map on scale 1 : 10 000.

Table 3. Example of combining nine mapping units of the soil map on scale 1 : 10 000, into four units of the map 1 : 25 000 and one unit of the map 1 : 50 000.



	scale	number of mapping units
1	1 : 10 000	9
2	1 : 25 000	4
3	1 : 50 000	1

Simulation map

For the soil physical interpretation of the mapping units of the 1: 10 000 soil map, use is made of descriptions of the 'representative profiles'. The separate soil horizons of the 'representative profiles' are characterized by measured soil physical characteristics (Wösten et al, (1985). When pedologically different horizons behave physically identically, they are combined into one soil physical horizon or 'building block'.

Simulation

The following data are necessary for the simulation of changes in yield:

- moisture retention curves (h(Θ)-relation) for surface and subsurface horizons;
- hydraulic conductivities (K(h)-relation) for subsurface horizons;
- MHW- and MLW-values in the present and former situation;
- thickness of the rootzone;
- a function describing the uptake of water from the rootzone (the so called 'sink term');

- precipitation and evaporation from a free water surface (Eo);

- a crop coefficient for calculating the potential evapotranspiration (Epot) from the evaporation from a free water surface (Eo).

The simulation model being used was proposed by De Laat (1980). It calculates a water balance for successive 10-day periods. Calculations were made for the period 1956-1985 for a grass vegetation with a growing season of 180 days from April 1 till September 30. The potential evapotranspiration (Epot) for the soil under grass cover is estimated as : Epot = 0.8 Eo. Moisture supply from the rootzone was assumed to decrease linearly with increasing log/h/ values beyond a critical value of log/h/ = 2.7.

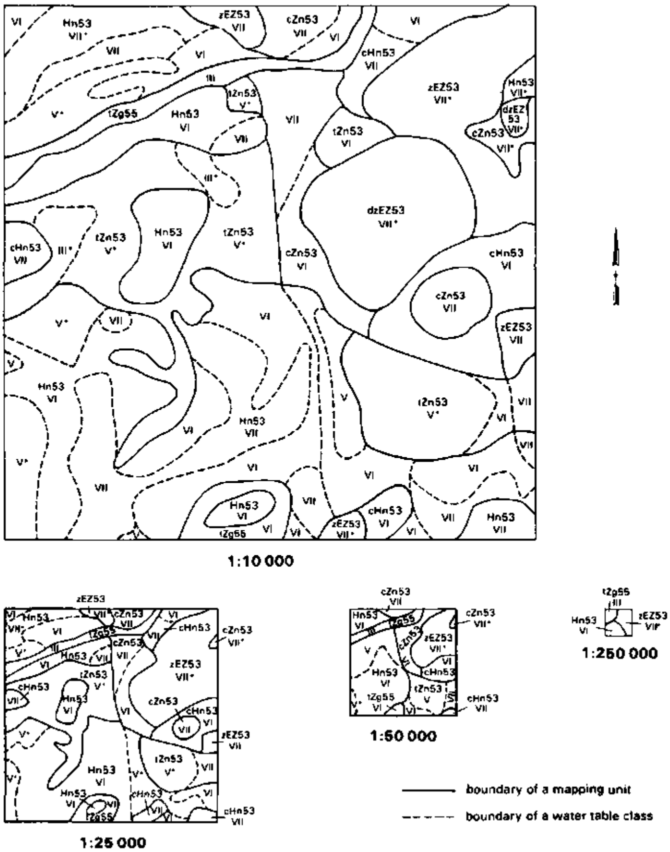


Figure 1. Segments of the soil maps of the sample area on four scales

Calculated moisture deficits for the two hydrological situations are transformed to relative yields by making use of the established relation between yield per mm moisture and potential production (Bohemen, 1981). The difference between relative yields in the former and present hydrological situation forms the yield reduction due to increased moisture deficit.

Water extraction may also lead to yield increase. This is the case for soils that were originally too wet. Increase in yield is derived from existing curves that relate changes in yield to MHW and MLW. Final results are presented as an average net change in yield, expressed as a yearly percentage, due to lowering of the water table. Negative percentages indicate a dominant effect of a lower water supply during the growing season. Positive percentages indicate a dominant effect of lower water tables during the wet periods of the year.

RESULTS

Soil map

Figure 1 shows a segment of the soil maps of the sample area on scales 1 : 10 000, 1 : 25 000, 1 : 50 000 and 1 : 250 000. The soil maps of the sample area of 404 ha contain 216, 126, 46 and 5 mapping units, respectively. According to common boring densities for the various scales, these maps would conceptually be based on approximately 606, 135, 54 and 0 borings respectively. Figure 2 shows costs as a function of map-scale for the sample area. Costs to calculate changes in yield can be divided into costs of the survey, costs of processing data and costs of simulation. Costs of soil survey are based on specific boring densities and on the number of hectares that can be surveyed per day. For the soil map on scale 1 : 250 000 no costs of survey are involved because this map has been compiled from existing 1 : 50 000 soil maps (Steur et al, 1985). Costs of processing data refer to transformation of the soil map into a simulation map, including costs of cartography. Costs of simulation refer to calculations for the 'representative profile' of each mapping unit for a period of 30 years in the former and present hydrological situation.

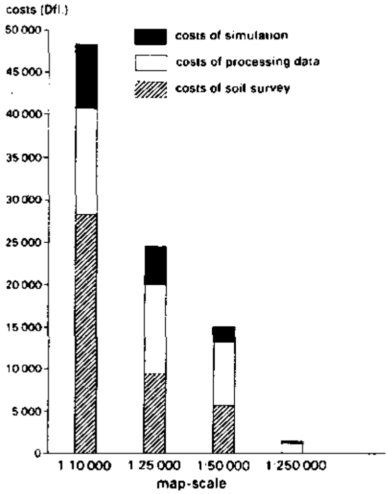


Figure 2. Relation between costs and the map-scale of the sample area of 404 ha

Figure 2 indicates that with decreasing scale, the number of borings and mapping units decreases strongly. So do the costs. Total costs per hectare for the sample area on scale 1 : 10 000, 1 : 25 000, 1 : 50 000 and 1 : 250 000 are Dfl 120, Dfl 60, Dfl 37 and Dfl 3 respectively. These values indicate the importance of choosing the right scale for specific problems. Answering a problem on scale 1 : 10 000 rather than on scale 1 : 25 000 doubles the cost. This may be unnecessary when the problem being addressed can be answered on scale 1 : 25 000.

Changes in yield due to lowering of the water table

Figure 3 shows a segment of maps of the sample area with changes in yield due to lowering of the water table calculated on four scales. For every mapping unit, the yearly average change in yield is indicated. The general pattern on these maps agrees well with what might be expected on the basis of the patterns on the soil maps.

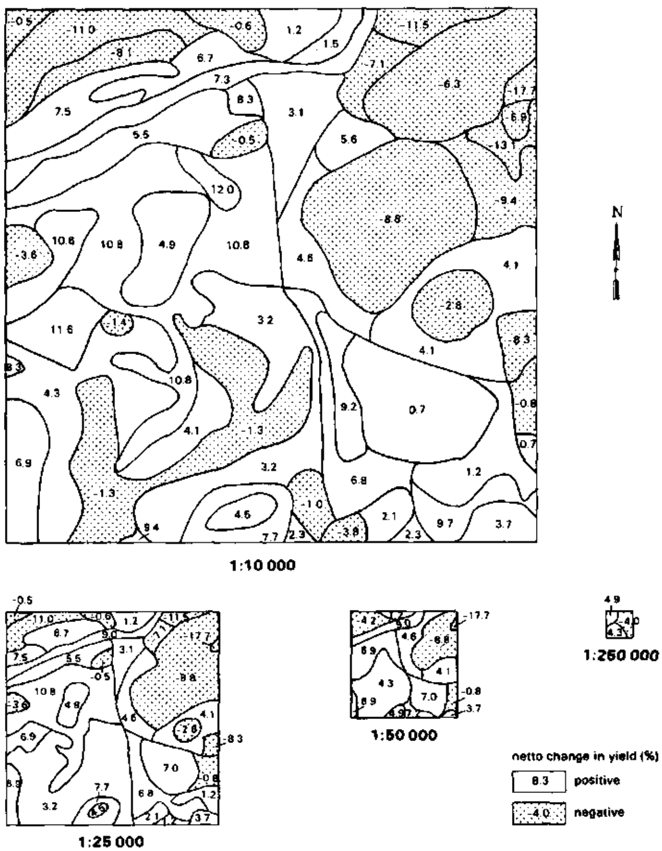


Figure 3. Segments of the maps of the sample area with changes in yield calculated on four scales

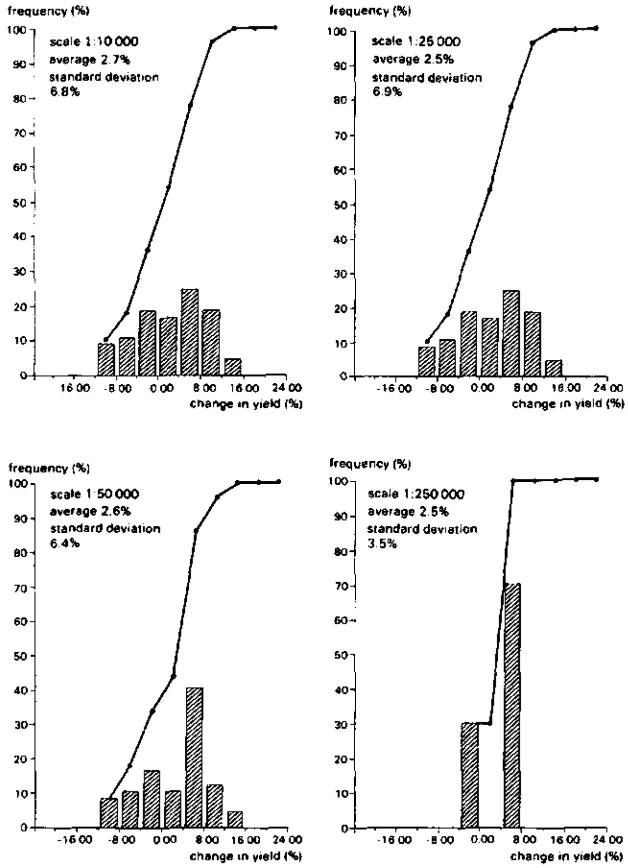


Figure 4. Relative and cumulative frequency distribution of changes in yield,weighted with the area,for the four scales

Figure 4 shows, for the sample area as a whole, the relative and cumulative frequency distribution of changes in yield, weighted with the area, for the four scales. The average change in yield and the standard deviation of the average for the entire area is respectively 2.7% (and 6.8%), 2.5% (and 6.9%), 2.6% (and 6.4%) and 2.5% (and 3.5%) for the four scales, going from small to large.

Table 4. Changes in yield of 20 separate parcels derived from maps on four scales. Changes in yield are expressed in percentages. Within brackets, the difference in change in yield is expressed in Dfl. per parcel per year as compared with scale 1 : 10 000.

parcel	area in ha.	reference 1:10 000	1:25 000	1:50 000	1:250 000
1	0.29	+ 5.5	+ 4.5 (-Dfl 15)	+ 4.5 (-Dfl 15)	- 4.1 (-Dfl 139)
2	1.60	- 8.8	- 8.8 (Dfl 0)	- 8.8 (Dfl 0)	- 5.3 (+Dfl 280)
3	1.20	- 9.4	- 8.8 (+Dfl 36)	- 8.8 (+Dfl 36)	- 4.0 (+Dfl 324)
4	1.14	- 5.2	- 4.0 (+Dfl 68)	- 4.0 (+Dfl 68)	- 4.0 (+Dfl 60)
5	1.21	+ 8.3	+11.5 (+Dfl 194)	+11.7 (+Dfl 206)	+ 6.6 (-Dfl 103)
6	1.14	+ 5.8	+ 5.4 (-Dfl 23)	+ 5.4 (-Dfl 23)	+ 4.3 (-Dfl 86)
7	0.88	+ 6.7	+ 6.7 (Dfl 0)	+ 6.7 (Dfl 0)	+ 4.7 (-Dfl 88)
8	0.80	-10.8	-10.8 (Dfl 0)	-10.8 (Dfl 0)	- 4.0 (+Dfl 272)
9	0.85	- 6.1	- 6.7 (-Dfl 26)	- 4.2 (+Dfl 81)	+ 0.2 (+Dfl 268)
10	0.90	+ 9.1	+ 9.1 (Dfl 0)	+ 8.9 (-Dfl 9)	+ 7.6 (-Dfl 68)
11	1.49	+ 7.4	+10.5 (+Dfl 231)	+ 9.1 (+Dfl 126)	+ 6.4 (-Dfl 75)
12	1.04	+ 7.5	+ 7.5 (Dfl 0)	+ 6.8 (-Dfl 36)	+ 4.9 (-Dfl 135)
13	1.09	- 0.7	- 0.9 (-Dfl 11)	+ 2.4 (+Dfl 169)	+ 4.9 (+Dfl 305)
14	0.90	+ 2.3	+ 2.3 (Dfl 0)	+ 7.2 (+Dfl 221)	+ 4.9 (+Dfl 117)
15	1.51	+ 4.7	+ 4.5 (-Dfl 15)	+ 5.3 (+Dfl 45)	+ 4.9 (+Dfl 15)
16	1.23	+ 8.0	+ 8.0 (-Dfl 0)	+ 6.3 (Dfl 105)	+ 5.9 (-Dfl 129)
17	1.36	+ 9.6	+ 9.6 (Dfl 0)	+ 7.8 (-Dfl 122)	+ 7.2 (-Dfl 163)
18	1.15	- 9.0	- 0.9 (Dfl 0)	- 0.9 (Dfl 0)	- 4.0 (+Dfl 288)
19	1.77	- 6.8	- 4.0 (+Dfl 248)	- 4.0 (+Dfl 248)	- 4.0 (+Dfl 248)
20	0.88	-10.8	-10.8 (Dfl 0)	-10.8 (Dfl 0)	- 4.0 (+Dfl 299)
average absolute change in yield per parcel			Dfl 43	Dfl 76	Dfl 173

These average values indicate that, for the sample area as a whole, the decrease in yield reduction due to excess of water is bigger than the increase in yield reduction due to moisture deficits. For the sample area as a whole, lowering the water table has a positive effect on the yield. The average values on the four scales are also about the same. This means that all maps are suitable if an average change in yield has to be obtained for the entire area. Calculation of this average change in yield may be useful for regional planning purposes. However, this procedure provides no information for changes in yield on the level of individual parcels of land. This information is crucial for individual farmers.

Changes in yield calculated for 20 parcels on four scales

In order to investigate the influence of scale on the accuracy of the calculated changes in yield for separate parcels of land, 20 parcels in the sample area were selected at random. For every parcel the change in yield expressed in Dutch guilders per year, was determined using maps on the various scales (Table 4). Results were compared with those of the 1 : 10 000 map because larger scale maps are not economically feasible, and this map was therefore considered to be the reference. Table 4 shows that differences among the changes in yield calculated for the various scales and scale 1 : 10 000 are considerable. Differences increase as the map scale decreases, illustrating a decreasing accuracy. Percentages in Table 4 were multipled by Dfl 50.- per percent change in yield (LAGO-report, 1984) to obtain a financial expression for the changes. These amounts were multiplied by the area of the parcels to obtain an average yearly change in yield.

Table 4 indicates that changes in yield of individual parcels can differ quite significantly depending on the scale of the map being used. Differences of Dfl 300 per parcel per year are possible. The average absolute change in yield, expressed in Dfl per parcel per year, increases strongly with decreasing scale.

CONCLUSIONS

The most appropriate scale at which to work is determined by the costs associated with working at a certain scale in relation to the desired accuracy of the presented information. In a water extraction area, different users will require different accuracies of information and will therefore require different scales. As a consequence they should also pay different prices.

A water extraction firm or a provincial waterboard may, for example, be provided with sufficient information while working on scale 1 : 50 000. If, however, information has to be provided on the level of parcels and this information has to be sufficiently accurate to facilitate an honest settlement of financial losses, then the 1 : 10 000 scale is necessary. The costs of survey at larger scales than 1 : 10 000 are prohibitive for the problems being studied here.

It is not possible to recommend any one particular scale. For every problem an evaluation is necessary, considering the desired accuracy of data and the associated costs.

This application focused on predicting the sensitivity of soils to lowering the water table. Comparable studies, however, are relevant for evaluating effects of scale on detail and costs of other environmental interpretations. In all cases, optimal spending of resources should be determined by defining the degree of detail of required data considering the problem being studied. This degree of detail is, in turn, associated with a particular mapping scale.

REFERENCES

Commissie Grondwaterwet Waterleiding bedrijven, Werkgroep Landbouwkundige Aspecten. 1984. Landbouwkundige aspecten van grondwateronttrekking, LAGO werkgroeprapport

De Laat, P.J.M. 1980. Model for unsaturated flow above a shallow water table, applied to a regional subsurface flow problem. Agricultural Research Report 895 (Pudoc, Wageningen)

De Bakker, H. & Schelling, J. 1966. Systeem van bodemclassificatie voor Nederland;de hogere niveaus (System of soil classification for The Netherlands; the higher levels), (Pudoc, Wageningen)

Soil Survey Staff. 1975. Soil Taxonomy: a basic system of soil classification for making and interpreting soil surveys. USDA-SCS Agric Handbook 436. US Government Printing Office, Washington, DC

Steur, G.G.L., de Vries, F. & van Wallenburg, C. 1985. Bodemkaart van Nederland 1 : 250 000. (Stichting voor Bodemkartering, Wageningen)

Stoffelsen, G.H. & van Holst, A.F. 1985. Bodemkundig-hydrologisch onderzoek in het waterwingebied Mander. Fase II: De bodemgesteldheid, de voormalige en huidige hydrologische situatie en de grondwaterstandsverlaging. Stiboka rapport nr. 1807

Van Bohemen, P.J.M. 1981. Toename van de produktie van grasland bij verbetering van de watervoorziening. ICW nota 1298

Van der Sluijs, P. & de Gruijter, J.J. 1985. Water-table classes: a method used to indicate seasonal fluctuation and duration of water tables on Dutch soil maps. Agricultural Water Management, 10(2), 109-125

Van Heesen, H.C. 1970. Presentation of the seasonal fluctuation of the water table on soil maps. Geoderma, 4, 257-278

Wösten, J.H.M. & Bouma, J. 1985. Using simulation to define moisture availability and trafficability for a heavy clay soil in The Netherlands. Geoderma, 35, 187-196

Wösten, J.H.M., Bouma, J. & Stoffelsen, G.H. 1985. Use of soil survey data for regional soil water simulation models. Soil Science Society of America Journal, 49, 1238-1244

3. USE OF SOIL SURVEY DATA FOR REGIONAL SOIL WATER SIMULATION MODELS

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Use of Soil Survey Data for Regional Soil Water Simulation Models[1]

J. H. M. Wösten, J. Bouma, and G. H. Stoffelsen[2]

ABSTRACT

A detailed soil survey of an area of 650 ha was used to obtain basic soil physical data for a simulation model of the water regime in the unsaturated zone. Nine major soil horizons were defined for the entire survey area in terms of pedological classification and easily measurable characteristics such as texture, structure, organic matter content, and bulk density. In these horizons multiple measurements of hydraulic conductivity and moisture retention curves were made, yielding average curves. Statistical comparison of the curves showed that only five of the nine major soil horizons were different from a soil-physical point of view. Next, representative soils for the mapping units were transformed into soils composed of a characteristic sequence of some of these five horizons. In this way a simulation map, containing data to be used for simulation of the soil water regime, was formed from the soil map. For a sample area of 125 ha the simulation map contained 41 delineated areas as compared to 110 delineations on the soil map. Sixty independent test borings indicated 80 ± 6% purity comparing borings and the legend of the simulation map. A simulation run for one pedon with the model SWATRE showed excellent agreement between measured and calculated actual evapotranspiration for the years 1976, 1977 and 1978. Major soil horizons, rather than individual points of observation, were used in this study as carriers of soil physical information, allowing extrapolations based on a limited number of measurements.

Additional Index Words: **Soil mapping, soil horizons, hydraulic conductivity, moisture retention.**

Wösten, J.H.M., J. Bouma, and G.H. Stoffelsen, 1985. Use of soil survey data for regional soil water simulation models. Soil Sci. Soc. Am. J. 49:1238–1244.

Water tables are being lowered in several areas of the Netherlands due to agricultural water management practices, sprinkler irrigation, and water extraction for municipal water supply. The effects on soil moisture regimes and associated land qualities need to be measured quantitatively for modern land evaluation.

Computer simulation of soil water regimes offers an effective means of making the necessary quantitative predictions of the effects of changes in agricultural water management practices (e.g., Busoni et al., 1983; Bouma and De Laat, 1981). There is a problem, however, as to how to collect and manipulate soil physical data when one deals with areas of land. An earlier study (Bouma et al., 1980) used "representative" soils from soil surveys, with corresponding physical data, for cells in a grid pattern, and distinguished four phases of data gathering. Use of grid-cells implies that a soil unit is selected which occupies the largest area within a cell. Other units that occur are ignored. Furthermore, the selection of soil units (phase II in Bouma et al., 1980) is based on pedological criteria that mainly emphasize characteristics of the surface soil even though subsurface soil characteristics are very important in the hydrological behaviour of soils. Also, in these procedures, the existing soil map is used without an analysis of soil variability.

The collection of soil physical data is time consuming and costly, often exceeding practical possibilities (phase IV in Bouma et al., 1980). Geostatistics can be used, in principle, to optimize and thus reduce sampling programs (e.g., Vachaud, 1982). However, even

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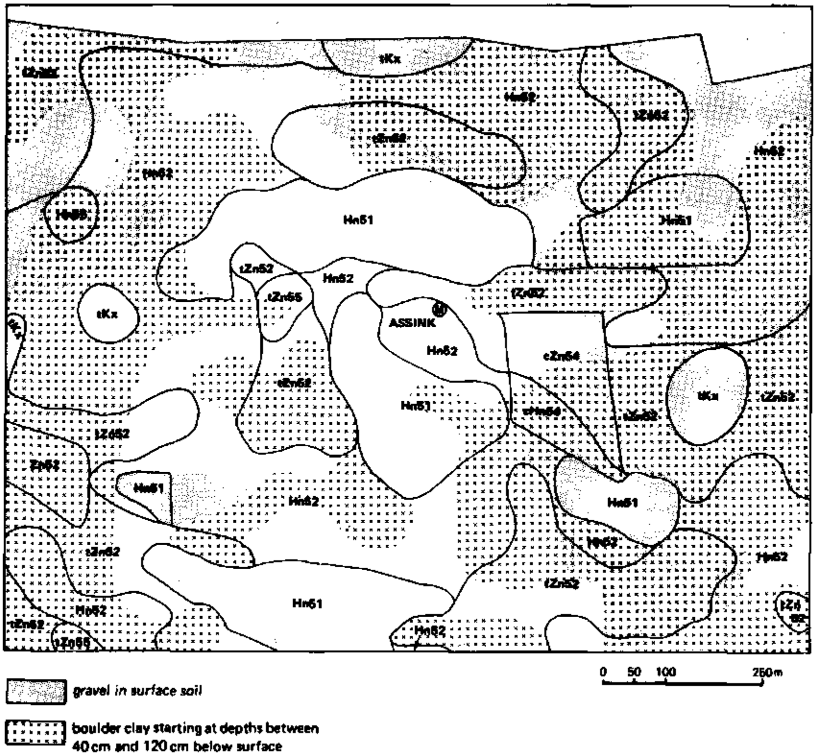

Fig. 1. Soil map of the sample area of 125 ha. The legend is explained in Table 1.

then, the amount of work to be done may be prohibitive from an operational and economical point of view.

This study focuses on operational aspects associated with the application of simulation models for the soil moisture regime. Specific objectives are: (i) to present an approach for using soil survey data in an efficient soil physical measurement program in terms of selection of measurement sites and of extrapolation of data obtained, and (ii) to partially evaluate the simulation results obtained.

MATERIALS AND METHODS

Soil Mapping

The survey area "Hupselse-Beek" covered 650 ha, and was located in the eastern part of The Netherlands near Groenlo. Out of this survey area a sample area of 125 ha (Fig. 1) was considered in this study. The area is underlain by Miocene clay sediments starting between 20-cm and 10-m below the present land surface. The area was affected by glaciers which deposited boulder clay (glacial till) in an early Pleistocene period. Later, aeolian sands were deposited over the entire area forming a surface relief that is quite different from the relief of the underlying boulder clay surface.

The soil survey was made at a scale of 1:5000, with an average observation density of two borings per hectare. Borings extended to a maximum depth of 2 m, to the upper surface of the boulder clay, or the Miocene clay. Existing mapping units and classifications of the Dutch soil survey (Table 1) mainly reflect properties of the horizons near the

Table 1. Mapping units of the detailed soil map (Fig. 1) and their classification according to *Soil Taxonomy* (Soil Survey Staff, 1975).

Mapping unit	Thickness A1 or Ap (cm)	% < 50 μm in surface soil	Classification (*Soil Taxonomy*)
Hn51	< 30	0–10	Sandy, siliceous, mesic Typic Haplaquods
Hn52	< 30	10–17.5	Sandy, siliceous, mesic Typic Haplaquod
Hn55	< 30	17.5–32.5	Sandy, siliceous, mesic Typic Haplaquods
cHn54	30–50	10–32.5	Sandy over loamy, siliceous, mesic Plaggeptic Haplaquods
tZn52	15–30	10–17.5	Sandy, siliceous, mesic Typic Haplaquepts
tZn55	15–30	17.5–32.5	Sandy, siliceous, mesic Typic Haplaquepts
cZn54	30–50	10–32.5	Sandy, siliceous, mesic Plaggeptic Umbraquepts
Zn52	< 15	10–17.5	Sandy, siliceous, mesic Typic Udipsamments
tKx	15–25	–	Loamy or clayey, siliceous, mesic Typic Haplaquepts

Table 2. Texture analysis of nine major horizons which occur in the study area.

Horizon	Organic matter, g kg⁻¹	Soil texture (% fraction < 2000 μm)									% total dry soil > 2000 μm	Bulk density (kg m⁻³)
		<2	2–16	16–50	50–75	75–105	105–150	150–210	210–2000	M50 (μm)†		
A1	46 ± 17*	2.7 ± 0.6	2.7 ± 0.9	6.0 ± 2.5	3.8 ± 1.6	10.4 ± 1.7	22.9 ± 1.6	24.0 ± 3.4	37.4 ± 2.5	167 ± 7	–	1420 ± 20
With gravel	45 ± 18	6.2 ± 1.1	4.5 ± 1.0	7.8 ± 1.6	3.4 ± 1.5	7.3 ± 1.5	10.8 ± 1.7	13.9 ± 2.9	46.1 ± 8.4	417 ± 73	50	1450 ± 200
Ap + Aan	66 ± 28	3.8 ± 1.4	3.8 ± 2.3	7.1 ± 2.8	4.1 ± 1.6	9.6 ± 1.2	22.1 ± 2.0	22.6 ± 4.0	27.1 ± 3.3	168 ± 8	–	1260 ± 80
B2	29 ± 18	2.6 ± 0.5	1.9 ± 1.1	3.5 ± 3.2	3.3 ± 2.1	10.8 ± 2.2	23.7 ± 2.9	25.7 ± 3.7	28.4 ± 3.4	168 ± 10	–	1480 ± 70
C11	5 ± 3	2.2 ± 0.5	0.9 ± 0.3	3.1 ± 1.6	4.8 ± 1.2	13.4 ± 2.3	28.0 ± 3.0	25.1 ± 2.3	22.4 ± 4.3	153 ± 11	–	1640 ± 60
C12	1 ± 1	2.2 ± 0.5	0.8 ± 0.5	3.3 ± 1.1	6.5 ± 2.0	16.1 ± 2.4	24.5 ± 3.3	23.4 ± 2.7	23.4 ± 5.2	152 ± 14	–	1690 ± 20
D1	12 ± 14	19.9 ± 13.7	7.6 ± 2.8	8.2 ± 8.1	2.0 ± 0.7	5.9 ± 5.8	11.8 ± 12.2	7.4 ± 1.6	37.1 ± 31.6	584 ± 407	10	1490 ± 360
						Boulder clay						
D2	13 ± 1	37.4 ± 2.4	20.3 ± 1.8	29.2 ± 1.8	1.6 ± 0.2	0.7 ± 0.2	1.3 ± 0.3	1.8 ± 0.3	7.7 ± 1.0	220 ± 13	–	1410 ± 70
						Miocene clay						
Gravel	5 ± 2	6.6 ± 1.0	6.2 ± 0.9	3.0 ± 0.6	1.3 ± 0.5	2.8 ± 1.0	4.9 ± 0.9	6.3 ± 2.6	68.9 ± 10.4	906 ± 184	45	1720 ± 300

* Standard deviation. † The term M50 refers to median sand-grain size.

soil surface. However, when the soil survey is to be used for hydrologic interpretations, it should also include data on those subsurface soil horizons expected to be important for soil physical properties. Observations were made, therefore, of rooting depths of the dominant grass vegetation, and of textures and structures of *all* horizons, including subsoil horizons. Rooting depths were needed for the simulation model. In addition, the occurrence was recorded of gravel in surface and subsurface soil horizons, as well as of boulder clay or Miocene clay that started at depths between 40 and 120 cm below surface. The mapping unit tKx contains boulder clay starting at a level above 40 cm (Table 1). This depth classification was adopted from existing mapping criteria in the Dutch soil survey. However, a more detailed depth classification to be discussed later was devised for the simulation map.

The following procedure deviated from a standard Dutch soil survey only by its deeper borings. The mapping units occurring in the study area represent major units of sandy soils in The Netherlands. Mappings units as used in detailed Dutch soil surveys, are shown in Fig. 1 and Table 1, which also includes classifications according to Soil Survey Staff (1975).

Soil Physical Measurements

Field and laboratory measurement techniques for water retention (h-Θ) and hydraulic conductivity (K-h) were selected emphasizing use of relatively simple and rapid techniques (e.g., Bouma, 1983). Hydraulic conductivities of soil above the water-table were measured by:

1. The column method for $K_{sat\text{-}vert}$ (e.g., Bouma, 1982).
2. The crust-test for K_{unsat} down to approximately $h = -50$ cm (latest version of the method reported by Bouma et al., 1983).
3. The sorptivity method for lower K_{unsat}-values in the case of sand (Dirksen, 1979).
4. The hot air method for lower K_{unsat}-values in the case of loam and clay (Arya et al., 1975).

Moisture retention curves were obtained by slow evaporation of wet, undisturbed samples in the laboratory, as reported by Bouma et al. (1983). In these samples pressure heads were periodically measured with transducer-tensiometers, at which time subsamples were also taken to determine moisture contents. Thus, points relating h and Θ were obtained. Moisture contents corresponding with pressure heads < -800 cm, were obtained by conventional methods using air pressure (Richards, 1965). Six replicate measurements were made of the nine major soil horizons and results are expressed by regression analysis in terms of average values and their variability as described elsewhere (e.g., Baker and Bouma, 1976; Baker, 1978). Statistical testing of differences among soil horizons was based on Wilcoxon's test, using data points within small successive pressure-head increments.

Simulation Map

The term "simulation map" is used here to describe a map which presents a soil physical interpretation of the soil map, providing data for simulation of the soil water regime.

Basic soil data, necessary for simulation of soil water regimes as used in this study by the SWATRE model, include: moisture retention data (h-Θ) of surface and subsurface soil horizons; hydraulic conductivity data (K-h) for subsurface soil horizons; rooting depth and a function which defines water uptake from the root zone (the so called "sink term"). In addition, environmental data are needed in terms of precipitation, potential evapotranspiration (for grass), as well as water-table levels (e.g., De Laat, 1980; Bouma et al., 1980). Criteria on which the legend of the soil map are based, are often only indirectly related to basic soil physical data needed for simulation. Much work has been done elsewhere relating simple soil characteristics, such as texture, bulk density, and organic matter content, to h-Θ and K-h relations. This can be done by various calculation schemes (e.g., Brooks and Corey, 1964; Bloemen, 1980a, b). However, these procedures include approximations and may yield inaccurate results. More important, only point data are obtained and extrapolation to areas of land poses a problem. A different approach was therefore followed in this study. Physical measurements of K-h and h-Θ were made in major pedological soil horizons as defined in Table 2 in terms of texture, structure, organic matter content, and bulk density. These soil horizons were derived from the soil map. The extrapolation problem is thus reduced because major horizons occur in limited numbers in well defined patterns in the landscape, as indicated by the soil map. The total number of major soil horizons was such that all soils in the area could be represented by a sequence of these major horizons. Six measurements were made in each horizon. Sampling locations were chosen at random within the various delineated areas of the soil map. Next, those major horizons were distinguished whose physical properties differed significantly. Their number is usually lower than the number of horizons being distinguished by pedological classification, because pedological differences do not necessarily correspond with differences in physical properties. For example, subsurface soil horizons that are identical from a physical point of view may occur in different mapping units.

