Description of the regional groundwater flow model SIMGRO

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Report 7

The WINAND STARING CENTRE, Wageningen, The Netherlands (1989)

ABSTRACT

Querner, E.P. and P.J.T. van Bakel. 1989. Description of the regional groundwater flow model SIMGRO. Wageningen, The Netherlands, The Winand Staring Centre for Integrated Land, Soil and Water Research. Report 7.
44 p.; 22 figs.; 8 tables; 5 annexes.

The groundwater model SIMGRO has been developed to simulate regional groundwater flow and water levels in the surface water system. It includes also aspects such as the retention of water in the unsaturated zone, sprinkling, evapotranspiration, water supply and subsurface irrigation.
The unsaturated zone has been modelled per land use, characterized as different agricultural areas (crops), built-up areas, forests and nature reserves. The surface water system is considered as a single reservoir per subregion. Water is extracted from it for sprinkling and water level control is possible.

Keywords: regional groundwater flow, unsaturated zone, evapotranspiration, sprinkling, surface water system, finite element method, land use, drainage

ISSN 0924-3062

This report has earlier been announced as Report 26 (1988) of the Institute for Land and Water Management Research (ICW), Wageningen, the Netherlands.

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The WINAND STARING CENTRE is continuing the research of: the Institute for Land and Water Management Research (ICW), Institute for Pesticide Research, Environment Divison (IOB), Dorschkamp Research Institute for Forestry and Landscape Planning, Divison of Landscape Planning (LB), and Soil Survey Institute (STIBOKA).

Project 100.21 [REP/rep7]

CONTENTS

SUMMARY

The groundwater model SIMGRO described in this report has been developed to simulate the flow of water in the saturated zone, the unsaturated zone and the surface water. It simulates the hydrological process within a region. It includes also the effect of irrigation and its impact on the water requirements of the surface water system. The model has been constructed in such a manner that it is sufficient accurate without requiring too many input data and excessive computer time.

The saturated zone has been modelled by means of the finite element method. The region is divided into a finite number of elements. Quasi three-dimensional flow is considered, which means horizontal flow in water bearing layers and vertical in less-permeable layers. The groundwater levels and fluxes are calculated per nodal point. The unsaturated zone has been modelled by means of two reservoirs, one for the root zone and one for the subsoil. For the root zone the storage of water in it is considered, with extractions and inflows. From the water balance of the subsoil the height of the phreatic surface is calculated using a storage coefficient. The unsaturated zone has been modelled per land use on a subregional level. The surface water system of a subregion, in reality a network of small channels, is modelled as a single reservoir with criteria for, among others, water supply, discharge characteristics, water level control, extractions for sprinkling. A number of nodes of the finite element network represent a subregion. This means that soil properties and hydrological conditions should be relatively homogeneous within a subregion. The categories of land use consist of agricultural areas, built-up areas, forests and nature reserves. The agricultural technologies can apply additional water by means of sprinkling from groundwater or surface water.

The usefulness of the regional groundwater flow model is demonstrated for a region. The results of the unsaturated zone model are analysed by comparing it with results from a more accurate model. Measured and calculated groundwater levels are compared. A sensitivity analysis on the geohydrological parameters gave an idea of the variation on the results.

1. INTRODUCTION

In rural areas different types of activities may be present having an impact on the environment. Especially the impacts on regional water management caused by agriculture and water supply may interfere. These activities often interfere with the interest of nature conservation as well. The need for research on these aspect has resulted in a study on: optimization of regional water management in areas with conflicting interests (WERKGROEP OPTIMALISERING REGIONAAL WATERBEHEER, 1988). The objective was to develop a system of models to generate and evaluate alternatives for regional water management.

In the optimization model all effects of production, land-use, water movements, etc have to be related in certain criteria and/or constraints. With these constraints the optimal and/or feasible solutions (scenarios) for the area can be calculated. Due to the large amount of variables this model requires very simple relationships for all the criteria, otherwise the calculation method will be too complicated and the cost for running such a program excessive. For the optimization model linear programming is used. The input for this model have to describe all the effects related to the region (eg. relations for production, costs, labour, water movements, etc). Because of this linearization of all relations the results from the optimi-zation model should be verified with more accurate models. The comprehen-sive models should then describe certain processes (eg. agri-cultural pro-duction, water quantity, water quality, etc), more accurately. These models are for the verification of the outcome from the optimization model, and can be used to estimate input parameters more accurately.

The groundwater flow model to be described in this report has been developed to simulate the flow of water in the saturated zone, the unsaturated zone and the surface water. Within a region often the hydrological conditions can vary, which require an integrated model to simulate all processes involved. It includes also the effect of irrigation and its impact on the water requirements of the surface water system. The aim was to simulate the rather complex processes involved in such a manner that it is accurate enough without requiring too many input data and too much computer time.

The schematization of land use and hydrological system is discussed in Chapter 2. In Chapter 3 the calculation method for the saturated zone, the unsaturated zone and the surface water is given in detail. The required input data, an example and a sensitivity analysis is discussed in Chapter 4.

2. GROUNDWATER FLOW MODEL SIMGRO

2.1. INTRODUCTION

Shallow groundwater prevails in many parts in The Netherlands. For the purpose of land drainage dense channel systems are present. These systems are often used for water supply as well and therefore levels in the channels vary during the year. In winter levels are kept low to allow a quick drainage and in spring levels are raised to conserve water for dry periods in summer. Because these variations have a significant impact on agricultural production and the hydrological conditions in nature reserves, it is important that the levels are manipulated in an optimal manner. If certain changes for a regional system is suggested, the important question arises what is the effect of these changes on groundwater levels and evapotranspiration. These values or the changes will then give an indication of variation of crop production in the region or conditions in nature reserves.

The groundwater flow model, SIMGRO (SIMulation of GROundwater flow and surface water levels), has been developed to simulate regional groundwater flow and to calculate the requirements for water supply, sprinkling, subsurface irrigation and water level control. The computerprogram FEMSAT (VAN BAKEL, 1978) was extended to simulate all the hydrological processes involved. The unsaturated zone and surface water system formerly not present in this model has been included, which has resulted in the program SIMGRO. A user's manual is available (QUERNER,1988a) as well as a program manual (QUERNER, 1988b). A flow chart of the programme is given in Annex B.

The saturated zone has been modelled by means of the finite element method and groundwater levels and fluxes are calculated per nodal point. The unsaturated zone has been modelled per land use on a subregional level. Subregions, represented by a number of nodal points, are choosen in such a way that they have relatively homogeneous soil properties and hydrological conditions. The surface water system of a subregion, in reality a network of small channels, is modelled as a single reservoir.

A typical schematization of the hydrological system showing all the various water movements within a subregion is given in Figure 1.

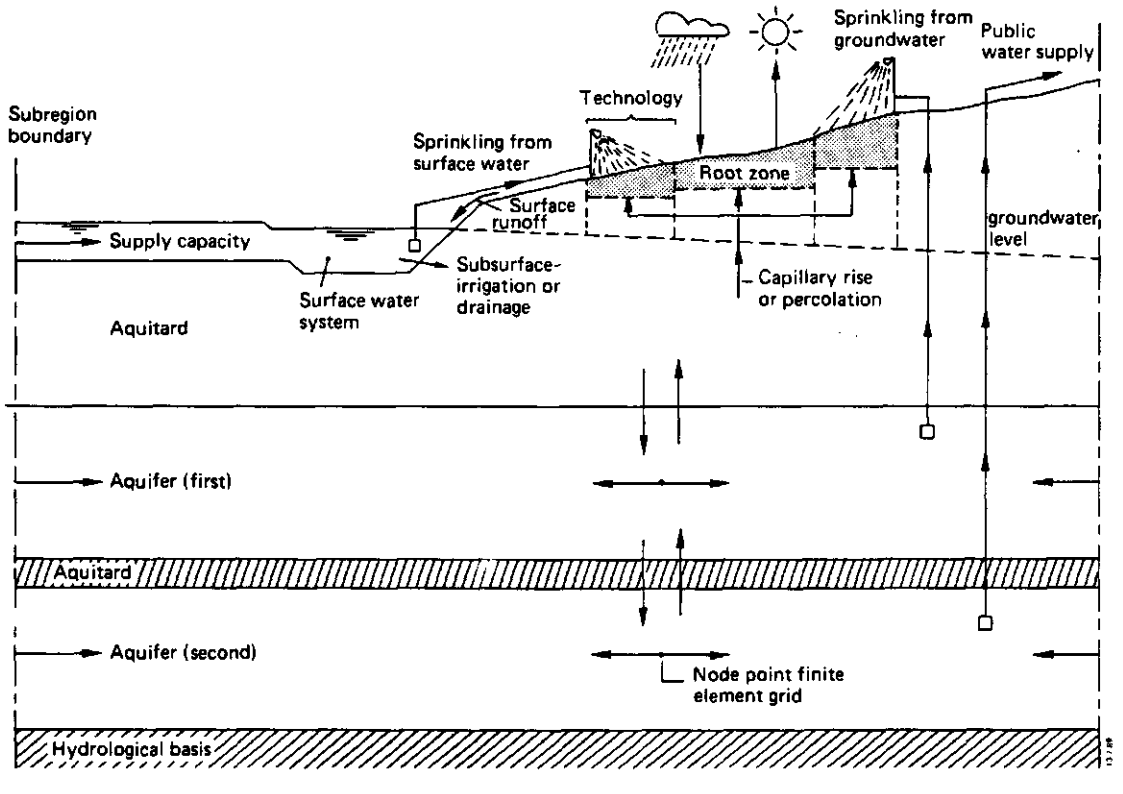

Figure 1. Schematization of hydrological system within a subregion by means of an integration of saturated zone, unsaturated zone and the surface water system. Technology is a specific agricultural activity

2.2. SCHEMATISATION OF LAND USE

2.2.1. Subregion

The unsaturated zone has been modelled per land use activity on a subregional level. The division of the region in a number of subregions is therefore very important. Within subregions homogeneous conditions must be present with respect to soil type, class referring to the depth of the groundwater level below surface and in relation to characteristics of the surface water system. Of the soil type capillary rise and moisture content of the root zone are the important factors. The groundwater level class, characterized by minimum and maximum groundwater levels over the years, should not vary considerable within a subregion. The average depth of the groundwater level below surface is used for the unsaturated zone (for computing capillary rise, storage coefficient and equilibrium moisture coeffi-

cient). The surface water system consists of a single reservoir per subregion. The water management is another important factor to select subregions (e.g. water supply, water level control, target levels).

2.2.2. Technologies

A subregion is subdivided into different areas characterized by its land use. The area involved is therefore defined by an agricultural activity of growing and processing a certain crop, or grassland with livestock. These areas are called technologies that use land. From each technology only the area involved is known as a percentage of the subregion, and not its geometrical position. The total area of a technology may be present as numerous portions of land scattered over a subregion.
The region that must be analysed is schematically subdivided into four main categories of land use which are important for the calculations of the various water movements. They are: agricultural areas, built-up areas, nature reserves, and forests.

In the model the water balance of a subregion should take into account all different categories of land use that play a role in the hydrological cycle. Similar criteria as for the agricultural technologies can be defined for built-up areas, nature reserves and forests, so that they can be incorporated in the computer model. The built-up areas are further split up into areas with an impermeable surface (e.g. houses, streets, etc) and the rest. For the impermeable surface areas there is no connection with the unsaturated zone. These areas can be disregarded, because the runoff from these areas is generally collected by a combined stormwater and foul sewer system. The permeable areas in the towns are considered to have the same characteristics as grassland.

Nature reserves have a vegetation of grass. Forests are distinguished because they have quite different evapotranspiration values and thickness of the root zone. The agricultural technologies are also characterized with respect to a water availability condition by means of a production level. A high production level means a higher water demand in the growing season. The demand is met by means of sprinkling, where each production level has its own criterium for applying the sprinkler irrigation in terms of available moisture in the root zone.