Finally, delineated areas of the soil map were transformed into areas of the simulation map, which were each defined in terms of a sequence of major horizons, that were significantly different from a soil physical point of view (Table 3). Rooting depths were also defined for each area, assuming the presence of a grass crop. Rooting depths, which were

Table 3. Characteristics for 18 units being distinguished on the simulation map.†

Unit of simula-map	Thickness of root-zone (cm)	Presence of horizon	Starting depth (cm below surface)		
			Gravel	Boulder clay	Miocene clay
4	15	C11		40–120	
5	26	B2 C11 C12	>120		
6	25	(B2) C11 C12		120–200	
7	25	(B2) C11 C12			
8	25	(B2) C11 C12			120–200
10	25	B2 C11	<40	40–120	
12	25	(B2) C11	40–80		
13	25	(B2) C11	40–80	80–120	
14	25	B2 C11	40–120	120–200	
15	25	B2 C11	80–120		
18	25	(B2) C11		40–120	
23	25	C11	<40		
29	25	C11		<40	
32	25	C11			40–120
33	25		<40	40–120	
35	35	B2 C11 C12			
42	35	C11	80–120		
45	35	C11		40–120	

† The (B2) notation indicates that a B2 horizon may or may not be present. The original numbers for the units, as used for the survey area of 650 ha, have been maintained. Units 35, 42 and 45 have moisture retention curves for the Aan + Ap, all other units have the curve for A1 + A1 with gravel. *Physical properties of the various horizons are shown in Fig. 2 and 3.*

observed in the field, were relatively shallow due to high levels of chemical fertilization. Only minor differences occur among the various soils.

The procedure followed here is the same, in principle, as that used by USDA-SCS (1971), which assigns estimated permeabilities (K_{sat}), and measured moisture retention data to major soil horizons. The difference, however, consists of the use of *measured K* values, including unsaturated conductivity, and the grouping of statistically identical horizons, which reduces the total number of different horizons to be distinguished for simulation.

Natural fluctuations of the water-table during the year were also determined following Dutch soil survey procedures defining "ground water class" (*Gt*-values). These are essential for simulation purposes. In this paper, however, attention is focused on soil physical characteristics of the unsaturated zone.

Validation Procedures

Validation includes two tests: (i) an independent test of the occurrence of major horizons in units of the simulation map, and (ii) comparison of simulated data (obtained by using the simulation map) with measured data. The first test deals with the accuracy of the mapping process, the second with use of the simulation map.

Mapping accuracy was tested by 60 test borings in the sample area of 125 ha, using statistical procedures described by Marsman and de Gruyter (1985). They determine the starting point and the direction of a five-point regular grid at random. Locations of the borings are included in Fig. 4. The soil description made at each boring was compared with the description according to the legend of the simulation map (Table 3). This test is also partly a test of the soil map. The second test consisted of running the simulation model SWATRE (Feddes et al., 1978; Belmans et al., 1983), using basic physical data as indicated by the simulation map. The model rendered actual evapotranspiration values, which were compared with measured ones as reported by Stricker (1981). Calculations were made for a randomly selected pedon near the Assink Experimental Station (soil unit Hn 52/simulation unit no. 7) where measured evapotranspiration values were available. This calculation is used as an example. Measured evapotranspiration values were not available for other locations. However, measured and calculated moisture contents of the soil—which are less attractive for validation purposes—are also being compared at several randomly selected sites.

RESULTS AND DISCUSSION

Soil Map

The soil map (Fig. 1) contains 110 delineated areas. Nine major soil units are distinguished (Table 1). The topography of the boulder clay surface in the subsurface soil is independent from patterns formed by various soils in the sandy surface layers. Hence, a complex combined pattern results. Gravelly A horizons are only found in part of the area, adding another element of variation. The total number of mapping units, including occurrence of gravel and boulder clay, is 21.

Physical Data

Nine major soil horizons were distinguished and defined in terms of organic matter content, texture, and bulk density (Table 2). Results of multiple physical measurements of moisture retention and hydraulic conductivity are presented in Fig. 2 and 3.

The *K-h* curves of the B2, C11, and C12 horizons were not significantly different (Fig. 3). Curves for the boulder clay (D1) and the Miocene clay (D2) formed one population, but were different from the other three curves. Various types of moisture retention curves were obtained for the boulder clay which contained irregular sandy layers. All measured data fitted between the two lines shown in Fig. 2 (D1 + D2). They form one (variable) population with data for the Miocene clay. Two types of retention curves could be distinguished for the A horizons (Fig. 2). A1 horizons with a high gravel content (see Table 2) were not significantly different from the A1 horizons without gravel. The moisture retention curves for the B2, C11 and C12 were not significantly different.

In summary, nine major pedological soil horizons were distinguished (surface: A1, A1 with gravel, Aan + Ap; subsurface: B2, C11, C12, D1, D2, gravel). The physical measurements allowed distinction of five different moisture retention curves (A1 + A1 with gravel, Aan + Ap, B2 + C11 + C12, D1 + D2, subsurface gravel), for all horizons, and three hydraulic conductivity curves (B2 + C11 + C12, D1 + D2, subsurface gravel) for subsurface soil horizons. Hence, only five populations were found for the nine major soil horizons that were significantly different from a soil physical point of view.

Simulation Map

The simulation map (Fig. 4) was defined in terms of a sequence of major soil horizons, forming 18 units as defined in Table 3. The simulation map contains 41 delineated areas compared with 110 on the detailed soil map. The reduction of the number of areas is due to the following reasons: (i) moisture retention data for surface soils could be divided into only two types (Fig. 2); more distinctions were made for the soil map

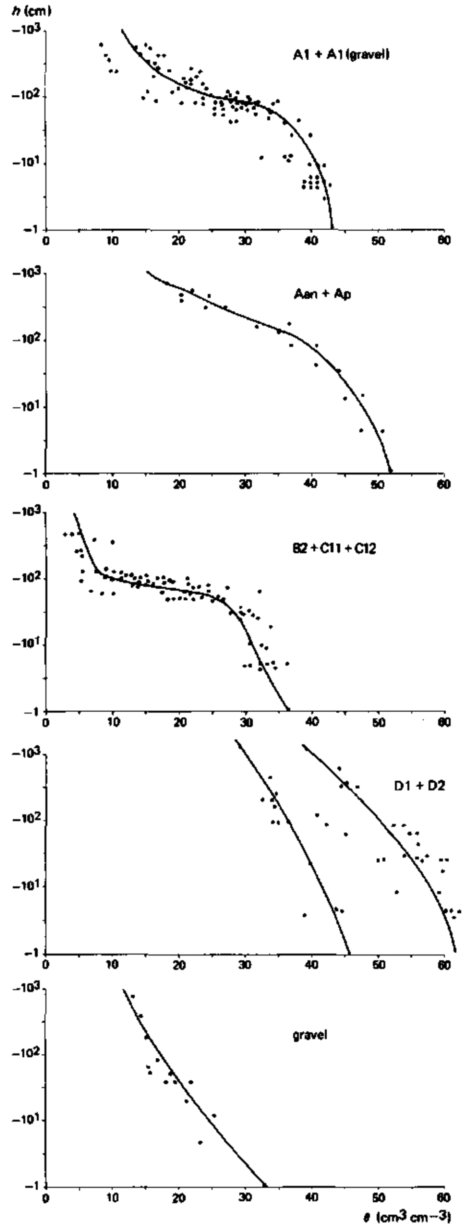

Fig. 2. Moisture retention curves for the major soil horizons. (θ = moisture content: $cm^3\ cm^{-3}$ and h = pressure head (cm in water) Combined curves are shown that are significantly different. These curves characterize soil horizons in Table 3.

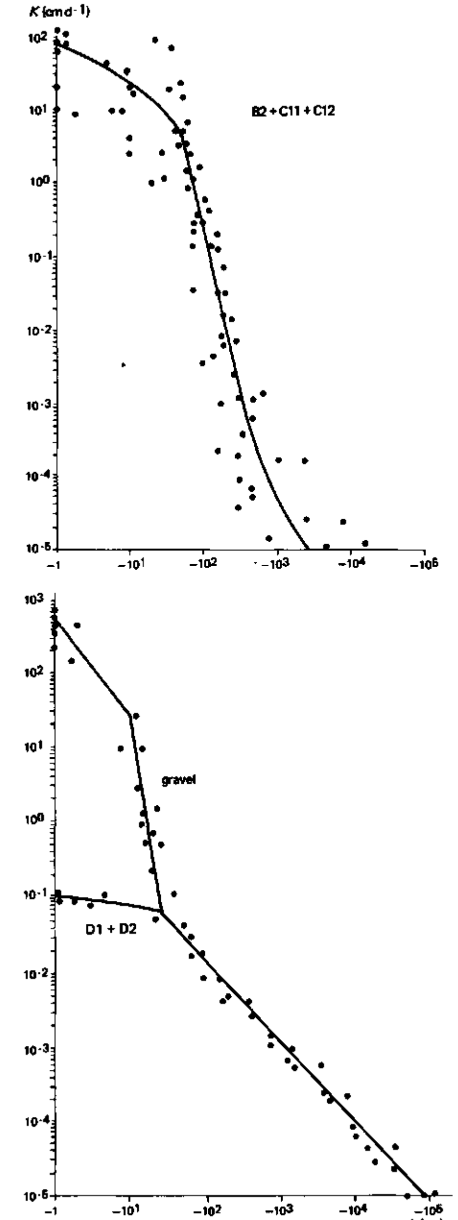

Fig. 3. Hydraulic conductivity for the major soil horizons. (k = hydrolic conductivity: $cm\ d^{-1}$ . . .) Combined curves are shown that are significantly different. These curves characterize soil horizons in Table 3.

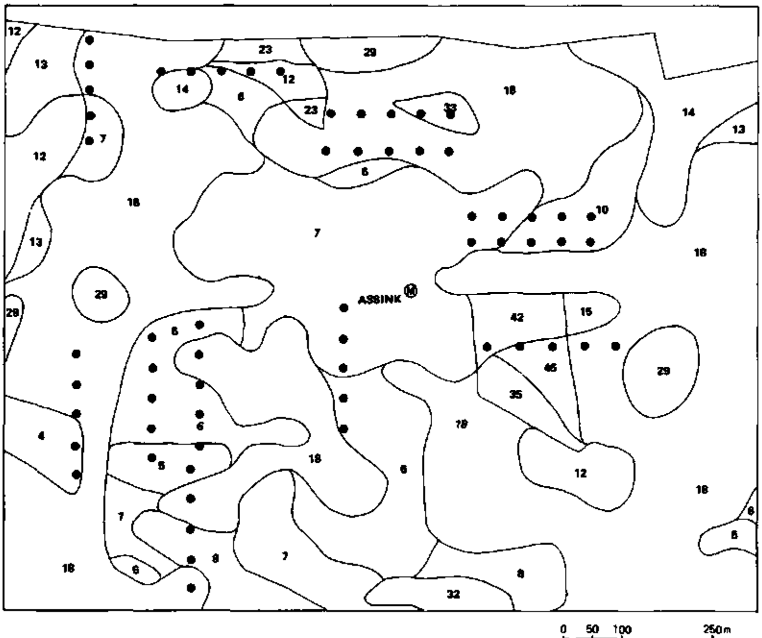

Fig. 4. Simulation map of the area of 125 ha (legend in Table 3). The locations of 60 independant validation borings are indicated.

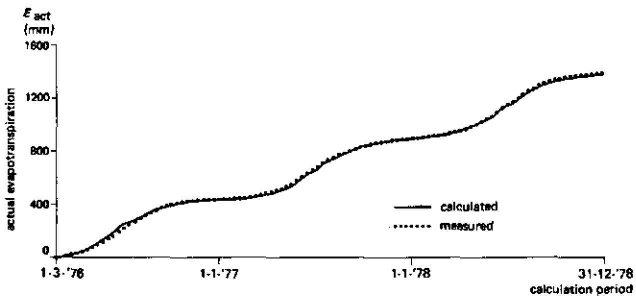

Fig. 5. Calculated and measured actual evapotranspiration (E_{act}) for a grass vegetation on a pedon near the Assink Experimental Station for the years 1976, 1977, and 1978 (after Feddes and de Graaf, 1984).

in terms of surface-soil textures (Table 1), and (ii) hydraulic conductivities of the B2, C11, and C12 horizons were statistically identical. The C horizons are part of eight different soil units being distinguished (Table 1). Only one soil unit will be distinguished on the simulation map if surface and subsurface soil characteristics are physically identical. An overall reduction in the number of delineated areas occurs, even though more classes are distinguished for the depth to gravel and boulder clay, which were defined in fewer depth classes on the soil map (comparing legend of Fig. 1 and Table 3).

The thickness of the rootzone was classified in three

Table 4. Validation of the simulation map, based on a comparison between data from 60 independent borings (locations indicated in Fig. 4) and the corresponding units of the simulation map.

Feature	Purity, %	The 90% confidence interval
Presence of B2 horizon	83 ± 4	76–91%
Presence of C11 horizon	100	–
Presence of C12 horizon	85 ± 5	75–95%
Starting depth of gravel	85 ± 5	75–95%
Starting depth of boulder clay	77 ± 6	65–89%
Complete legend	63 ± 9	
Complete legend except presence of B2 horizon	80 ± 6	68–92%

relatively shallow classes, which in part reflect soil profile characteristics, but mainly reflect the effect of high chemical fertilization that has induced shallow rooting patterns of the grass vegetation. Thickness of the rootzone is therefore not included in the validation analysis.

Validation

The results of 60 independent test borings are summarized in Table 4, in terms of percent agreement between boring data and data according to the legend of the simulation map. Consideration of all aspects together resulted in 63% purity. However, when excluding the presence of a B2 horizon, purity was 80%. This exclusion is justified because its physical properties do not differ significantly from the underlying C horizons (Fig. 2 and 3). Agreement for the highly variable depth to gravel and to boulder clay also avg 80%. This percentage is considered satisfactory.

Results of the simulation (Feddes and De Graaf, 1984) (Fig. 5) indicate excellent agreement between measured and calculated values of actual evapotranspiration of the grass for a period of 3 yr for the one pedon being considered. Calculations should be made for more pedons, but measured evapotranspiration data are very difficult to obtain.

EVALUATION

This study focused on the distinction of major soil horizons, defined pedologically as well as in terms of easily obtainable soil characteristics. These horizons were then characterized in terms of hydraulic conductivity and moisture retention.

This procedure differs in two ways from procedures used before: 1. Major horizons as derived from the soil map are tested for physical behaviour in terms of *K-h* and *h-Θ* relations, which are essential for simulation. Separate distinctions according to purely pedological criteria are avoided. For example, the C11 horizon occurring in Typic Haplaquods is not separated from the one in Aquic Haplumbrepts when their physical properties are not significantly different. Thus, a reduction is obtained of the total number of horizons to be distinguished. This simplifies calculations.

2. Major horizons, thus defined, rather than individual borings, are used as "carriers" of soil physical information. As discussed, earlier procedures are available to calculate *K-h* and *h-Θ* relations for each point-observation. This procedure offers scientific problems and leaves the extrapolation problem unsolved.

Distinction of major horizons is a form of sample stratification and limits the number of measurements to be made. The occurrence of these major horizons is derived from the soil map. Hence, the extrapolation problem is significantly reduced.

ACKNOWLEDGMENTS

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REFERENCES

Arya, L.M., D.A. Farrell, and G.R. Blake. 1975. A field study of soil water depletion patterns in presence of growing soybean roots. I. Determination of hydraulic properties of the soil. Soil Sci. Soc. Am. Proc. 39:424–430.

Baker, F.G. 1978. Variability of hydraulic conductivity within and between nine Wisconsin soil series. Water Resour. Res. 14(1):103–108.

Baker, F.G., and J. Bouma. 1976. Variability of hydraulic conductivity in two subsurface horizons of two silt loam soils. Soil Sci. Soc. Am. J. 40:219–222.

Belmans, C., J.C. Wesseling, and R.A. Feddes. 1983. Simulation model of the water balance of a cropped soil: SWATRE. J. Hydrol. 63:271–286.

Bloemen, G.W. 1980a. Calculation of hydraulic conductivities of soils from texture and organic matter content. Zeitschrift für Pflanzenernährung und Bodenkunde, 143(5):581–615.

Bloemen, G.W. 1980b. Calculation of steady state capillary rise from the groundwater table in multi-layered soil profiles. Zeitschrift für Pflanzenernährung und Bodenkunde, 143(6):701–719.

Bouma, J. 1982. Measuring the hydraulic conductivity of soil horizons with continuous macropores. Soil Sci. Soc. Am. J. 46:438–441.

Bouma, J. 1983. Use of soil survey data to select measurement techniques for hydraulic conductivity. Agric. Water Manage. 6(2/3):177–190.

Bouma, J., and P.J.M. de Laat. 1981. Estimation of the moisture supply capacity of some swelling clay soils in The Netherlands. J. Hydrol. 49:247–259.

Bouma, J., C. Belmans, L.W. Dekker, and W.J.M. Jeurissen. 1983. Assessing the suitability of soils with macropores for subsurface liquid waste disposal. J. Environ. Qual. 12:305–310.

Bouma, J., P.J.M. de Laat, R.C.H.M. Awater, A.F. van Holst, H.C. van Heesen, and Th.J.M. van de Nes. 1980. Use of soil survey data in simulation model for predicting regional soil moisture regimes. Soil Sci. Soc. Am. J. 44:808–814.

Brooks, R.H., and T. Corey, 1964. Hydraulic properties of porous media. Colorado State Univ., Hydrol. Pap. 3.

Busoni, E., P.J. Kowalik, and G. Sanesi. 1983. Effect of drainage conditions on winter wheat production. Agric. Water Manage. 7:425–438.

De Laat, P.J.M. 1980. Model for unsaturated flow above a shallow water table, applied to a regional sub-surface flow problem. Agric. Res. Rep. 895. Pudoc, Wageningen, The Netherlands.

Dirksen, C. 1979. Flux-controlled sorptivity measurements to determine soil hydraulic property functions. Soil Sci. Soc. Am. J. 43:827–834.

Feddes, R.A., and M. de Graaf. 1984. Simulation results of SWATRE for the Hupselse Beek area p. 105–118. Committee for Hydrological Research TNO Rep. no. 13. The Hague, The Netherlands.

Feddes, R.A., P.J. Kowalik, and H. Zaradny. 1978. Simulation of field water use and crop yield. Simulation monogr. Pudoc, Wageningen, The Netherlands.

Marsman, B.A., and J.J. de Gruijter. 1985. Quality of soil maps. A comparison of survey methods in a sandy area. Soil Survey Pap. 15. Soil Survey Institute Wageningen, The Netherlands (in press).

Richards, L.A. 1965. Physical condition of water in soil. *In* C.A. Black et al. (ed.) Methods of soil analysis, Part I. Agronomy 9:131–137.

Soil Survey Staff. 1975. Soil taxonomy. A basic system of soil classification for making and interpreting soil surveys. USDA-SCS Agric. Handb. 436. U.S. Government Printing Office, Washington, DC.

Stricker, J.N.M. 1981. Methods of estimating evapotranspiration from meteorological data and their applicability in hydrology. p. 59–76. Committee for Hydrological Research TNO Proc. no. 28. The Hague, The Netherlands.

USDA-SCS. 1971. Guide for interpreting engineering uses of soils, no. 0107-0332. U.S. Government Printing Office, Washington, DC.

Vachaud, G. 1982. Soil physics research and water management. Whither soil research. p. 32–59. *In* Proc. Int. Soc. Soil Sci.-Congress. New Dehli, India.

4. USING TEXTURE AND OTHER SOIL PROPERTIES TO PREDICT THE UNSATURATED SOIL HYDRAULIC FUNCTIONS

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Using Texture and Other Soil Properties to Predict the Unsaturated Soil Hydraulic Functions

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ABSTRACT

The unsaturated hydraulic properties of soils are important but difficult to measure. Rather than measuring the hydraulic properties directly, we followed a different approach by fitting analytical expressions for the soil water retention and hydraulic conductivity functions to experimental data for a wide range of soils in the Netherlands. Analysis of the data shows the flexibility of the analytical expressions and also gives insight into how the different model parameters affect the calculated unsaturated hydraulic functions. Regression analyses are used to relate the estimated model parameters to more easily measured soil properties, such as bulk density and percentages silt, clay and organic matter. After calibration, the relations are used to predict the hydraulic functions of soils for which only the more easily measured soil properties are available. Accuracy of the predictions is analyzed in terms of functional criteria which are relevant to practical management problems. The predictive regression models are useful for estimating the unsaturated soil hydraulic properties of large areas of land, but need improvement for application to specific sites.

Soil hydraulic behavior is characterized by the soil water retention curve which defines the volumetric water content (θ) as a function of the soil water pressure head (h), and the hydraulic conductivity curve which relates the hydraulic conductivity (K) to water content or pressure head. These curves are crucial ingredients for predicting water flow and solute transport in the vadose zone, and for evaluating alternative soil-water-crop management practices.

Even though new field or laboratory techniques for measuring the unsaturated soil hydraulic functions have been developed (e.g. Bouma, 1983; Klute, 1986), the methods have remained relatively cumbersome and costly, especially for the hydraulic conductivity. The measurement problem is complicated by results of studies (e.g., Nielsen et al., 1973; Russo and Bresler, 1981) which show that soils exhibit significant temporal and spatial variabilities in their hydraulic properties. This variability implies that numerous samples may be needed to properly characterize a given field. To partially circumvent the measurement problem, several investigators have proposed theoretical pore-size distribution models that predict the hydraulic conductivity from more easily measured soil water retention data (Millington and Quirk, 1959; Mualem, 1976). Others have approximated the hydraulic properties from soil textural, bulk density or other soil data (Bloemen, 1980; Rawls et al., 1982; Haverkamp and Parlange, 1986).

One of the proposed unsaturated hydraulic models is described by van Genuchten (1980) who combined an empirical S-shaped curve for the soil water retention function with the pore-size distribution theory of Mualem (1976) to derive a closed-form analytical expression for the unsaturated hydraulic conductivity. Except for the saturated hydraulic conductivity, K_s, the resulting conductivity function contains parameters that can be estimated from measured soil water retention data. An advantage of this model, like previous ones by Brooks and Corey (1964) and others, is that the hydraulic properties are expressed in the form of analytical (nontabular) functions, a feature that facilitates their efficient inclusion into numerical simulation models and also enables the rapid comparison of the hydraulic properties of different soils. Van Genuchten's unsaturated hydraulic functions have been shown in several recent studies (e.g., Stephens and Rehfeldt, 1985; van Genuchten and Nielsen, 1985; Hopmans and Dane, 1986) to give good descriptions of observed retention and/or conductivity data for a large number of soils. Hopmans and Overmars (1986) recently also demonstrated the applicability of those functions in a hydrological research project.

The above studies rely on available soil water retention data for predicting the unsaturated hydraulic conductivity. Sometimes, as is the case in this study, directly measured unsaturated hydraulic conductivity data may also be available. To make use of all available data, we will apply in this paper nonlinear least-squares parameter estimation techniques to fit van Genuchten's functions simultaneously to observed water retention and hydraulic conductivity data from a large number of soils in the Netherlands. Once estimated, the optimized model parameters will be used in a multiple regression analysis to investigate relationships between the fitted parameters and such more easily measured basic soil properties as bulk density, soil texture, and organic matter content. The regression model, in turn, is used to predict the hydraulic functions of three soils using only their basic soil properties. Rather than focusing on a statistical analysis of the fitted hydraulic functions, we will compare the predicted and the measured hydraulic functions in terms of functional criteria (Wösten et al., 1986) which relate directly to practical applications (e.g., travel time from the soil surface to the groundwater table).

Specific purposes of this study are hence (i) to fit the analytical functions of van Genuchten (1980) to a large set of measured soil water retention and hy-

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draulic conductivity data, (ii) to explore the possibility of predicting the hydraulic functions from basic soil properties, and (iii) to compare measured and predicted hydraulic functions on the basis of functional criteria which are relevant to soil-water management practices.

MATERIALS AND METHODS

Soil water retention and hydraulic conductivity curves were measured for a large number of soils in the Netherlands. The soils were classified according to soil texture (as used by the Netherlands Soil Survey Inst.), and type of horizon, being either topsoil (A horizon) or subsoil (B and C horizons). This classification resulted in 20 different soil groups (Table 1) comprising a total of 197 individual curves. Tabulated forms of the geometrically averaged curves for the 20 soil groups are given elsewhere (Wösten et al., 1987a). Because of wide ranges in soil texture, the individual curves within each group are quite variable. As a set, the curves form a unique data base covering a broad spectrum of soils in the Netherlands.

Hydraulic conductivities were estimated using a combination of the following five methods:

1. The column method (e.g., Bouma, 1982) for the vertical saturated hydraulic conductivity, K_s.
2. The crust-test (an updated version of the method described by Bouma et al., 1983) for unsaturated conductivities when the pressure head, h, is between 0 and -50 cm.
3. The sorptivity method (Dirksen, 1979) for conductivities of coarse-textured soils when $h \leq -50$ cm.
4. The hot-air method (Arya et al., 1975) for conductivities of medium- and fine-textured soils when $h \leq -50$ cm.
5. The evaporation method (Boels et al., 1978) for hydraulic conductivities when h is between 0 and -800 cm.

Soil water retention curves were obtained by slow evaporation of wet, undisturbed samples in the laboratory as reported by Boels et al. (1978) and Bouma et al. (1983). In this method, pressure heads are periodically measured with transducer-tensiometers while at the same time subsamples are taken to determine water contents, thus yielding points relating to h and θ. Water contents corresponding with pressure heads lower than -800 cm were obtained by conventional methods using air pressure (Richards, 1965). For relatively fine-textured soils, a staining technique was applied to record the effects of horizontal cracks on the upward flux of water from the water table to the rootzone (Bouma, 1984).

DATA ANALYSIS

The volumetric soil water content, θ, as a function of pressure head, h, is described with the following empirical equation of van Genuchten (1980)

$$\theta = \theta_r + \frac{\theta_s - \theta_r}{[1 + |\alpha h|^n]^m} \qquad (\theta_r \leq \theta \leq \theta_s) \qquad [1]$$

where the subscripts r and s refer to residual and saturated values, and where α, n and m are parameters which determine the shape of the curve. The residual water content, θ_r, refers to the water content where the gradient $d\theta/dh$ becomes zero ($h \to -\infty$). In practice, θ_r is the water content at some large negative value of the soil water pressure head. The dimensionless parameter n determines the rate at which the S-shaped retention curve turns toward the ordinate for large negative values of h, thus reflecting the steepness of the curve, while α (cm^{-1}) equals approximately the inverse of the pres-

Table 1. Number of individual curves, ranges in measured soil properties, and average values for the optimized parameters α, n and l for 20 soil groups.

Soil	Number of curves	Silt 2–50 μm	Clay < 2 μm	Organic matter	Bulk density	M50†	α	n	l
		%	%	%	g/cm^3	μm	cm^{-1}		
				Coarse-textured soils (105 samples)					
Topsoil									
B1	5	4–7	–‡	1–4	1.4–1.6	140–170	0.0169	1.83	0.826
B2	11	11–17	–	3–10	1.3–1.5	130–170	0.0203	1.52	−0.432
B3	9	19–29	–	4–13	1.1–1.5	130–165	0.0290	1.44	−0.152
B4	5	37–49	–	2–5	1.1–1.5	130–160	0.0157	1.58	0.225
Subsoil									
01	49	1–9	–	0.1–2	1.4–1.8	150–180	0.0551	2.43	0.310
02	7	10–16	–	0.3–2	1.5–1.7	150–175	0.0201	2.05	0.666
03	14	21–32	–	0.2–2	1.4–1.8	130–170	0.0278	1.75	0.229
04	5	37–47	–	0.3–1	1.4–1.7	130–165	0.0154	1.82	0.153
				Medium-textured soils (43 samples)					
Topsoil									
B7	5	–‡	10–12	2–6	1.2–1.7		0.0347	1.27	−0.543
B8	10	–	12–16	2–3	1.3–1.6		0.0624	1.25	−3.30
B9	3	–	18–22	2–8	1.3–1.6		0.0536	1.15	−4.59
Subsoil									
08	7	–	9–11	0.4–1	1.5–1.6		0.0316	1.32	−1.14
09	12	–	12–16	0.3–1	1.4–1.7		0.0424	1.28	−2.43
010	6	–	18–22	0.1–1	1.3–1.5		0.0400	1.21	−3.26
				Fine-textured soils (49 samples)					
Topsoil									
B10	8	–	26–34	2–5	1.1–1.5		0.133	1.12	−8.36
B11	5	–	35–50	4–15	1.1–1.7		0.138	1.12	−7.66
B12	5	–	51–77	2–5	0.9–1.3		0.141	1.08	−8.30
Subsoil									
011	7	–	28–33	0.9–2	1.4–1.6		0.0510	1.13	−5.50
012	12	–	37–47	0.1–2	1.0–1.5		0.0602	1.14	−6.37
013	12	–	52–77	0.1–2	1.0–1.4		0.0740	1.08	−9.38

† Median sand particle size. ‡ Not included in the regression analysis.

sure head at the inflection point where $d\theta/dh$ has its maximum value.

Assuming $m = 1-1/n$, van Genuchten combined Eq. [1] with the following theoretical pore-size distribution model which was derived by Mualem (1976)

$$K(h) = K_s S^l [\int_0^S \frac{1}{h(x)} dx \,/\, \int_0^1 \frac{1}{h(x)} dx]^2 \qquad [2]$$

where l is an unknown parameter, x is a dummy variable, and S is relative saturation ($0 \leq S \leq 1$):

$$S = (\theta - \theta_r)/(\theta_s - \theta_r)\,. \qquad [3]$$

Combining Eq. [1] and [2] leads to (van Genuchten, 1980)

$$K(S) = K_s S^l\,[1 - (1-S^{1/m})^m]^2 \qquad [4]$$

or in terms of the soil water pressure head

$$K(h) = K_s \frac{\{[1 + |\alpha h|^n]^m - |\alpha h|^{n-1}\}^2}{[1 + |\alpha h|^n]^{m(l+2)}}\,. \qquad [5]$$

Equations [1] and [5] are the analytical functions describing the $\theta(h)$ and $K(h)$ relationships, respectively. Although l is presumably a soil specific parameter, Mualem (1976) concluded from an analysis of 45 soil hydraulic data sets that l should be on the average about 0.5. In this study l is not fixed but considered to be one of the experimental unknowns.

Figure 1 shows how the parameters n, α, θ_r and l affect the shapes of the calculated soil water retention (Eq. [1]) and unsaturated hydraulic conductivity (Eq. [5]) functions. Unless noted otherwise, all curves were obtained with $\theta_r = 0.10$, $\theta_s = 0.50$, $\alpha = 0.005$ (cm^{-1}), $n = 2.5$ and $l = 0.5$. Figure 1c shows that the parameter θ_r affects only the shape of the retention curve while leaving the relative conductivity ($K_r = K/K_s$) function unaffected. The parameter l, on the other hand, only affects the hydraulic conductivity and leaves the retention curve unchanged (Fig. 1d). These properties follow immediately from Eq. [1] and [5]. Figure 1 clearly demonstrates the flexibility of the analytical functions in generating different shapes of the soil water retention and hydraulic conductivity curves. We note that, except for the parameter l, similar plots of the hydraulic functions, with different scales and from different perspectives, have been presented and discussed also by Stephens and Rehfeldt (1985) and Hopmans and Overmars (1986). Van Genuchten and Nielsen (1985) in addition discussed the computationally more complicated case when the parameters m and n in Eq. [1] are independent.

A nonlinear least-squares optimization program (SOHYP) was previously constructed (van Genuchten, 1978) to estimate the parameters θ_r, α and n from observed soil water retention data. The SOHYP code was modified to yield the RETC code (van Genuchten, 1986; unpublished) which allows some or all unknown coefficients (θ_r, θ_s, α, n, l and/or K_s) in Eq. [1] and [5] to be estimated simultaneously from measured soil water retention and hydraulic conductivity data. As before, the least squares parameter estimation process is based on Marquardt's maximum neighborhood method (Marquardt, 1963). The objective function, $O(b)$, minimized in RETC is of the general form

$$O(b) = \sum_{i=1}^{M} \{w_i[\theta_i^* - \theta_i(b)]\}^2 + \sum_{i=M+1}^{N} \{W_1 W_2 w_i[\ln(K_i^*) - \ln(K_i(b))]\}^2 \qquad [6]$$

where θ_i^* and K_i^* are the measured water contents and hydraulic conductivities, respectively, $\theta_i(b)$ and $K_i(b)$ are the predicted responses for a given parameter vector $\{b\}$ of unknown coefficients, w_i are weighting coefficients for the individual observations (set to unity in this study), M is the number of observed retention data and N is the total number of observed retention and conductivity data. The parameters W_1 and W_2 are also weighting coefficients. Parameter W_2 is calculated internally in the program as follows

$$W_2 = \frac{1}{M}\sum_{i=1}^{M} w_i\theta_i^* \,/\, \frac{1}{N-M}\sum_{i=M+1}^{N} w_i|\ln(K_i^*)|\,, \qquad [7]$$

which gives the water content data approximately the same weight as the $\ln(K)$ data. Parameter W_1 is an independent input parameter which adds extra flexibility in assigning more or less weight to the water content data as compared to the water conductivity data. Because hydraulic conductivities often show more scatter than retention data, it is sometimes beneficial to assign a relatively small value to W_1. In our study a value of 1.0 was assigned to W_1, except for the sandy soils which required a much smaller value of 0.1 to ensure visibly better fits of the typically steep part of the retention curve at the inflection point.

In the most general case, the parameter vector $\{b\}$ will contain all coefficients, i.e.: $\{b\} = \{\theta_r, \theta_s, \alpha, n, l, K_s\}$. The parameters θ_s and K_s in this study were fixed at their independently measured values, thus reducing the number of unknowns to four. The parameter θ_r was initially also considered to be an unknown. Since only a few soil water retention and hydraulic conductivity data points were available in the dry range, we restricted our analysis to pressure heads in the range from 0 to -5000 cm. Because of this limited range in pressure heads, θ_r was estimated to be zero, or close to zero, for a majority of the data sets. Following the example of Greminger et al. (1985), we consequently fixed θ_r at zero for all soils, and then reanalyzed the data with only three unknown parameters: $\{b\} = \{\alpha, n, l\}$.

After the parameter estimation step, multiple regression techniques were used to investigate the dependency of the estimated model parameters on more easily measured basic soil properties. Model parameters for the coarse-textured soils were found to be correlated to the quantitative variables percent silt, percent organic matter, bulk density and median sand particle size (M50), and also to the qualitative variable "topsoil or subsoil." For medium- and fine-textured soils, the regressed variables were percent clay, percent organic matter, bulk density and again topsoil or subsoil. Linear, reciprocal and exponential relationships of these basic soil properties were used in the regression analysis, while also possible interactions among the soil properties themselves were investigated. Thus, the resulting regression model consisted of various basic soil properties, and their interactions, which contributed significantly to the optimized soil hydraulic parameters. Once established, the regression models were used to predict the unknown hydraulic parameters of three typical fine-, coarse- and medium-textured soils using their basic soil properties only.

Finally, predicted and measured hydraulic functions of the three soils will be compared on the basis of functional criteria which have immediate relevance to management-type applications Wösten et al. (1986) introduced for this purpose three functional criteria:

1. Travel time, T_r, of water from the soil surface down to a water table at depth L, assumed to be 1 m. The travel time is approximated by $T = \bar{\theta}L/\bar{q}_d$, where $\bar{q}_d$ the average daily vertical *downward* flux for Dutch winter conditions (0.14 cm d^{-1}), and $\bar{\theta}$ is the profile-averaged water content corresponding to $\bar{q}_d$ as derived from the calculated or measured $K(h|\theta)$ curve.
2. Depth of water table, L_c, which can sustain a given upward flux of water to the soil surface or to the bottom of the root zone. As shown by Gardner (1958), this depth may be obtained by integrating Darcy's law

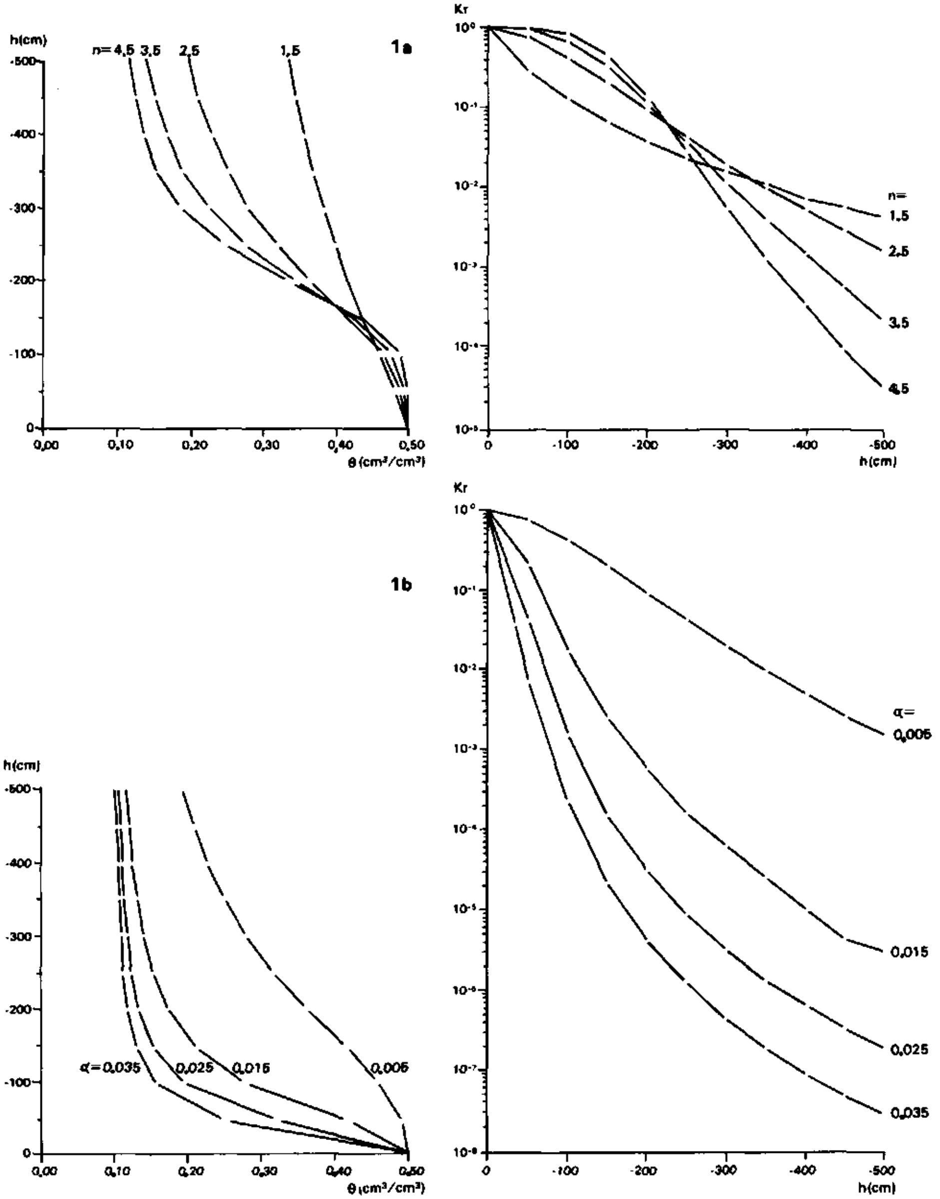

Fig. 1. Effect of the parameters n, α, θ_r and l on the calculated soil water retention, $\theta(h)$, and relative hydraulic conductivity, $K_r(h)$, curves (Fig. 1a, b, c, d, respectively). Continued on next page.

to give

$$L_c = \int_0^{h_c} \frac{1}{1 + \bar{q}_u/K(h)}\, dh \qquad [8]$$

where h_c is the imposed soil water pressure head (taken as -500 cm) at the bottom of the root zone, and $\bar{q}_u$ is the steady-state *upward* water flux density (assumed to be 0.2 cm d^{-1}).

3. Downward flux of water, q_a, corresponding to a minimum soil air content, θ_a (taken here to be 5%) which is required (FAO, 1985) to maintain adequate aeration in the root zone for maximum root activity and crop growth. The q_a follows immediately from the $K(h|\theta)$ function at a water content which is 0.05 $cm^3\,cm^{-3}$ less than the saturated value.

We refer to Wösten et al. (1986) for additional discussions of the above functional criteria. Analysis of those criteria as calculated with the measured and predicted hydraulic functions yields practical information about the accuracy of the predictions.

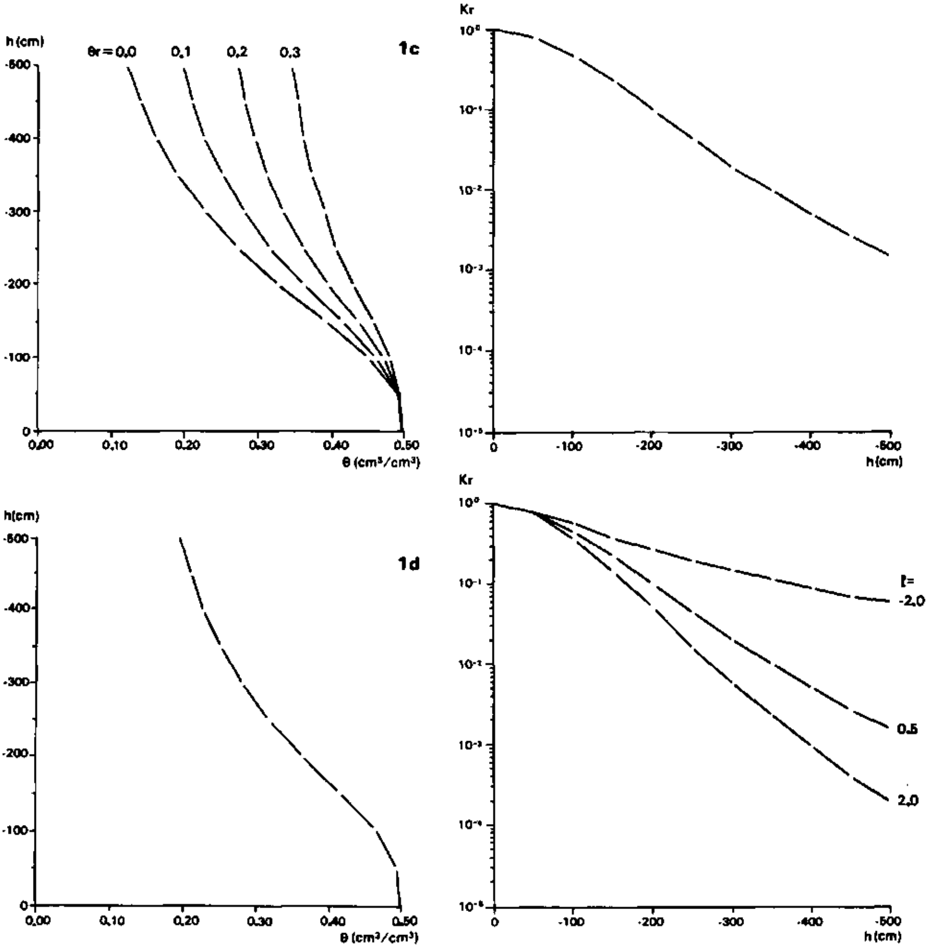

Figure	θs	θr	α	n	ℓ
1a	0.5	0.1	0.005	1.5, 2.5, 3.5, 4.5	0.5
1b	0.5	0.1	0.005, 0.015, 0.025, 0.035	2.5	0.5
1c	0.5	0, 0.1, 0.2, 0.3	0.005	2.5	0.5
1d	0.5	0.1	0.005	2.5	0.5, -2.0, 2.0

Fig. 1. Continued.

RESULTS AND DISCUSSION

Table 1 summarizes results of the simultaneous fit of Eq. [1] and [5] to the measured soil water retention and hydraulic conductivity data. Note that the data sets are divided into three soil textural groups (coarse, medium, fine) and also according to "topsoil versus subsoil." The table gives for each group of soil horizons the average value of the estimated parameters α, n and l. The R^2 (R is the correlation coefficient) between measured and fitted data ranged from 0.94 to 0.99 for all individual curves.

Figure 2 shows good agreement between measured and fitted curves for three typical soils selected from the three soil textural groups. The three soils were classified as a sandy, siliceous, mesic, Typic Haplaquod (01); a fine loamy, mixed, mesic, Typic Fluvaquent (B8); and a fine clayey, mixed, mesic, Typic Fluvaquent (013), respectively (Soil Survey Staff, 1975). The R^2-values between measured and calculated data were 0.97, 0.99 and 0.99, respectively.

Gupta and Larson (1979), Poelman and van Egmond (1979) and Rawls et al. (1982) previously presented several regression equations for predicting water contents at a limited number of pressure head values from readily available basic soil properties. Saxton et al. (1986) just recently derived three functions for the soil water retention curve covering the full range of pressure heads from saturation to $h = -15\,000$ cm. Hence, they avoided the need to interpolate between (or extrapolate) tabular water contents and pressure head data. Saxton et al. (1986) found good agreement between their water retention functions and those calculated with equations of Rawls et al. (1982) A similar prediction of the hydraulic conductivity curve was less successful, however. Our study deviates from the above approaches in that the fitted parameters n, α and l are used for both the soil water retention and the hydraulic conductivity curves.

Multiple regression of each of the model parameters with the basic soil properties yielded six different regression equations. The parameter l for the coarse-textured soils was found to be uncorrelated ($R^2 = 0.07$) with the basic soil properties. Therefore, the average value of 0.22 for l was used for all coarse-textured soils irrespective of texture, bulk density or median sand particle size. The 95% confidence limits of l were -0.16 and 0.60, which indicates that the average value of 0.22 for these coarse-textured soils is in agreement with Mualem's (1976) finding that l could be fixed at 0.5. For the medium- and fine-textured soils l was clearly related to percent silt, bulk density, and topsoil or subsoil, with $R^2 = 0.68$ and l for the individual curves ranging from -16.0 to 2.2. Fixing l at 0.5 in these cases did not result in acceptable fits of the experimental data (see also Fig. 1d).

The regression models for the parameters α and n revealed R^2-values which ranged from 0.42 to 0.76, which is much lower than the values of 0.99 reported by Saxton et al. (1986). This discrepancy is of course partly due to the fact that our model parameters pertain simultaneously to the water retention and hydraulic conductivity curves over a broad range of pressure heads. Still, the relatively low R^2-values indicate a need to improve the regression models. One possible way to achieve this is to use a much larger hydraulic data set involving a broader array of soils than used here. Alternative approaches to relate the hydraulic data to basic soil properties may also be needed. Possible approaches are indicated by Williams et al. (1983) who also included soil structure and clay mineralogy in their analysis, and by Haverkamp and Parlange (1986) who used the cumulative particle-size distribution instead of specific particle size classes such as percent sand, silt or clay.

The regression models established in this study were used to predict the unsaturated soil hydraulic functions from soil texture and other properties in conjunction with the measured values of θ_s and K_s. Figure 2 shows the measured and predicted curves for three soils: 01, B8 and 013. The 90% confidence intervals for the predicted curves were calculated from the standard deviations of the differences between measured and predicted values of θ and $\log K$ at 12 values for h. This was done for all soil hydraulic functions of both the coarse-textured soil group and the medium- and fine-textured soil group. Because these differences were normally distributed, and because of the relatively large number of hydraulic functions in the two soil groups, the confidence limits could be approximated by taking the predicted values plus and minus 1.65 times the respective standard deviations. The calculated confidence intervals are relatively wide, indicating that predictions made with the regression models of this study will show considerable dispersion.

Table 2 compares the values of the three functional criteria (T_r, L_c, and q_a) as calculated from the measured and predicted hydraulic functions. The hydraulic functions hence are compared in terms of specific management problems and the desired accuracy of the answers for these problems. For example, predicting the travel time T_r for the three soils with an accuracy of ± 8 wk (2 months) is often good enough for most management problems on relatively detailed scales (e.g., for scales of 1:10 000 or greater; Breeuwsma et al., 1986). Depth to water table (L_c) and downward flux density (q_a) appear also accurate enough for most management problems on detailed scales in the case of coarse-textured soils. However, their accuracy for medium- and fine-textured soils may permit only answers on a more general scale (e.g., for scales of 1:50 000 or smaller; van der Sluijs and De Gruijter, 1985). This finding is consistent with the results of Fig. 2 which showed better agreement between measured and predicted hydraulic properties for coarse-textured soils as compared to medium- and fine-textured soils. Our contention that evaluation of the regression results should be based on their intended

Table 2. Functional criteria as calculated from the measured and predicted hydraulic properties of three typical coarse-, medium- and fine-textured soils.

	Coarse-textured soil		Medium-textured soil		Fine-textured soil	
	Measured	Predicted	Measured	Predicted	Measured	Predicted
Travel time, T_r (day)	143	121	300	257	407	421
Depth to water table, L_c, (cm)	145	120	138	65	22	12
Downward flux density, q_a (cm d^{-1})	30	23	0.2	0.5	0.002	0.010

use is consistent with the conclusion of de Jong (1982) that water retention parameters of soil textural groups as derived by Clapp and Hornberger (1978) are useful for modeling large areas of soils but fail to accurately approximate soil water characteristics of specific sites. In this respect we note a soil survey mapping study by Wösten et al. (1987b) who showed that the scale of application of a given model should depend on the accuracy by which answers must be known, and on the mapping costs involved.

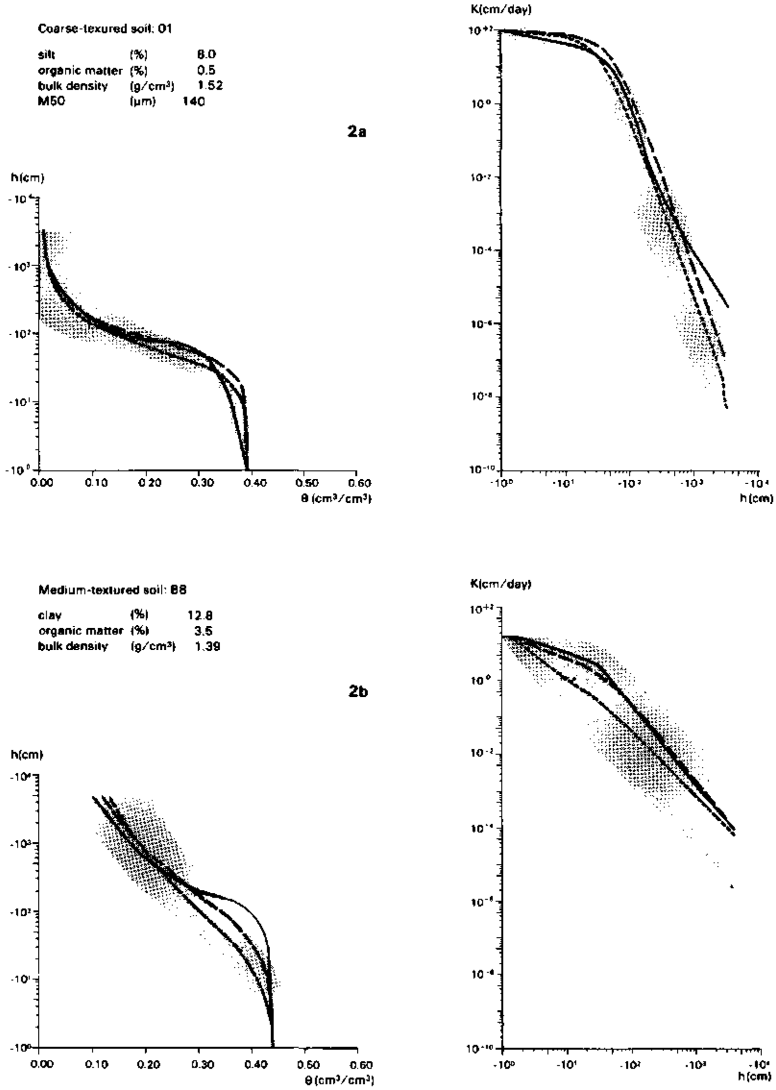

Fig. 2 Measured, fitted and predicted soil water retention, $\theta(h)$, and hydraulic conductivity, $K(h)$, curves for three typical coarse-, medium- and fine-textured soils. Continued on next page.

SUMMARY AND CONCLUSIONS

The analytical hydraulic functions of van Genuchten (1980) were fitted simultaneously to observed soil water retention and hydraulic conductivity data for a wide variety of soils in the Netherlands. The parameter θ_r was fixed at zero without significantly affecting the accuracy of the results. Soil water retention data was weighted more than hydraulic conductivity data (W_1 in Eq. [6] was set at 0.1) to provide acceptable fits of the retention curves for the coarse-textured soils. The hydraulic properties of these soils were not dependent upon the parameter l, yielding a constant value of 0.22 ($\pm$0.38) which agreed well with Mualem's (1976) average value of 0.5. The conductivity and retention data for the medium- and fine-textured soils were weighted equally, leading to l-values which varied between -16.0 to 2.2 as a function of soil texture and other factors. The optimization procedure resulted in reasonably accurate analytical approximations for the hydraulic properties, thus avoiding the cumbersome and computationally inefficient handling of tabulated data in mathematical models of unsaturated flow.

An approach that relates the estimated parameters to easily measured soil properties is described. Once established, the resulting expressions allow prediction of the hydraulic functions of unsaturated soils for which only soil texture and other data (organic matter content, bulk density) have been measured. Predicted hydraulic properties were compared in terms of functional criteria which relate directly to practical management applications. We believe that any judgement about the accuracy of the predicted hydraulic functions should be based on the desired accuracy of those criteria.

Results indicate that the predictive regression models established in this study can be used to derive soil hydraulic functions for large areas of land when working on scales of 1:50 000 or smaller. The models need improvement for site-specific applications at scales of 1:10 000 or larger, however. The use of other data bases, or the inclusion of different soil textural properties such as the cumulative particle-size distribution (Haverkamp and Parlange, 1986) needs to be considered.

ACKNOWLEDGMENT

Soil hydraulic data used in this study included data made available by the Inst. of Land and Water Management Research, Wageningen. Their permission to use those data is appreciated.

APPENDIX–SYMBOLS USED

$\{b\}$ = unknown parameter vector
O = least-squares objective function given by Eq. [6]
h = soil water pressure head (cm of water)
h_c = imposed soil surface boundary condition for the pressure head (cm of water)
K = unsaturated hydraulic conductivity (cm d^{-1})
K_r = relative hydraulic conductivity
K_s = saturated hydraulic conductivity (cm d^{-1})
l = parameter in Eq. [2]
L = water table depth (cm)
L_c = critical water table depth (cm)
m = parameter in Eq. [1]: $m = 1 - 1/n$
M = number of soil water retention data in least-squares fit

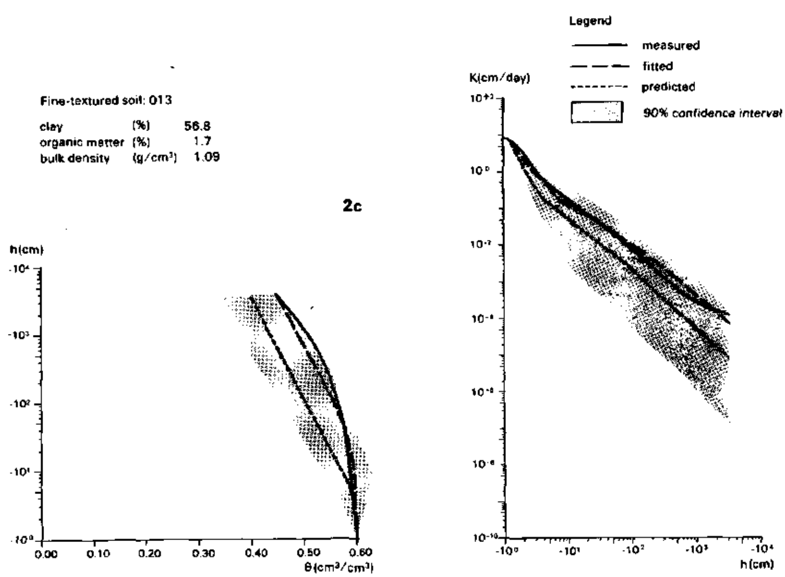

Fig 2. Continued.

M50 = median sand particle size (μm)
n = parameter in Eq. [1]
N = total number of data in least-squares fit
q_a = water flux density at given air content (cm d^{-1})
$\bar{q}_d$ = downward water flux density (cm d^{-1})
$\bar{q}_u$ = upward water flux density (cm d^{-1})
S = reduced water content
T_r = travel time in the unsaturated zone
w_i = weighting coefficients in Eq. [6]
W_1 = Weighting coefficient in optimization process
W_2 = Weighting coefficient defined by Eq. [7]
α = parameter in Eq. [1] (cm^{-1}).
θ = volumetric water content (cm^3 cm^{-3})
$\bar{\theta}$ = profile-averaged water content
θ_a = volumetric air content of soil
θ_r = residual water content (cm^3 cm^{-3})
θ_s = saturated water content (cm^3 cm^{-3}).

REFERENCES

Arya, L.M., D.A. Farrell, and G.R. Blake. 1975. A field study of soil water depletion patterns in presence of growing soybean roots. I. Determination of hydraulic properties of the soil. Soil Sci. Soc. Am. Proc. 39:424–430.

Bloemen, G.W. 1980. Calculation of hydraulic conductivities of soils from texture and organic matter content. Z. Pflanzenernaehr Bodenkd. 143(5):581–615.

Boels, D., J.B.H.M. Van Gils, G.J. Veerman, and K.E. Wit, 1978. Theory and system of automatic determination of soil moisture characteristics and unsaturated hydraulic conductivities. Soil Sci. 126:191–199.

Bouma, J. 1982. Measuring the hydraulic conductivity of soil horizons with continuous macropores. Soil Sci. Soc. Am. J. 46:438–441.

Bouma, J. 1983. Use of soil survey data to select measurement techniques for hydraulic conductivity. Agric. Water Manage. 6(2/3):177–190.

Bouma, J. 1984. Using soil morphology to develop measurement methods and simulation techniques for water movement in heavy clay soils. p. 298–310. *In* J. Bouma and P.A.C. Raats (ed.), Proc. of the ISSS Sympos. on Water and solute movement in heavy clay soils. Wageningen, The Netherlands. 27–31 Aug. 1984. Publ. 37. Int. Inst. for Land Reclamation and Improvement (ILRI), Wageningen.

Bouma, J., C. Belmans, L.W. Dekker, and W.J.M. Jeurissen. 1983. Assessing the suitability of soils with macropores for subsurface liquid waste disposal. J. Environ. Qual., 12(3):305–311.

Breeuwsma, A., J.H.M. Wösten, J.J. Vleeshouwer, A.M. van Slobbe, and J. Bouma. 1986. Derivation of land qualities to assess environmental problems from soil surveys. Soil Sci. Soc. Am. J. 50:186–190.

Brooks, R.H., and C.T. Corey. 1964. Hydraulic properties of porous media. Hydrology Paper 3, Colorado State Univ., Fort Collins, CO.

Clapp, R.B., and G.M. Hornberger. 1978. Empirical equations for some hydraulic properties. Water Resour. Res. 14:601–604.

de Jong, R. 1982. Assessment of empirical parameters that describe soil water characteristics. Can. Agric. Eng. 24:65–70.

Dirksen, C. 1979. Flux-controlled sorptivity measurements to determine soil hydraulic property functions. Soil Sci. Soc. Am. J., 43:827–834.

FAO. 1985. Guidelines: Land evaluation for irrigated agriculture. FAO Soils Bulletin 55, Rome.

Gardner, W.R. 1958. Some steady state solutions of the unsaturated moisture flow equation with application to evaporation from a water table. Soil Sci. 85:228–232.

Greminger, P.J., Y.K. Sud, and D.R. Nielsen. 1985. Spatial variability of field-measured soil water characteristics. Soil Sci. Soc. Am. J. 49:1075–1082.

Gupta, S.C., and W.E. Larson. 1979. Estimating soil water retention characteristics from particle size distribution, organic matter percent and bulk density. Water Resour.. Res. 15:1633–1635.

Haverkamp, R., and J.Y. Parlange. 1986. Predicting the water-retention curve from particle-size distribution: 1. Sandy soils without organic matter. Soil Sci. 142:325–339.

Hopmans, J.W., and J.H. Dane. 1986. Temperature dependence of soil hydraulic properties. Soil Sci. Soc. Am. J. 50:4–9.

Hopmans, J.W., and B. Overmars. 1986. Predicting and application of an analytical model to describe soil hydraulic properties. J. Hydrol. 87:135–143.

Klute, A. (ed.) 1986. Methods of soil analysis. Part 1. 2nd ed. Agronomy 9.

Marquardt, D.W. 1963. An algorithm for least-squares estimation of nonlinear parameters. J. Soc. Ind. Appl. Math. 11:431–441.

Millington, R.J., and J.P. Quirk. 1959. Permeability of porous media. Nature 183:387–389.

Mualem, Y. 1976. A new model for predicting the hydraulic conductivity of unsaturated porous media. Water Resour. Res. 12:513–522.

Nielsen, D.R., J.W. Biggar, and K.T. Erh. 1973. Spatial variability of field measured soil-water properties. Hilgardia 42:215–259.

Poelman, J.N.B., and Th. van Egmond. 1979. Water retention curves for sea- and river-clay soils derived from easily measured soil properties. Report 1492, (in Dutch.) Soil Survey Inst., Wageningen, The Netherlands.

Rawls, W.J., D.L. Brakensiek, and K.E. Saxton. 1982. Estimation of soil water properties. Trans. ASAE 25(5):1316–1320,1328.

Richards, L.A. 1965. Physical condition of water in soil. *In* C.A. Black et al. (ed.) Methods of soil analysis, Agronomy 9:131–137.

Russo, D., and E. Bresler. 1981. Soil hydraulic properties as stochastic processes: I. An analysis of field spatial variability. Soil Sci. Soc. Am. J. 45:682–687.

Saxton, K.E., W.J. Rawls, J.S. Romberger, and R.I. Papendick. 1986. Estimating generalized soil-water characteristics from texture. Soil Sci. Soc. Am. J. 50:1031–1036.

Soil Survey Staff. 1976. Soil taxonomy, A basic system of soil classification for making and interpreting soil survey. USDA-SCS Handb. 436. Washington, DC.

Stephens, D.B., and K.R. Rehfeldt. 1985. Evaluation of closed-form analytical models to calculate conductivity in a fine sand. Soil Sci. Soc. Am. J. 49:12–19.

van der Sluijs, P., and J.J. de Gruijter. 1985. Water table classes: A method used to indicate seasonal fluctuation and duration of water tables on Dutch soil maps. Agric. Water Manage. 10:109–125.

van Genuchten, M.Th. 1978. Calculating the unsaturated hydraulic conductivity with a new closed-form analytical model. Research Rep. 78-WR-08. Water Resources Program, Princeton Univ., Princeton, NJ.

van Genuchten, M.Th. 1980. A closed-form equation for predicting the hydraulic conductivity of unsaturated soils. Soil Sci. Soc. Am. J. 44:892–898.

van Genuchten, M.Th., and D.R. Nielsen. 1985. On describing and predicting the hydraulic properties of unsaturated soils. Ann. Geophys. (Gauthier-Villars) 3(5):615–628.

Williams, J., R.E. Prebble, W.T. Williams, and C.T. Hignett. 1983. The influence of texture, structure and clay mineralogy on the soil moisture characteristic. Aust. J. Soil Res. 21:15–32.

Wösten, J.H.M., M.H. Bannink, J.J. de Gruijter, and J. Bouma. 1986. A procedure to identify different groups of hydraulic-conductivity and moisture-retention curves for soil horizons. J. Hydrol. 86:133–145.

Wösten, J.H.M., M.H. Bannink, and J. Beuving. 1987a. Water retention and hydraulic conductivity characteristics of top- and subsoils in The Netherlands: The staring series. Report 1932, Soil Survey Inst., Wageningen, The Netherlands.

Wösten, J.H.M., M.H. Bannink, and J. Bouma. 1987b. Land evaluation at different scales: You pay for what you get! Soil Surv. Land Eval. 7:13–24.

5. A PROCEDURE TO IDENTIFY DIFFERENT GROUPS OF HYDRAULIC-CONDUCTIVITY AND MOISTURE-RETENTION CURVES FOR SOIL HORIZONS

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A PROCEDURE TO IDENTIFY DIFFERENT GROUPS OF HYDRAULIC-CONDUCTIVITY AND MOISTURE-RETENTION CURVES FOR SOIL HORIZONS

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ABSTRACT

Wösten, J.H.M., Bannink, M.H., De Gruijter, J.J. and Bouma, J., 1986. A procedure to identify different groups of hydraulic-conductivity and moisture-retention curves for soil horizons. J. Hydrol., 86: 133–145.

Saturated and unsaturated hydraulic-conductivity and moisture-retention data were measured in 25 C horizons with a sand texture and in 23 C horizons with a clay loam and silty clay loam texture. Measurements were made on large, undisturbed soil columns. The identification of the two types of C horizons was based on calculated functional properties rather than on physical characteristics themselves. Three functional properties were distinguished: (1) travel times from soil surface to water table, (2) water-table depth allowing a defined upward-flux density, and (3) downward-flux densities at a defined air content. The two types of horizons considered here were identified as two distinct groups by analysing the standard errors of prediction of the three functional properties. Graphs are presented that allow an estimate of the desired accuracy (expressed in terms of a prediction interval) as a function of sample size and measurement costs. The procedure which is illustrated here for two textural classes will in future be applied to all textures.

INTRODUCTION

Saturated and unsaturated hydraulic-conductivity and moisture-retention data are important soil-physical characteristics for models which simulate the water movement in the unsaturated zone. A major problem in applying simulation models is lack of data. Therefore, simple and reliable methods are needed allowing multiple measurements at many locations in a relatively short period of time (e.g. Bouma, 1983). In addition, soil maps are used to define pedological soil horizons according to soil-physical criteria forming groups of different pedological horizons with similar soil-physical properties. Identification of these groups allows reliable extrapolation of data and, therefore, fewer measurements (Wösten et al., 1985).

Groups of horizons can be identified on the basis of a statistical analysis of physical data of the individual horizons (e.g. Baker and Bouma, 1976). However, it is more attractive to identify these groups on the basis of functional properties related directly to practical applications. Then, the identifications will be governed by accuracy of prediction and not by statistical significance of differences between groups. This approach requires calculations using both hydraulic-conductivity and moisture-retention data. In this study, three functional properties are introduced:

(1) calculated travel times of water from the soil surface to a defined water table;

(2) calculated water tables which allow a defined upward-flux density to a defined level;

(3) calculated downward-flux densities that correspond with a defined air content in the soil.

In order to calculate these functional properties, certain assumptions have to be made and resulting values are, therefore, not necessarily associated with real field conditions. The relation between the sample size (n) and the standard error of prediction (SEP) of the functional properties can be used to minimize the number of measurements needed to obtain a defined accuracy of prediction.

Considering the above, the purpose of this study is to:

(1) identify two groups of soil horizons, as derived from Dutch soil survey criteria, using the three functional properties defined above;

(2) define the minimal number of measured individual soil horizons that are needed to obtain a desired accuracy of prediction.