2.3. SCHEMATISATION OF HYDROLOGICAL SYSTEM

2.3.1. Saturated zone

There is a schematization of the groundwater system in a number of layers. Quasi three-dimensional flow is considered, which means horizontal flow in water bearing layers (aquifers) and vertical flow in less-permeable layers (aquitards). The aquifer may be unconfined when it has a free water surface or confined when it is enclosed above and below by aquitards. For each layer geohydrological information is required, such as hydraulic transmissivity, vertical resistance, layer thickness, specific storage, etc. Boundary conditions can be in the form of a flux or hydraulic head. For the modelling of the horizontal water flow in an aquifer, an accurate representation of the geohydrological situation is required. Therefore the region to be modelled is subdivided into finite elements. In this manner we can describe with relative simple element shapes, easily complex geometrical configurations. The finite element method describes the hydraulic head at every point in an element by means of linear interpolation functions.
The smaller ditches spread even over the subregion are primary involved in the interaction between surface water and groundwater. These are commonly called the secondary and tertiary water courses. Aditionally to this a so-called channel system can be present (Fig. 2).

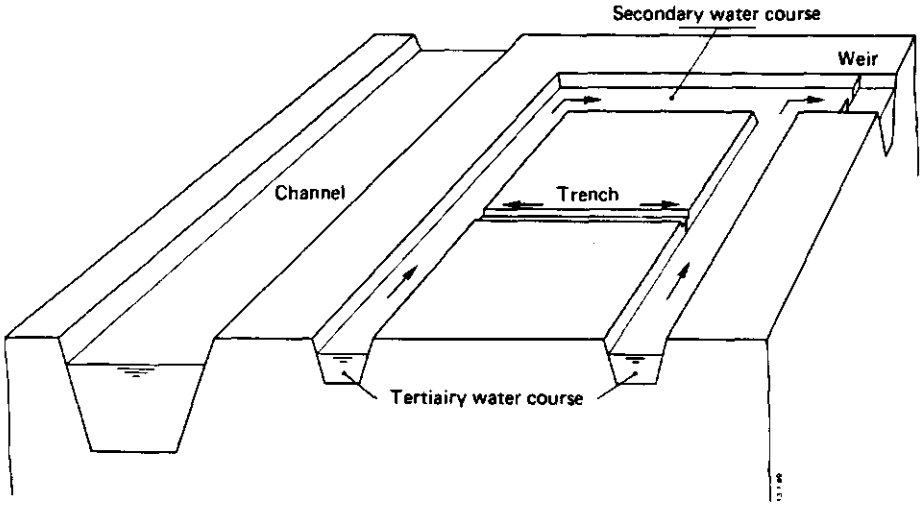

Figure 2. Schematization of interaction groundwater and surface water in four different categories. In the secondary and tertiary water courses the water level fluctuates, in the channel system it remains constant

The secondary system consists of the larger water courses that always contain water. The tertiary system consists of shallow ditches which are intermittently filled with water. This secondary and tertiary system is the schematization of the surface water system on a subregional level. It is assumed that these ditches are spread evenly over the area and are present in all nodes. The channel system can be present as well, but in specific nodes. It should represent the larger channels having its own special influence, often locally, on the groundwater system. It is not related to the area of a node, but the length of channels must be given.

2.3.2. Unsaturated zone

The unsaturated zone has been modelled by means of two reservoirs, one for the root zone and one for the subsoil (unsaturated zone between root zone and groundwater level). The reservoir for the root zone simulates the storage of moisture in the root zone with inflow and extractions as rainfall, evapotranspiration, and capillary rise or percolation. If a certain equilibrium moisture content is exceeded, the excess percolates to the saturated zone. If the moisture content is below the equilibrium moisture content, then the result will be a capillary rise from the saturated zone. From the water balance of the subsoil the height of the phreatic surface is calculated, using a storage coefficient.

Ideally the flow and retention of water in the unsaturated zone should be calculated for each nodal point and per technology separately because:

- the soil physical properties and the groundwater level differ per nodal point
- the potential evapotranspiration differs per technology
- the actual evapotranspiration depends on the soil physical unit, technology and hydrological conditions
- the capillary rise depends on the soil physical unit and the depth of the groundwater level
- the root zone depth may be different per technology

To take all these specific relations and different flow behaviour in the root zone into account would require per nodal point and per technology a model to simulate the unsaturated zone. This would require a great amount

of input data and a heavy demand on both computer time and storage. Therefore, a simplification has been introduced that per subregion and per technology one model (reservoir) is used to calculate moisture content, evapotranspiration and capillary rise (or percolation), using average hydrological conditions. For example the amount of capillary rise in a subregion is now dependent on the average depth of the groundwater level. Because the schematization of the subregions is based on more or less homogeneous conditions with respect to groundwater levels and soil types this simplification is justified. This also means that only one soil physical unit per subregion can be present.

2.3.3. Surface water system

The functioning of the surface water system is different for the summer and winter situation, and therefore require each to be modelled separately according to its special characteristics.
The summer situation is in general characterized by a supply of water. This supply is governed by a certain maximum capacity. Water is extracted from the system for sprinkling and subsurface irrigation. In the winter situation drainage dominates and surface runoff can also occur regularly. The ground level over a subregion can vary by some meters. Taking this into account would mean that for each nodal point one model is required, to simulate the interaction between surface water and groundwater. This would involve a large amount of input data and a heavy demand on computer time. Therefore the surface water system is modelled per subregion as one reservoir with a weir situated at the outlet. In reality a number of weirs may be present within one subregion, but it is assumed that these weirs keep the water level at a constant distance below ground level (Fig. 3). Then the schematization of the surface water system as one reservoir is justified. For each node the water level as a depth below ground level can be translated to a water level relative to the reference datum. The interaction between groundwater and surface water is often very quick responding, which results in rapid water level fluctuations. To overcome this problem the timestep used for calculating the water levels must be small. For the groundwater system the timestep can be much larger. Therefore the surface water system has its own timestep. In between two timesteps of the ground-

water module a number of timesteps of the surface water module are performed. The groundwater level remains constant during those time intervals and the interaction groundwater surface water is simply added up.

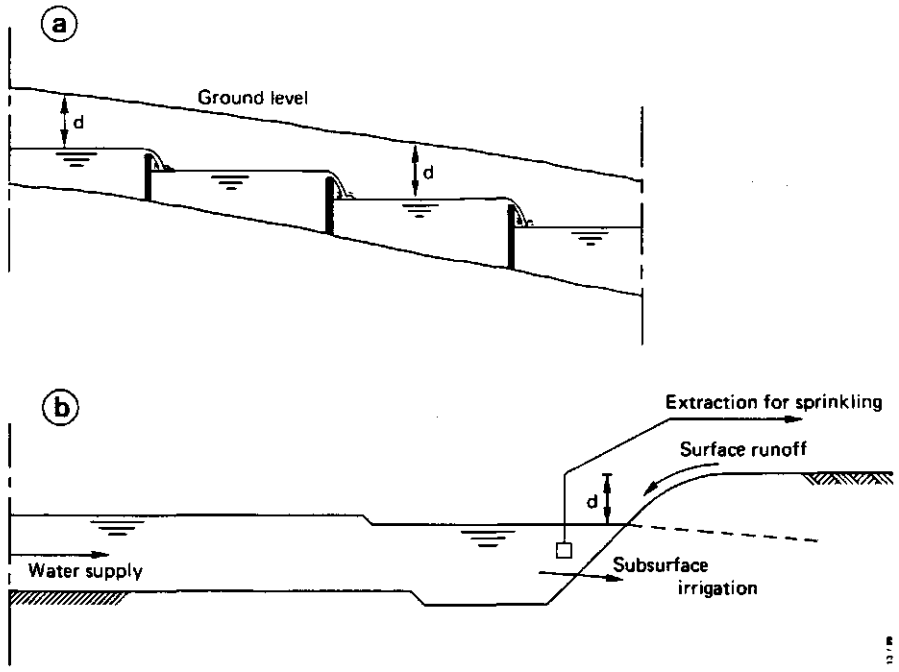

Figure 3. Schematization of surface water system per subregion (d = distance of water level below ground level)
a. in reality, with a number of weirs;
b. schematization as one reservoir per subregion and related to the ground level per node

The various water transport and storage processes are thus simulated by three different submodels. They represent the saturated zone, the unsaturated zone, and the surface water system. The various water movements allowed for within the schematization of a subregion and between the three submodels is shown in Figure 1.

3. CALCULATION METHOD

3.1. SATURATED ZONE

3.1.1. Finite element method

Finite elements have been used for centuries. For example the calculation of the circumference and area of a circle was approximated by dividing the circle in finite elements by the ancient Egyptians. The modern finite element method was originally introduced as a method for analyzing continuous structural members. The method was later formulated in terms of a variational approach and shown to be an extension of the classical Ritz or Galerkin method (ZIENKIEWICZ, 1971).
In the variational formulation, the problem is to find the unknown functions, which maximize, minimize or make stationary a functional, subject to a given boundary condition. In the finite element process the displacements or flux terms are usually specified by functions per element, each nodal parameter influencing only adjacent elements. The region is thus an assemblage of individual elements of either triangular or quadrilateral in shape, as shown in Figure 4.

The system is discretized into irregularly shaped elements. With the finite element method we describe the head at every point (x,y) of an element by means of a linear interpolating function. Within an element with nodes i, j and k numbered counterclockwisely we define a local coordinate system. The head within an element can be written in terms of the nodal values as:

$$h(x,y) = \xi_i h_i + \xi_j h_j + \xi_k h_k \qquad (1)$$

where: ξ are the element interpolation functions that depends on the coordinates of a nodal point. In Figure 5 the interpolation function for a node i is shown, being equal to unity at node i and zero at the other nodes.

To express the elemental contributions to nodal flows we define the flows across the boundary as shown in Figure 6. The flow going through node i can be found as:

$$1/2\ Q_i = -\ (1/2\ Q_k + 1/2\ Q_j) \qquad (2)$$

Per element the flows in terms of nodal values can be written in matrix form as:

$$[L_e]\ \{h_e\} = \{f_e\} \qquad (3)$$

where: the $[L_e]$ matrix is a square coefficient matrix containing the transmissivity between the nodes of the element, the $\{h_e\}$ vector contains the nodal heads and the vector $\{f_e\}$ contains the boundary conditions. For the entire region the other elements follow an identical pattern. To obtain the overall transmissivity matrix $[T_{ij}]$ it is simply necessary to add the contributions of all elements, as:

$$[T_{ij}] = \sum_{m}^{\Gamma} [L_e] \qquad (4)$$

The boundary conditions are either prescribed heads, which means the head remains constant, or a prescribed flux. The flow across a boundary is divided up evenly between adjacent boundary nodes. A distributed re(dis)-charge over an element is incorporated by distributing the element value evenly over the three nodes.

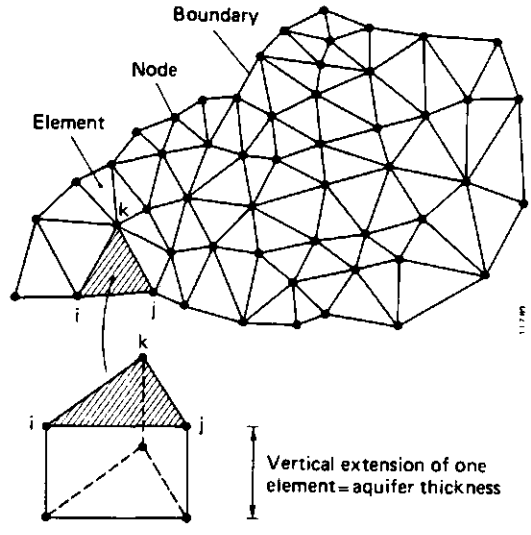

Figure 4. Subdivision of solution domain in finite (triangular) elements. Each element extends over the layer thickness and has nodes i, j and k numbered counterclockwise.

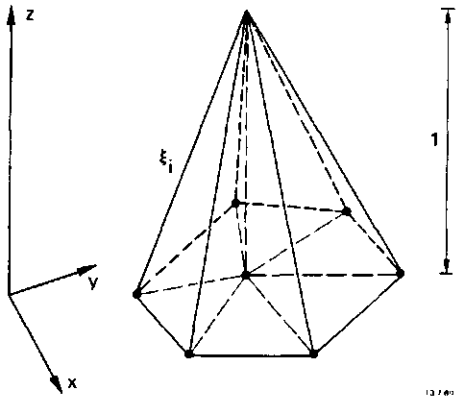

Figure 5. Interpolating function for node i, equal to unity at node i and zero at the adjacent nodes (ξ_i is the element interpolation function)

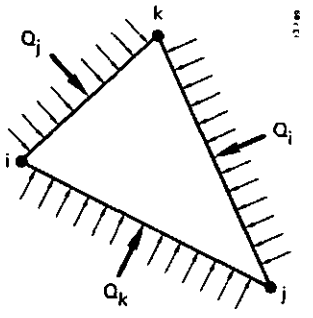

Figure 6. Definition of flows across element boundaries. The flow, Q_i, through node i is equal to half of the flow across the adjacent sides

3.1.2. Calculation scheme and general flow equation

The original programme FEMSAT was based on an explicit calculation scheme (VAN BAKEL, 1978) which means that all external flows imposed on a layer used for the calculations at a certain timestep were taken from the previous time level (Fig. 7). However, the external flows such as discharge to the surface water system or capillary rise also depend on the unknown hydraulic heads at the present time level.

Explicit methods turn out to be stable if the rate which information from a certain point spreads in the computational scheme is smaller than

the physical velocity of propagation of a disturbance. This often means a severe restriction on the timestep to be used.

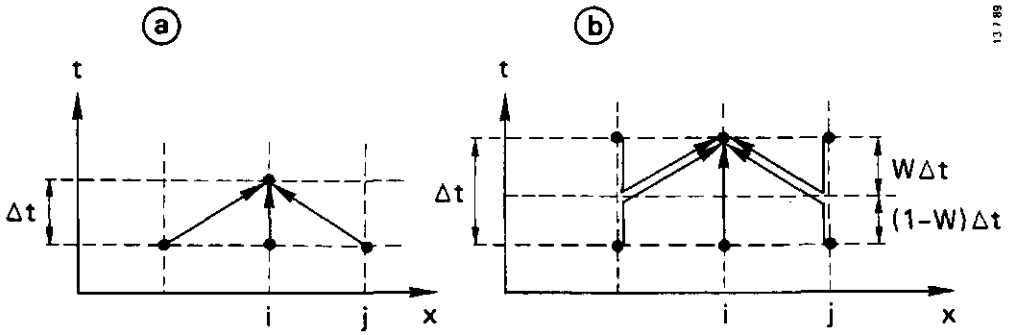

Figure 7. Explicit (a) and implicit (b) calculation scheme.
The implicit scheme is based on the Crank-Nicholson approximation (W = weighting parameter, Δt = timestep)

In an implicit method all factors affecting the flow at a certain time level are used for calculating the flow at that particular time. The timestep can be chosen independently of the spacing of the nodal points. It uses in addition to the flows from the present time level, the flows from the previous time level (Fig. 7). This method is in general unconditionally stable and will not impose restrictions on the timestep to be used. If the weighting parameter W is 0.5, it means a straightforward averaging between the two time levels (Crank-Nicolson approximation).

The equation of motion can be obtained by considering an aquifer layer and applying the principle of linear resistance (Darcy's law) and conservation of mass. According to Darcy's law the rate of flow through a porous medium is proportional to the gradient of the hydraulic head. For instance for flow in the x-direction it can be written as:

$$q_x = -k_x \frac{\partial h}{\partial x} \qquad (5)$$

where: q_x is the volume flux of water passing through a unit area per unit time and k_x is the hydraulic conductivity. The coefficient k in the Darcy

equation is taken to be a constant, depending on both the properties of the porous medium and the fluid.

Unsteady flow conditions will indicate that during a time interval from t to t+Δt a nett quantity of water will flow to or from a node i. The amount of water involved will result in a rise or fall of the hydraulic head. Therefore one can write the continuity equation per node and layer as:

$$A_i\mu \frac{\Delta h}{t} = W \left| \sum_j Q_{ji} + Q \right|^{t+\Delta t} + (1-W) \left| \sum_j Q_{ji} + Q \right|^{t} \qquad (6)$$

where: W is a weighting parameter between the time levels t and t+Δt, A_i is the area of node i and μ is the storage coefficient. The term Q_{ji} is flow from node j to node i and Q are all fluxes from adjacent layers, the unsaturated zone, interactions with the surface water and extractions. The convention is that flow towards a node is assumed to be positive. Equation (6) can be rewritten as:

$$A_i\mu \frac{\Delta h}{\Delta t} = \sum_j Q_{ji}^{t} + Q_e^{t} + W \left| \sum_j \Delta Q_{ji} + \frac{dQ_e}{dh} \Delta h \right| + Q_t \qquad (7)$$

assumming that:

$$Q^{t+\Delta t} = Q^t + \Delta Q \qquad (8)$$

and

$$h^{t+\Delta t} = h^t + \Delta h \qquad (9)$$

where: Q_e is the boundary flow dependent on the groundwater level (leakage and drainage to large channels) and Q_t are constant fluxes such as drainage to the secondary and tertiary system, capillary rise and extractions for water supply and sprinkling.

The first two terms on the right-hand side of Equation (7) represent the flows to or from node i at time t and the third and fourth term are the

actual change in flow over the considered timestep. Equation (7) is solved by the Gauss-Seidel iterative method and using an over-relaxation factor (see Annex A).
The linearization of the equations has been done to avoid re-evaluating certain level dependent parameters after each iteration. To include the unkown boundary flows, it would require also to include their contributions in the equation of motion. Therefore the equation for a boundary condition must be written as a function of the unknown hydraulic head.

The flow between two nodes in an aquifer can be written as:

$$Q_{ji} = T_{ij} (h_j - h_i) \tag{10}$$

where: T_{ij} contains the transmissivity parameters for horizontal flow in a water bearing layer and explained in Paragraph 3.1.1. The flow between two nodes is linearly related to the difference in hydraulic head. Equation (10) can therefore also be used to define a change in flow given the changes in hydraulic head between two adjacent nodes. The flow towards a node is taken as positive and from a node as negative.

The external flow is composed of the following flux terms:

$$Q_e = A_i q_l + L_i q_s \tag{11}$$

and the constant flux term is:

$$Q_t = A_i \left(- q_c - q_e + \sum_{\frac{\Delta t}{\Delta t_s}} q_w\right) \tag{12}$$

In which A_i is the area of the node, q_s is the flow from the channel system per unit length, l_i is the total length of the channels for node i, q_l is the leakage, q_w is the flow from the surface water system, q_c is the capillary rise and q_e are the extractions for public water supply as well as for sprinkling. For the calculation of the water levels in the surface water water system a timestep Δt_s is used, which is in general smaller than the timestep for the groundwater system. The drainage for a timestep of the groundwater system is computed as the sum of the drainage for the time steps of the surface water submodel.

If we consider a water bearing layer enclosed between two less-permeable layers, then the vertical flux for layer L can be expressed as:

$$q_l = \frac{h_{i,L+1} - h_i}{0.5\ c_{i,L+1}} + \frac{h_{i,L-1} - h_i}{0.5\ c_{i,L-1}} \qquad (13)$$

where: c_i is the hydraulic (vertical) resistance of the less-permeable layer defined as:

$$c_i = \frac{D_i}{k_i} \qquad (14)$$

where: D_i is the vertical thickness of the layer and k_i the hydraulic conductivity.

3.1.3. Drainage

The ditches that actually are involved in the interaction between surface and groundwater are commonly the combined secondary and tertiary water courses, together with the shallow trenches. Additional to this a so-called channel system can be present (see also Fig. 2).
The secondary system consists of the larger water courses that always contain water. The tertiary system consists of shallow ditches which are intermittently filled with water. The secondary and tertiary system together forms the schematisation of the surface water system on a subregional level. It is assumed that these ditches are spread evenly over the area and are present in all nodes. The interaction between groundwater and surface water is only possible with the toplayer. The channel system can be present as well, but only in specific nodes. It should represent the larger channels having their own special influence, often locally, on the groundwater system. It is not related to the area of a node, but the length of channels must be given for certain nodes. An interaction with all layers defined in the model is possible.

The interaction between the three subsystems (2nd, 3rd and trenches) and the groundwater system can be described as (ERNST, 1978):

$$q_w = \sum_{n=1,3} \frac{(h_w - h_i)}{\alpha_s \, T_s} \qquad (15)$$

where: h_w is the water level or invert level of a subsystem, α_s is a geometry factor dependent on the shape of the groundwater table and T_s is the so-called drainage resistance. Here is h_w the highest value of either the invert level or the water level. The geometry factor α_s is necessary, when using the average level for a nodal point, instead of the level midway between two ditches. The drainage resistance is either kept constant or used as a function of the depth of the groundwater; in the latter case it is dependent on the number of ditches per unit area and the geohydrological properties of the soil. The drainage resistance can then be written as:

$$T = a \exp (b * h_{min}) \qquad (16)$$

with:

$$h_{min} = \min (h_g - h_i \, , \, h_g - h_w) \qquad (17)$$

where: a and b are constants, h_{min} is the minimum distance between ground level h_g and the groundwater level h_i or surface water level h_w. This means that when the groundwater level is lower than the surface water level infiltration takes place.

For the channel system the equation per unit length of ditch is:

$$q_s = \beta \, (h_c - h_i) \qquad (18)$$

with

$$\beta = - \frac{1}{R_r + R_e \, / \, P} \qquad (19)$$

in which R_r is the radial resistance, R_e the entrance resistance and P is the wetted perimeter of the channel.

The entry resistance R_e is strongly dependent on local conditions and cannot be calculated explicitly, but must be measured on site and is part of the input of the model.

The radial resistance, R_r can be calculated with the equation:

$$R_r = \frac{1}{\pi k} \ln \frac{\alpha_c d}{\pi P} \qquad (20)$$

in which k is the hydraulic conductivity, d is the thickness of the saturated layer and α_c is a coefficient depending on size of the channel. If the channel is small (depth less than 2.50 m), then α_c is 4.0 to 5.0, and for larger channels α_c is 6.0 to 7.0. The upper limit can be selected for empty stage and the lower limit for bankfull channels.

3.1.4. Interaction with unsaturated zone

For the characteristics of the unsaturated zone certain parameters are related to the depth of the groundwater level (see Paragraph 3.2). Because the unsaturated zone is considered on a subregional level the average ground level and depth of the groundwater level are required. After the hydraulic head in each node is calculated with Equation (7) and (9), the average head per subregion is calculated as the weighted average of the area per node as:

$$\bar{h}_r = \frac{\sum_{nr} h_i * A_i}{A_r} \qquad (21)$$

where: A_i is the area of node i and A_r is the area of subregion r, and n_r is the number of nodes per subregion. For the average ground level of a subregion the same procedure has been followed.

The storage coefficient used in Equation (7), is dependent on the average depth of the groundwater level in a subregion and the soil physical unit. Within a subregion an average root zone depth is calculated as the weighted average of the area per technology. This average depth is used for the calculation of the storage coefficient.

The extractions from groundwater for irrigation, and the percolation or capillary rise, are calculated on the aggregation level of a subregion. Subsequently the fluxes are attributed to the nodes of a subregion by multiplication with the relative areas of the respective nodes. Therefore the flux to/from the unsaturated zone can be calculated for each node in a subregion as:

$$q_c = \frac{A_i}{A_r} \left(\sum_j q_c^j * A_j \right) \qquad (22)$$

where: A_j is the area of land allocated to technology j within each subregion. In a similar manner the extraction from groundwater for sprinkling is allocated to the nodes.

The groundwater extractions for the public water supply are attributed to a single node per subregion.

3.2. UNSATURATED ZONE

3.2.1. Water balance of root zone

The unsaturated zone is modelled by means of two reservoirs, one for the root zone and one for the subsoil (i.e. the unsaturated zone between root zone and groundwater level). The reservoir for the root zone simulates the storage of moisture in it with inflow and outflows as precipitation, evapotranspiration, and capillary rise or percolation. If a certain equilibrium moisture content (moisture content corresponding with a steady situation where capillary rise is zero) is exceeded, the excess will percolate to the saturated zone. If the moisture content is below the equilibrium moisture content, then the result will be a capillary rise from the saturated zone. From the water balance of the subsoil the height of the phreatic surface is calculated, using a storage coefficient which is dependent on the depth of the groundwater level.