MATERIALS AND METHODS

Soils

In this study, two groups of C horizons were used: one with a sand texture and one with a clay loam and silty clay loam texture, as distinguished in the Dutch soil survey. Twenty-five C horizons with a fine sand texture (Soil Survey Staff, 1975) originated from pleistocene cover-sand sediments with an organic-matter content less than 1% and a silt content ranging from 0 to 8%. The clay content varied between 0 and 3%. Average and standard deviation of the bulk density were 1620 kg m^{-3} and 30 kg m^{-3}. The horizons were structureless (single grain). The 25 soil profiles containing these horizons were classified according to Soil Survey Staff (1975), yielding 18 sandy, siliceous, mesic, Typic Haplaquods; 5 sandy, siliceous, mesic, Plaggepts and 2 sandy, siliceous, mesic, Typic Humaquepts. This classification indicates that similar subsoil horizons may occur in different soil types (see also Wösten et al., 1985).

Twenty-three clay loam and silty clay loam C horizons had developed in holocene fluviatile or marine deposits with an organic-matter content less than 1% and a clay content ranging from 25 to 35%. Average and standard deviation of the bulk density were 1420 kg m^{-3} and 40 kg m^{-3}. The horizons had moderately developed prisms, parting to a weak, fine, subangular blocky structure.

Horizons were both calcareous and non-calcareous. The 23 soil profiles containing these horizons were classified according to Soil Survey Staff (1975), yielding 19 fine silty, mixed, mesic, Typic Fluvaquents and 4 fine silty, mixed, mesic, Fluventic Eutrochrepts.

Physical methods

Relatively simple and rapid laboratory techniques (e.g. Bouma, 1983) were used to measure the water retention ($\theta - h$) and the hydraulic conductivity ($k - h$). Hydraulic conductivities of soil above the water table were measured by:

(1) the column method for vertical k_{sat} (e.g. Bouma, 1982);

(2) the crust-test for k_{unsat} down to approximately $h = -50$ cm (latest version of the method reported by Bouma et al., 1983);

(3) the sorptivity method for lower values of k_{unsat} in sand (Dirksen, 1979); and

(4) the hot-air method for lower values of k_{unsat} in clay loam and silty clay loam (Arya et al., 1975).

Moisture-retention curves were obtained by slow evaporation of wet, undisturbed samples in the laboratory, as reported by Bouma et al. (1983). In these samples, pressure heads were periodically measured with transducer-tensiometers and at the same time subsamples were taken to determine moisture contents. Thus, points relating h and θ were obtained. Moisture contents corresponding with pressure heads lower than -800 cm were obtained by conventional methods using air pressure (Richards, 1965). In the clay loam and silty clay loam soils a staining technique was applied to record the effects of horizontal cracks on the upward flux of water from the water table to the rootzone (k_{macro}; Bouma, 1984).

Calculation of travel times from soil surface to a defined water table

The travel time (T), as mentioned in the introduction, is the time that it takes water to travel from the soil surface to the water table. Assuming that all the water in the unsaturated zone is mobile and that piston-flow occurs, T is calculated as follows:

$$T = \frac{D \cdot \theta}{N} \tag{1}$$

where T = travel time of water from the soil surface to the water table (day), D = thickness of the unsaturated zone (m), θ = average moisture content of the unsaturated zone ($m^3 m^{-3}$), and N = average yearly precipitation surplus expressed as a daily rate for a winter period of 6 months (for The Netherlands $N = 1.4 \times 10^{-3}$ m day^{-1}). The thickness of the unsaturated zone is assumed to be 1 m.

The average moisture content is calculated using as input data:

(1) hydraulic conductivity and moisture retention data;

(2) the steady downward flux of $1.4 \times 10^{-3}\,\mathrm{m\,day^{-1}}$.

The steady flux is transformed into a corresponding pressure head, using the $k - h$ curve for the C horizon assuming unit-gradient flow in a semi-infinite porous medium. The calculated h value is transformed into moisture content (θ) using the corresponding $\theta - h$ curve, yielding an average moisture content for the unsaturated zone. This value of θ is used in eqn. (1), assuming $D = 1\,\mathrm{m}$.

Calculation of water table allowing a defined upward-flux density

Calculation of the water table, as mentioned in the introduction, is based on the Darcy equation for steady, upward, vertical flow:

$$v = -k\left(1 + \frac{\mathrm{d}h}{\mathrm{d}z}\right) \tag{2}$$

where $\mathrm{d}h/\mathrm{d}z$ = gradient of the pressure head ($\mathrm{m\,m^{-1}}$), k = hydraulic conductivity ($\mathrm{m\,day^{-1}}$), and v = flux density ($\mathrm{m\,day^{-1}}$). Integration (Gardner, 1958) yields:

$$z_n = -\int_0^{h_n} \frac{\mathrm{d}h}{1 + v/k} \tag{3}$$

where z_n = depth (m) below a reference layer, such as the lower boundary of the rootzone at which boundary a pressure head of $-h_n$ (m) is experienced. By choosing a steady upward-flux density (v) and by reading appropriate hydraulic-conductivity values from the $k - h$ relation at a specific h, a complete graph of z versus v may be plotted. Calculations are made by a computer (e.g. De Laat, 1980). For this test, v is assumed to be $0.002\,\mathrm{m\,day^{-1}}$ and $h = -500\,\mathrm{cm}$.

Calculation of downward-flux densities at a defined air content

In order to allow root activity and plant growth, the aeration status of the soil profile should be such that the profile contains at least 5% air by volume. This value has been proposed as a general criterium (FAO, 1985). The critical, steady, downward flux allowing 5% air by volume in the soil profile is calculated using the measured soil-physical characteristics.

The critical pressure head h at which the water content is 5% lower than the water content at saturation ($h = 0\,\mathrm{cm}$) is derived from the moisture-retention curve. This pressure head h is transformed into a flux density using the hydraulic-conductivity curve, assuming unit-gradient flow where the flux density is equal to the hydraulic conductivity. Thus, one value for the flux density is obtained, as discussed in the introduction.

Statistical analysis

The mean hydraulic conductivity and the mean moisture-retention curves for sand and for clay loam and silty clay loam were calculated by using data measured for the individual horizons. The 25 values of k and θ for sand were averaged at 13 selected values of h. The same procedure was followed for the 23 values for clay loam and silty clay loam.

Mean values for travel time, water table and downward-flux density were calculated by averaging the 25 values calculated for the individual sand horizons and the 23 values calculated for the individual clay loam and silty clay loam horizons.

The standard errors of prediction (SEP) of travel time, water table and downward-flux density were calculated from the standard deviations (S) of these properties and the sample size (n), according to:

$$\mathrm{SEP} = S\sqrt{1 + 1/n} \tag{4}$$

With increasing sample size, the standard error of prediction approaches the standard deviation. This minimal value of the standard error of prediction reflects the variability of the population with respect to the functional property, plus the error in measuring this property.

Approximate 90% confidence intervals for the means and half widths of 90% prediction intervals for the functional properties; (1) travel times from soil surface to water table, and (2) water-table depth allowing a defined upward-flux density were calculated, assuming that both data sets were random samples from normally distributed populations. For the functional property; (3) downward-flux densities at a defined air content, both data sets were transformed because they showed a log-normal distribution. In fact, the samples were purposive instead of random and the actual frequency distribution will deviate from normality and log-normality. The results may therefore be considered as rough approximations only, but they fulfil the primary purpose of illustrating the approach.

RESULTS

Basic physical data

The hydraulic-conductivity curves for sand and for clay loam and silty clay loam are shown in Figs. 1 and 2. The latter expresses the effect of horizontal cracks. Corresponding moisture-retention curves are presented in Figs. 3 and 4. The figures show mean relations and their upper and lower 90% confidence limits. In the following sections, comparisons between the curves will be based on the three functional properties discussed earlier.

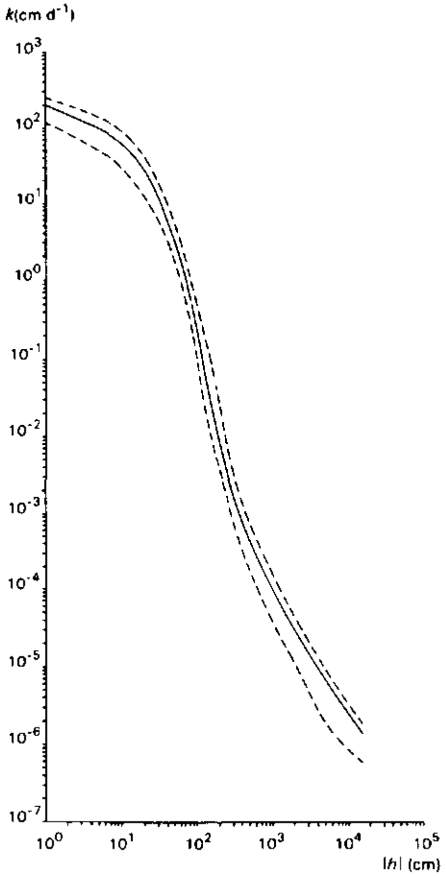

Fig. 1. Mean k–h relation and its 90% confidence limits for sand.

Travel times from soil surface to a water table depth of 1 m

The mean of the travel times for sand is 135 days, the 90% confidence limits of the mean are 118 and 152 days, and the half width of the 90% prediction interval is 89 days. The mean of the travel times for clay loam and silty clay loam is 298 days, the 90% confidence limits of the mean are 285 and 311 days and the half width of the 90% prediction interval is 64 days. The higher mean values for the clay loam and silty clay loam are primarily due to the higher average moisture content of the unsaturated zone in clay loam and silty clay loam.

The half width of the 90% prediction interval of the travel time as a function of the sample size (n) is presented in Fig. 5. The smaller the prediction interval the higher the accuracy.

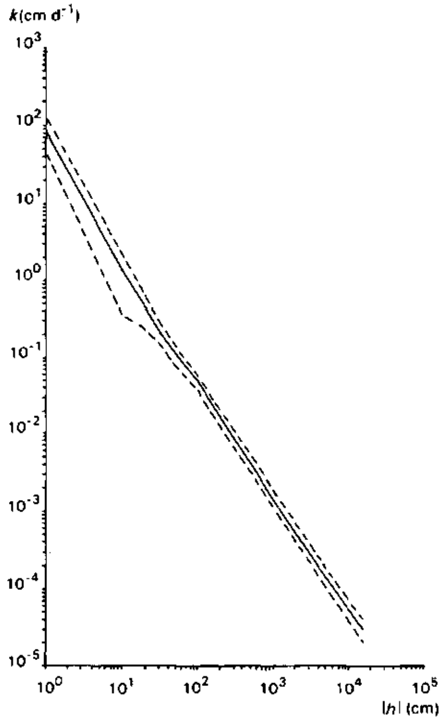

Fig. 2. Mean k–h relation and its 90% confidence limits for clay loam and silty clay loam with horizontal cracks.

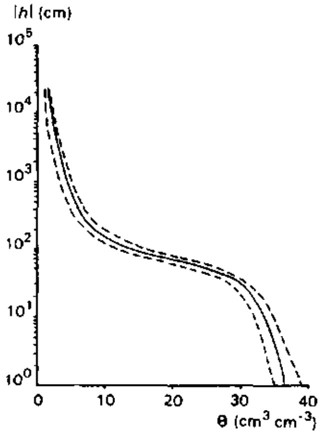

Fig. 3. Mean θ–h relation and its 90% confidence limits for sand.

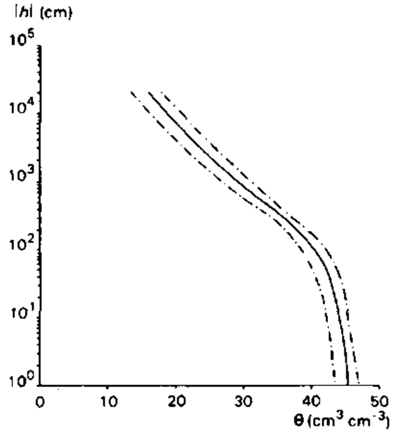

Fig. 4. Mean θ–h relation and its 90% confidence limits for clay loam and silty clay loam with horizontal cracks.

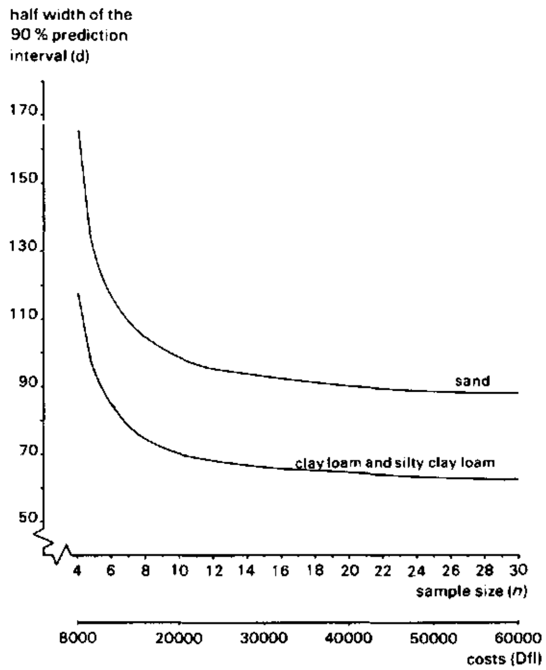

Fig. 5. Half width of the 90% prediction interval of the travel time from soil surface to a water-table depth of 1 m as a function of the sample size (n).

Sampling effort can also be expressed in terms of costs involved. Measurement of $\theta - h$ and $k - h$ relations of one soil horizon costs approximately Dfl. 2000.–($700.–). Figure 5 enables the user to determine the sample size that is for instance needed to calculate travel times in classes of half a year for a soil map at a scale of 1:50,000 (Breeuwsma et al., 1986).

Water tables allowing an upward-flux density of 0.002 m day^{-1} towards a reference layer with $h = -500$ cm

The mean of the water tables for sand is 94 cm, the 90% confidence limits of the mean are 87 and 101 cm, and the half width of the 90% prediction interval is 37 cm. The mean of the water tables for clay loam and silty clay loam is 64 cm, the 90% confidence limits of the mean are 55 and 73 cm, and the half width of the 90% prediction interval is 46 cm. The deeper water tables for sand are due to the higher k-values for sand down to a pressure head of $h = -500$ cm.

The half width of the 90% prediction interval of the water table as a function of the sample size (n) is presented in Fig. 6. This figure enables the user to choose the desired degree of accuracy of the calculated water table, again also considering the costs involved.

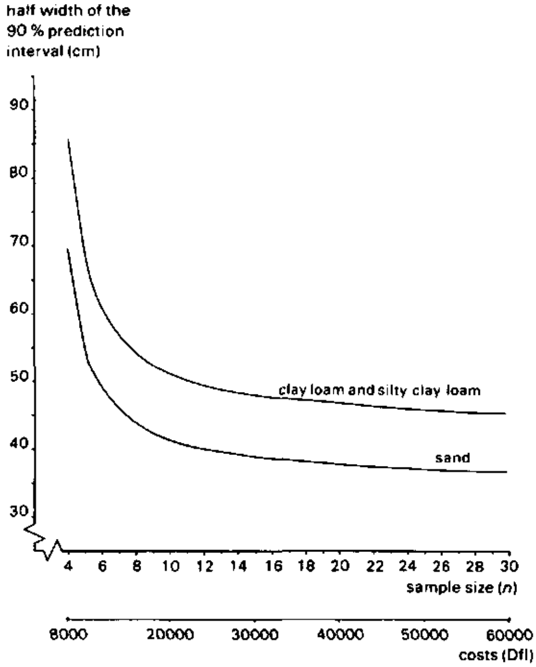

Fig. 6. Half width of the 90% prediction interval of the water table allowing an upward-flux density of 0.002 m day^{-1} as a function of the sample size (n).

Natural fluctuations of the water tables in The Netherlands are determined according to Dutch soil survey procedures (Van der Sluijs and De Gruijter, 1985) defining water table classes (Gt-values). The best accuracy for a water table that can be obtained, when using Gt-values, is a 90% prediction interval of approximately 80 cm. Figure 6 enables the user to determine the sample size if this accuracy is considered to be sufficient.

Downward-flux densities at an air content of 5% and with unit-gradient flow

The data are transformed logarithmically and the results of the statistical analysis are transformed back to the original scale. The geometric mean of the downward fluxes for sand is 198 mm day^{-1}, the 90% confidence limits of the mean are 122 and 323 mm day^{-1}, and the half width of the 90% prediction interval is 1185 mm day^{-1}. The geometric mean of the downward fluxes for clay loam and silty clay loam is 0.6 mm day^{-1}, the 90% confidence limits of the mean are 0.4 and 0.9 mm day^{-1}, and the half width of the 90% prediction interval is 2.4 mm day^{-1}. The higher values for sand are due to the higher critical pressure head and corresponding higher flux density in sand. The half width of the 90% prediction interval of the flux density as a function of the sample size (n) is presented in Figs. 7 and 8 for sand and for clay loam and silty clay loam respectively. This figure enables the user to choose the desired degree of accuracy of the calculated downward-flux density, again also considering the costs involved.

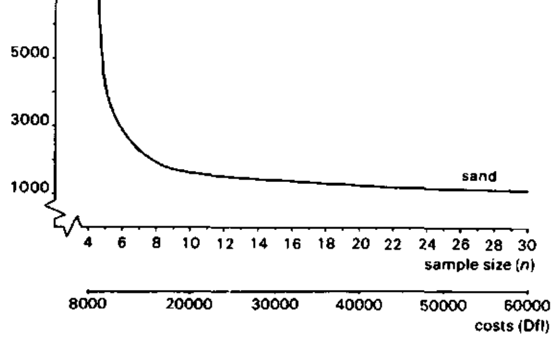

Fig. 7. Half width of the 90% prediction interval of the downward-flux density for sand at an air content of 5% and with unit-gradient flow as a function of the sample size (n).

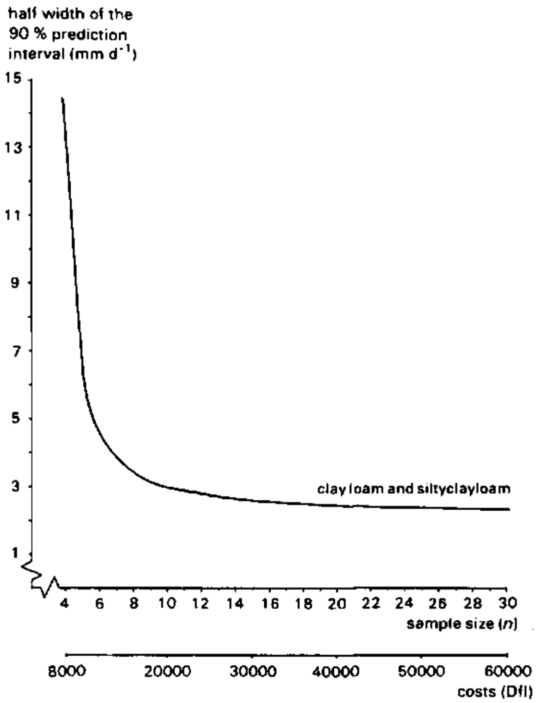

Fig. 8. Half width of the 90% prediction interval of the downward-flux density for clay loam and silty clay loam at an air content of 5% and with unit-gradient flow as a function of the sample size (n).

With sprinkler irrigation, this downward-flux density can be manipulated in such a way that an adequate aeration status is created. However, natural rainfall cannot be manipulated and it is therefore likely that under Dutch weather conditions the aeration status of clay loam and silty clay loam is often insufficient whereas sand normally has no problems in this respect.

DISCUSSION

Measurements of $k - h$ and $\theta - h$ relations in pedological horizons have indicated that different soil horizons are not always associated with significantly different physical characteristics (e.g. Wösten et al., 1985). However, the question should be raised which properties are to be used to judge differences in soil-physical characteristics between soil horizons. In this paper, a procedure is presented to identify soil horizons by means of three practical soil-physical interpretations that yield characteristic numbers, rather than by comparing the conductivity and retention functions as such. The procedure has been demonstrated for horizons in two contrasting soil-texture classes, as used in the Dutch soil survey at a scale of 1:50,000. The procedure will now also be applied to horizons that belong to other texture classes. The connection with the texture classes is important, because data obtained will have to be used for calculations of land areas as delineated on soil maps. The connection also implies that the different groups of horizons cannot be chosen freely, as they correspond with the existing soil-texture classification as used in soil-map legends.

The principles presented in this paper will be used to examine whether different existing texture classes can be combined or not. If the standard error of prediction for the combined texture class is markedly larger than that for the separate classes, the classes will not be joined; if this is not the case, classes may be combined. The critical value of the standard error of prediction, determining whether or not classes are combined, depends on the desired accuracy of prediction.

Once a group of soil horizons has been identified, the question arises as to how many individual samples should be taken to arrive at a required accuracy of prediction. This level of accuracy is a function of the intended use of the data, which may vary, and of funds available. A graph is therefore presented that allows an estimate of the accuracy as a function of the number of samples and the associated costs. The desired accuracy should not exceed the limited degree of detail that is often adequate for specific interpretations. This aspect deserves special attention as to avoid over-accuracy.

REFERENCES

Arya, L.M., Farrell, D.A. and Blake, G.R., 1975. A field study of soil water depletion patterns in presence of growing soybean roots. I. Determination of hydraulic properties of the soil. Soil Sci. Soc. Am. Proc., 39: 424–430.

Baker, F.G. and Bouma, J., 1976. Variability of hydraulic conductivity in two subsurface horizons of two silt loam soils. Soil Sci. Soc. Am. J., 40: 219–222.

Bouma, J., 1982. Measuring the hydraulic conductivity of soil horizons with continuous macropores. Soil Sci. Soc. Am. J., 46(2): 438–441.

Bouma, J., 1983. Use of soil survey data to select measurement techniques for hydraulic conductivity. Agric. Water Manage., 6(2/3): 177–190.

Bouma, J., 1984. Using soil morphology to develop measurement methods and simulation techniques for water movement in heavy clay soils. In: J. Bouma and P.A.C. Raats (Editors), Water and Solute Movement in Heavy Clay Soils. Proc. of an ISSS Symposium, ILRI, Wageningen.

Bouma, J. et al., 1983. Assessing the suitability of soils with macropores for subsurface liquid waste disposal. J. Environ. Qual., 12(3): 305–310.

Breeuwsma, A. et al., 1986. Derivation of land qualities to assess environmental problems from soil surveys. Soil Sci. Soc. Am. J., 50: 186–190.

De Laat, P.J.M., 1980. Model for unsaturated flow above a shallow water table, applied to a regional sub-surface flow problem. Agric. Res. Rep. 895. Pudoc, Wageningen.

Dirksen, C., 1979. Flux-controlled sorptivity measurements to determine soil hydraulic property functions. Soil Sci. Soc. Am. J., 43: 827–834.

FAO, 1985. Guidelines: land evaluation for irrigated agriculture. FAO Soils Bull. 55, Rome.

Gardner, W.R., 1958. Some steady state solutions of the unsaturated moisture flow equation with application to evaporation from a water-table. Soil Sci., 85: 228–232.

Richards, L.A., 1965. Physical condition of water in soil. In: C.A. Black (Editor), Methods of Soil Analysis I, Agronomy 9, Am. Soc. Agron., Madison, Wisc., pp. 131–137.
Soil Survey Staff, 1975. Soil Taxonomy. A Basic System of Soil Classification for Making and Interpreting Soil Survey. Agric. Handb., 436. USDA–SCS, Washington, D.C., 754 pp.
Van der Sluijs, P. and De Gruijter, J.J., 1985. Water table classes: a method used to indicate seasonal fluctuation and duration of water tables on Dutch soil maps. Agric. Water Manage., 10(2): 109–125.
Wösten, J.H.M., Bouma, J. and Stoffelsen, G.H., 1985. The use of soil survey data for regional soil water simulation models. Soil Sci. Soc. Am. J., 49: 1238–1244.

6. FUNCTIONAL SENSITIVITY ANALYSIS OF FOUR METHODS TO GENERATE SOIL HYDRAULIC FUNCTIONS

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Functional Sensitivity Analysis of Four Methods to Generate Soil Hydraulic Functions

J. H. M. Wösten,* C. H. J. E. Schuren, J. Bouma, and A. Stein

ABSTRACT

Rapid advances in model building have led to the understanding that applicability of future simulations depends, to a great extent, on the availability of accurate soil hydraulic functions obtained with efficient methods rather than on new models. In this study, four different methods were used to generate hydraulic functions: Method A, direct on-site measurement; Method B, use of measured hydraulic functions averaged on a regional scale; Method C, use of measured hydraulic functions averaged on a national scale; and Method D, use of van Genuchten parameters correlated with soil texture and organic-matter content. Accuracy of these methods was tested by comparing the simulated water storage with the measured water storage of the upper 0.5 m of three soil profiles over a period of 7 yr. Differences in performance of the four methods were not significant. Agreement between measured and simulated water storage was best, however, when directly measured hydraulic functions (Method A) were used. Next best agreement was obtained when continuous (Method D) and two types of class pedotransfer functions (Methods B and C) were used, which relate textures and soil horizons to physical characteristics. Costs involved in obtaining directly measured soil hydraulic functions are prohibitively high, compared with costs for the other methods. With regard to both accuracy and costs, the development of a data base of measured soil hydraulic functions and use of this information to derive continuous and class pedotransfer functions assures, in many cases, optimal spending of limited available resources.

SIMULATION MODELS have become indispensible research tools for describing movement of water and solutes into and through the unsaturated zone. Models, ranging from very simple to highly complex, are being used increasingly to evaluate effects of management practices on crop yield and groundwater quality (Dumanski and Onofrei, 1989; Addiscott and Wagenet, 1985; Penning de Vries and van Laar, 1982). This use of models for research and management purposes has led to the understanding that different problems ask for different approaches. Sometimes an in-depth approach is needed to study a problem in detail and sometimes a simple approach is appropriate when addressing a general question (Bouma, 1989).

The availability of input data is pertinent to the use of models. As our ability to simulate more complex systems increases, the accuracy of future simulations may well depend on the availability of accurate input data. Lack of accurate soil hydraulic functions, in particular, is often considered to be a major obstacle for making progress (van Genuchten et al., 1989). Formulated in a positive way, using available data is acceptable as long as we are aware of the uncertainty of our predictions. Use of "perfect" data, if it exists at all, is often not necessary because many problems do not ask for exact solutions. At the same time, use of "perfect" data obtained by direct measurement of soil hydraulic functions is often prohibitive because of the cost and time involved. For every problem, the challenge is to identify a modeling approach that provides a quantitative estimate of system behavior from a relatively simple and cheap data set while providing an indication of the uncertainty of predictions.

In this study, a functional sensitivity analysis was conducted in which the effects on simulation of soil water storage were evaluated for four different methods to generate soil hydraulic functions. Methods to generate soil hydraulic functions included Method A, relatively expensive direct on-site measurement; Method B, use of measured hydraulic functions averaged by soil horizon on a regional scale; Method C, use of measured hydraulic functions averaged by soil horizon on a national scale; and Method D, use of van Genuchten parameters correlated with soil texture and organic-matter content. Simulated water storage in the upper 0.5 m of three soil profiles was compared with measured water storage for a period of 7 yr. The SWATRE model being used has been validated for this watershed in a previous study (Wösten et al., 1985), as well as for other studies (Feddes et al., 1988).

MATERIALS AND METHODS

Site Characteristics

The Hupselse Beek watershed is situated in the eastern part of the Netherlands and has been an experimental study area of the National Dutch Water Service for 20 yr. The area covers 650 ha, and its altitude varies between 22 and 33 m above mean sea level. Land use is predominantly agricultural: 80% pasture, 12% arable land, and 8% forest. The area is underlain by Miocene clay sediments starting between 0.2 and 10.0 m below the present soil surface. The area was affected by glaciers that deposited boulder clay (glacial till) in an early Pleistocene period. Later, aeolian sands were deposited over the entire area, forming a surface relief that is quite different from the relief of the underlying boulder-clay surface. Wösten et al. (1985) presented a detailed soil map of the area.

In this study, water storage was simulated for three sites in two dominant mapping units in the area. Figure 1 shows the soil profiles for the three sites. The soil horizons A, B, and C indicate sandy horizons; D1 is boulder clay and D2 is Miocene clay. According to Soil Survey Staff (1975), Sites 1 and 3 are classified as sandy, siliceous, mesic Typic Haplaquods and Site 2 as a sandy, siliceous, mesic Plaggept. The boulder clay at Site 3 is variable in composition, with sand veins adjacent to heavy clay. At the same time, the starting depth of the boulder clay varies strongly over short distances (Bouma et al., 1989). As a result, profiles at different sites in this mapping unit may differ from the profile at Site 3. Groundwater levels for Site 1 range from 0.4 m below the soil surface in winter to 1.4 m in summer. Ranges are from 2.2 m in winter to 2.9 m in summer for Site 2, and from 0.6 m in winter to 1.3 m in summer for Site 3.

Generating Soil Hydraulic Functions

Soil water-retention and hydraulic-conductivity curves were generated for each soil horizon using the four different methods.

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A. Direct Measurement. Hydraulic conductivities of the different soil horizons were measured using a combination of the following four methods: (i) the column method (Bouma, 1982) for the vertical saturated hydraulic conductivity (K_s), (ii) the crust test (latest version of the method reported by Bouma et al., 1983) for unsaturated conductivities when the pressure head, h, was between 0 and −5 kPa, (iii) the sorptivity method (Dirksen, 1979) for conductivities in the case of sand when $h < -5$ kPa, and (iv) the hot-air method (Arya et al., 1975) for conductivities in the case of loam and clay when $h < -5$ kPa. Soil water retention was obtained by slow evaporation of wet undisturbed samples in the laboratory, as reported by Bouma et al. (1983). In these samples, pressure heads were periodically measured with transducer-tensiometers while, at the same time, subsamples were taken to determine water contents, θ. This procedure yielded points relating h and θ. Water contents corresponding with $h < -80$ kPa were obtained by conventional air-pressure methods (Klute, 1986). A set of 13 measured soil water-retention and hydraulic-conductivity curves were used for Method A to provide model input for all horizons of the three soils, as illustrated in Fig. 1.

B. Use of Measured Soil Hydraulic Functions Averaged by Soil Horizon on a Regional Scale. In the study by Wösten et al. (1985), the hydraulic functions of all soil horizons distinguished in the Hupselse Beek watershed were measured in sixfold by using the described measurement techniques. Sampling locations were chosen at random within the various delineated areas of the soil map. Replicate measurements for each horizon were used to calculate the average hydraulic functions for every horizon. After comparison of the hydraulic functions of the different horizons, only those horizons were distinguished whose soil hydraulic functions differed significantly. Their number was lower than the number of horizons that was distinguished by pedological classification, because pedological differences do not necessarily correspond with difference in hydraulic functions.

Hydraulic functions averaged on a regional scale were used in Method B to obtain soil hydraulic data for the soil profiles at the three sites. A set of four different soil water-retention and hydraulic-conductivity curves was distinguished: A1, Aan, B2 + C11 + C12 + C11g, and D1 + D2.

C. Use of Measured Soil Hydraulic Functions Averaged by Soil Horizon on a National Scale. The described techniques have been used to measure hydraulic functions for a large number of soil horizons in the Netherlands. For application on a national scale, soils were classified according to soil texture (as used by the Netherlands Soil Survey Institute) and type of horizon, either topsoil (A horizon) or subsoil (B and C horizons). This classification resulted in 20 different soil groups comprising a total of 197 individual curves. Tabulated forms of the averaged curves for the 20 soil groups were presented by Wösten et al. (1987). As a set, the curves form a unique data base covering the broad spectrum of soils in the Netherlands. This set is increasingly being used to simulate regional soil water regimes. In Method C, the averaged curves for three soil groups from the national set were used in this study to provide hydraulic functions for the A1 + Aan, B2 + C11 + C12 + C11g, andD1 + D2 horizons.

D. Use of Soil Properties to Predict Soil Hydraulic Functions. Analytical expressions of van Genuchten (1980) for the water-retention and hydraulic-conductivity curves were fitted simultaneously to the set of measured hydraulic functions for a wide range of soils in the Netherlands (Wösten et al., 1987). Wösten and van Genuchten (1988) used regression analysis to relate estimated model parameters to more easily measured soil properties such as bulk density and percentages of silt, clay, and organic matter. The resulting functions predict the hydraulic parameters for all horizons shown in Fig. 1 from their corresponding texture values.

Methods B, C, and D qualify as pedotransfer functions, which relate different land and soil characteristics with one another and to land qualities. Methods B and C can be described in terms of class pedotransfer functions, relating soil horizons to associated hydraulic functions. Method D can be described in terms of a continuous pedotransfer function, relating, e.g., soil texture, bulk density, and percentage organic matter to associated hydraulic functions (Bouma and van Lanen, 1987; Bouma, 1989).

Statistical Data Analysis

Two statistical properties were calculated to evaluate the differences between measured and calculated water storage. The mean residual error (ME) was defined as

$$\mathrm{ME} = (1/n) \sum_{i=1}^{n} (x_i - y_i) \qquad [1]$$

and the mean squared residual error (MSE) was defined as

$$\mathrm{MSE} = (1/n) \sum_{i=1}^{n} (x_i - y_i)^2 \qquad [2]$$

where n = number of data points of measured and calculated water storage, x = measured water storage, and y = calculated water storage.

Calculated water storages are regressed vs. x and are presented in a graph that includes the 1:1 line.

Mean residual error is a measure for the bias in the simulation results. Values close to zero indicate that measured and calculated water storages do not differ systematically from each other or, equivalently, that there is no consistent bias. Values that differ greatly from zero indicate the presence of systematic deviation or bias.

Mean squared error is a measure for the scatter of the data points around the 1:1 line. Low MSE values indicate little scatter, and high MSE values indicate large scatter. Low MSE values also imply low ME values.

The variance of the differences between measured and calculated water storages (VAR) was estimated as

$$\mathrm{VAR} = \frac{n(\mathrm{MSE} - \mathrm{ME}^2)}{n - 1} \qquad [3]$$

With the assumption of normal distribution and independence of differences between measured and calculated water storages, the half width of the 95% confidence interval for ME was calculated as

$$t\left(\frac{\mathrm{VAR}}{n}\right)^{1/2} \qquad [4]$$

where t = the value of the Student's t distribution for α = 0.05 and $n - 1$ degrees of freedom. This half width of the confidence interval is used to examine, for each of the four methods, whether differences between measured and calculated water storages are statistically significant.

An analysis of variance (ANOVA) revealed if there were statistically significant differences between the four methods with respect to their approximation of the measured water storages. With the assumption of independence of both ME and MSE values, ANOVA was carried out on the twelve ME and twelve MSE values calculated (four methods, with the three sites as repetitions).

Expenses of the Different Methods

The four methods used to generate soil hydraulic functions differed considerably in terms of costs. In Method A, hydraulic functions were measured for each of the 13 horizons, which constitute the soil profiles at three sites in the watershed. These sites represented a pedon area of 10 m² each. Measurement of the hydraulic functions took an estimated labor cost of 3 d per horizon, or a total of 39 d of operational activities. Hydraulic functions averaged on a regional scale (Method B) were obtained from replicate measurements of identical horizons at different locations within the delineated areas. As a consequence, initial investment to obtain average functions was, in principle, the same as the cost of the total number of measurements made in the area. The nine horizons distinguished in the Hupselse Beek watershed were measured in sixfold to obtain average functions (Wösten et al., 1985). This required a total of 9 × 6 × 3 = 162 d. Once these average curves were obtained, on-site visits to establish the type of soil horizons were satisfactory to generate hydraulic functions for the entire watershed area of 650 ha. This required little operational activity, perhaps 1 d.

Hydraulic functions averaged on a national scale (Method C) were derived from a data base comprising a total of 197 individual curves. The initial investment to obtain these average curves was therefore 197 × 3 = 591 d. The functions averaged on a national scale provide information for the entire Netherlands, an area of about 3 000 000 ha. Analogous to conditions on a regional scale, operational activities are restricted to on-site visits to establish the type of soil horizons. This takes an estimated 5 d of operational activities. The continuous pedotransfer functions used in Method D were derived from the same data base as used for Method C. Therefore, initial investment and area covered were the same as with Method C. In this case, however, operational activities were not restricted to on-site horizon identification, but also included measurement of soil texture of the identified horizons. This took an estimated 10 d of operational activities for 13 horizons.

Thus, direct measurement (Method A) implies a very high investment per unit area, compared with Methods B, C, and D. After making a major initial investment for the determination of a standard set of measurements, Methods B, C, and D can be executed with little effort. Besides the fact that Method A is prohibitively expensive, when average project budgets are considered, the measurements are site-specific and cannot be extrapolated. The next project will again require identical measurements. Since Method A is not a practical possibility, research has to focus on whether investments required to execute Methods B, C, and D are justified. This would be the case if water storages are equally well described by calculating them using data obtained with Methods B, C, and D, compared with data obtained with Method A.

Simulation of Soil Water Storage

Soil water flow was simulated with the model SWATRE (Soil Water Actual Transpiration Extended; Feddes et al., 1978; Belmans et al., 1983), which is a one-dimensional, finite-difference model that describes transient, unsaturated water flow in a heterogeneous soil/root system that may or may not be under groundwater influence. In this study, soil physical input data for every horizon of the three sites were obtained with four different methods. Calculations were made for a grass crop by using a sink term to simulate water uptake by plant roots. Rooting depth was 0.2 m for Site 1 and 0.25 m for Sites 2 and 3.

The meteorological station of the Hupselse Beek watershed provided daily precipitation and daily potential evapotranspiration values for the period March 1976 to December 1982. Potential evapotranspiration values were calculated according to Thom and Oliver (1977). Groundwater levels for Site 1, which is located at the meteorological station, were measured daily. Groundwater levels for Sites 2 and 3 were measured every week. After correlation of the weekly levels for Sites 2 and 3 to the daily levels of Site 1, daily groundwater levels for Sites 2 and 3 were estimated. These daily groundwater levels for the period March 1976 to December 1982 served as the lower boundary condition for the SWATRE simulations.

The SWATRE model was used to make daily calculations for an uninterrupted 7-yr period from March 1976 to December 1982. To a depth of 0.5 m, the soil profiles were schematized into five compartments with a thickness of 0.10 m each and, deeper than 0.5 m, into eight compartments with a thickness of 0.20 m each. By accumulating water contents of the top five compartments, cumulative water storages of the first 0.5 m of the soil profiles were calculated.

In a previous study for Site 1, van Vuuren (1984) demonstrated that hysteresis affected the results. Use of an "assumed" adsorption water-retention curve for simulation of wetting after the exceptionally dry conditions of 1976 gave results that agreed better with measured water contents when compared with results obtained with the use of a measured desorption curve. However, measured adsorption curves were not available. Therefore, data points for this very dry period, which has a probability of occurrence of only 2%, were omitted from the data analysis in this study.

Measurements of Water Content with the Neutron Probe

As part of ongoing research in the Hupselse Beek watershed, water contents at different locations have been measured since 1976 with a neutron probe. Calibration curves were established for the different soil horizons in the watershed to provide an accurate conversion of neutron counts into water contents.

Distances between locations where neutron-probe measurements were made and locations where soil samples were taken for measurement of hydraulic functions were about 10 m for each site in order not to disturb the neutron-probe readings. At the three sites considered in this study, neutron-probe measurements were made once every 2 wk to a depth of 2 m below the soil surface. In the topsoil, measurements were made at 0.25-, 0.35-, and 0.45-m depth. Water contents were such that the radius of the sphere of influence for neutrons was about 0.25 m (Gardner, 1986). Therefore, water contents at these depths could be used to calculate the water storage of the first 0.5 m of each of the three soil profiles. The large quantity of measured values of soil water storage over the period 1976 to 1982 form an exceptional data base for validation of values of soil water storage as simulated with the four different methods discussed above.

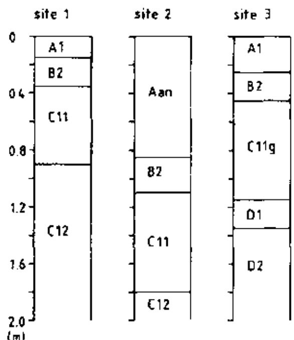

Fig. 1. Soil profiles at the three sites.

RESULTS AND DISCUSSION

Measured and calculated water storages show a pattern that agrees with what might be expected from the prevailing weather conditions over the years 1976 to 1982. The dry year 1976 and, to a lesser extent, also

1982 showed, as expected, a strong decrease in water storage in the summer period while, in the wet years 1977 and 1979, changes in water storage over the seasons were less pronounced. The general trend of water storages calculated with the four methods over the years agrees well with measured values for all sites. For reasons of brevity, Fig. 2 shows only results obtained with Method A, for the characteristic period March 1976 to March 1978, of measured and simulated water storages in the upper 0.5 m of Site 1. Drying of the upper 0.5 m of the soil profile in 1976 was well predicted by simulation. There was also good agreement between measured and calculated values over the period March 1977 to March 1978. However, wetting of this top layer in the winter 1976–1977 was systematically overestimated with simulation. The main reason for this discrepancy is that the desorption water-retention curve was used in the simulation model, while use of an adsorption curve would have been more realistic for wetting under these dry conditions (see discussion above).

Table 1 summarizes the results of the statistical analysis of the differences between measured and calculated water storages for the four methods and the three sites. For each site and for each method, values are presented for the number of data points, the ME, the half width of the 95% confidence interval of the ME, and the MSE. Figure 3 shows the results for Site 1 of the regression of calculated water storage vs. measured values for the four methods. The 1:1 line, as well as the regression line, are included. The figure provides a good visual reflection of results of the statistical analysis in terms of bias and scatter of data points. Comparable expressions were obtained for the other two sites.

The slopes of the regression lines varied from 0.61 for Method D at Site 2 to 1.40 for Method C at Site 3. Variation in r^2 (r is the correlation coefficient) was from 0.46 for Method D at Site 2 to 0.85 for Method B at Site 3.

For Sites 1 and 2, ME values for Methods A and D are relatively close to zero, compared with the values for Methods B and C. For Site 3, there is not such a trend. When ME values differ from zero, this indicates that measured and calculated water storages deviate in most cases. The ME values, together with the calculated half widths of the 95% confidence interval, indicate that these deviations are statistically significant for *all* methods and *all* sites except for Methods A and D for Site 2. Only in these two cases is the hypothesis that measured and calculated values are the same (i.e., ME = 0) not rejected. The positive ME values for Site 1 indicate that simulated water storages are systematically lower than measured values. This result is supported by Fig. 3, where the majority of data points are found below the 1:1 line. The negative ME values for Site 3 indicate an opposite effect; in this case, simulated water storages are systematically higher than measured values. In many simulation-model studies, such systematic deviations are corrected in the model-calibration phase, thereby eliminating their negative effect on model output. In this study, we have chosen to present the real output data because emphasis is on an accurate simulation as well as on comparison of the four methods.

Low MSE values indicate little systematic deviation between measured and calculated values, as well as absence of a systematic trend in the scatter of data points around the 1:1 line. While systematic deviations, as also expressed by high ME values, can be corrected by model calibration, trends in the scatter of data points are not corrected and are undesirable. Therefore, low MSE values are desirable. Comparison of the MSE values in Table 1 shows that, for Sites 1 and 2, Method A had the least dispersion around the 1:1 line. Next, Method D performed best, while Methods B and C showed the most dispersion. For Site 3, MSE values for the methods are about the same.

The ANOVA for both ME and MSE shows that differences between methods were not statistically significant. Since agreement between measured and calculated water storage is not highly influenced by the choice of the method used to generate soil hydraulic functions, this choice can be based on other considerations such as cost or ease of use.

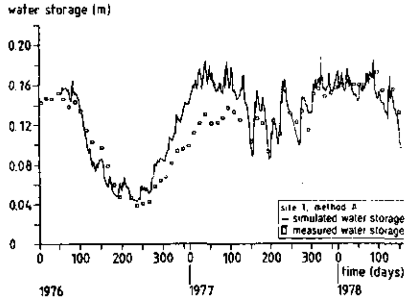

Fig. 2. Yearly trend in measured and calculated water storages of the upper 0.5 m at Site 1 when using Method A (direct measurement).

Table 1. Results of the statistical analysis, for each of the four methods, of the differences between measured and simulated water storages.

Method	n	Mean residual error (ME)	Half width of the 95% confidence interval of ME	Mean squared residual error
		cm	cm	cm^2
Site 1				
A	132	1.329	0.317	5.118
B	132	2.433	0.319	9.315
C	132	2.615	0.347	10.857
D	132	2.010	0.295	6.953
Site 2				
A	154	0.1556	0.317	3.954
B	154	−1.8966	0.353	8.494
C	154	1.6840	0.341	7.391
D	154	−0.2437	0.332	4.383
Site 3				
A	155	−1.223	0.312	3.475
B	155	−1.242	0.211	3.304
C	155	−1.044	0.250	3.533
D	155	−1.497	0.181	3.534

SUMMARY AND CONCLUSIONS

Four methods were used to generate soil hydraulic functions for use in a soil-water simulation model.

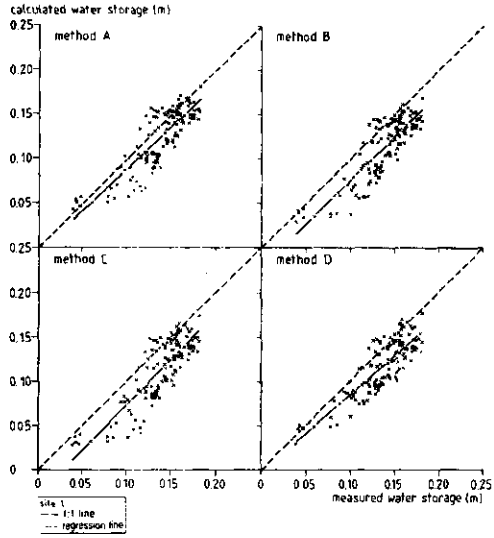

Fig. 3. Comparison between measured and calculated water storages at Site 1 for a 7-yr period, using the four methods.

Calculated water-storage values for the upper 0.5 m of three sandy soils were compared, for a period of 7 yr, with water-storage values measured with a neutron probe. Comparison of the different methods showed that, in most cases, calculated values deviate systematically from measured values. However, when a well-tested simulation model is used, this systematic deviation can be corrected in the model-calibration phase. Evaluating the four methods in terms of ME and MSE indicates that direct measurement of hydraulic functions (Method A) gave the best results, followed by continuous pedotransfer functions (Method D) and averaged regional and national functions (Methods B and C). Analysis of variance shows that differences between methods with respect to their approximation of measured water storages were not significant. Compared with Methods B, C, and D, the costs of using Method A are very high, making it an unrealistic alternative. Methods B, C, and D require a substantial initial investment for the determination of a standard set of measured hydraulic functions. Once such a set is established, however, its use is attractive in terms of accuracy and cost.

ACKNOWLEDGMENTS

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REFERENCES

Addiscott, T.M., and R.J. Wagenet. 1985. Concepts of solute leaching in soils: A review of modelling approaches. J. Soil Sci. 36:411–424.

Arya, L.M., D.A. Farrell, and G.R. Blake. 1975. A field study of soil water depletion patterns in presence of growing soybean roots. I. Determination of hydraulic properties of the soil. Soil Sci. Soc. Am. Proc. 39:424–430.

Belmans, C., J.G. Wesseling, and R.A. Feddes. 1983. Simulation model of the water balance of a cropped soil: SWATRE. J. Hydrol. (Amsterdam) 63:271–286.

Bouma, J. 1982. Measuring the hydraulic conductivity of soil horizons with continuous macropores. Soil Sci. Soc. Am. J. 46:438–441.

Bouma, J. 1989. Land qualities in space and time. p. 3–13. *In* J. Bouma and A.K. Bregt (ed.) Proc. ISSS Symp. on land qualities in space and time, Wageningen, the Netherlands. 22–26 Aug. 1988. Pudoc, Wageningen.

Bouma, J., C. Belmans, L.W. Dekker, and W.J.M. Jeurissen. 1983. Assessing the suitability of soils with macropores for subsurface liquid waste disposal. J. Environ. Qual. 12:305–311.

Bouma, J., A.G. Jongmans, A. Stein, and G. Peek. 1989. Characterizing spatially variable hydraulic properties of a boulder clay in the Netherlands. Geoderma 45:19–29.

Bouma, J., and H.A.J. van Lanen. 1987. Transfer functions and threshold values: From soil characteristics to land qualities. p. 106–111. *In* K.J. Beek, P.A. Burrough, and D.E. McCormack (ed.) Proc. ISSS/SSSA workshop on quantified land evaluation procedures. Int. Inst. for Aerospace Surv. and Earth Sci. Publ. no. 6. Enschede, the Netherlands.

Dirksen, C. 1979 Flux-controlled sorptivity measurements to determine soil hydraulic property functions. Soil Sci. Soc. Am. J. 43:827–834.

Dumanski, J., and C. Onofrei. 1989. Crop yield models for agricultural land evaluation. Soil Use Manage. (in press).

Feddes, R.A., P.J. Kowalik, and H. Zaradny. 1978. Simulation of field water use and crop yield. Simulation Monogr. Pudoc, Wageningen, the Netherlands.

Feddes, R.A., M. de Graaf, J. Bouma, and C.D. van Loon. 1988. Simulation of water use and production of potatoes as affected by soil compaction. Potato Res. 31:225–239.

Gardner, W.H. 1986. Water content. p. 493–544. *In* A. Klute (ed.) Methods of soil analysis. Part 1. 2nd ed. Agron. Monogr. 9. ASA and SSSA, Madison, WI.

Klute, A. 1986. Water retention: laboratory methods. p. 635–662. *In* A. Klute (ed.) Methods of soil analysis. Part 1. 2nd ed. Agron. Monogr. 9. ASA and SSSA, Madison, WI.

Penning de Vries, F., and H. van Laar (ed.). 1982. Simulation of plant growth and crop production. Simulation Monogr. Pudoc, Wageningen, the Netherlands.

Soil Survey Staff. 1975. Soil taxonomy: A basic system of soil classification for making and interpreting soil surveys. USDA-SCS Agric. Handb. 436. U.S. Gov. Print. Office, Washington, DC.

Thom, A.S. and H.R. Oliver. 1977. On Penman's equation for estimating regional evaporation. Q.J.R. Meteorol. Soc. 103:345–357.

van Genuchten, M.Th. 1980. A closed-form equation for predicting the hydraulic properties of unsaturated soils. Soil Sci. Soc. Am. J. 44:892–898.

van Genuchten, M.Th., F. Kaveh, W.B. Russell, and S.R. Yates. 1989. Direct and indirect methods for estimating the hydraulic properties of unsaturated soils. p. 61–72. *In* J. Bouma and A.K. Bregt (ed.) Proc ISSS Symp. on land qualities in space and time, Wageningen, the Netherlands. 22–26 Aug. 1988. Pudoc, Wageningen.

van Vuuren, W.E. 1984. Validation of the agrohydrological model DEMGEN (Demand Generator) on point data from the Hupselse Beek area in the Netherlands. p. 829–839. *In* P. Udluft et al. (ed.) Proc. Int. Symp. on recent investigations in the zone of aeration, Munich, W. Germany. October 1984. Technical Univ. Munich.

Wösten, J.H.M. and M.Th. van Genuchten. 1988. Using texture and other soil properties to predict the unsaturated soil hydraulic functions. Soil Sci. Soc. Am. J. 52:1762–1770.

Wösten, J.H.M., J. Bouma, and G.H. Stoffelsen. 1985. Use of soil survey data for regional soil water simulation models. Soil Sci. Soc. Am. J. 49:1238–1244.

Wösten, J.H.M., M.H. Bannink, and J. Beuving. 1987. Water retention and hydraulic conductivity characteristics of top- and subsoils in the Netherlands: The Staring series. Rep. 1932. Soil Surv. Inst., Wageningen, the Netherlands (in Dutch).

7. USE OF PRACTICAL ASPECTS OF SOIL BEHAVIOUR TO EVALUATE DIFFERENT METHODS TO GENERATE SOIL HYDRAULIC FUNCTIONS

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USE OF PRACTICAL ASPECTS OF SOIL BEHAVIOUR TO EVALUATE DIFFERENT METHODS TO GENERATE SOIL HYDRAULIC FUNCTIONS

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ABSTRACT

Soil hydraulic functions can be obtained with methods that range from complex and costly to simple and cheap. Decisions as to which is the most appropriate method for a specific application have to be based on a comparison of generated hydraulic functions. This comparison should preferably be based on a statistical comparison of practical applications calculated with the different hydraulic functions rather than on a statistical comparison of the functions themselves. In this study four different methods were used to generate hydraulic functions: (A) direct on-site measurement, (B) measurement in soil horizons in the area, (C) use of a national data set, and (D) use of Van Genuchten parameters correlated with soil texture and organic matter content. The four methods were compared by their effect on two practical aspects of soil behaviour: (1) evapotranspiration deficit and (2) flux through a plane at 30 cm below soil surface. These two aspects are highly relevant for agricultural and environmental use. However, direct measurement is not feasible. A validated simulation model was used for the calculations and results obtained with method A were taken as a reference. Calculations were performed for three soil profiles for a period of seven years. Deficits and fluxes, calculated with the four methods to generate hydraulic functions, were not significantly different using the data of the seven-year period. However, methods were significantly different when rainfall deficits were used as a covariable. This is true with the exception of downward fluxes in the period October until March which are most important for leaching of pollutants. The user has to decide whether differences between methods are sufficiently large to justify repeated, expensive on-site measurements (method A) or whether an investment will be made to make standard series of curves to be used everywhere (methods C and D).

KEY WORDS Hydraulic functions Simulation model Evapotranspiration deficit Fluxes

INTRODUCTION

Soil hydraulic functions, which are crucial for modelling water flow in the unsaturated zone, can be obtained with different methods. One approach is direct measurement of the functions. Methods to do so were reviewed by Klute and Dirksen (1986) and Green *et al.* (1986). Use of parameter estimation methods to calculate hydraulic functions (Kool *et al.*, 1987) is an inverse method that has recently drawn much attention. Use of theoretical methods to predict hydraulic conductivity from more easily measured field or laboratory soil water retention data is another approach that was reviewed by Mualem (1986). Approaches in which the soil hydraulic functions are statistically related to soil texture and to other soil data that are relatively easy to measure were presented by Vereecken *et al.* (1988) and Wösten and van Genuchten (1988). The different methods mentioned differ considerably in complexity and cost. Direct measurement methods, including those based on recently developed parameter estimation procedures, are complicated and costly, while indirect theoretical and statistical methods to estimate hydraulic functions are relatively simple and often quite cost-effective (e.g. Wösten *et al.*, 1989).

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A decision as to which is the appropriate method for a specific application has to be based on a comparison of methods. In principle, a comparison can be based on a statistical anaylsis of the functions themselves (e.g. Baker and Bouma, 1976). However, soil hydraulic functions are not an aim in themselves but serve as input data for simulations. It is therefore attractive to compare the different methods on the basis of practical applications. Differences among results of calculations, each of which being based on use of basic data obtained with one of the methods, allow comparison among methods.

In this study, four different methods, ranging from complex and costly to simple and cheap, were used to generate soil hydraulic functions. These functions are compared by their effect on two practical aspects: (1) calculated evapotranspiration deficit and (2) flux through a plane at 30 cm below soil surface. Evapotranspiration deficit is a relevant property for agricultural use because it is directly related to yield. The flux through a plane at 30 cm depth is relevant for environmental studies that deal with migration of pesticides and fertilizers from the rootzone towards the ground water table. Obviously, downward movement of pollutants can only occur when downward fluxes occur.

Calculations for these two practical aspects were made for three sites in a watershed for a period of seven years. The SWATRE model being used has been validated for this watershed in previous studies (Wösten *et al.*, 1985; Wösten *et al.*, 1989) as well as for other studies (Feddes *et al.*, 1988).

MATERIALS AND METHODS

Soil map

Calculations are made for three sites in the watershed 'Hupselse Beek', which is situated in the eastern part of The Netherlands and has been an experimental study area of the National Dutch Water Service for 20 years. The area covers 650 ha and its altitude varies between 22 and 33 m above mean sea level. Landuse is predominantly agricultural: 80 per cent pasture, 12 per cent arable land, and 8 per cent forest. The area is underlain by Miocene clay sediments starting between 0·2 and 10·0 m below the present soil surface. The area was affected by glaciers which deposited boulder clay (glacial till) in an early Pleistocene period. Later, aeolian sands were deposited over the entire area forming a surface relief that is quite different from the relief of the underlying boulder clay surface. Wösten *et al.* (1985) presented a detailed soil map of the area.

In this study evapotranspiration deficit and flux through a plane at 30 cm depth, were simulated for three sites in two dominating mapping units in the area. Figure 1 shows the soil profiles for the three sites. The soil horizons A, B, and C indicate sandy soil horizons; D1 boulder clay and D2 Miocene clay. According to Soil Survey Staff (1975) sites 1 and 3 are classified as sandy, siliceous, mesic Typic Haplaquods and site 2 as a sandy, siliceous, mesic, Plaggept. Groundwater levels for site 1 range from 0·4 m below soil surface in winter to 1·4 m below soil surface in summer. Ranges for site 2 are from 2·2 m in winter to 2·9 m in summer and for site 3 from 0·6 m in winter to 1·3 m in summer.

Generating soil hydraulic functions

The soil hydraulic functions were generated using four different methods which are described in detail by Wösten *et al.* (1989). In summary these methods are:

A. Direct measurement of hydraulic functions of the horizons of every soil profile. Measurements were made using a combination of the following four techniques: (1) the columm method (Bouma, 1982) for the vertical saturated hydraulic conductivity (K_s); (2) the crust-test (latest version of the method reported by Bouma *et al.*, 1983) for unsaturated conductivities when the pressure head, h, is between 0 and -50 cm; (3) the sorptivity method (Dirksen, 1979) for conductivities in the case of sand when $h < -50$ cm and; (4) the hot air method (Arya *et al.*, 1975) for conductivities in the case of loam and clay when $h < -50$ cm. Soil water retention was obtained by slow evaporation of wet undisturbed samples in the laboratory as reported by Bouma *et al.* (1983). In these samples, pressure heads are periodically measured with transducer-tensiometers while at the same time subsamples are taken to determine water contents, thus yielding points relating h and θ. Water contents corresponding with pressure heads lower than -800 cm were obtained by conventional methods using air pressure (Richards, 1965). A set of 13 measured soil water retention and hydraulic

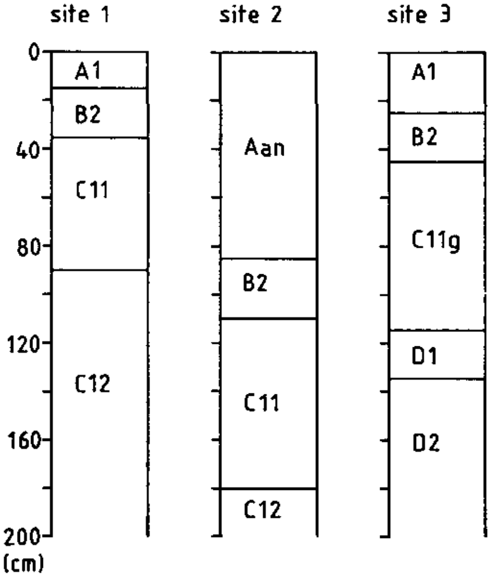

Figure 1. Soil profiles at the three sites

conductivity curves were used in method A to provide model input for all horizons of the three sites as illustrated in Figure 1.

B. Measured hydraulic functions averaged on a regional scale. Using the combination of different methods, hydraulic functions of all soil horizons distinguished in the Hupselse Beek watershed were measured in sixfold. Sampling locations were chosen at random within the various delineated areas of the soil map. Replicate measurements for each horizon were used to obtain the average hydraulic functions for every horizon. After comparison of the hydraulic functions of the different horizons only those horizons were distinguished whose soil hydraulic functions differed significantly. Their number was lower than the number of horizons being distinguished on the basis of pedological classification, because pedological differences do not necessarily correspond with differences in hydraulic functions.

Hydraulic functions averaged on a regional scale were used in method B to obtain soil physical data for the soil profiles at the three sites. A set of four different soil water retention and hydraulic conductivity curves was distinguished: A1, Aan, B2 + C11 + C12 + C11g, and D1 + D2 (Wösten *et al.*, 1985).

C. Measured hydraulic functions averaged on a national scale. Using the combination of different methods, hydraulic functions have been measured for a large number of soil horizons in the Netherlands. For application on a national scale, soils were classified according to soil texture and type of horizon. This classification resulted in 20 different soil groups comprising a total of 197 individual curves. Tabulated forms of the averaged curves for the 20 soil groups were presented by Wösten *et al.* (1987). This set is increasingly being used to simulate regional soil water regimes. In method C, the averaged curves for three soil groups from the national set are used to provide hydraulic functions for the A1 + Aan, B2 + C11 + C12 + C11g, and D1 + D2 horizons.

D. Use of soil properties to predict hydraulic functions. Analytical expressions of van Genuchten (1980) for the water retention and hydraulic conductivity curves were fitted simultaneously to the set of measured hydraulic functions described under method C. Wösten and van Genuchten (1988) used regression analysis to relate estimated model parameters to more easily measured soil properties, such as bulk density and percentages of silt, clay, and organic matter. The resulting functions predict the hydraulic parameters for all horizons shown in Figure 1 using their texture values.

Methods B, C, and D qualify as pedotransfer functions which relate different land and soil characteristics with one another and to land qualities. Methods B and C can be described in terms of class pedotransfer functions, relating soil horizons to associated hydraulic functions. Method D can be described in terms of a continuous pedotransfer function, relating, for example, soil texture, bulk density, and percentage organic matter to associated hydraulic functions (Bouma and van Lanen, 1987; Bouma, 1989).

The four methods to generate soil hydraulic functions differ considerably in terms of costs. Method A in which the hydraulic functions of specific soil horizons are directly measured, implies a very high investment per unit area as compared with methods B, C, and D. After making a major initial investment for the determination of a standard set of measurements, methods B, C, and D can be executed at relatively low costs.

Analysis of meteorological data

The meteorological station of the Hupselse Beek watershed provided daily precipitation (N) and daily potential evapotranspiration (E_{pot}) values for the period March 1976 until December 1982. Potential evapotranspiration values were calculated according to Thom and Oliver (1977). Table I summarizes the meteorological data for the summer half years (April–September) of the seven year period. Rainfall deficits ($E_{pot} - N$) are also expressed in terms of a probability (P) of exceeding these deficits thereby indicating the degree of drought of the summer half years. Table I shows that the summer of 1976 was very dry with a probability of exceeding its rainfall deficit of only 2 per cent; the summer of 1982 was relatively dry as well.

Meteorological data do affect the fluxes and deficits calculated in this study. In wet months actual evapotranspiration equals potential evapotranspiration and no evapotranspiration deficit will develop. In dry months actual values may be lower than potential values, creating an evapotranspiration deficit. In wet months the flux through a plane 30 cm depth will be fairly independent of the soil hydraulic functions and will equal the rainfall surplus. In dry months the flux gives an indication of the amount of capillary rise from the ground water to the rootzone.

Figure 2 shows rainfall deficits, defined as potential evapotranspiration minus rainfall, for different months of the calculation period. Rainfall deficits are positive for many of the summer months and negative for the winter months, indicating a rainfall surplus. The years 1976 and 1982 show considerable rainfall deficits for many of the summer months.

Simulation of practical aspects of soil behaviour

Soil water flow was simulated with the model SWATRE (Soil Water Actual Transpiration Extended) (Feddes *et al.*, 1978; Belmans *et al.*, 1983). SWATRE is a one-dimensional model which describes transient, unsaturated water flow in a heterogeneous soil-root system. In this study, soil physical input data for every horizon of the three sites were obtained with the four different methods.

Table I. Meteorological data of the 'Hupselse Beek' area for the summer half years (April–September) 1976–1982

Year	E_{pot} (mm)	N (mm)	$E_{pot} - N$ (mm)	P (%)
1976	504	214	291	2
1977	364	392	−28	77
1978	374	355	19	63
1979	382	367	15	64
1980	414	405	9	67
1981	397	313	84	40
1982	444	254	190	12

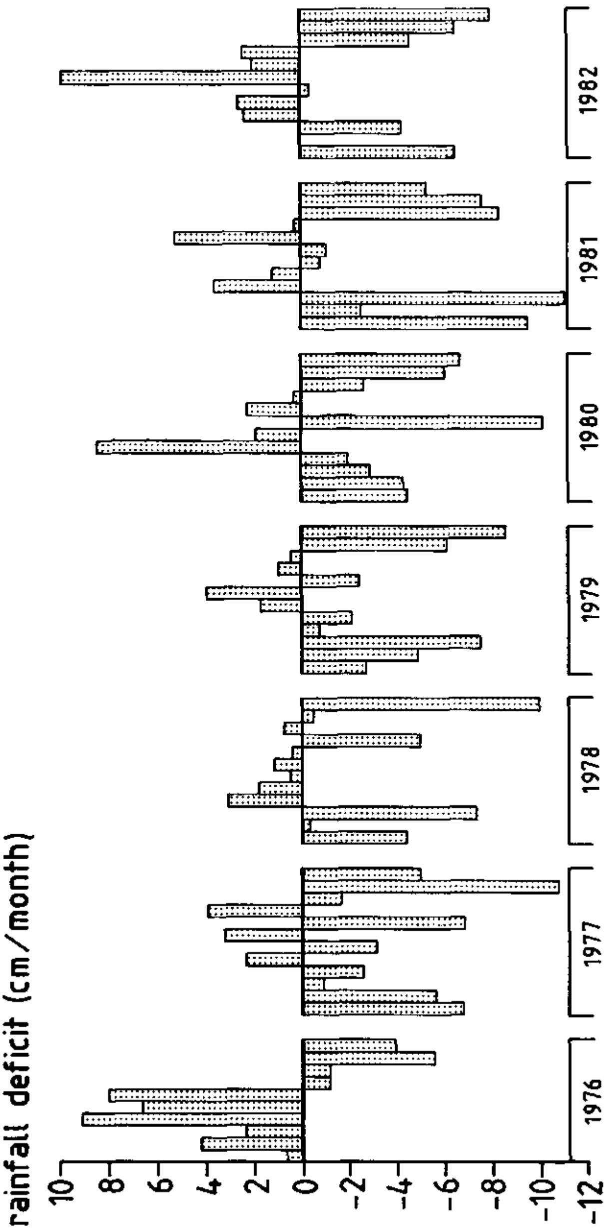

Figure 2. Rainfall deficits for the different months of the calculation period

Calculations were made for grass using a sink term to simulate water uptake by plant roots. Rooting depth for site 1 was 20 cm, for site 2 and 3 depths are 25 cm. These depths were kept constant during the whole year.