As stated in Chapter 2, ideally the flow and retention of water in the unsaturated zone should be calculated for each nodal point and per technology separately. The simplification introduced is that per subregion and per technology one model (reservoir) is used for calculating moisture con-

tent, evapotranspiration and capillary rise. In this case average hydrological conditions over the subregion are used. For example the amount of capillary rise in a subregion is now dependent on the average depth of the groundwater level. A schematization of the unsaturated zone is shown in Figure 8 and the major functions are summarized below.

The root zone for which the water balance is considered has a depth d_r, which is a function of the technology and the soil physical unit. Therefore:

$$d_r = f\ (j\ ,\ s) \tag{23}$$

In the model a constant root zone has been assumed all year round, with no changes during and over the years.

For the root zone of each technology with area A_j the change of moisture content ΔV_w over a timestep Δt due to net precipitation P_n, sprinkling P_s and evapotranspiration E is:

$$\Delta V_w = (P_n + P_s - E) * \Delta t \tag{24}$$

The rainfall is corrected for plant interception and has a maximum infiltration rate. Interception is assumed to be present in summer and dependent on the technology. If rainfall exceeds the maximum infiltration rate, this excess is considered to become surface runoff. Due to irregularity in sprinkling it has been assumed that 10% of the sprinkling is not stored in the root zone, but percolates directly to the saturated zone. Irrigation by means of sprinkling will take place if the condition for the considered technology is valid (see Paragraph 3.2.3). Evapotranspiration is computed as a function of the crop and moisture content in the root zone (see Paragraph 3.2.2). Without considering percolation or capillary rise the moisture content for the next timestep would be:

$$V = V_w^t + \Delta V_w \tag{25}$$

The moisture content of the root zone at equilibrium condition V_{eq} is calculated with the function:

$$V_{eq} = f\ (s\ ,\ d_r\ ,\ \bar{h}_r^*) \qquad (26)$$

where: $\bar{h}_r^*$ is the average depth of the groundwater level in a subregion. The equilibrium moisture content is required per soil physical unit and for root zone depths of 0.25 m, 0.50 m and 1.00 m (Annex C). If the moisture content V is above the equilibrium, percolation occurs, otherwise there is capillary rise. Therefore:

$$Q_c = f\ (s\ ,\ \bar{h}_r^*\ ,\ d_r) \qquad V < V_{eq} \quad \text{for capillary rise} \qquad (27)$$

or

$$Q_c = \frac{V_{eq} - V}{\Delta t} \qquad V > V_{eq} \quad \text{for percolation} \qquad (28)$$

The moisture content for the next timestep can now be calculated as:

$$V_w^{t+\Delta t} = V + \Delta t * Q_c \qquad (29)$$

The model concept for the unsaturated zone has been verified with results from a more accurate model SWATRE (BELMANS et al, 1983), as given in Paragraph 4.2.1.

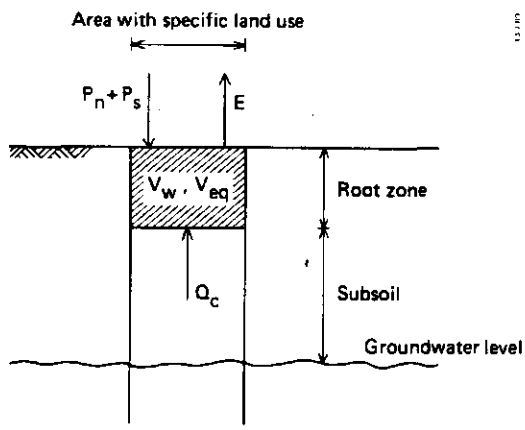

Figure 8. Schematization of unsaturated zone per land use.
Nomenclature used: P_n = net precipitation; P_s = sprinkling; E = evapotranspiration; V_w = moisture content; V_{eq} = equilibrium moisture content and Q_c = capillary rise

3.2.2. Evapotranspiration

From the meteorological data the potential evapotranspiration for grassland is calculated with the Makkink equation (DE BRUIN, 1987). The potential evapotranspiration for other crops or vegetation are calculated using crop factors (FEDDES, 1987). The potential evapotranspiration for pine-forest is calculated as the sum of transpiration and interception. An interception reservoir of 2.0 mm and 1.5 mm was taken for the summer and winter period respectively. The actual evapotranspiration E is calculated as:

$$E = \alpha E_p \tag{30}$$

with

$$\alpha = f\ (V_w\ /\ V_{eq}^o) \tag{31}$$

where: E_p is the potential evapotranspiration, V_w is the actual moisture content of the root zone and V_{eq}^o is the equilibrium soil moisture content in the root zone for groundwater level equal to ground level. In this way the effect of water-logging can be incorporated. The relative evapotranspiration can be determined from Figure 9.

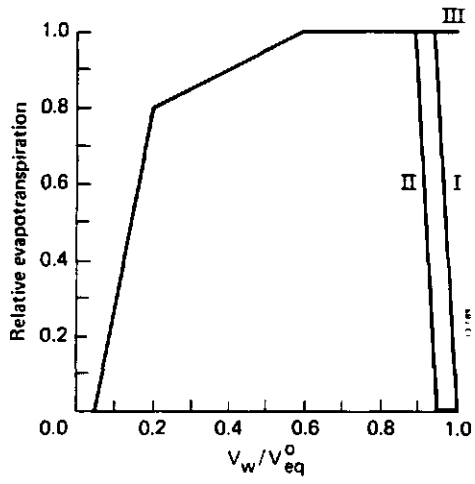

Figure 9. Relationship for calculation relative evapotranspiration from soil moisture conditions (I = standard curve, II = plants sensitive to water logging, III = plants in nature areas)

Figure 9 shows that root water uptake is zero when V_w/V_{eq}^o is below 0.05 (wilting point). When this ratio is equal to unity (anaerobiosis point) certain plants will have zero rootwater uptake, which is shown by line I in Figure 9. Line II is for plants which are very sensitive to water-logging (e.g. potatoes). Line III is for nature areas, because it has been assumed that natural vegetation is especially suited for these wet conditions and a reduction in evapotranspiration will not occur.

3.2.3. Sprinkling

In practice sprinkling is operated following a rotation scheme along separate fields. The sprinkling is continued as long as the soil moisture content is below a certain level. The groundwater flow model cannot allow for a fully realistic simulation of sprinkling according to a rotation scheme, but depending on the production levels of the technologies the sprinkling is operated. A rotation scheme of say 7 or 14 days can be selected. This means that in 7 or 14 days the total area of the specific land use is sprinkled with 25 mm. A loss of 5% is taken into account and 10% of the water is not stored in the root zone but percolates directly to the saturated zone, due to irregularities in the distribution of the sprinkling (Eq. 24). For a high production level a high water demand is necessary, which results in frequent sprinkling. The threshold values for applying sprinkling depend therefore on the production level and the relative moisture content in the root zone. The relative moisture content is defined here as the present moisture content divided by the equilibrium moisture content for a depth of the groundwater level of 1.0 m (V_w/V_{eq}'). In Table 1 the criteria for sprinkling are given. Subsequent to the starting of sprinkling a check is made to see if the moisture content does not exceed the criterium for stopping.

In general the extraction for sprinkling from surface water is preferential to the extraction from groundwater. But due to limitations on the availability extraction from groundwater is possible as well. The maximum extraction from groundwater and surface water per subregion must be given. It reflects the availability of surface water within the subregion.

Table 1. Criteria for sprinkling related to production level (high production level means a high water demand and the relative moisture content (V_w) is the present moisture content divided by the equilibrium moisture content for a depth of the groundwater of 1.0 m (V'_{eq})

	Relative moisture content (V_w/V'_{eq})	
Production level	start	stop
0	no sprinkling	
1	0.60	0.75
2	0.70	0.85
3	0.80	0.95

Land more than say 150-250 m from the water courses uses in general groundwater. Also dependent on the water supply and the amount of subsurface irrigation the extraction from the surface water system can be limited.

The maximum extraction from surface water is allowed when the supply is enough to maintain the summer target level. When the supply is not enough the water level becomes less than the target level. If it becomes below a certain level, the extraction for sprinkling is reduced.

The capacity available for sprinkling becomes now the supply minus the subsurface irrigation. When this occurs, it will result in a reduction of the sprinkling rate according to this reduced capacity.

3.3. SURFACE WATER SYSTEM

The secondary and tertiary water courses are considered to represent the surface water system. It is often a very quickly responding system with rapid water level fluctuations. Therefore the timestep used for calculating water levels must be small, say between 0.2 and 1.0 day.

For the groundwater system the timestep can be much larger, say between 1 and 7 d. Therefore the surface water system has its own timestep. In between two timesteps of the groundwater module a number of timesteps of the surface water module are performed. The groundwater level remains constant during that time and the interaction between groundwater and surface water is added up. The next time the groundwater module is activated this amount of drainage or subsurface irrigation is used for calculating a new groundwater level.

The surface water system in a subregion is modelled as a single reservoir with a weir situated at the outlet. In reality a number of weirs may be present. But under the assumption that the weir levels are a constant distance below ground level (Fig. 3). The schematization of the surface water system as a single reservoir is justified. Within a subregion it is possible that in the higher parts subsurface irrigation occurs and in the lower parts drainage. This drainage water cannot fully be reused in the same subregion. Simplifying the surface water system as a single reservoir would mean lifting up water from the lower parts, to the higher parts of the subregion. So in the model only a fraction of this drainage water can be reused in the same subregion.

The water level is calculated from the stage discharge or stage-storage relation, depending on the hydrological situation. Included in the stage-discharge relation is the flow resistance within the system, this is also dependent on the drainage rate. When subsurface irrigation is taking place the storage capacity of the system is taken into account. From the supply, the subsurface irrigation, the extractions (sprinkling) and the stage-storage relation a new water level is calculated. This water level in a subregion is a distance below ground level. From this distance the actual water level for all nodal points is calculated.

Depending on the time of the year and the hydrological situation the water level h_w of the surface water is calculated as:

drainage

h_w = f (Q-h relation) for winter period

h = target level for summer period

subsurface irrigation

h_w = f (S-h relation)

where: Q-h is the stage-discharge curve for the weir that includes the flow resistance and S-h is the stage-storage curve. Drainage situations in summer do not effect the water level. It is assumed that automatically the weirs are lowered to maintain the same level. In drainage situations the storage capacity is very small and plays a minor role. Often the storage of the system is equal to the drainage rate for one to three days. For drainage situations the discharge curve is therefore used for calculating the water

level (only winter period). The storage capacity is only important for the summer period when there is an external source of supply. When the supply is smaller than the subsurface irrigation plus the extraction for sprinkling, the storage in the system is used for calculating the lowering of the water level. The characteristics of the surface water system must be given in the form of a stage-storage-discharge relation. For the summer and winter period a target level must be specified. In the summer period, from 1st of April until a specified time, water level control is possible by manipulation of weir levels. Water level control in the region can be differentiated into categories, each having its own criteria of weir level manipulation in relation to the groundwater level. This means that in the early summer, when groundwater levels are relative high, the target level for the winter period is maintained. When the groundwater lowers the target levels are raised to levels that are allowed in the summer situation.

The maximum supply capacity per subregion must be seen as the maximum capacity for which the system is designed. For instance the watercourses are dimensioned in such a way that a supply of say 1 mm per day is possible. The supply capacity for the whole region can have a limit as well. External factors such as pumping stations or inlet works can be a limit for the total supply. If this capacity is reached, then the maximum supply capacity per subregion is reduced.

The model does not take into account the surface water system of the region as a network. Their is no influence between subregions, so a discharge from a subregion is not transported through other subregions.

4. MODEL APPLICATIONS

4.1. INPUT DATA

For the operation of the program, four fundamental sets of data are required in addition to the control parameters. These are:
- node and element characteristics
- boundary conditions
- unsaturated zone characteristics
- time dependent data

For each nodal point and per layer specific geohydrological data are required. Other input data per node are the coordinates, elevation of the ground surface and the invert levels of the surface water system. The configuration of the elements is given by the node numbers.