Precipitation and potential evapotranspiration serve as upper boundary conditions for the SWATRE simulations. Groundwater levels for site 1, which is located at the meteorological station, were measured daily. Groundwater levels for sites 2 and 3 were measured every week. After correlation of the weekly levels for sites 2 and 3 with the daily levels of site 1, daily groundwater levels for sites 2 and 3 were obtained by estimation. These daily groundwater levels for the period March 1976 until December 1982, serve as lower boundary condition for the simulations.

The evapotranspiration deficit is calculated as the difference between potential evapotranspiration, which is an input parameter, and calculated actual evapotranspiration. The flux is calculated through a plane at 30 cm below soil surface which is 5 cm below the deepest root zone of the three sites. Both evapotranspiration deficit and flux were calculated with the SWATRE model for an uninterrupted seven year period from March 1976 until December 1982. The model generates daily output for the two aspects. However, in order to present more general trends rather than fluctuating short-time results, the calculated deficits and fluxes are presented in this paper in terms of total monthly values.

Analysis of differences between methods to generate soil hydraulic functions

Comparison among the four methods to generate soil hydraulic functions was based on the analysis of differences among simulated deficits and fluxes. An analysis of variance (ANOVA) was carried out to reveal whether these differences were statistically significant.

Calculated monthly values for deficits and fluxes show a clear seasonal variation (Figures 4 and 6 respectively). In this case, deficits and fluxes calculated for one particular month cannot be considered to be independent of the value calculated for other months in the same year. However, when deficits and fluxes are calculated as an average monthly value for a season, the values can be assumed to be independent.

Based on the monthly values for the period 1976 to 1982 an average monthly deficit was calculated for the season from May until September for each year. For the months October until April no deficits occurred. Average monthly fluxes were calculated for the season from April until September and for the season from October until March. As a result, ANOVA of the deficits and of the two fluxes was carried out on values for three sites, four methods, and seven seasonal repetitions. This yielded a total of 84 values for each of the three properties.

Use of a covariable offers the opportunity to account for effects of other factors than differences in methods on simulated deficits and fluxes. Meteorological data strongly affected the two simulated values. Therefore, rainfall deficit was used as a covariable in this study. This covariable partly explained the variability of the two simulated values. The remaining variability was assigned to differences between the four methods to generate soil hydraulic functions. After elimination of meteorological influences by using a covariable, calculated monthly deficits and fluxes could be used. As a result, ANOVA of the deficits was carried out on the 420 values calculated for three sites, four methods, five months, and seven years. For the fluxes in the months April until September, 504 values calculated for three sites, four methods, six months, and seven years were analysed. Calculations started in March 1976, therefore, 504 minus the 24 values calculated for three sites, four methods, and two months were analysed for the fluxes in the months October until March.

RESULTS AND DISCUSSION

Evapotranspiration deficits

Figure 3 shows monthly evapotranspiration deficits per method averaed over the period 1976–1982 and averaged over the three sites. In Figure 4 the same evapotranspiration deficits are shown but now averaged over only the dry years 1976 and 1982.

The figures indicate that evapotranspiration deficits do occur in the months May until September. Outside this period actual evapotranspiration equals potential evapotranspiration. The order of decrease in calculated evapotranspiration deficits is in most cases from methods A to C to B and to D.

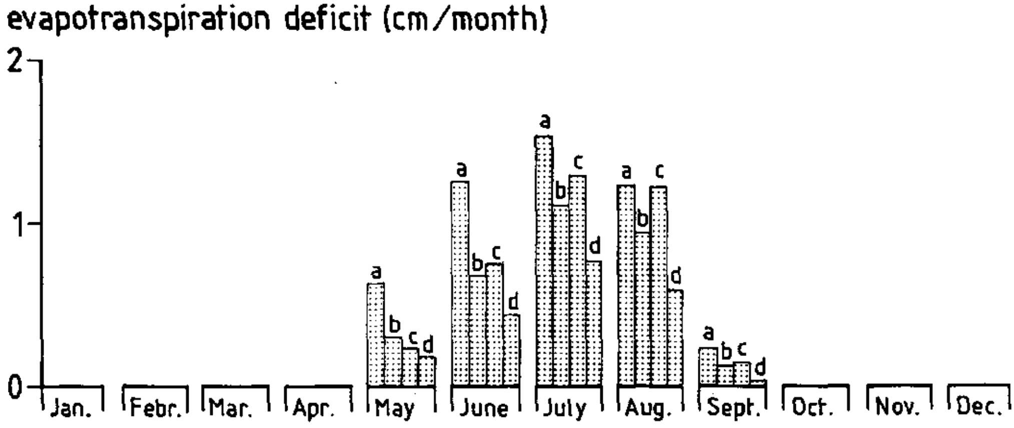

Figure 3. Monthly evapotranspiration deficits per method averaged over the period 1976 to 1982 and averaged over the three sites

The figures indicate that when calculation of the monthly evapotranspiration deficit is restricted to dry years this deficit is about three times larger than when all data of the seven years are used. Differences in evapotranspiration deficits calculated with methods A and D for the growing seasons (May–September) of the dry years 1976 and 1982 are 92 mm and 52 mm respectively. Under Dutch conditions 1 mm evapotranspiration deficit is equivalent to a net loss of $4 per hectare per year (van Boheemen, 1981). Therefore, in a very dry year such as 1976 (with a probability of only 2 per cent of exceeding its rainfall deficit) use of method A or D would make a difference of $368 per hectare per year. In 1982 (with a probability of 12 per cent of exceeding its rainfall deficit) using method A or D amounts to a difference of $208 per hectare per year. These amounts are substantial. The user has to decide whether the difference is large enough to justify expensive measurements *in situ* (method A) which are the most reliable.

Evapotranspiration deficits were statistically analysed by using the 84 values of the average monthly deficits calculated for the season from May until September. When a covariable was used, the 420 values of the months May until September were analysed.

Table II shows the results of the analysis without and with the use of a covariable. The high variance ratio for the covariable indicates that rainfall deficit strongly influences the calculated deficits. The variance ratio, or F-statistic, shows that differences among methods are not significant at an $\alpha = 0{\cdot}01$ level when no covariable is used. However, differences between methods are significant when rainfall deficit is used as a covariable. In this case, the statistical analysis reveals also that deficits calculated with each method are significantly different from defícts calculated with each of the other methods.

When comparing evapotranspiration deficits among sites, values are highest for site 2 followed by sites 1 and 3. The deep water table of more than 3 m below soil surface at site 2, prevents water supply by capillary rise and therefore evapotranspiration depends completely on the limited amount of available water in the rootzone. The somewhat deeper ground water table in summer and the shallower rootzone of site 1 as compared to site 3 make evapotranspiration deficits for site 1 higher than deficits for site 3.

Flux through a plane at 30 cm depth

A similar procedure as for the evapotranspiration deficit was followed for the practical aspect: flux through a plane at 30 cm below soil surface. Figure 5 shows monthly fluxes averaged per method over the period 1976 to 1982 and averaged over the three sites. Figure 6 shows the same fluxes but now averaged over only the dry years 1976 and 1982. Negative fluxes indicate downward and positive indicate upward movement of water. Figures 5 and 6 show that downward fluxes in winter months are not affected by using either data of the

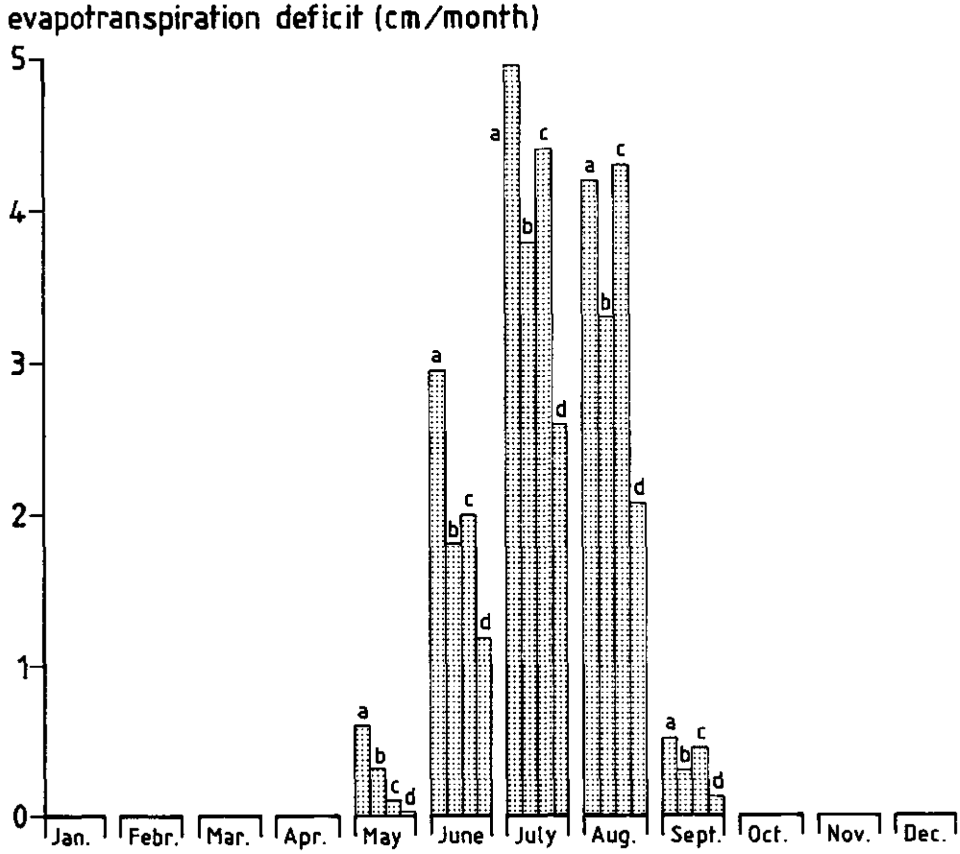

Figure 4. Monthly evapotranspiration deficits per method averaged over the years 1976 and 1982 and averaged over the three sites

Table II. Statistical analysis of evapotranspiration deficits calculated with four methods for three sites for the seasons May until September of the years 1976 to 1982

Source of variation	Degrees of freedom	Mean squares	Variance ratio
Without covariable			
Method	3	1·22	1·01
Site	2	4·36	3·62
Residual	78	1·21	
With rainfall deficit as covariable			
Method	3	6·01	3·82*
Site	2	19·81	12·61*
Covariable	1	289·72	184·45*
Residual	413	1·57	

* Significant at an $\alpha = 0{\cdot}01$ level

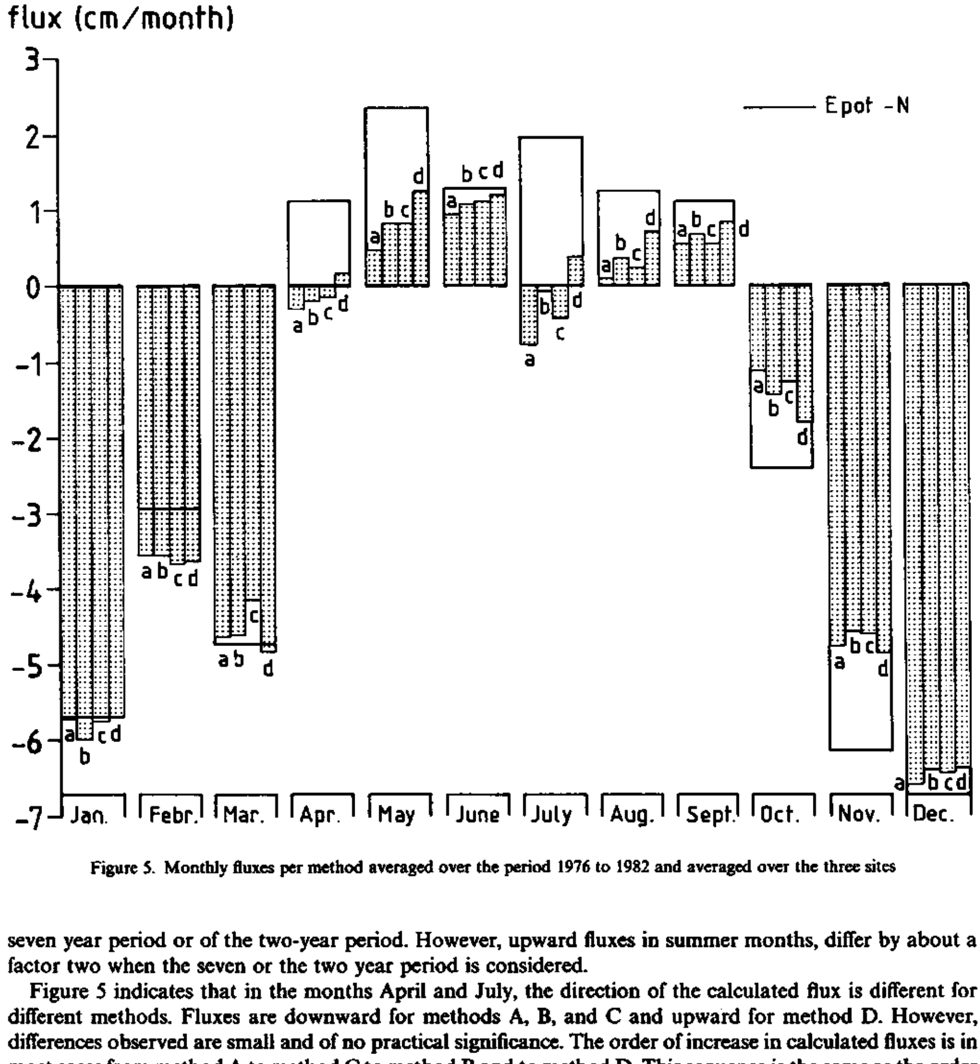

Figure 5. Monthly fluxes per method averaged over the period 1976 to 1982 and averaged over the three sites

seven year period or of the two-year period. However, upward fluxes in summer months, differ by about a factor two when the seven or the two year period is considered.

Figure 5 indicates that in the months April and July, the direction of the calculated flux is different for different methods. Fluxes are downward for methods A, B, and C and upward for method D. However, differences observed are small and of no practical significance. The order of increase in calculated fluxes is in most cases from method A to method C to method B and to method D. This sequence is the same as the order of decrease in evapotranspiration deficits (compare Figures 6 and 4). This may be expected since a high upward flux implies a relatively good water supply and, therefore, a relatively low evapotranspiration deficit.

Values for the rainfall deficit which are included in Figure 5, indicate that deficits in the winter months January, March, and December, when actual evapotranspiration equals potential evapotranspiration, virtually correspond with downward fluxes. In the summer months April until September, rainfall deficits are higher than upward fluxes indicating that other sources such as changes in water storage of the upper 30 cm of the soil contribute to water supply. At the same time, a water deficit might occur when upward fluxes together with change in water storage are not sufficient to maintain actual evapotranspiration at the potential level. In October and November, when the downward flux is less than the rainfall surplus, part of

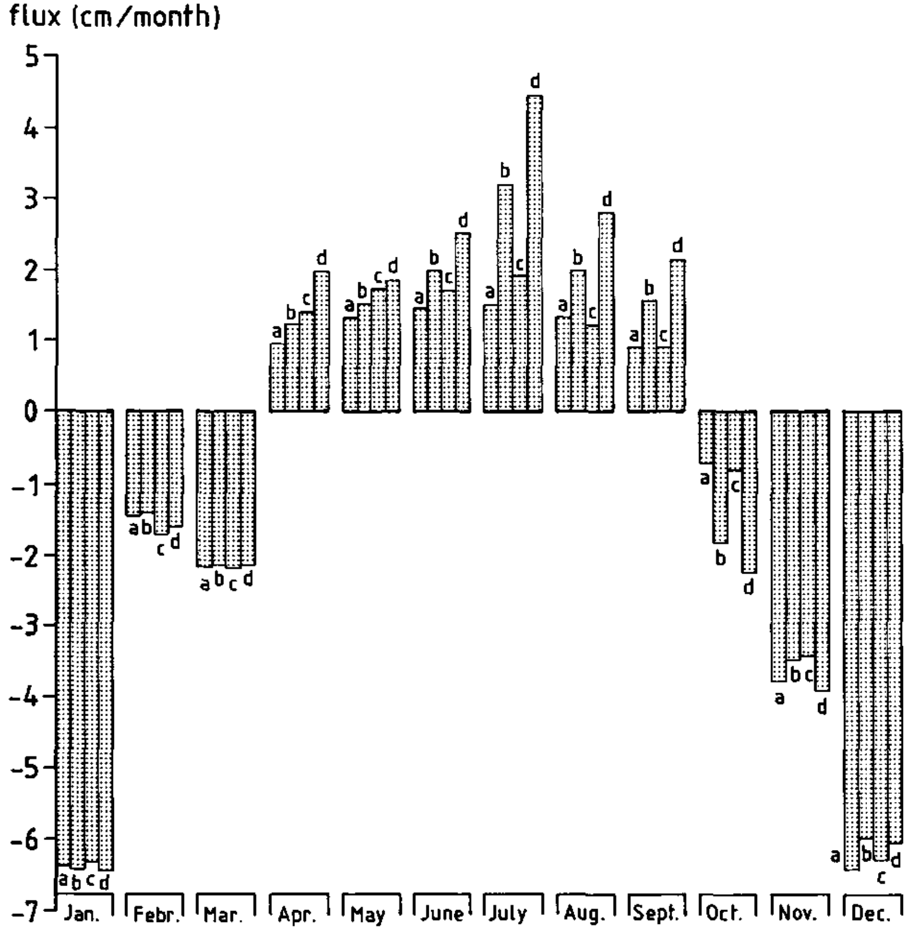

Figure 6. Monthly fluxes per method averaged over the years 1976 and 1982 and averaged over the three sites

the surplus is used to wet the upper 30 cm of the soil. A higher downward flux as compared to rainfall surplus in February is caused by draining of the originally saturated upper 30 cm.

Similar to evapotranspiration deficits, differences in fluxes calculated for the period 1976 to 1982 were statistically analysed by using the 84 average monthly fluxes for the season from April until September (which are dominantly upward) and by using the 84 average monthly fluxes for the season from October until March which are downward only. When a covariable was used, the 504 values of the months April until September and the 480 values of the months October until March were analysed.

Table III shows the results for fluxes of the months April until September without and with the use of a covariable. As with the evapotranspiration deficit, the high variance ratio for the covariable indicates that rainfall deficit strongly influences the calculated fluxes. The variance ratio shows that differences among methods are not significant at an $\alpha = 0{\cdot}01$ level when no covariable is used. However, differences between

Table III. Statistical analysis of fluxes calculated with four methods for three sites for the seasons April until September of the years 1976 to 1982

Source of variation	Degrees of freedom	Mean squares	Variance ratio
Without covariable			
Method	3	1·24	0·96
Site	2	3·42	2·63
Residual	78	1·30	
With rainfall deficit as covariable			
Method	3	7·56	3·80*
Site	2	20·53	10·33*
Covariable	1	2136·99	1074·96*
Residual	497	1·99	

*Significant at an $\alpha = 0{\cdot}01$ level

methods are significant when rainfall deficit is used as a covariable. The statistical analysis reveals also that, in this case, fluxes calculated with each method are signicantly different from fluxes calculated with each of the other methods.

ANOVA of fluxes of the months October until March, without and with using rainfall deficit as a covariable, shows in both cases no significant differences among methods for these winter months when fluxes generally equal rainfall surplusses. This conclusion is of most practical significance because downward fluxes occur in these winter periods and they may be associated with groundwater pollution if the water carries pollutants that are not being adsorbed.

Differences in upward fluxes between sites are again a reflection of differences in ground water table (e.g. site 1 and 3 shallow, site 2 deep).

SUMMARY AND CONCLUSIONS

Four different methods to generate soil hydraulic functions were evaluated in terms of their effect on calculation of two practical aspects of soil behaviour: evapotranspiration deficit and flux through a plane at 30 cm depth. These two aspects cannot directly be measured and were therefore calculated with a validated simulation model. Calculations were made for three sites for a seven-year period from 1976 to 1982.

Differences among the four methods to generate soil hydraulic functions in terms of calculated deficits and fluxes were not significant. This result agrees with findings of the study by Wösten *et al.* (1989) in which the same four methods were compared in terms of calculated water storages of the upper 50 cm of the same soil profiles at the three sites. These calculations could be validated by measured water contents using neutron probes.

Meteorological data strongly affect calculated evapotranspiration deficits and fluxes. An analysis of results using rainfall deficit as a covariable, indicated that the four methods differ significantly. Largest differences in evapotranspiration deficits are calculated when comparing results of method A and D, where method A (measured on-site values) can be considered as a reference. These differences amount to a maximal financial difference of $368 per hectare per year for the growing season of the very dry year 1976. Also largest differences in fluxes are calculated using method A and D, but these differences are difficult to express in financial terms. The average monthly upward flux calculated with method A for the month July of the dry years 1976 and 1982 is 1·51 cm. Using method D an upward flux of 4·46 cm per month is calculted. However, differences among fluxes for the wet period of the year (October to March) are not significant and this is important because downward transport of pollutants occurs in this period.

The user has to decide which method to use when obtaining hydraulic functions. The short- and long-term costs and benefits have to be balanced on the basis of the degree of accuracy required for the results. When considering the important downward fluxes in the period October to March, methods yield comparable results. In this case, it would be advisable to make a major investment in a program that results in a national standard series of hydraulic functions as described for method C. Once the standard series is available, estimates can be made very rapidly for each new site.

This study demonstrates the magnitude of differences obtained when using methods A, B, C, and D to calculate fluxes and deficits, while method A is taken as a reference. The user has to decide whether these differences are large enough to justify making very expensive on-site measurements.

ACKNOWLEDGEMENTS

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REFERENCES

Arya, L. M., Farrell, D. A., and Blake, G. R. 1975. 'A field study of soil water depletion patterns in presence of growing soybean roots. I. Determination of hydraulic properties of the soil', *Soil Sci. Soc. Am. Proc.*, **39**, 424–430.

Baker, F. G. and Bouma, J. 1976. 'Variability of hydraulic conductivity in two subsurface horizons of two silt loam soil', *Soil Sci. Soc. Am. J.*, **40**, 219-222.

Belmans, C., Wesseling, J. G., and Feddes, R. A. 1983. 'Simulation model of the water balance of a cropped soil: SWATRE'. *J. Hydrol.*, **63**, 271-286.

Bouma, J. 1982. 'Measuring the hydraulic conductivity of soil horizons with continuous macropores', *Soil Sci. Soc. Am. J.*, **46**, 438-441.

Bouma, J. 1983. 'Use of soil survey data to select measurement techniques for hydraulic conductivity', *Agric. Water Manage.*, **6**, 177-190.

Bouma, J. 1989. 'Land qualities in space and time', in Bouma, J. and Bregt, A. K. (Eds), *Proceedings of the ISSS Symposium on Land Qualities in Space and Time.*, Pudoc, Wageningen, The Netherlands, 22-26 Aug. 1988, 3-13.

Bouma, J., Belmans, C., Dekker, L. W., and Jeurissen, W. J. M. 1983. 'Assessing the suitability of soils with macropores for subsurface liquid waste disposal', *J. Environ. Qual.*, **12**, 305-311.

Bouma, J. and van Lanen, H. A. J. 1987. 'Transfer functions and threshold values: from soil characteristics to land qualities', in Beek, K. J., Burrough, P. A., and McCormack, D. E. (Eds), *Proceedings of the Workshop by ISSS/SSSA on Quantified Land Evaluation Procedures*, ITC publication No. 6, Enschede, The Netherlands, 106-111.

Dirksen, C. 1979. 'Flux-controlled sorptivity measurements to determine soil hydraulic property functions', *Soil Sci. Soc. Am. J.*, **43**, 827-834.

Feddes, R. A., de Graaf, M., Bouma, J., van Loon, C. D. 1988. 'Simulation of water use and production of potatoes as affected by soil compaction', *Potato Research*, **31**, 225-239.

Feddes, R. A., Kowalik, P. J., and Zaradny, H. 1978. 'Simulation of field water use and crop yield' Simulation Monograph, Pudoc, Wageningen, The Netherlands, 189 pp.

Green, R. E., Ahuja, L. R., and Chong, S. K. 1986. 'Hydraulic conductivity, diffusivity, and sorptivity of unsaturated soils: field methods' in Klute, A. (Ed.), *Methods of Soil Analysis*, 1. *Physical and Mineralogical Methods. Agronomy*, **9**, 771-798, 2nd ed., Am. Soc. Agron. Madison, Wisc.

Klute, A. and Dirksen, C. 1986. 'Hydraulic conductivity and diffusivity: Laboratory Methods', in Klute, A. (Ed.), *Methods of Soil Analysis*, 1. *Physical and Mineralogical Methods. Agronomy*, **9**, 2nd ed., Am. Soc. Agron. Madison, Wisc.

Kool, J. B., Parker, J. C., and van Genuchten, M.Th. 1987. 'Parameter estimation for unsaturated flow and transport models. A review', *J. Hydrol.*, **91**, 255-293.

Mualem, Y. 1986. 'Hydraulic conductivity of unsaturated soils: Prediction and formulas' in Klute, A. (Ed.), *Methods of Soil Analysis*, 1. *Physical and Mineralogical Methods. Agronomy*, **9**, 799-823, 2nd ed., Am. Soc. Agron. Madison, Wisc.

Richards, L. A. 1965. 'Physical condition of water in soil', in Black, C. A. *et al.* (Eds), *Methods of Soil Analysis. Agronomy*, **9**, 131-137.

Soil Survey Staff 1975. 'Soil taxonomy. A basic system of soil classification for making and interpreting soil surveys', *USDA-SCS Agric. Handb.*, **436**, U.S. Government Printing Office, Washington, DC.

Thom, A. S. and Oliver, H. R. 1977. 'On Penman's equation for estimating regional evaporation', *Q.J.R. Meteorol. Soc.*, **103**, 345-357.

Van Boheemen, P. J. M. 1981. 'Increase in grassland production as affected by improvement in water management', *Report nr. 1298 of the Institute for Land and Water Research*, Wageningen, The Netherlands (in Dutch).

Van Genuchten, M.Th. 1980. 'A closed-form equation for predicting the hydraulic properties of unsaturated soils', *Soil Sci, Soc. Am. J.*, **44**, 892-898.

Vereecken, H., Maes, J., and Feyen, J. 1988. 'Estimating the unsaturated hydraulic conductivity from easy to measure soil properties', *Soil Science* (in press).

Wösten, J. H. M., Bannink, M. H., and Beuving, J. 1987. 'Water retention and hydraulic conductivity characteristics of top- and sub-soils in the Netherlands: The Staring series', *Report 1932, Soil Survey Inst.*, Wageningen, The Netherlands (in Dutch).

Wösten, J. H. M., Bouma, J., and Stoffelsen, G. H. 1985. 'Use of soil survey data for regional soil water simulation models', *Soil Sci. Soc. Am. J.*, **49**, 1238-1244.

Wösten, J. H. M., Schuren, C. H. J. E., Bouma, J., and Stein, A. 1989. *Soil Sci. Soc. Am. J.* (in press).

Wösten, J. H. M. and van Genuchten, M.Th. 1988. 'Using texture and other soil properties to predict the unsaturated soil hydraulic functions', *Soil Sci. Soc. Am. J.*, **52**, 1762-1770.

8. USE OF SCALING TECHNIQUES TO QUANTIFY VARIABILITY IN HYDRAULIC FUNCTIONS OF SOILS IN THE NETHERLANDS

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USE OF SCALING TECHNIQUES TO QUANTIFY VARIABILITY IN HYDRAULIC FUNCTIONS OF SOILS IN THE NETHERLANDS

J.H.M. Wösten

Quantification of variability in hydraulic functions of different soils allowed calculation of the variability in output of models which use these functions as input. Variability of an output variable defines the uncertainty of a particular calculation and is therefore indispensable for applications which need a measure of accuracy. Meaningful description of variability requires a minimum number of measured hydraulic functions. Defined accuracies of calculated functional criteria which are practical aspects of soil behavior, were used as references to classify soils, for which insufficient hydraulic functions are known in larger soil groups. For the Netherlands this classification resulted in three large soil groups; coarse-textured, medium-textured and fine-textured, each comprising at least 30 measured functions. Scaling was successfully used to reduce the variation in measured hydraulic functions into a narrow band around the scaled mean hydraulic function for each soil group. Distribution functions of scale factors were used to transform the scaled mean hydraulic conductivity and moisture retention functions into 100 new hydraulic functions for each soil group. Variability in measured and in newly generated functions was compared on the basis of calculated functional criteria. Mean values of functional criteria are close for the measured and newly generated sets of functions. The percentage of variation in functional criteria explained by the newly generated hydraulic functions varied from 15% to 96%. Although scaling is an attractive technique to simplify description of the complex hydraulic heterogeneity of soils in the Netherlands, it underestimates the variability.

1. INTRODUCTION

Soil water retention and hydraulic conductivity functions are crucial ingredients for the estimation of rates of soil water flow. During the past decade hydraulic functions have been measured for a wide variety of soils in The Netherlands. The 197 measured functions were classified into 20 different soil groups according to soil texture (as used by the former Netherlands Soil Survey Institute), and type of horizon, being either topsoil (A horizon) or subsoil (B and C horizons). Tabulated forms of the geometrically averaged functions for the 20 soil groups were presented by Wösten et al. (1987). These average functions facilitate soil physical interpretations of soil maps which are relevant for land evaluation and environmental and hydrological studies that require soil physical information for the unsaturated zone of areas of land (e.g. Wösten et al., 1985; Breeuwsma et al., 1986).

Individual functions within each of the 20 soil groups show considerable variability due to differences in soil texture, organic matter content and bulk density. Similar variability in hydraulic functions measured in the field is reported by other investigators (e.g. Nielsen et al., 1973; Hopmans, 1987). Until now, existing variability in hydraulic functions of soil groups distinguished in the Netherlands has not been quantified. Consequently model output, so far, does not reflect the effects of this variability. Presentation of model output in the form of a mean and variance becomes increasingly attractive because it reflects and quantifies the uncertainty of a particular calculation. In order to provide this measure of accuracy of model output, quantification of variability in hydraulic functions is a necessity (e.g. Bresler and Dagan, 1979; Van der Zee and Van Riemsdijk, 1987).

Quantification of variability in hydraulic functions of a soil group is only meaningful if a minimum number of functions is measured for that particular soil group. Warrick et al. (1977a) showed that the accuracy of estimated mean soil water fluxes strongly increased with increasing number of samples. If in turn a certain accuracy in calculated functional criteria is taken as a reference, the number of samples can be determined. Soil groups containing insufficient samples are combined into larger, new groups.

Once sufficiently large soil groups are established, variability in hydraulic functions is quantified with the scaling theory of similar media. The theory of scaling (Miller and Miller, 1955, 1956) assumes that similar media differ only in the scale of their internal microscopic geometries, and therefore have equal porosities. Scaling relates the hydraulic functions $h(\theta)$ and $K(\theta)$ for soils at different locations to a representative mean through a single stochastic variable, the scale factor. The complexity of variability in hydraulic functions is thus simplified to the description of distribution functions of scale factors. Methods to determine scale factors are described by Warrick et al. (1977b), Simmons et al. (1979) and Russo and Bresler (1980). Sharma and Luxmoore (1979) and Clapp et al. (1983) used scaling techniques to investigate the practical consequences of soil heterogeneity on water budget components in a watershed. Similar to studies by Ahuja et al. (1984), and Hopmans and Stricker (1989), distribution functions of scale factors are used, in this study, to generate a new set of hydraulic functions for each soil group. Variability in measured and newly generated hydraulic functions is compared by comparing calculated functional criteria which relate to practical applications.

The specific objectives of this study were hence (1) to form soil groups that contained sufficient individual hydraulic functions to allow quantification of variability in these functions, (2) to apply scaling as a technique for quantification of variability within each soil group, and (3) to examine the effectiveness of scaling in describing variability in measured hydraulic data.

2. MATERIALS AND METHODS

A total of 197 water retention and hydraulic conductivity functions were measured for different soils in the Netherlands. The measured functions for the soils, which were classified in 20 different soil groups, form a unique data base covering a broad spectrum of soils.
Hydraulic conductivities were measured using a combination of the following five methods :

(1) The column method (e.g., Bouma, 1982) for the vertical saturated hydraulic conductivity, K_S.
(2) The crust-test (Bouma et al., 1983) for unsaturated conductivities when the pressure head, h, is between 0 and -50 cm.
(3) The sorptivity-method (Dirksen, 1979) for conductivities of coarse-textured soils when h < -50 cm.
(4) The hot-air method (Arya et al., 1975) for conductivities of medium- and fine-textured soils when h < -50 cm.
(5) The evaporation method (Boels et al., 1978) for hydraulic conductivities when h is between 0 and -800 cm.