The groundwater flow model was developed and tested for a region, approximately 35,000 ha, in the South Western part of The Netherlands (Southern Peel region). Based on the soil physical properties six different soil physical units where distinguished, all of them typical sandy soils. This in relation with the hydrological situation of the groundwater fluctuations within the area resulted in 31 subregions (SMIDT, 1983). In Figure 10 the finite element network is shown for the region, 748 elements with 404 nodal points. The boundary conditions were given in the form of fluxes per half year (WIT, 1986).

In the following paragraphs, typical aspects, such as soil physical properties, drainage resistances and surface water system characteristics are discussed.

4.1.1. Hydrological parameters

From field measurements it has been found that the top layer in the Southern Peel region can be modelled as an aquitard. The second and fourth layers are aquifers, and the third layer is an aquitard.

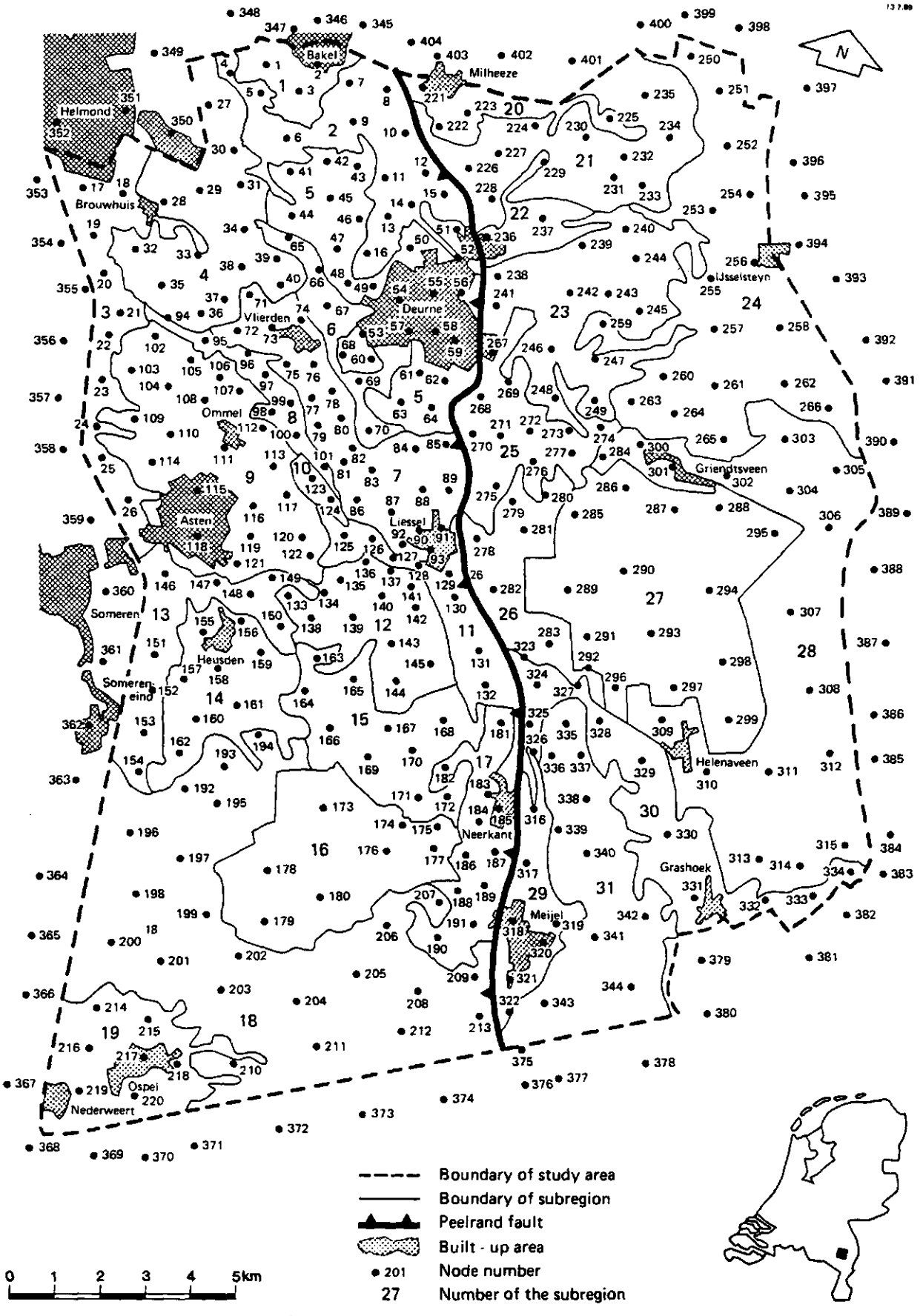

Figure 10. Finite element network of Southern Peel region with 404 nodes, 748 elements and 31 subregions. The Central Slenk is left of the fault and the Horst is on the right. The location of the study area in the Netherlands is indicated at the foot of the figure

These four layers are present in the Central Slenk area which is on the west side of the Peelrand fault (Fig. 11). On the Peel Horst the third and fourth layer are not present and the hydrological basis is below the second layer. The soil properties of each layer in the Central Slenk and Peel Horst area are given
in Table 2 (REES VELLINGA and BROERTJES, 1984; WIT, 1986).

In Table 2 the specific storage is the volume of water released or stored in an aquifer or aquitard by a change in hydraulic head.

Table 2. Typical geohydrological properties of the Southern Peel region

Layer	Layer thickness (m)	Vertical resistance (d)	kD ($m^2.d^{-1}$)	Specific storage (m^{-1})
Central Slenk				
1	25	100-2,500		2.10^{-5}
2	45-50		750-3,500	2.10^{-5}
3	110	1,500-20,000		2.10^{-5}
4	160		5500	2.10^{-5}
Horst				
1	4-25	1,000-2,000		2.10^{-5}
2	4-34		200-2,000	2.10^{-5}

4.1.2. Drainage resistance

For computing the interaction between groundwater and surface water Equation (15) uses a so-called drainage resistance. This resistance can be derived from the density of the surface water network as:

$$\gamma = \frac{L_d^2}{8\ k\ D} + L_d\ R_r \qquad (32)$$

with L_d as the average distance between the ditches, kD the transmissivity and R_r the radial resistance (see also Eq. 20). For various depths of the groundwater level and different channel sizes the resistance can be calculated. It is assumed that all the ditches are free draining.

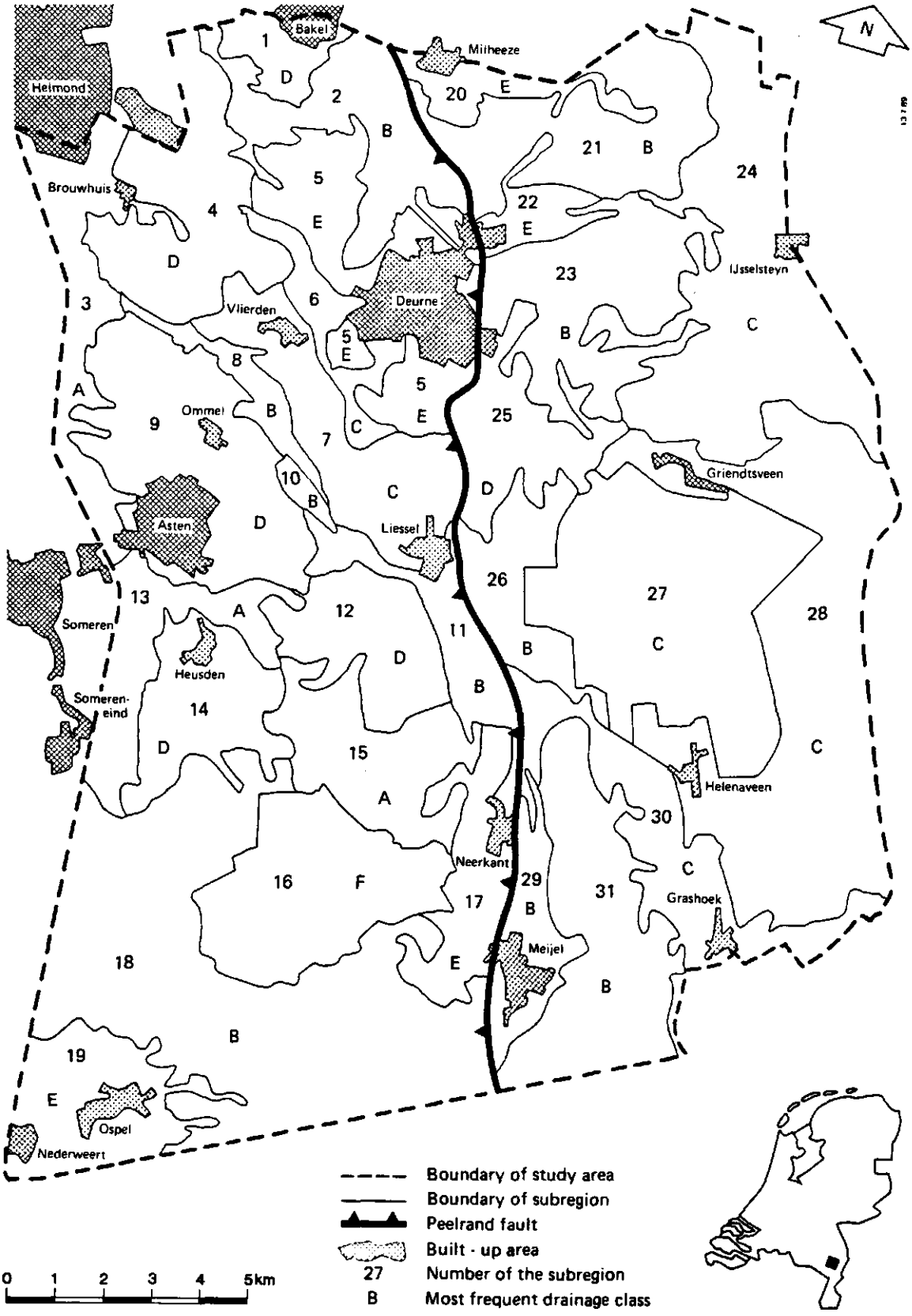

Figure 11. Subregions in the Southern Peel region given the hydrological and soil physical characteristics, together with the adopted drainage classes A to F. The classes refer to the drainage resistance curves as shown in Figure 12

To simplify the derivation of the drainage resistance as a function of the depth of the groundwater level, six different classes of drain density have been distinguished (classes A to F). The classes A to F refer to an overall density of ditches and brooks per subregion. Class A has a dense drainage system and class E has hardly any drainage. Class F refers to the two nature reserves in subregion number 16 and 27. The selected class per subregion is shown in Figure 11. In some subregions there is quite a variation of ditch density. In these cases the most frequent drainage class has been selected.

The ditches and brooks were classified in relation to the depths (ERNST, 1978). For each average depth per ditch category the drainage resistance was calculated with Equation (32). An equivalent drainage resistance for all the categories was calculated for specific depths. For these values the following exponential functions were derived:

$$\Upsilon = a * \exp(b * h^*) \qquad (33)$$

where: Υ is the drainage resistance, h^* is the depth of the groundwater level, and the constants a and b are dependent on the classes A to F (Table 3). The derived drainage resistance as a function of groundwater level classes is given in Figure 12.

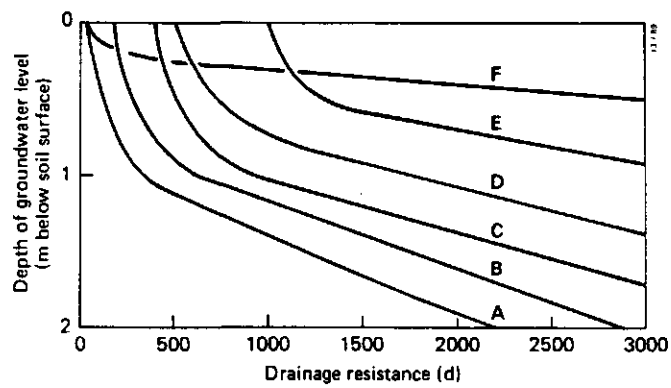

Figure 12. Drainage resistance characteristics in the classes A to F (see also Table 3)

Table 3. Values of coefficients a and b for calculation of drainage resistance with Equation (33): $T = a * \exp(b * h^*)$

Class	a	b
A	50	1.94
B	165	1.45
C	325	1.20
D	500	1.20
E	1000	1.20
F	25	14.00

4.1.3. Surface water system

From the surface water system the following characteristics were required per subregion:

- storage capacity
- maximum supply capacity
- water level (target) for summer and weir level for winter period
- stage-discharge relation for drainage situation

The storage capacity and discharge could be derived from the defined ditch density per subregion (Class A to F). The maximum supply capacity was determined from field measurements and information from local Water Boards.