Soil water retention functions were obtained by slow evaporation of wet, undisturbed samples in the laboratory as reported by Boels et al. (1978) and Bouma et al. (1983). In this method, pressure heads were periodically measured with transducer-tensiometers while at the same time subsamples were taken to determine water contents, thus yielding points relating to h and θ. Water contents corresponding with pressure heads lower than -800 cm of water were obtained by conventional methods using air pressure (Richards, 1965). For relatively fine-textured soils, a staining technique was applied to record the effects of horizontal cracks on the upward flux of water from the water table to the rootzone (Bouma, 1984).
The measured water retention and hydraulic conductivity functions were fitted in a previous study (Wösten and Van Genuchten, 1988) with the following closed-form analytical expressions of Van Genuchten (1980):

$$S = \frac{\theta - \theta_r}{\theta_s - \theta_r} = \left[1 + |\alpha h|^{n}\right]^{-m} \tag{1}$$

and

$$K(S) = K_s S^{\iota}\left[1 - \left(1 - S^{1/m}\right)^{m}\right]^2 \tag{2}$$

The parameter S is the degree of saturation. The subscripts r and s refer to residual and saturated values of the volumetric water content θ. K_S is the saturated hydraulic conductivity and α, n, m and ι are parameters which determine the shape of the functions. A modified form of a nonlinear least squares optimization program developed by van Genuchten (1978) was used to estimate the unknown parameters (θ_r, θ_S, α, n, ι and K_S) simultaneously from measured water retention and hydraulic conductivity data. Following Wösten and Van Genuchten (1988), the parameter θ_r was fixed at zero and the parameters θ_s and K_s were fixed at their independently measured values. Consequently, only three unknown parameters (α, n and ι) were needed to fully describe the individual functions. In this study the fitted hydraulic functions were used.

3. DATA ANALYSIS

The minimum number of functions required for a meaningful description of variability in each soil group was estimated using the data of the largest of the presently distinguished 20 soil groups. The selected soil group comprised 49 measured functions and had the following ranges in soil properties: silt 1-9% ; organic matter 0.1-2% ; bulk density 1.4-1.8 g cm^{-3}. The 49 functions were used to calculate three functional criteria which are practical aspects of soil behavior that have immediate relevance to management-type applications (Wösten et al., 1986):

(1) Travel time (T) of water from the soil surface down to a water table assumed to be at a depth (D) of 1 m. The travel time is approximated by $T = \theta D/q_d$, where q_d is the average daily vertical downward flux for Dutch winter conditions (0.14 cm d^{-1}), and θ is the profile-averaged water content corresponding to a flux q_d.θ is derived from the K-θ function.

(2) Depth of water table (L) which can sustain a given upward flux of water to the soil surface or to the bottom of the root zone. As shown by Gardner (1958), this depth may be obtained by integrating Darcy's law to give:

$$L = \int_0^h \frac{1}{1 + q_u/K(h)}\, dh \qquad (3)$$

where h is the imposed soil water pressure head (arbitrarily taken as -500 cm) at the bottom of the root zone, and q_u is the steady-state upward water flux density, assumed to be 0.2 cm d-1.

(3) Downward flux of water (F) corresponding to the minimum soil air content, θ_a (taken to be 5%) which is required to maintain adequate aeration in the root zone for maximum root activity and crop growth (FAO, 1985). Downward flux of water follows immediately from the retention and conductivity function at a water content which is 0.05 cm^3 cm^{-3} less than the saturated value.

In this study emphasis was placed on the accuracy of predicting new values of the functional criteria rather than on the accuracy of predicting mean values. Accuracy of predicting new values, as expressed by the Standard Error of Prediction (SEP), depends on the standard deviation (sd) of the functional criterion and the sample size (n), according to:

$$SEP = sd\sqrt{1 + 1/n}$$

SEP (Montgomery and Peck, 1982) values are used to establish the relation between the width of the 90% prediction interval and sample size. This width indicates that new values of the functional criteria are with 90% certainty within the limits indicated by the width. In turn, the relation between width and sample size can be used to determine the minimum number of samples required to obtain a defined accuracy.

Combination of soil groups containing insufficient hydraulic functions is based on an evaluation of the discriminative power of the two soil properties; "soil texture and type of horizon" presently used to classify the 20 soil groups. Only the most discriminative soil property (e.g. soil texture or type of horizon) was used to re-classify the insufficiently large soil groups into larger, new groups.

Variability in hydraulic functions of the sufficiently large soil groups was quantified with scaling. According to scaling theory the dimensionless scale factor, a_r, is defined as the ratio of the microscopic characteristic length l_r of a soil at location r and the characteristic length l_m of a reference soil:

$$a_r = l_r / l_m \tag{5}$$

If a reference soil water pressure head h_m and conductivity K_m together with the scale factors are known, the water pressure head (h_r) and hydraulic conductivity (K_r) at given water contents at any location r is calculated according to:

$$h_r = h_m / a_r \tag{6}$$

and

$$K_r = K_m a_r^2 \tag{7}$$

Because soils do not have identical porosities, h and K are written as functions of the effective saturation S (Eq. 1) rather than functions of the volumetric water content θ. The saturated hydraulic conductivity K_s is not included in the scaling procedure because it is often controlled by flow in macropores, whereas the unsaturated hydraulic conductivity is controlled by the entire continuum of pore sizes (Jury et al., 1987a). Specific scaling methods differ according to how the reference $h_m(S)$ and $K_m(S)$ functions are determined. Hopmans (1987) concluded from a comparison of five different scaling methods that the method described by Warrick et al. (1977b) was the most promising. Therefore, only the latter scaling method was used in this study.

After establishing the distribution function of the calculated scale factors for each soil group, this distribution function was used, in turn, to generate randomly a set of 100 new scale factors. Next, the new scale factors were used to transform the or reference soil water retention and hydraulic conductivity functions into 100 new hydraulic functions according to Eqns. (6) and (7). Variability in these newly generated hydraulic functions was analysed on the basis of the variability of the calculated functional criteria described earlier. Comparison of the means and variances of functional criteria calculated from the newly generated and the measured hydraulic functions allowed an evaluation of the effectiveness of scaling in describing variability in soil hydraulic data.

When functional criteria and scale factors were normally distributed, the mean (m_1) and variance (sd_1^2) of these criteria and factors were estimated using untransformed data according to:

$$m_1 = \frac{1}{n}\sum_{i=1}^{n} x_i \tag{8}$$

and

$$sd_1^2 = \frac{1}{n-1}\sum_{i=1}^{n} (x_i - m_1)^2 \tag{9}$$

In these equations x_i is the ith observation and n is the sample size. However, if these same criteria were lognormally distributed, then their mean (m_2) and variance ($sd_2{}^2$) was estimated using transformed data according to:

$$m_2 = \exp(\mu + \sigma^2/2) \tag{10}$$

and

$$sd_2^2 = \exp(2\mu+\sigma^2)[\exp(\sigma^2)-1] \tag{11}$$

where

$$\mu = \frac{1}{n}\sum_{i=1}^{n} \ln(x_i) \tag{12}$$

and

$$\sigma^2 = \frac{1}{n-1}\sum_{i=1}^{n} (\ln(x_i) - \mu)^2 \tag{13}$$

The coëfficient of variation (CV) was calculated by dividing the standard deviation (sd) by the mean (m).

4. RESULTS AND DISCUSSION

Relations between half widths of 90% prediction intervals for three functional criteria and sample sizes are shown in Figure 1 for the soil group comprising 49 measured functions. The figure indicates that the accuracy of predicting the functional criteria, as expressed by the half width of the 90% prediction interval, increases with increasing sample size. In this study a difference of 5% between the half width of the 90% prediction interval and the true half width (where SEP $\rightarrow$ sd when $n \rightarrow \infty$) is used as a reference. Figure 1 indicates that, in order to meet this requirement, soil groups should comprise about 30 functions.

Linear regression of calculated functional criteria versus soil texture and type of horizon, for all 197 individual functions, showed that the discriminative power of the soil property; "soil texture" is large compared to the discriminative power of the soil property; "type of horizon". This result of the regression analysis combined with a required sample size of at least 30 functions led

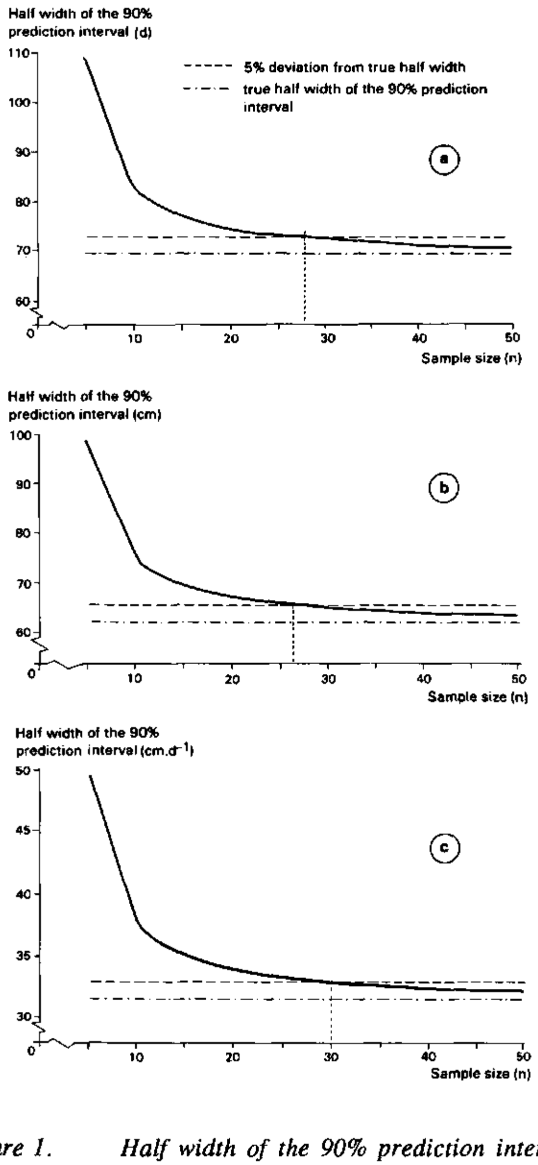

Figure 1. Half width of the 90% prediction interval as a function of the sample size for the functional criteria: a) travel time, b) depth of water table and c) downward flux of water. For the calculation of these data the used soil group comprised 49 measured functions which had the following ranges in soil properties: silt 1-9%; organic matter 0.1-2%; bulk density 1.4-1.8 g cm^{-3}

to consolidation of the original, distinguished 20 soil groups into 3 new, sufficiently large groups; coarse-textured, medium-textured and fine-textured soils. The number of functions and ranges in soil texture, organic matter content, bulk density and median of the sand fraction (M50) for each of the three new soil groups are presented in Table 1.

Table 1. Number of individual functions, ranges in soil texture, organic matter content, bulk density and M50 for three, newly generated soil groups.

soil group	number of functions	silt, 2-50 μm (%)	clay, >2μm (%)	organic matter (%)	bulk density (g/cm^3)	M50[1] (μm)
coarse-textured	105	4-49		0.1-13	1.1.-1.8	130-180
medium-textured	43		9-22	0.2-6	1.2-1.7	
fine-textured	49		26-77	0.1-15	0.9-1.7	

[1] median sand fraction

Figures 2, 3 and 4 show the unscaled and scaled water retention data and hydraulic conductivity data for the coarse-textured, medium-textured and fine-textured soils, respectively. Scaled data resemble the characteristic shapes of hydraulic functions normally measured for sandy, loamy and clay type soils. Percentage reductions in the Sum of Squares (SS) of deviations between scaled mean hydraulic function and individual hydraulic data for each soil group, before and after scaling, are presented in Table 2. A high reduction in SS indicates that scaling was successful. Table 2 and Figures 2, 3 and 4 illustrate that scaling was indeed effective in reducing the scatter of the original data points into a narrow band around the mean scaled functions. Effectiveness was greater for water retention as compared to conductivity data. Because K_s was not included in the scaling procedure, its variation did not decrease after scaling.

The fact that scaling is more effective in reducing variability of the moisture retention function as compared to the hydraulic conductivity function agrees with findings of Warrick et al. (1977b) and Hopmans (1987). This result is attributed to the fact that changes in the degree of saturation (S) have usually a greater effect on K than on h, as well as to the fact that measurement of h is less complicated and therefore more reliable than measurement of K.

The assumption that the three soil groups could be considered as similar media was tested by comparing the scale factors calculated from h data, i.e. $a_r(h)$, with those calculated from K data, i.e. $a_r(K)$, for all three soil groups together. Because individual soil groups are quite variable in soil texture (Table 1), and because no distinction is made between type of horizon it can not be expected that the two sets of scale factors follow exactly the 1:1 line. However, the two sets of scale factors were correlated significantly with R^2 equals 0.74 (R is the correlation coefficient). The slope of the regression line, emanating from the origin was calculated to be 1.02. These results agree with those reported by Warrick et al. (1977b), Simmons et al. (1979), Russo and

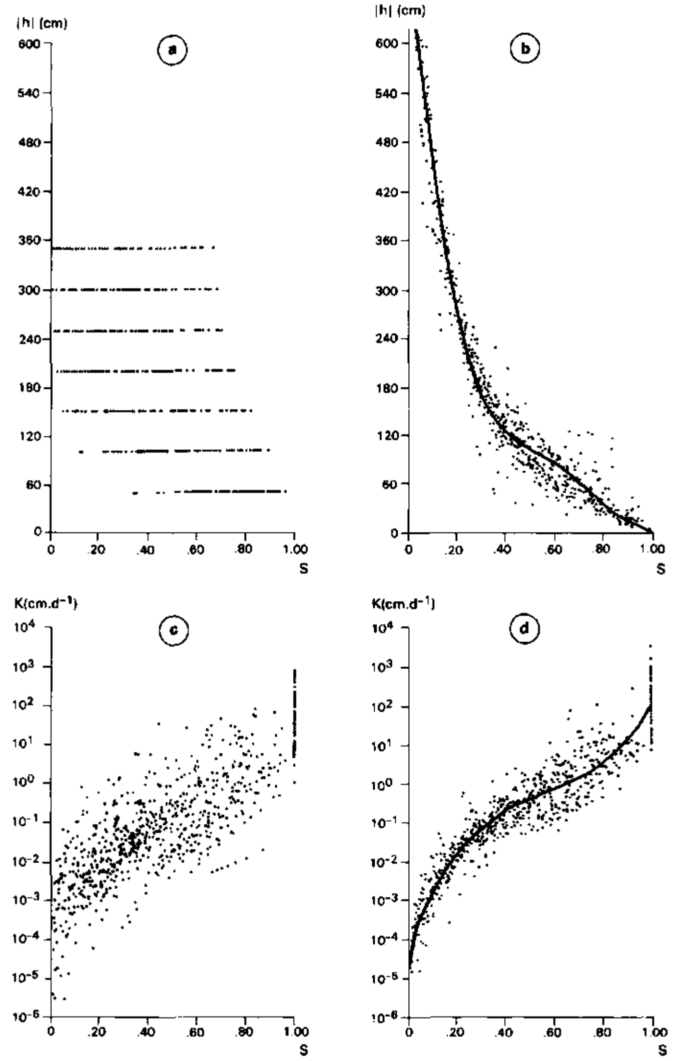

Figure 2. *Unscaled (a) and scaled (b) water retention data and unscaled (c) and scaled (d) hydraulic conductivity data for the coarse-textured soil group.*

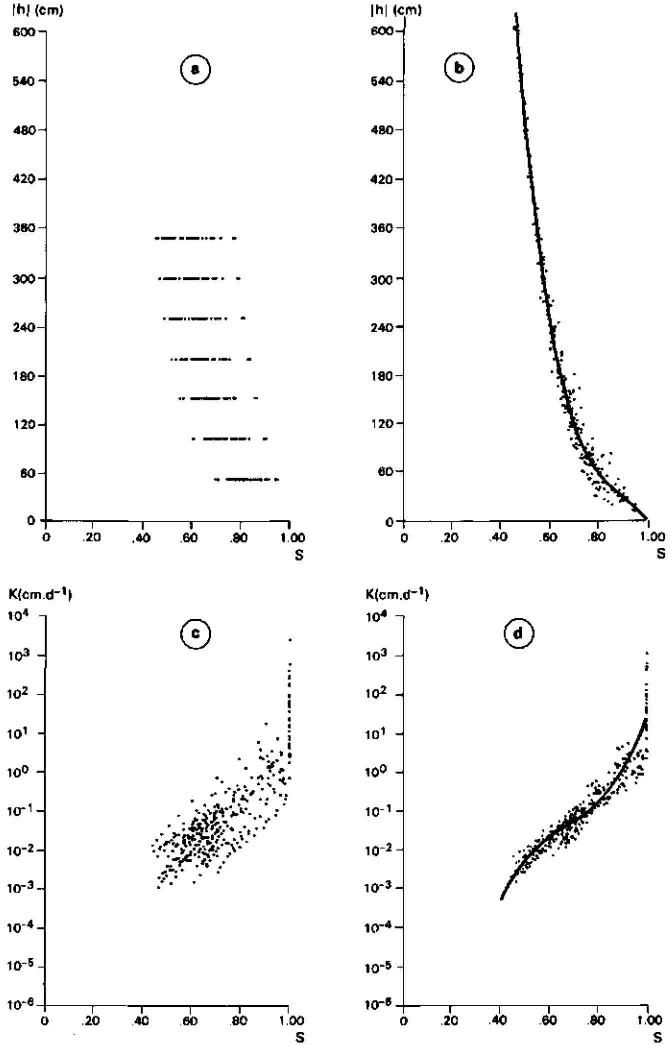

Figure 3. Unscaled (a) and scaled (b) water retention data and unscaled (c) and scaled (d) hydraulic conductivity data for the medium-textured soil group.

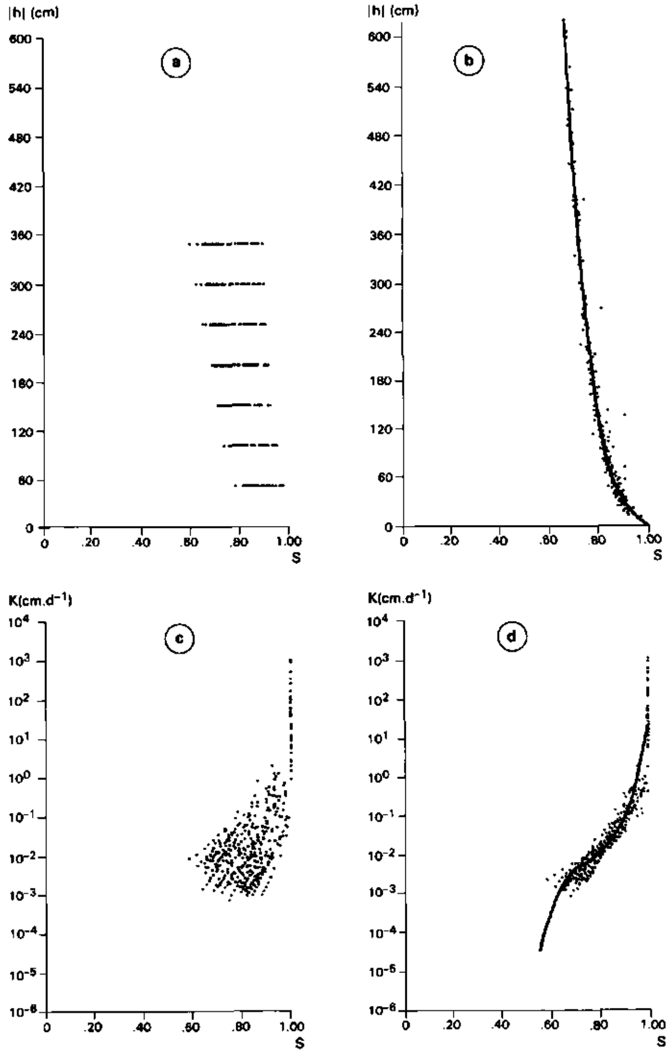

Figure 4. Unscaled (a) and scaled (b) water retention data and unscaled (c) and scaled (d) hydraulic conductivity data for the fine-textured soil group.

Table 2. Percent reduction in sum of squares (SS) of deviations between scaled mean hydraulic functions and individual hydraulic data points before and after scaling, for three soil groups.

soil group	reduction in SS (%) h-values	k-values
coarse-textured	80	53
medium-textured	93	35
fine-texutred	90	27

Bresler (1980) and Hopmans (1987). Lack of correlation between the two sets of scale factors reported by Rao et al. (1983) is probably due to the fact that they used soils that were classified by the same taxonomic name, but consisted of distinctly different horizons, consisting of sand and clay in the same profile. Based on the results of this study, distribution functions of scale factors calculated from h data can be used to describe variability of water retention as well as variability of hydraulic conductivity.

Scale factors were lognormally distributed which agrees with findings of Nielsen et al. (1973), Warrick et al. (1977b), Russo and Bresler (1980) and Hopmans (1987). Table 3 presents the mean, variance and coefficient of variation of scale factors calculated from h data for each soil group. The coefficients of variation are within the range reported by Warrick et al. (1977b), Russo and Bresler (1980) and Hopmans (1987). Jury et al. (1987a) and Hopmans (1987) reported spatial correlations for both hydraulic conductivity and water retention over distances in the order of about 10 m. This implies that in the analysis of scale factors of nearby measurements spatial correlations of these scale factors should be taken into account (Jury et al., 1987b). However, because the hydraulic functions used in this study were measured at locations all over the Netherlands, it was assumed that scale factors were uncorrelated and statistically independent.

Table 3. Mean (m_2), variance (sd_2^2) and coefficient of variation of scale factors calculated from moisture retention data.

soil group	mean	variance	coefficient of variation
coarse-tectured	1.00	0.56	0.75
medium-textured	1.02	0.37	0.60
fine-textured	1.04	1.07	0.99

Distribution functions of scale factors (Table 3) were used to generate 100 new hydraulic functions for each soil group. The three functional criteria which were described earlier, were calculated for these new functions. This procedure is comparable to studies (e.g. Clapp et al. 1983) which use Monte Carlo simulations to calculate the distribution of soil moisture using scale factors as random variables. However, rather than assuming a certain distribution of scale factors the actually calculated distribution (Table 3) is used in this study. Functional criteria calculated on the basis of newly generated hydraulic functions, and on the basis of measured hydraulic functions are summarized in Table 4. Comparison among the three soil groups shows that all functional criteria are different, indicating differences in soil hydraulic functions of the three soil groups. Mean values of functional criteria using unscaled and scaled hydraulic functions are fairly close. However, the functional criteria show smaller variations, as expressed by standard deviations, for scaled hydraulic functions as compared to unscaled hydraulic functions. This is especially true for Depth of water table (L) and Downward flux of water (F). Ranges in the percentage variation in functional criteria of unscaled hydraulic functions which would be explained by variation in functional criteria of scaled hydraulic functions are from 96% for Travel time (T) in the fine-textured soil group to 15% for Downward flux of water (F) in the coarse-textured soil group. The ranges in percentage explained variation for different functional criteria within the same soil group was due to the fact that effectiveness of scaling in reducing scatter of the original data differed for different h- and K values along the h(S)- and K(S) functionline (Figures 2, 3 and 4). Since different functional criteria use different h(S)- and K(S) values, the percentage of explained variation differs.

Although scaling is clearly an attractive procedure to describe variability in hydraulic functions in the form of a single random variable, it underestimates at the same time the variation in functional criteria. Explained variation in functional criteria is maximal 100% only when scaling is 100% effective and scaled hydraulic functions would coincide completely with scaled mean hydraulic functions.

Table 4. Mean (m), standard deviation (sd) and percentage variation (pv) of the functional criteria explained by the scaled hydraulic functions.

soil group	number of functions	travel time (d)			depth of water table (cm)			downward flux of water (cm.d-1)		
		m	sd	pv	m	sd	pv	m	sd	pv
coarse-textured										
unscaled	105	118.7	62.3		146.4	52.3		38.6	163.6	
scaled	100	123.5	40.3	65	147.5	28.3	54	31.1	24.8	15
medium-textured										
unscaled	43	237.4	36.7		90.5	36.0		1.4	3.5	
scaled	100	234.3	33.6	92	88.9	6.5	18	0.9	1.5	43
fine-textured										
unscaled	49	339.5	66.9		44.7	25.3		1.0	10.7	
scaled	100	335.8	63.9	96	50.5	9.4	37	0.7	2.7	25

5. CONCLUSIONS

Scaling was successfully used to reduce variation in measured hydraulic functions into a narrow band around the scaled mean hydraulic functions of a large data base containing water retention and hydraulic conductivity data of soils in the Netherlands. In order to be successful, soils have to be grouped in sufficient large groups. Scaling proved to be an attractive technique to simplify the description of variability in soil hydraulic functions in the form of scaled mean hydraulic functions and distribution functions of scale factors. However, at the same time scaling underestimates existing variability. Results obtained for the three soil groups can be used to estimate the minimal variability in model output. Expansion of the data base with new experimental hydraulic functions will facilitate the distinction of different subgroups within the three soil groups distinguished in this study.

REFERENCES

Ahuja, L.R., J.W. Naney and D.R. Nielsen, 1984: *Scaling soil water properties and infiltration modeling*. Soil Sci. Soc. Am. J., **48**, 970-973.

Arya, L.M.,D.A. Farrell and G.R. Blake, 1975: *A field study of soil water depletion patterns in presence of growing soybean roots. I. Determination of hydraulic properties of the soil*. Soil Sci. Soc. Am. Proc., **39**, 424-430.

Boels, D., J.B.H.M. Van Gils, G.J. Veerman and K.E. Wit, 1978: *Theory and system of automatic determination of soil moisture properties and unsaturated hydraulic conductivities*. Soil Sci., **126**, 191-199.

Bouma, J., 1982: *Measuring the hydraulic conductivity of soil horizons with continuous macropores*. Soil Sci. Soc. Am. J., **46**, 438-441.

Bouma, J., 1984: *Using soil morphology to develop measurement methods and simulation techniques for water movement in heavy clay soils*. p. 298-310. In J. Bouma and P.A.C. Raats (eds.), Proc. of the ISSS Sympos. on Water and solute movement in heavy clay soils. Wageningen, The Netherlands, 27-31 Aug. 1984. Publ. 37 Int. Inst. for Land Reclamation and Improvement (ILRI), Wageningen.

Bouma, J., C. Belmans, L.W. Dekker and W.J.M. Jeurissen, 1983: *Assessing the suitability of soils with macropores for subsurface liquis waste disposal*. J. Environ. Qual., **12**, 305-311.

Breeuwsma, A., J.H.M. Wösten, J.J. Vleeshouwer, A.M. van Slobbe and J. Bouma, 1986: *Derivation of land qualities to assess environmental problems from soil surveys*. Soil Sci. Soc. Am. J., **50**, 186-190.

Bresler, E. and G. Dagan, 1979: *Solute dispersion in unsaturated hetero-geneous soil at field scale: II. Applications*. Soil Sci. Soc. Am. J., **43**, 467-472.

Clapp, R.B., G.M. Hornberger and B.J. Cosby, 1983: *Estimating spatial variability in soil moisture with a simplified dynamic model*. Water Resour. Res., **19**, 739-745.

Dirksen, C., 1979: *Flux-controlled sorptivity measurements to determine soil hydraulic property functions*. Soil Sci. Soc. Am. J., **43**, 827-834.

FAO, 1985: *Guidelines: land evaluation for irrigated agriculture*. FAO Soils Bulletin, **55**, Rome.

Gardner, W.R., 1958: *Some steady state solutions of the unsaturated moisture flow equation with application to evaporation from a water-table*. Soil Sci., **85**, 228-232.

Hopmans, J.W., 1987: *A comparison of various methods to scale soil hydraulic properties*. J. of Hydrol., **93**, 241-256.

Hopmans, J.W. and J.N.M. Stricker, 1989: *Stochastic analysis of soil water regime in a watershed*. J. of Hydrol., **105**, 57-84.

Jury, W.A., D. Russo, G. Sposito and H. Elabd, 1987a: *The spatial variability of water and solute transport properties in unsaturated soil. I. Analysis of property variation and spatial structure with statistical models*. Hilgardia, **55**, 1-32.

Jury, W.A., D. Russo and G. Sposito, 1987b: *The spatial variability of water and solute transport properties in unsaturated soil. II. Scaling models of water transport*. Hilgardia, **55**, 33-56.

Miller, E.E. and R.D. Miller, 1955: *Theory of capillary flow: I. Practical implications*. Soil Sci. Soc. Am. Proc., **19**, 267-271.

Miller, E.E. and R.D. Miller, 1956: *Physical theory for capillary flow phenomena*. J. Appl. Phys., **27**, 324-332.

Montgomery, D.C. and F.A. Peck, 1982: *Introduction to linear regression analysis*. John Wiley & Sons, New York. pp. 504.

Nielsen, D.R., J.W. Biggar and K.T. Erh, 1973: *Spatial variability of field-measured soil-water properties*. Hilgardia, **42**, 215-259. (Plus separate appendices).

Rao, P.S.C., R.E. Jessup, A.C. Hornsby, D.K. Cassel and W.A. Pollans, 1983: *Scaling soil microhydrological properties of Lakeland and Konawa soils using similar media concepts*. Agric. Water Manage, **6**, 277-290.

Richards, L.A., 1965: *Physical condition of water in soil*. In C.A. Black et al. (ed.) Methods of soil analysis, Agronomy, **9**. 131-137.

Russo, D. and E. Bresler, 1980: *Scaling soil hydraulic properties of a heterogeneous field*. Soil Sci. Soc. Am. J., **44**, 681-684.

Sharma, M.L. and R.J. Luxmoore, 1979: *Soil spatial variability and its consequences on simulated water balance*. Water Resour. Res., **15**, 1567-1573.

Simmons, C.S., D.R. Nielsen and J.W. Biggar, 1979: *Scaling of field-measured soil-water properties*. Hilgardia, **47**, 74-154.

van der Zee, S.E.A.T.M. and W.H. van Riemsdijk, 1987: *Transport of reactive solute in spatially variable soil systems*. Water Resour. Res., **23**, 2059-2070.

van Genuchten, M.Th., 1978: *Calculating the unsaturated hydraulic conductivity with a new closed-form analytical model*. Research Rep. 78-WR-08. Water Resources Program, Princeton Univ., Princeton, NJ.

van Genuchten, M.Th., 1980: *A closed-form equation for predicting the hydraulic conductivity of unsaturated soils*. Soil Sci. Soc. Am. J., **44**, 892-898.

Warrick, A.W., G.J. Mullen and D.R. Nielsen., 1977a: *Predictions of the soil water flux based upon field-measured soil-water properties*. Soil Sci. Soc. Am. J., **41**, 14-19.

Warrick, A.W., G.J. Mullen and D.R. Nielsen, 1977b: *Scaling field measured soil hydraulic properties using a similar media concept*. Water Resour. Res., **13**, 355-362.

Wösten, J.H.M., J. Bouma and G.H. Stoffelsen, 1985: *The use of soil survey data for regional soil water simulation models*. Soil Sci. Soc. Am. J., **49**, 1238-1244.

Wösten, J.H.M., M.H. Bannink, J.J. De Gruijter and J. Bouma, 1986: *A procedure to indentify different groups of hydraulic-conductivity and moisture-retention curves for soil horizons*. J. Hydrol., **86**, 133-145.

Wösten, J.H.M., M.H. Bannink and J. Beuving, 1987: *Water retention and hydraulic conductivity characteristics of top- and subsoils in the Netherlands: The Staring series*. Report No. 1932, Soil Survey Institute, Wageningen, The Netherlands.

Wösten, J.H.M. and M.Th. Van Genuchten, 1988: *Using texture and other soil properties to predict the unsaturated soil hydraulic functions*. Soil Sci. Soc. Am. J., **52**, 1762-1770.

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9. USING SIMULATION TO DEFINE MOISTURE AVAILABILITY AND TRAFFICABILITY FOR A HEAVY CLAY SOIL IN THE NETHERLANDS

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USING SIMULATION TO DEFINE MOISTURE AVAILABILITY AND TRAFFICABILITY FOR A HEAVY CLAY SOIL IN THE NETHERLANDS

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ABSTRACT

Wösten, J.H.M. and Bouma, J., 1985. Using simulation to define moisture availability and trafficability for a heavy clay soil in The Netherlands. Geoderma, 35: 187—196.

Computer simulation was used in assessing moisture availability and trafficabillity of a heavy clay soil in grass. Calculations were made of the soil-moisture regime over a period of 30 years using measured meteorological and water-table data. Soil-physical input data consisted of moisture-retention and hydraulic-conductivity data, which were determined on large, undisturbed soil columns. The effect of vertical and horizontal soil cracks was assessed by independent quantitative measurements. Results of the calculations were expressed in terms of the probability of occurrence of: (1) precipitation deficits, (2) moisture deficits considering the calculated soil-moisture regime, and (3) adequate trafficability. Results referred to successive 10-day periods during the year. The magnitude of the moisture deficit was also expressed in terms of its yearly probability. Simulation models are essential tools for a flexible characterization of land qualities in terms of probability estimates which are useful for planning purposes.

INTRODUCTION

Moisture availability and trafficability are two important land qualities for modern grassland management. Soil survey organizations of different countries use different approaches in estimating these and other land qualities from soil survey information (e.g. McKeague et al., 1984). Though useful, these interpretations do not include probability estimates for the occurrence of moisture deficits or inadequate trafficability. Such probability estimates would provide a basis for rational decisions on agricultural management practices. This study focused on using computer simulation for obtaining probability estimates for moisture availability and trafficability in a heavy clay soil for which water-table data were available for 30 years.

MATERIALS AND METHODS

Soil

The soil type being characterized was a very fine clayey, mixed, mesic Typic Fluvaquent (Soil Survey Staff, 1975) with the following major characteristics (De Bakker, 1979): A1g, 0—8 cm: dark greyish brown mottled clay; moderate, medium, compound prisms parting to strong fine, subangular blocky peds; B1g, 8—25 cm: dark-grey mottled clay; structure intermediate between adjacent horizons; B21g, 25—50 cm: dark-grey mottled clay; strong, coarse, compound prisms parting to strong, medium, angular blocky peds; A11b, 50—64 cm: dark-grey clay (old surface layer); B22bg, 64—94 cm: dark-grey clay; strong very coarse smooth prisms; A12b, 94—106 cm: buried surface horizon; B23bg, 106—120 cm: grey mottled clay with a compound prismatic structure parting to strong, fine angular blocky peds.