4.1.4. Land use and extractions for public water supply

The land use in the form of technologies is based on field observations (REINDS, 1985). In this way seven agricultural technologies where distinguished, each having a fraction with sprinkling and a fraction no sprinkling.

The extractions for public water supply are situated near Vlierden in subregion 7 and near Ospel in subregion 18 (Fig. 11). The pumping station situated in subregion 7 extracts water from the shallow aquifer (second layer), and the pumping station in subregion 18 extracts water from the deep aquifer (i.e. fourth layer). The average capacities of these pumping stations are:

Vlierden	-	9630	$m \cdot d^{-1}$
Ospel	-	5900	$m \cdot d^{-1}$

4.2. APPLICATIONS

In the model the hydrological processes are modelled as realistically as possible. The constraints are in general a lack of data and required computational effort, which can influence the results of some processes to a certain degree. The results from the unsaturated zone in the form of evapotranspiration and capillary rise is compared with results from a more accurate model. The hydrological schematization and the input parameters are compared with field measurements. A sensitivity analysis on the geohydrological parameters has been done to determine the accuracy of the results. All these aspects are discussed in the following Paragraphs 4.2.1 to 4.2.3.

4.2.1. Unsaturated zone

The simplified calculation method proposed for the water movements in the unsaturated zone (Paragraph 3.2.) has been tested as a stand-alone program (module SIMUNS). In this way a good comparison with results of other models was possible. For the saturated part a relation is defined to describe the flow to the surface water system and the seepage (Fig. 13). The computed results of this model could be compared with results from the SWATRE model. This model is a transient one-dimensional finite-difference model for the unsaturated zone with water uptake by roots (BELMANS et al, 1983).

In the present discussion the comparison will be restricted to the hydraulic heads and water balance terms of the unsaturated zone, calculated for the hydrological year 1975 (1st Oct 1974 to 30th Sept 1975). The results of the two models are given in Table 4, from where it can be seen that there is a reasonable agreement of the calculated results. The simplified approach has the tendency to yield less evapotranspiration (9-32 mm) and less capillary rise (0-18 mm). The storage coefficient used for the saturated zone is in the program assumed to be dependent only on the depth of the groundwater level. It should actually also be dependent on the

magnitude of the capillary rise or percolation. However, if we consider the extent of the introduced simplifications in this model, these results are quite satisfactory.

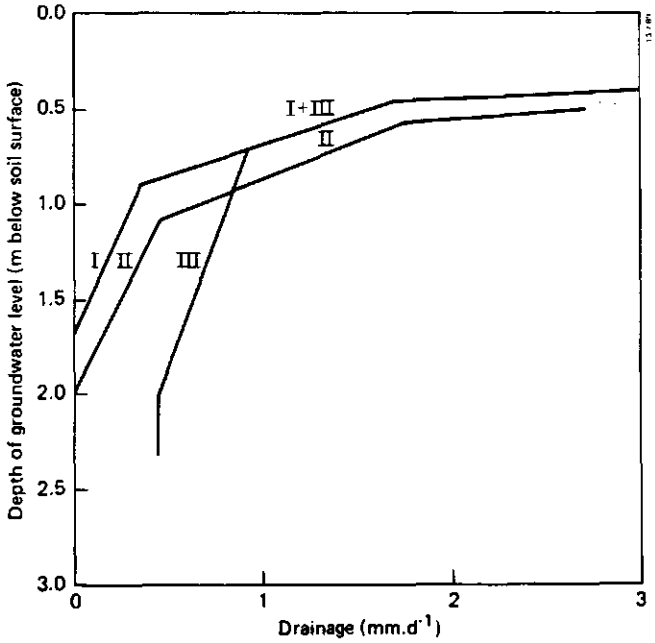

Figure 13. Imposed typical sink terms as the boundary condition for the saturated zone as found in the Southern Peel region (I = standard curve, II = plants sensitive to water logging, III = plants in nature areas)

Table 4. Comparison of model results from SIMUNS and SWATRE for grassland with root zone of 0.25 m for two soil physical units found in the Southern Peel region (sandy soil) (A = SIMUNS and B = SWATRE, potential evapotr. = 478 mm; for boundary conditions see Figure 13)

Soil unit	Bound. cond.	Groundwater level 1-04-75 (m)	1 - 10 - 75 A	1 - 10 - 75 B	Evapotranspiration (mm) A	Evapotranspiration (mm) B	Capillary rise (mm) A	Capillary rise (mm) B
5	I	0.54	1.23	1.26	406	420	41	40
8	I	0.47	1.66	1.70	447	459	87	83
5	II	0.68	1.34	1.37	389	398	28	20
8	II	0.65	1.73	1.78	416	448	60	76
5	III	0.54	1.40	1.52	394	407	27	28
8	III	0.53	1.86	2.14	412	444	50	68

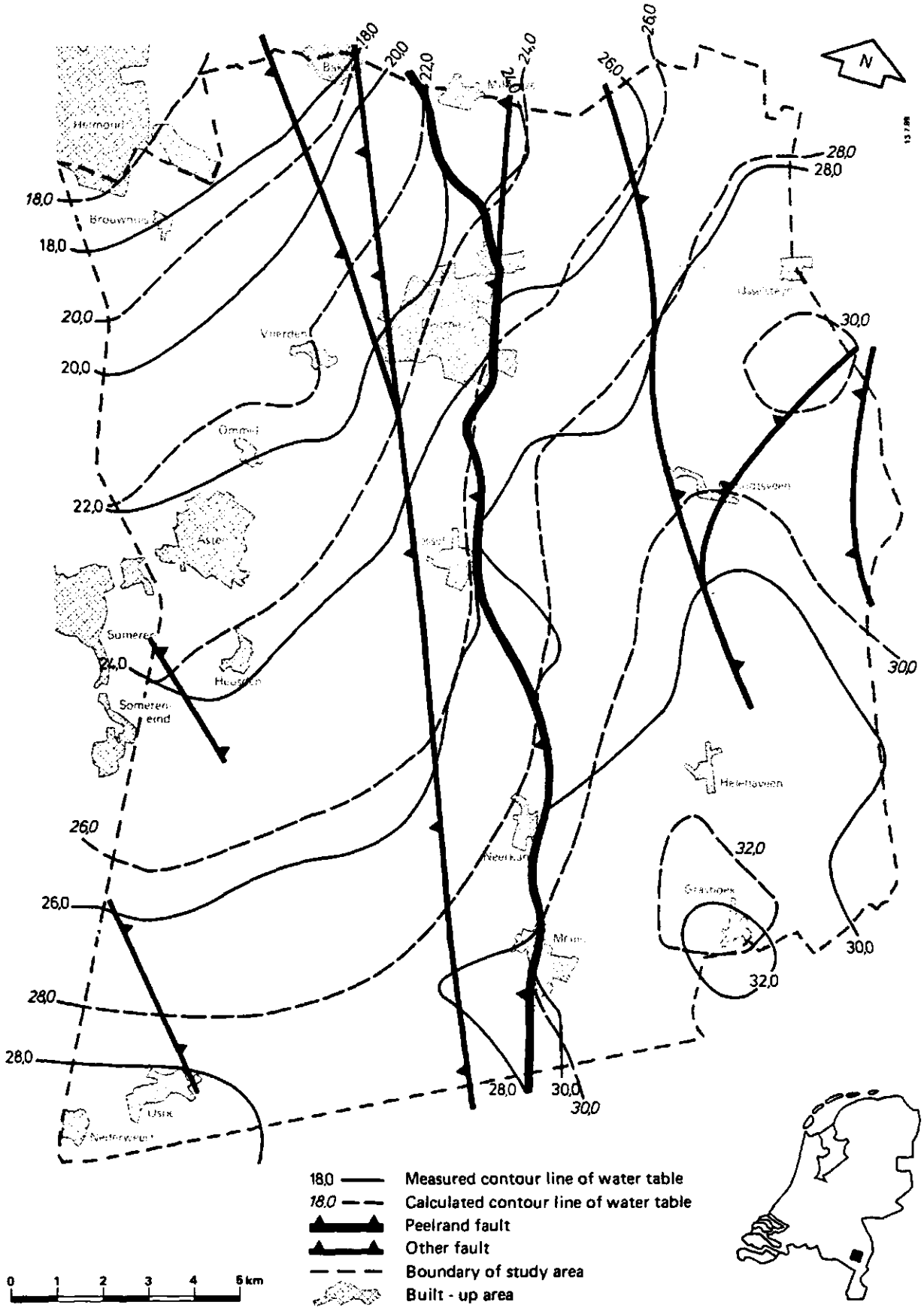

Figure 14. Measured and calculated hydraulic heads from august 1982, for first aquifer of the Southern Peel region (major fault divides Central Slenk, left side, from the Horst area)

4.2.2. Groundwater levels

For a rough verification of the entire model the computed results, in the form of groundwater levels, are compared with measured data. This comparison was undertaken for the year 1982, of which the actual land use and available sprinkling capacity are known.

The meteorological data were obtained from the Royal Dutch Meteorological Office (KNMI, personal communication).

Groundwater levels from eight measuring stations during the year 1982 were compared. These results are discussed in Annex D, from which it can be concluded that the calculated results in the Central Slenk area resemble the measured data reasonably well and that in the Horst area some differences occur (see also Figure D3 and D4 of Annex D for some results).

The calculated levels in the first aquifer for August 1982 are compared with the measured values. In Figure 14 the patterns of the contour line of water tables are given.

The calculated contour lines show a more regular pattern, because in the case of measured values there may be all kinds of local anomalies and also errors in the measurements. Another difference is the more smooth transition in the calculated values in the neighbourhood of the Peelrand fault. This is caused by the relatively coarse nodal network.

In general, however, the resemblance between calculated and measured isoline patterns seems satisfactory. In the Horst area the difference between calculated and measured levels is very small (0.1 - 0.4 m). For the Central Slenk area the same applies as for the Horst area, except in the North-West corner near the region boundary where the differences become greater closer to the boundary (up to 1.0 m). A hydrological aspect which perhaps is not included in the model or else an error in the boundary condition could be the possible cause of this difference.

4.2.3. Sensitivity analysis

Various parameters have been varied in order to analyse the effects of these variations on the results. The discussion of the results has been restricted to the average standard deviation as discussed in Annex D. The results of the sensitivity analysis are given in Annex E. The conclusions drawn from these results are:

- The variation of depth of the groundwater level at the beginning of the summer half year is more pronounced than at the end of the period.
- The depth of the groundwater level at the beginning of summer is dominated by the drainage resistance.
- The soil physical properties are very important for the correct estimation of the results at the end of the summer period.
- Variation of the geohydrological parameters has hardly any effect on the total sprinkling, actual evapotranspiration, and capillary rise.

5. CONCLUSIONS

The results of a model approach to evaluate hydrological effects are presented. The calculation method discussed in Chapter 3 still uses simple relations to describe the affected variables. It is now possible to simulate the hydrological processes in the saturated zone and the unsaturated zone in an integrated manner. The relations between change in groundwater level and capillary rise (or percolation), which are non-linear and time dependent could be taken into account.

The aim was to simulate the rather complex processes involved in such a manner that it is accurate enough without requiring too many input data and too much computer time. In this respect the use of subregions for the schematisation of the unsaturated zone and the surface water system has the advantage of reducing required input data and computational time. One can now simulate regional groundwater flow problems easily and a large number of alternatives can be calculated and compared. Using simulation models one can perform experiments on paper. For instance the effect of certain radical changes in the regional system, by means of land consolidation, water level control, etc. The effect of these changes within the agricultural area and for instance adjacent nature reserves can be quantified.
It is possible to simulate surface water management and the effect on the groundwater system, by means of water level control.

As shown by the examples the model can be used for all kinds of hydrological problems. Other topics not discussed in this paper are the effect of water level manipulations on crop production and water conservation (VAN BAKEL, 1988, QUERNER and FEDDES, 1988). The groundwater model can also be used in connection with a model to describe the flow in surface water networks (QUERNER, 1986). This connection can be necessary for problems more related to the surface water system, such as water distribution and maintenance.

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LIST OF SYMBOLS

A_i	-	area representative for node i	m^2
A_j	-	area per technology	m^2
A_r	-	area for subregion r	m^2
a	-	constant	-
b	-	constant	-
c	-	hydraulic (vertical) resistance	d
d_r	-	root zone depth	m
D	-	thickness of aquifer or aquitards	m
E	-	actual evapotranspiration	$m.d^{-1}$
E_p	-	potential evapotranspiration	$m.d^{-1}$
(f_e)	-	vector with boundary conditions per element	$m^3.d^{-1}$
h_a	-	mean standard deviation	m
h_c	-	water level in channel system	m
h_g	-	ground level at node i	m
h_i	-	hydraulic head for node i	m
h_m	-	measured hydraulic head	m
h_w	-	water level in surface water system	m
h*	-	depth of the groundwater level below ground surface	m
j	-	technology considered	-
k	-	hydraulic conductivity of aquifer layers	$m.d^{-1}$
$[L_e]$	-	transmissivity matrix per element e	$m^2.d^{-1}$
L	-	layer index	-
L_d	-	average distance between ditches	m
m	-	number of observations	-
n	-	iteration index	-
n_r	-	number of nodes per subregion	-
P_n	-	net precipitation	$m.d^{-1}$
P_s	-	sprinkling	$m.d^{-1}$
P	-	wetted perimeter of channel	m
q_c	-	capillary rise	$m.d^{-1}$
q_e	-	extraction for public water supply and sprinkling	$m.d^{-1}$
q_l	-	leakage	$m.d^{-1}$
q_s	-	flow from channel system	$m.d^{-1}$
q_w	-	flow from the surface water system	$m.d^{-1}$

Q_e	-	external flow	$m^3.d^{-1}$
Q_t	-	constant flux	$m^3.d^{-1}$
R_e	-	entry resistance	d
R_r	-	radial resistance	d
r	-	subregion number	-
s	-	index for soil physical unit	-
S	-	specific storage	m^{-1}
t	-	time	d
$[T_{ij}]$	-	transmissivity matrix node i	$m^2.d^{-1}$
V_w	-	moisture content in the root zone	m^3
V_{eq}	-	equilibrium moisture content	m^3
V_{eq}^{o}	-	equilibrium moisture content for groundwater level equal to ground level	m^3
$V_{eq}^{'}$	-	equilibrium moisture content for depth of the groundwater level of 1.0 m	m^3
W	-	weighting parameter	-
W_o	-	relaxation factor	-
W_{opt}	-	optimum relaxation factor	-
α	-	constant or geometry factor	-
ξ	-	interpolation function	-
μ	-	storage coefficient	1
Υ	-	drainage resistance	d
Δh	-	change in hydraulic head over a timestep	m
ΔQ_{ji}	-	change in flow between nodes over timestep	$m^3.d^{-1}$
ΔV_w	-	change in moisture content over timestep	m^3
Δt	-	timestep	d

ANNEX A

SOLUTION PROCEDURE

In Equation (7) all the flow contributions must be written in terms of the unknown hydraulic head. This means that an iteration process is required, because the hydraulic head at time t is calculated from information of the present and previous time level. For each node the hydraulic head is obtained with the Gauss-Seidel iterative method with successive over-relaxation. The Gauss-Seidel method is a point iterative solution method which uses updated information whenever it is possible. This means that the latest calculated value of Δh_i is used for the solution of levels in all other nodes affected by node i in the same iteration.

During the iteration process one can decrease the number of iterations by using an over-relaxation technique. The calculated change in hydraulic head for the considered timestep at iteration number n is modified by considering the change from the previous iteration (over-relaxation). This can be written as:

$$\Delta h_i^n = \Delta h_i^{n-1} + W_o \, (\Delta h_i^n - \Delta h_i^{n-1}) \qquad \text{(A1)}$$

in which n is the iteration presently considered and W_o is the relaxation factor or acceleration parameter. The value of W_o depends on the size of the solution domain and the actual maximum changes in the hydraulic heads per iteration and can vary between 0 and 2.

A practical and simple method for approximating the value of W_o is given by Carre (REMSON, 1971). For the first iteration use $W_o = 1.375$, for successive iterations it can be calculated with the equations:

$$W_{opt} = 2 \left[1 + \left[1 - \frac{(\lambda_m + W_o - 1)^2}{W_o} \right]^{\frac{1}{2}} \right]^{-1} \qquad \text{(A2)}$$

in which λ_m is the ratio of maximum change in hydraulic head between the present iteration n and the previous iteration and given as:

$$\lambda_m = \frac{(\Delta h_i^n - \Delta h_i^{n-1})}{(\Delta h_i^{n-1} - \Delta h_i^{n-2})} \qquad \text{(A3)}$$

Equation (A3) gives the optimum (also maximum) value for the over-relaxation that is posible. Carre suggests then to use a value somewhat lower than the value as calculated with Equation (A3). The new relaxation factor to be used in the next iteration becomes:

$$W_o = W_{opt} - (\frac{2 - W_{opt}}{4}) \qquad \text{(A4)}$$

Reference

REMSON, I., G.E. HORNBERGER and F.J. MOLZ, 1971. Numerical methods in subsurface hydrology. John Wiley and Sons, New York, 389 pp.

ANNEX B

FLOW CHARTS OF PROGRAM

The computerprogram SIMGRO can be divided into a number of modules. Figure B1 gives a broad outline of the calculation process of the model.

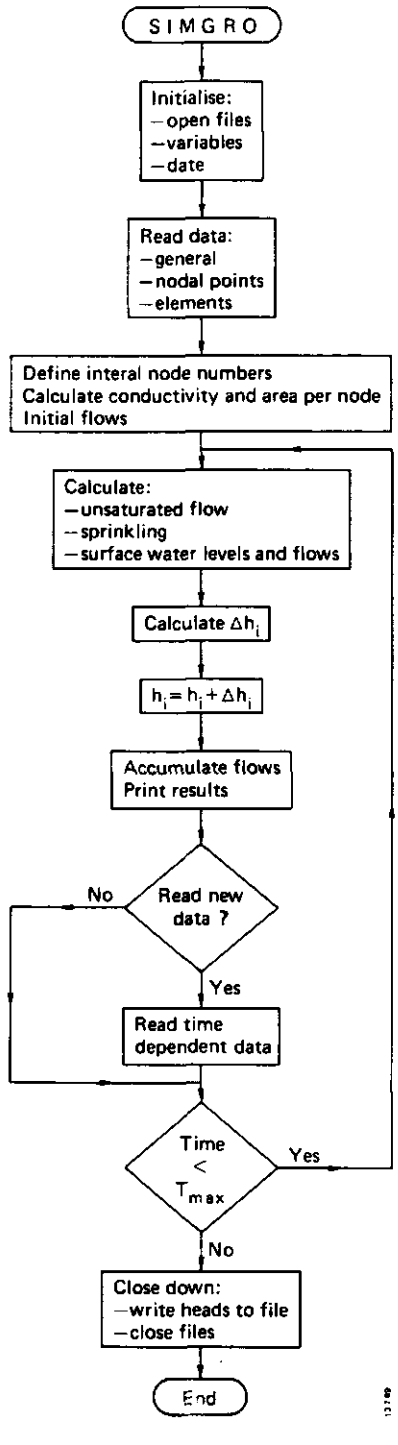

Figure B1. Flow chart of program SIMGRO

Figure B2 gives slightly more detail of the modelling concept used for the unsaturated zone. The program is built up in such a way that extensions can be included very easily.

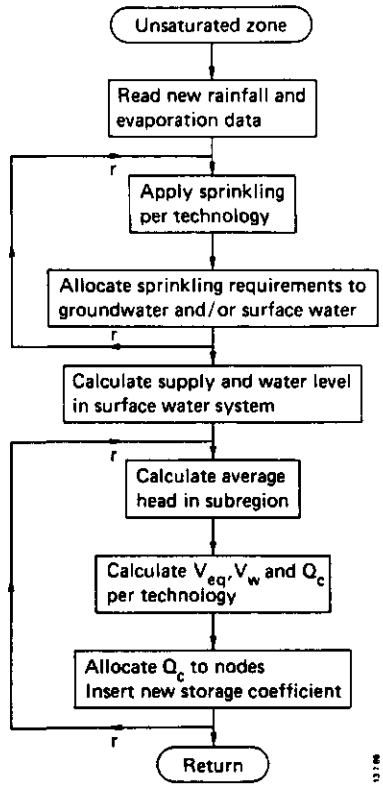

Figure B2. Flow chart of the unsaturated zone: a module of the program SIMGRO (r = number of subregions; V_{eq} = equilibrium moisture content root zone; V_w = moisture content and Q_c = capillary rise

ANNEX C

SOIL PHYSICAL PROPERTIES

Six different soil physical units are distinguished for the Southern Peel region (SMIDT, 1983). For each soil physical unit the equilibrium moisture content, capillary flux, and storage coefficient have been calculated and these are input data. The capillary rise and storage coefficient are calculated with the program CAPSEV (WESSELING et al., 1984). The equilibrium moisture content is calculated from the soil profile data. The values shown in the Figures C1 - C3 are based on a root zone depth of 0.25 m.

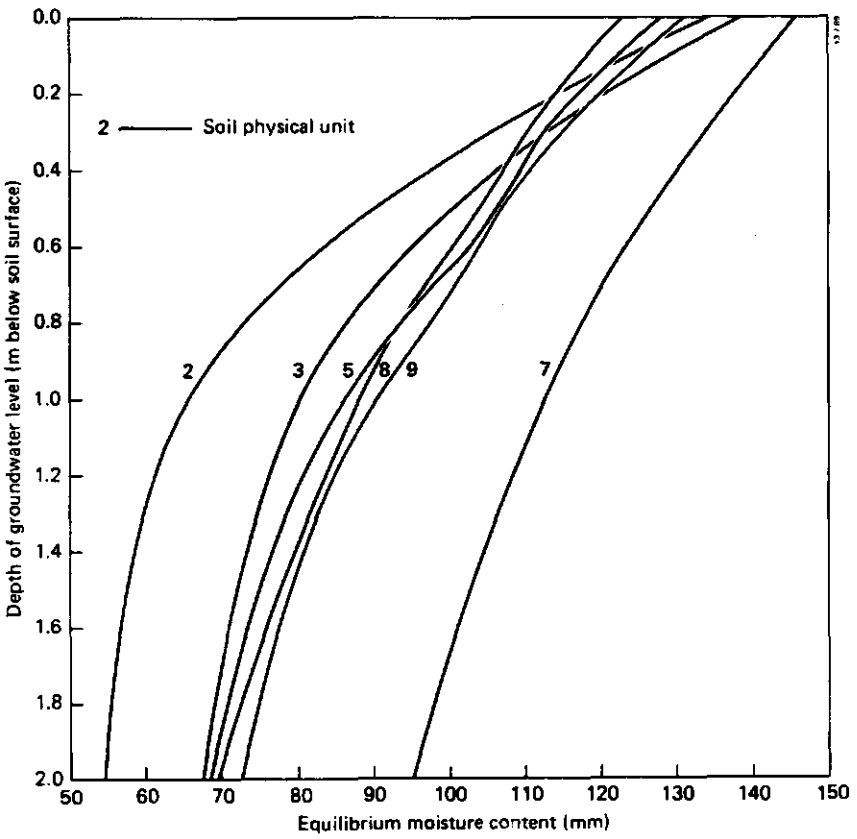

Figure C1. Equilibrium moisture conten for a root zone of 0.25 m for six typical sandy soils present in the Southern Peel region

Equilibrium moisture content

To account for different root zone depths the equilibrium moisture content is given per soil physical unit for depths of 0.25 m, 0.50 m and 1.00 m.

For other root zone depths the moisture content is obtained by linear interpolation.