Methods

The simulation model being used was proposed by De Laat (1980). It calculates a water balance for successive 10-day periods, using effective rainfall and potential evapotranspiration data to define the upper-boundary condition and water-tables for the lower-boundary condition.

Calculations were made for the period 1954 to 1983, using water-tables measured every fortnight at an experimental farm (Fig. 1). Precipitation and evaporation from a free water surface (E_O) were derived from the De Bilt meteorological station. Potential evapotranspiration (E_{pot}) for the soil under grass cover is estimated as: $E_{pot} = 0.8\ E_O$ for the growing season and $E_{pot} = 0.3\ E_O$ for the winter period with transitional values for spring and fall. The soil is schematized into a rootzone and a subsoil, which are separated by the effective rooting depth. The latter, which has a value of 20 cm, in this soil, is defined as the depth above which 80% of the roots occur. Moisture supply from the rootzone decreases linearly with increasing log-h values beyond a critical value of log-h = 2.7. Moisture-retention data (h—θ) are needed for the rootzone and the subsoil, and the hydraulic conductivity (K—h) is needed for the subsoil only. The symbol h describes the pressure head (m) and θ = volumetric moisture content ($m^3\ m^{-3}$).

Physical procedures included the column method, the crust test and the hot-air method for measuring the K—h relation (see Bouma, 1983). A staining technique was applied to record the effects of horizontal cracks on the upward flux of water from the water-table to the rootzone (K_{macro}). K_{micro} characterizes K for the peds (Bouma and De Laat, 1981).

Values for bypass-flow (earlier called: "short circuiting") were derived from meteorological data on the frequency of heavy and very heavy rains and from direct measurement. Bypass-flow refers to the vertical movement of free water along macropores through an unsaturated soil matrix (Bouma, 1984). Bypass-flow was assumed to be 10% of rainfall in April and September and 20% in the months of May to August. No bypass-flow was assumed to occur in the other months (Bouma and De Laat, 1981; Bouma et al., 1981). Effective rainfall for the summer period was defined as actual rainfall minus bypass-flow.

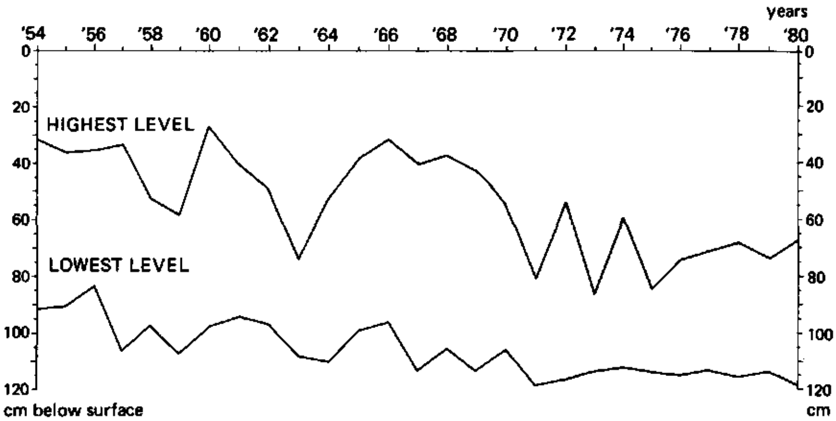

Fig. 1. Highest and lowest water-tables in a heavy clay soil during a 27-year period.

Moisture-retention data were measured using a series of tensiometer readings in a large, drying soil sample in the laboratory and corresponding gravimetric moisture samples (e.g. Bouma et al., 1983). Measurements of penetrometer resistance as a function of the pressure head in the surface soil (0—10 cm) were made during several seasons. Observation of corresponding compaction patterns resulted in the distinction of a critical pressure head of −90 cm (Bouma, 1981), which was reduced to $h = -100$ cm in this study to include a small safety margin. Thomasson (1982) reported lower critical pressure heads for comparable soils, when considering soil tillage practices.

RESULTS

Physical measurements and precipitation deficits

Results of the K—h measurement are presented in Fig. 2 and show a strong effect of horizontal cracks on K_{unsat} for the upward flux, as expressed by K_{macro}. The moisture-retention curves for the root-zone and the subsoil were not different (Fig. 3). The occurrence of a precipitation deficit, which is equal to the excess of evapotranspiration over precipitation, is presented as a probability graph in Fig. 4. This graph indicates the probability of occurrence of a precipitation deficit for any 10-day period over 30 years considering only rainfall and evapotranspiration data. For example, the probability of having a precipitation deficit in the period May 1 to May 10 inclusive is 77% with 90% and 75% as upper and lower 95% confidence limits. The graph does not allow an estimate of the magnitude of the deficit.

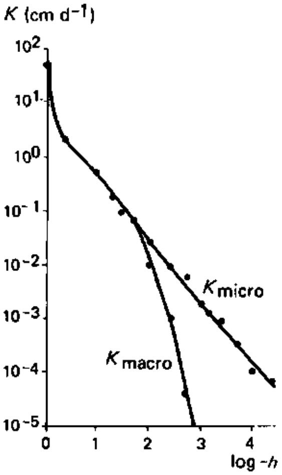

Fig. 2. K-curve for the subsoil (20 cm below surface to water-table). K_{micro} applies to the peds and K_{macro} defines the upward flux from the water-table to the rootzone, which is reduced by horizontal cracks.

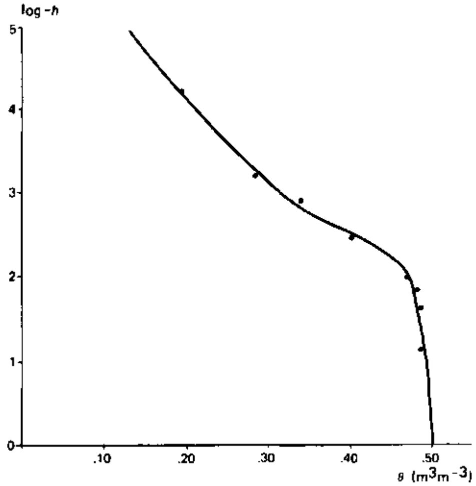

Fig. 3. Moisture-retention curve for the rootzone (0—20 cm) and the subsoil.

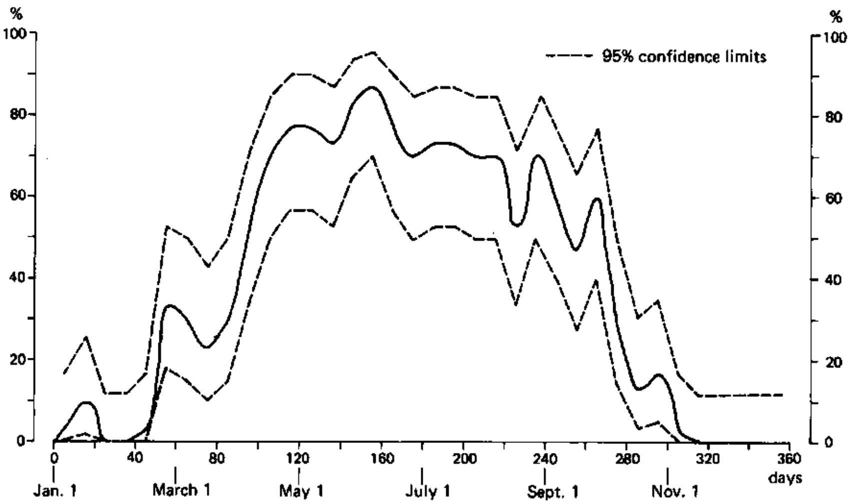

Fig. 4. Probability of occurrence of precipitation deficits, which are the differences between evapotranspiration and precipitation, for every 10-day period.

Validation of the model

Field measurements and simulation calculations of h and θ are, as an example, presented in Table I for five selected dates during the growing season of 1979. Only simulation results which incorporated both K_{macro} and bypass-flow corresponded well with field measurements. Both aspects, reflecting the effects of horizontal and vertical cracks, should therefore be considered when simulating water movement in a clay soil (e.g. Bouma and De Laat, 1981).

TABLE I

Validation of the simulation model comparing measured field data at a depth of 20 cm below surface with calculated data for the same depth. Calculations reflect the effects of soil cracking as explained in the text

Date (1979)	Simulation		Field measurement	
	h (cm)	θ ($cm^3\ cm^{-3}$)	h (cm)	θ ($cm^3\ cm^{-3}$)
June 25	−120	0.50	−100	0.50
July 19	−2 400	0.43	−8 000	0.41
August 1	−16 000	0.38	−10 000	0.40
August 13	−2 000	0.43	−2 000	0.43
August 30	−2 600	0.43	−2 000	0.43

Moisture availability

Results of the runs with the simulation model are shown in four ways: (1) as a diagram showing the yearly moisture deficits (Fig. 5); (2) as probabilities of occurrence of yearly moisture deficits, using classes of 30 mm (Table II); (3) as a graph showing the occurrence of a moisture deficit for every 10-day period, without noting its magnitude (Fig. 6) and (4)

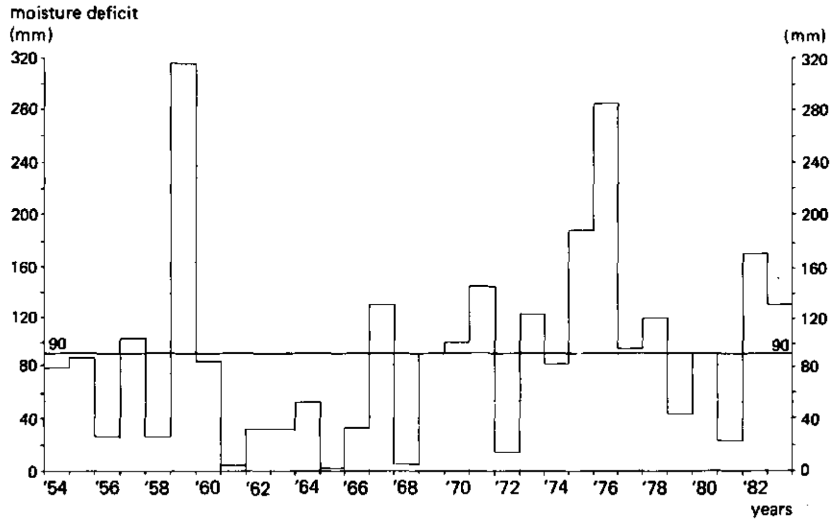

Fig. 5. Yearly moisture deficits for the heavy clay soil being studied (the average value is 90 mm).

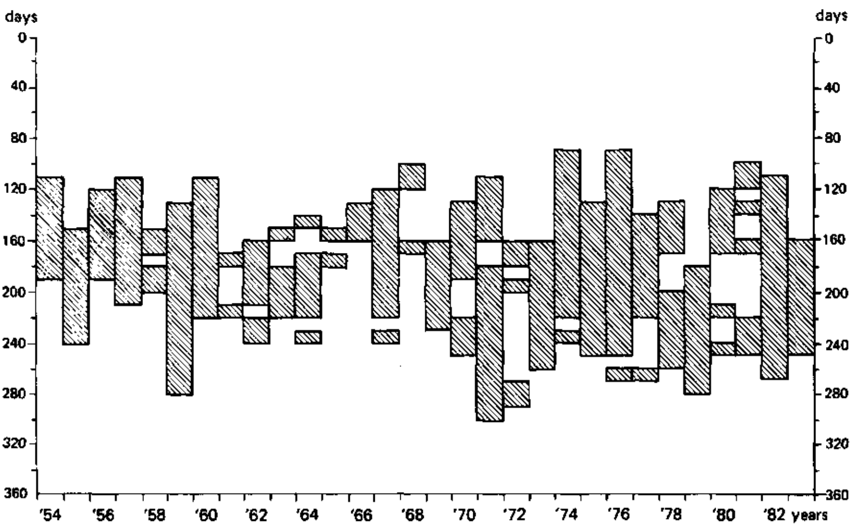

Fig. 6. Periods with moisture deficits during 30 years. Day zero corresponds with January 1.

as a graph indicating the probability of the occurrence of a moisture deficit for every 10-day period (Fig. 7).

The diagrams for each of the 30 years (Fig. 5) show very large differences between years, ranging from a moisture deficit of 320 mm in 1959 to 0 mm in 1965. Average values of moisture deficits (here 90 mm) are therefore of little practical significance. The expression in terms of a probability of a certain moisture deficit is more relevant. The data in Table II indicate that moisture deficits (in classes of 30 mm) up to 150 mm have an average probability of occurrence of approximately 17% for each class. Deficits higher than 150 mm have a much lower probability. It is not only the magnitude of the yearly deficit that is important but also the time of oc-

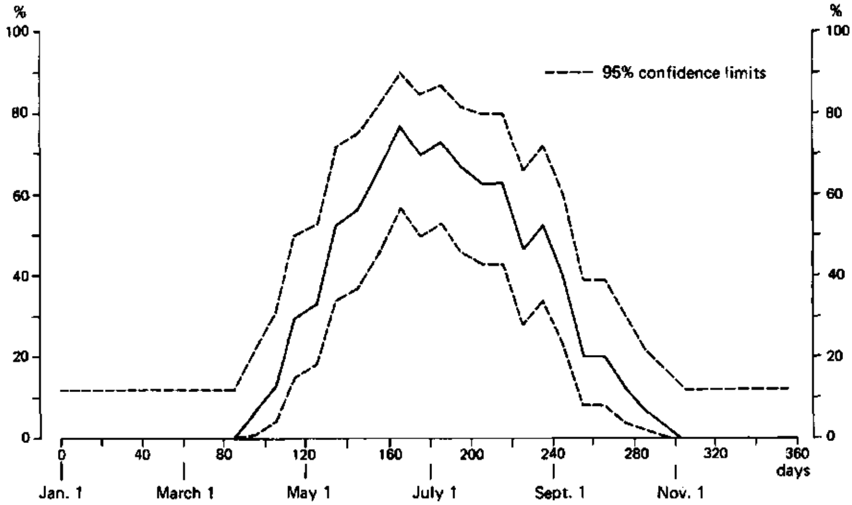

Fig. 7. Probability of occurrence of moisture deficits for every 10-day period.

TABLE II

Probability of occurrence of a yearly moisture deficit

Yearly moisture deficit (mm)	Probability of occurrence with 95% confidence limits (%)
0— 30	23.3 (10—43)
30— 60	16.7 (5—35)
60— 90	13.3 (4—31)
90—120	20.0 (8—39)
120—150	13.3 (4—31)
150—180	3.3 (0—17)
180—210	3.3 (0—17)
210—240	0 (0—12)
240—270	0 (0—12)
270—300	3.3 (0—17)
300—330	3.3 (0—17)

currence. Figs. 6 and 7 indicate that the months of June and July (days 150 to 210) have the highest probabilities for the occurrence of a moisture deficit (approximately 70%). The average graphs from Figs. 4 and 7 are combined in Fig. 8. The difference indicates the soil contribution to water being supplied to the grass crop. For example, the probability of having a moisture deficit in the period of May 1 to May 10 is 33%. This percentage is lower than the 77% probability of a precipitation deficit. The difference is therefore due to moisture supplied by water in the rootzone (0—20 cm) and by water flowing from the water-table to the rootzone.

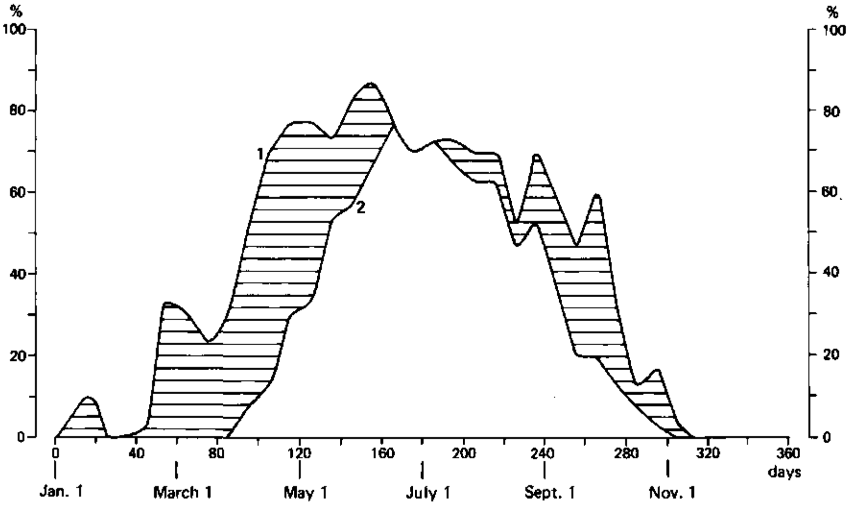

Fig. 8. Probability of occurrence of precipitation (*1*) and moisture (*2*) deficits, combining average graphs from Figs. 4 and 7.

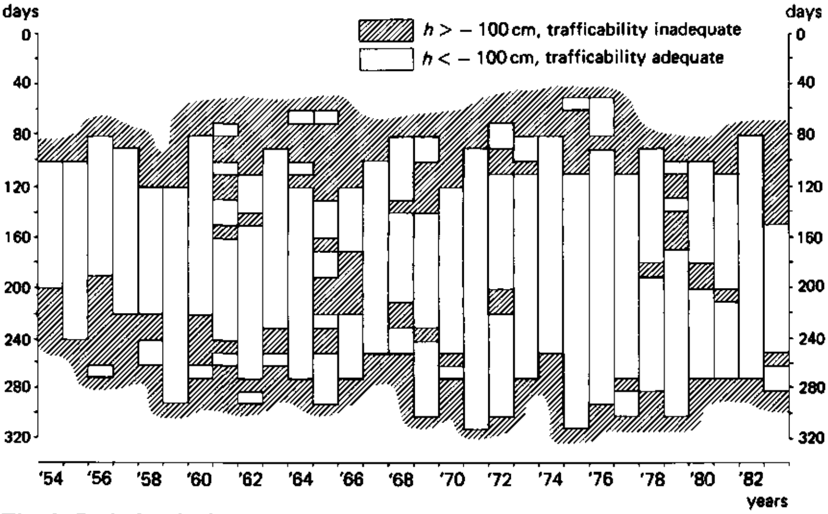

Fig. 9. Periods of adequate and inadequate trafficability during 30 years.

Trafficability

The pressure head in the 20 cm thick rootzone is produced as output by the model. If the critical value of $h = -100$ cm is chosen, the pressure head for each 10-day period can be classified easily in terms of adequate or inadequate trafficability. Results for every 10-day period of the 30 years are shown in Fig. 9. These results demonstrate the large variability of data among the various years, which makes definition of "average" values less relevant. A probability graph, as shown in Fig. 10, was therefore developed indicating the probability for trafficability to be adequate for any 10-day period during the year. For example, this probability is approximately 80% for the period of May 1—10.

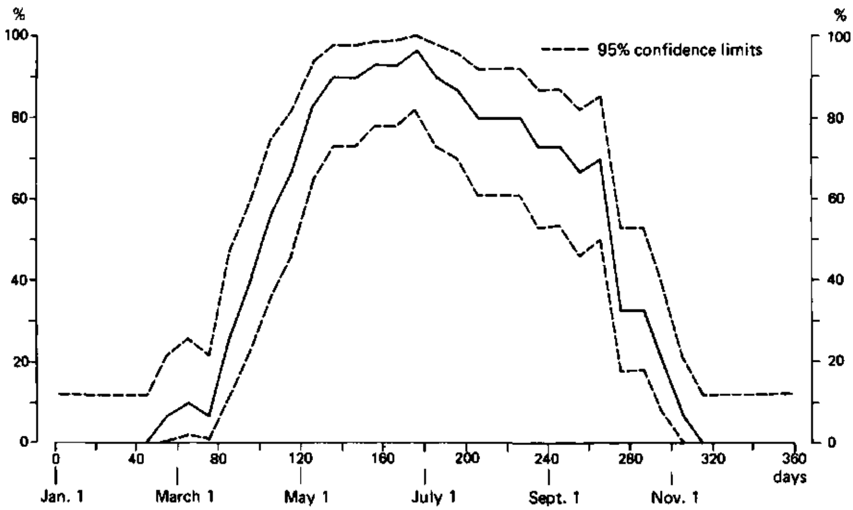

Fig. 10. Probability of adequate trafficability for every 10-day period.

DISCUSSION

Definition of land qualities in terms of probability of occurrence provides more relevant criteria for land and water management than definitions in terms of a fixed value for each soil which is based on somewhat arbitrarily defined conditions (e.g. McKeague et al., 1984). The user is, in general, confronted with a wide choice of possible management scenarios. Having the possibility of predicting the probable consequences associated with each of them is attractive in principle.

The simulation model being used is relatively simple and its application involves little expense. Simple models need relatively simple input data, but even these are not always available. Also, the model should not be more specific than the available data or than allowed by the spatial variability of the soils being considered. For example, water-tables in this study were measured only once every fortnight. Thus, there is no point in running a simulation model that requires daily input data. Use of relatively simple simulation models, such as the one used in this study, is therefore recommended for land evaluation studies because basic data either are often lacking or they have been obtained infrequently. In addition, one should realize that simulation models for soil-moisture regimes are based on soil-physical theory for stable, non-swelling soils. Use in clay soils imposes drastic changes due to the effects of soil cracking, as discussed in this paper and elsewhere. Rather than defining more detailed "sand" models, emphasis should be on the development of models which can properly characterize heterogeneous and anisotropic soils in the field (e.g. Bouma, 1984).

REFERENCES

Bouma, J., 1981. Soil survey interpretation: estimating use-potentials of a clay soil under various moisture regimes. Geoderma, 26 (3): 165—177.

Bouma, J., 1983. Use of soil survey data to select measurement techniques for hydraulic conductivity. Agric. Water Manage., 6 (2/3): 177—190.

Bouma, J., 1984. Using soil morphology to develop measurement methods and simulation techniques for water movement in heavy-clay soils. In: J. Bouma and P.A.C. Raats (Editors), Water and Solute Movement in Heavy-clay Soils. Proc. ISSS Symposium. ILRI Publ. 37, Wageningen, The Netherlands, 361 pp.

Bouma, J. and De Laat, P.J.M., 1981. Estimation of the moisture supply capacity of some swelling clay soils in The Netherlands. J. Hydrol., 49 (3/4): 247—259.

Bouma, J., Dekker, L.W. and Muilwijk, C.J., 1981. A field method for measuring short-circuiting in clay soils. J. Hydrol., 52 (3/4): 347—354.

Bouma, J., Belmans, C., Dekker, L.W. and Jeurissen, W.J.M., 1983. Assessing the suitability of soils with macropores for subsurface liquid waste disposal. J. Environ. Qual., 12 (3): 305—311.

De Bakker, H., 1979. Major Soils and Soil Regions in The Netherlands. Pudoc, Wageningen/Junk, The Hague, 203 pp.

De Laat, P.J.M., 1980. Model for Unsaturated Flow above a Shallow Water Table, Applied to a Regional Subsurface Flow Problem. Pudoc, Wageningen, Agric. Res. Rep., 895; 126 pp.

McKeague, J.A., Eilers, R.G., Thomasson, A.J., Reeve, M.J., Bouma, J., Grossman, R.B., Favrot, J.C., Renger, M. and Strebel, O., 1984. Tentative assessment of soil survey approaches to the characterization and interpretation of air-water properties of soils. Geoderma, 34: 69—100.

Soil Survey Staff, 1975. Soil Taxonomy; a Basic System of Soil Classification for Making and Interpreting Soil Surveys. U.S. Dept. Agric. Handbook No. 436. U.S. Govt. Print. Off., Washington, D.C., 754 pp.

Thomasson, A.J., 1982. Soil and climatic aspects of workability and trafficability. Proc. 9th Conf. Int. Soil Tillage Res. Organisation Osijek, Yugoslavia, pp.551—557.

10. CONCLUDING REMARKS

Simulation of water movement in soils has many environmental and agricultural applications. For example, movement of potentially harmful substances can be predicted and water management can be optimized. These simulations require a variety of input data. In this context, it is important to critically review each specific source of input data separately as well as in relation to the other sources. In doing this, a number of questions arise. For example, what is the most appropriate scale of the soil map needed to address a specific problem? Should it be a detailed soil map at a scale of 1 : 10 000 or is a more general soil map at a scale of 1 : 50 000 better? What sort of simulation model should be used: a model that simulates unsaturated flow of water by solving Richard's equation or a simple budget model? Are estimates of water tables in classes appropriate or are daily measured values needed? At what detail must the hydraulic functions be known? Can previously collected functions be used or do they have to be measured again on-site? To be able to answer these questions we must first identify the detail required in the answers to the problems posed.

For example, if a general answer is sufficient it can be concluded that the soil map at a scale of 1 : 50 000 is a good starting point to derive hydraulic functions for areas of land. Water tables measured in classes and previously collected hydraulic functions can be assumed to be adequate. Having made these choices there is no point in running a complex simulation model that would, for instance, require daily values for the water tables. In such cases, identification of the detail required in the answer provides the information that enables a decision to be made about the most appropriate approach. Only in this way can an optimal allocation of available limited resources be assured.

Fortunately, soil maps have proved to be a good starting point to derive hydraulic functions for areas of land. As a consequence, areas of land are not black boxes that should be analysed purely stochastically in order to derive their hydraulic functions. Therefore, use of the soil map is a form of sample stratification that limits the number of samples that have to be measured.

Many years of soil survey work have resulted in numerous soil maps and in an extensive body of basic soil data. Now that Geographical Information Systems (GIS) are readily available this information has become even more valuable because it can be stored, handled and manipulated. As a consequence, it is becoming increasingly attractive to focus on the hydraulic interpretation of the individual borings rather than on the interpretation of the "representative profile" identified for each mapping unit. Using the information from each boring means that the simulation of a specific aspect of soil behaviour is no longer restricted to the mean behaviour of the mapping unit; instead, information on the variability of the behaviour within the mapping unit can also be provided.

In generating soil hydraulic functions for areas of land it is important to realize that there is no single best method. Rather than attempting to give "the" method it is more attractive to be able to offer a variety of possible methods, each with particular advantages. An indication of the accuracy and costs of each method provides the potential user with information to make the choice most appropriate for his particular problem.

11. SAMENVATTING

Gebruik van gekarteerde bodemkenmerken om de berekening van watertransport in gronden te verbeteren.

De laatste jaren groeit het besef dat de kwaliteit van bodem en water verslechtert door tal van landbouwkundige en industriële activiteiten. In een poging deze situatie allereerst te beschrijven en vervolgens ook te veranderen, hebben wetenschappers computermodellen ontwikkeld. Doordat de kennis van de verschillende fysische, chemische en biologische processen in de onverzadigde zone van de bodem toeneemt, worden deze modellen steeds complexer. Er is nu een punt bereikt dat vooruitgang vooral afhankelijk is van de mogelijkheid om goede invoergegevens voor deze modellen te verzamelen en minder van de modellen zelf.

Voor de berekening van het watertransport in de onverzadigde zone moeten nauwkeurige waterretentie- en doorlatendheidskarakteristieken beschikbaar zijn. Tegelijkertijd moeten we erkennen dat het vaak niet mogelijk en misschien ook wel niet noodzakelijk is om ons volledig te concentreren op het verzamelen van steeds nauwkeuriger invoergegevens. Het probleem dat realistische voorspellingen moeten worden gedaan maar dat hiermee niet kan worden gewacht totdat alle onbekende grootheden nauwkeurig zijn gemeten, vormt een interessant onderwerp van studie.

In Hoofdstuk 2 van dit proefschrift wordt aangetoond dat het belangrijk is om zich bewust te zijn van het generalisatie-niveau waarop de modeluitkomsten moeten worden toegepast. Voor globale toepassingen kunnen we met een minder gedetailleerde en daardoor ook minder kostbare aanpak volstaan dan voor een specifieke vraagstelling. De bodemkaart die door de schaal van de kaart overeenstemt met een bepaald generalisatie-niveau is hierbij gebruikt als uitgangspunt voor het verkrijgen van bodemfysische karakteristieken van gebieden.

In Hoofdstuk 3 is de gevolgde procedure beschreven. Van de bodemhorizonten die tijdens de kartering zijn onderscheiden, zijn waterretentie- en doorlatendheidskarakteristieken gemeten. Het bleek dat niet alle bodemhorizonten zich ook in bodemfysisch opzicht verschillend gedroegen. Hierdoor kon een beperkt aantal "bodemfysische bouwstenen" ofwel horizonten met duidelijk verschillende bodemfysische karakteristieken, worden onderscheiden. De koppeling van bodemhorizont met "bodemfysische bouwsteen" vormt de sleutel waarmee vervolgens de bodemkaart is omgezet in een kaart met bodemfysische eenheden.

Het meten van bodemfysische karakteristieken is tijdrovend en kostbaar. Door in verschillende projecten bodemfysische metingen te verrichten, is evenwel in de loop der jaren een waardevol bestand aan gemeten karakteristieken opgebouwd. Uitgaande van dit bestand was het aantrekkelijk om te onderzoeken of relaties konden worden opgesteld tussen deze gemeten karakteristieken en bodemkenmerken waarop wordt gekarteerd. In Hoofdstuk 4 is beschreven hoe "bodemkundige vertaalfuncties" zijn opgesteld waarmee op eenvoudige en goedkope wijze, bodemfysische karakteristieken kunnen worden afgeleid uit bijvoorbeeld de percentages leem, lutum en organische stof, de dichtheid van de grond en de mediaan van de zandfractie.

De bodemfysische karakteristieken zijn geen doel op zich maar dienen als invoergegevens in modelberekeningen. Evaluatie van de verschillen tussen bodemfysische karakteristieken kan daarom nodig zijn. Het is daarbij beter de praktische bodemkenmerken die met deze karakteristieken zijn berekend te vergelijken, dan een statistische vergelijking te maken van de karakteristieken zelf.

In Hoofdstuk 5 zijn berekende kritieke grondwaterstand, verblijftijd en kritieke flux gebruikt om de bodemfysische karakteristieken van verschillende bodemhorizonten te vergelijken. In de Hoofdstukken 6 en 7 zijn de berekende vochtinhouden van de bovenste 50 cm van drie bodemprofielen, verdampingstekorten en fluxen door een vlak op 30 cm beneden maaiveld van dezelfde drie bodemprofielen gehanteerd als praktische bodemkenmerken. Door deze laatste kenmerken met elkaar te vergelijken, was een vergelijking mogelijk van vier verschillende methoden om bodemfysische karakteristieken te generen. Deze methoden waren: (A) directe meting ter plekke, (B) meting van bodemhorizonten in het gebied, (C) gebruik van het Nederlandse bestand aan gemeten karakteristieken en (D) gebruik van vertaalfuncties.

Modeluitkomsten zijn gevoelig voor variabiliteit in bodemfysische karakteristieken en ook voor variabiliteit in meteorologische gegevens en grondwaterstanden. In Hoofdstuk 8 is beschreven hoe de ruimtelijke variabiliteit in bodemfysische karakteristieken kan worden gekwantificeerd door gebruik te maken van de schaleringstechniek. Met deze techniek wordt de ruimtelijke variabiliteit op een handzame wijze beschreven door een gemiddelde en door een spreiding van schaalfactoren. Deze waarden zijn vervolgens gebruikt om nieuwe bodemfysische karakteristieken te generen. Hierbij bleek echter dat de spreiding in modeluitkomsten geringer was indien nieuw gegenereerde karakteristieken werden gebruikt in plaats van de oorspronkelijke, niet-geschaleerde karakteristieken. Het effect van temporele variabiliteit in meteorologische gegevens en grondwaterstanden op modeluitkomsten kon worden nagegaan door een periode van 30 jaar door te rekenen. De variabiliteit in berekende vochttekorten en bewerkbaarheden is weergegeven in grafieken die laten zien wat de kans is dat in een bepaalde periode een vochttekort voorkomt en wat de kans is dat in een bepaalde periode de grond bewerkt kan worden.

CURRICULUM VITAE

Jan Hindrik Maria Wösten werd op 3 november 1954 geboren in Emmen. In 1972 behaalde hij het diploma HBS-b aan het Katholiek Drents College in Emmen.

In 1972 begon hij met zijn studie bodemkunde en bemestingsleer aan de Landbouwhogeschool in Wageningen. De praktijktijd werd doorgebracht op het Volcani Center in Bet Dagan, Israël, en op de universiteit van Wisconsin in Madison, USA. In maart 1979 studeerde hij af als landbouwkundig ingenieur met als hoofdvakken bodemnatuurkunde en grondbewerking en met als bijvak theoretische teeltkunde.

Van april tot en met oktober 1979 werkte hij als tijdelijk medewerker van de afdeling theoretische teeltkunde op het project "Primaire Produktie in de Sahel" in Niono, Mali. Van november 1979 tot en met januari 1982 had hij een positie als postdoctoral fellow in het "Grain Legume Improvement Program" van het International Institute of Tropical Agriculture (IITA) in Ibadan, Nigeria. In deze periode selecteerde hij op verschillende locaties in het Sahel gebied, cowpea-variëteiten (Vigna unguiculata (L.) Walp.) op droogteresistentie.

In februari 1982 trad hij in dienst bij de Stichting voor Bodemkartering (Stiboka) waar hij werkzaam was op de afdeling Bodemfysica en Hydrologie. Bij de vorming van het Staring Centrum, waarin Stiboka is opgegaan, werd hij in april 1989 hoofd van het Bodemfysisch Laboratorium.

In 1982 trouwde hij met Liesbeth Scholten. Zij hebben drie dochters: Lidwien, Hedwig en Lilian.