Capillary rise

The capillary rise is calculated for a quasi steady-state condition, using a pressure head of -500 cm. The maximum flux is limited to 5 mm/d. The capillary rise for a root zone of 0.25 m is given as input data in the form of capillary fluxes for a number of groundwater levels (Fig. C2). For other root zone thickness the depth of the groundwater level is adjusted with the difference in root zone depth. For the capillary flux the distance between groundwater level and the underside of the rootzone is the important factor.

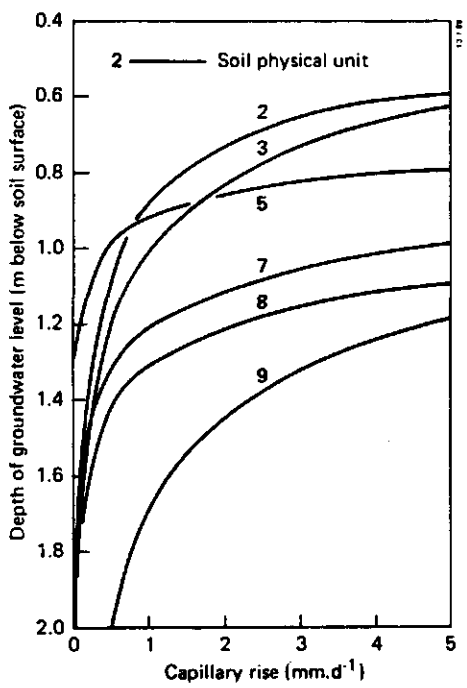

Figure C2. Capillary rise for a root zone depth of 0.25 m for six typical sandy soils present in the Southern Peel region

Storage coefficient

A typical relation for the storage coefficient of the zone between root zone and the grounwater level is shown in Figure C3. The storage coefficient becomes zero when the groundwater level reaches the underside of the root zone. If the groundwater level is at or above ground level, then the storage coefficient is equal to unity. If the groundwater level is in the root zone then pools of water on the surface will occur. To account for

this effect and to maintain numerical stability of the calculation process, the increase from underground storage to storage above the surface has been taken over the last 0.20 m, as shown by the dashed line in Figure C3.

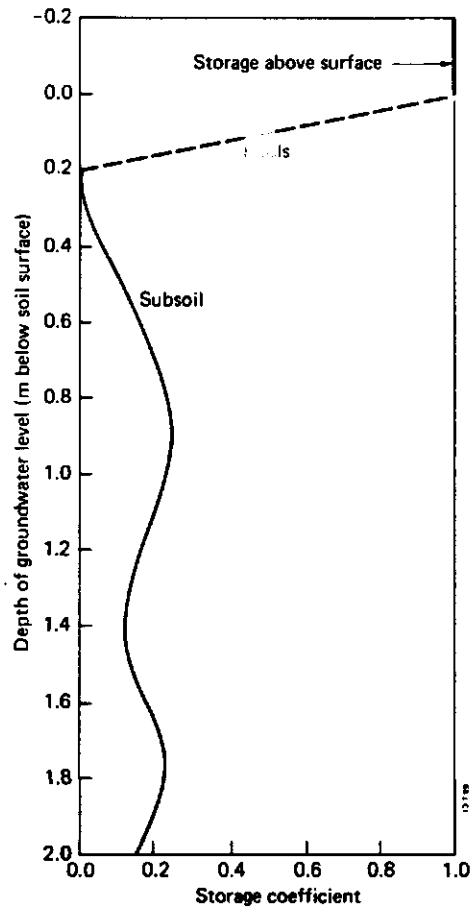

Figure C3. Typical relation for storage coefficient of the subsoil (between root zone and groundwater level). The root zone depth is 0.25 m

References

SMIDT, E.H., 1983. Explanation to the map of the subregions in the Southern Peel area. Nota 1420-II, Institute for Land and Water Management Research (ICW), Wageningen, The Netherlands: 62-66.

WESSELING, J.G., G.W. BLOEMEN and W.A.J.M. KROONEN, 1984. Computer program CAPSEV to calculate : I. Soil hydraulic conductivity from grain size distribution. II. Steady state water flow in layered soil profiles. Nota 1500, Institute for Land and Water Management Research (ICW), Wageningen, The Netherlands, 34 pp.

ANNEX D

COMPARISON OF CALCULATED AND MEASURED HYDRAULIC HEADS

From eight measuring stations the hydraulic head is compared with calculated results. From these stations time-hydraulic head curves were available; the locations are shown on Figure D1. The results are analysed by using the mean standard deviation as a measure for the agreement between the measured and calculated values.

From the model results and measured time-hydraulic head values the mean standard deviation has been calculated with the equation:

$$h_a = \sqrt{\left(\frac{1}{m}\sum_{m}(h_m - h_i + h_l)^2\right)} \qquad (D1)$$

where: h_a is the mean standard deviation, h_m is the measured hydraulic head, h_i is the calculated hydraulic head, h_l is a constant head to convert the measured levels for location i to nodal point n, and m is the number of observations over which the summation is taken. The measured levels are for a location i, and the calculated results correspond to the average hydraulic head for a nodal point. In order to compare the results h_l is used as a conversion. This factor should be time dependent, because it depends on the difference in head between the surface water and the groundwater level midway between two ditches. The position of the observation point in relation to the surface water system is also important, as is shown in Figure D2. These aspects have been ignored and the factor has been assumed to be independent of time.

The results of Equation (D1) for the eight stations are given in Table D1.

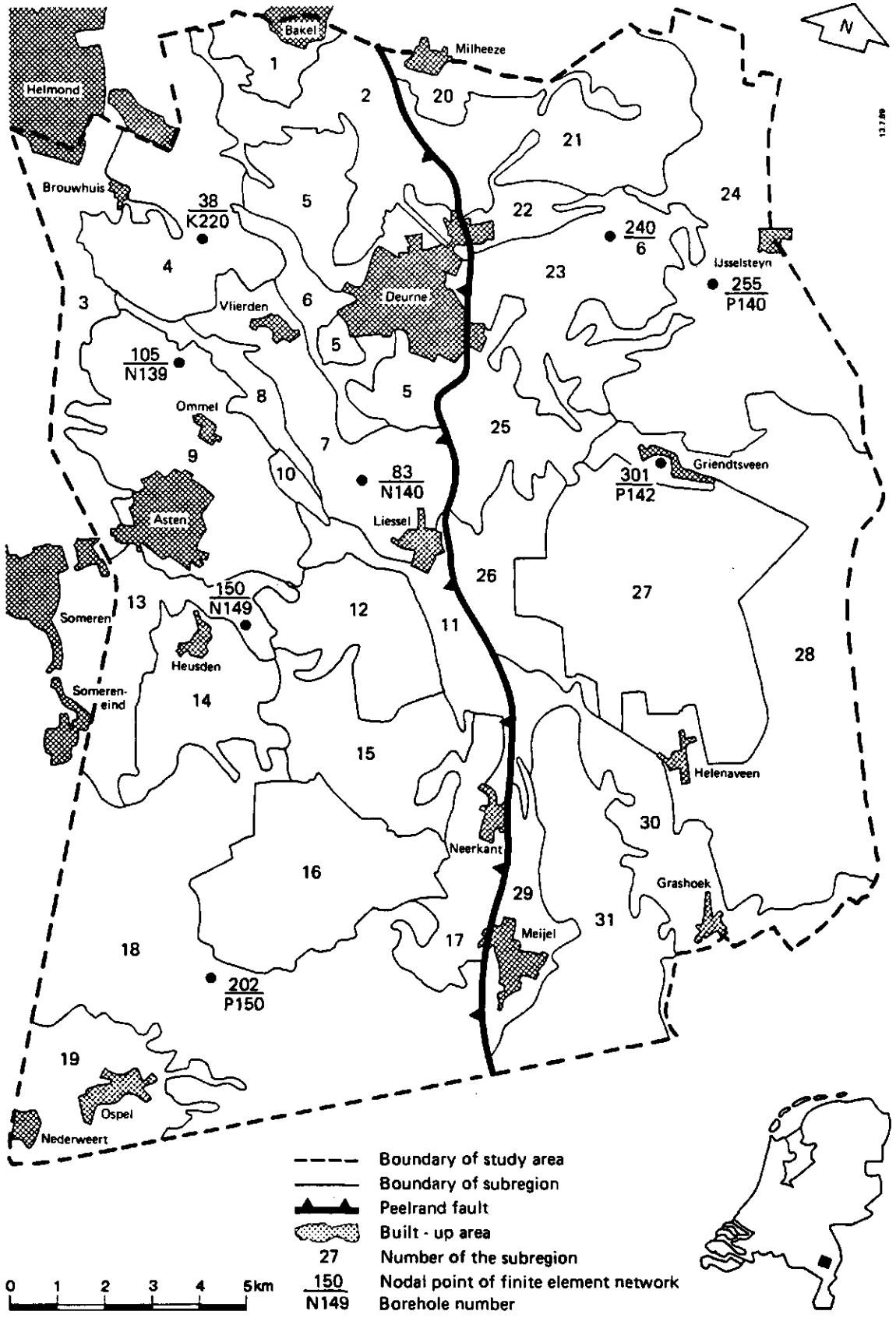

Figure D1. Location of eight measuring stations in the Southern Peel region for comparison of calculated and measured hydraulic heads

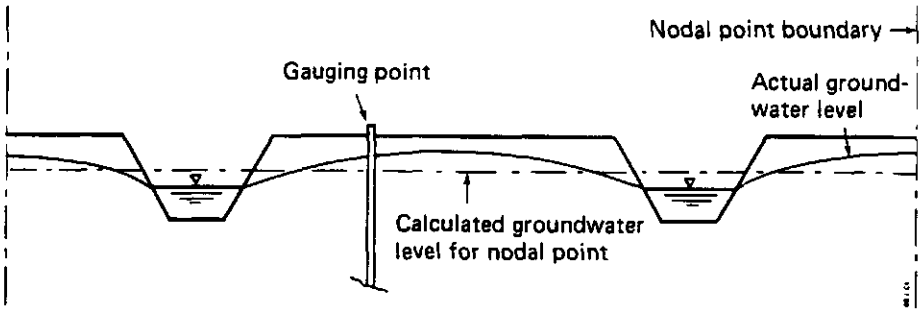

Figure D2. Correction to relate point measurements (solid line) to calculated average heads over a nodal point (dashed line)

Table D1. Mean deviation with minimum, maximum and average difference in hydraulic head (layer no 1 - aquitard ; layer no 2 - aquifer)

Nodal point	Layer no	h_a (m)	$h_m - h_i$ (m) min	max	average	h_l (m)
38	1	0.21	-0.33	0.02	-0.19	0
83	1	0.22	-0.48	0.07	-0.19	0
83	2	0.21	-0.43	0.09	-0.15	0
105	1	0.19	-0.38	0.43	-0.01	0
105	2	0.16	-0.34	0.13	-0.09	0
150	1	0.18	-0.37	0.25	-0.09	0
150	2	0.17	-0.46	0.13	-0.11	0
202	1	0.24	-0.49	0.11	-0.19	0
202	2	0.46	-0.66	0.02	-0.42	0
240	1	0.20	-0.25	0.39	0.05	0
255	1	0.42	-0.63	-0.13	-0.39	-0.11
255	2	0.29	-0.51	0.00	-0.26	-0.11
301	1	0.30	-0.61	0.18	-0.23	0
301	2	0.36	-0.65	0.09	-0.32	0
Average		0.26	-0.47	0.13	-0.19	

The points 38 through 202 are situated in the Central Slenk area (left hand side of Peelrand fault) and the others points are situated in the Horst area (Fig. D1). From Table D1 it can be seen that the mean standard deviation is smaller in the Central Slenk area than in the Horst area. The first water bearing layer in the Central Slenk has relatively uniform characteristics and can be modelled satisfactorily by the relative coarse nodal network. In the Horst area the characteristics of the water bearing layer are spatially very irregular, caused by the presence of small faults (REES VELLINGA and BROERTJES, 1984). The thickness of the water bearing layer for instance varies from 4 to 25 m.

The calculated and measured results are plotted and shown in Figure D3. From these figures and also the results in Table D1 it can be seen that in general the calculated heads are higher than the measured heads, especially in the summer period. The fact that point measured heads are compared with the average calculated heads for a nodal point contributes to part of these differences.

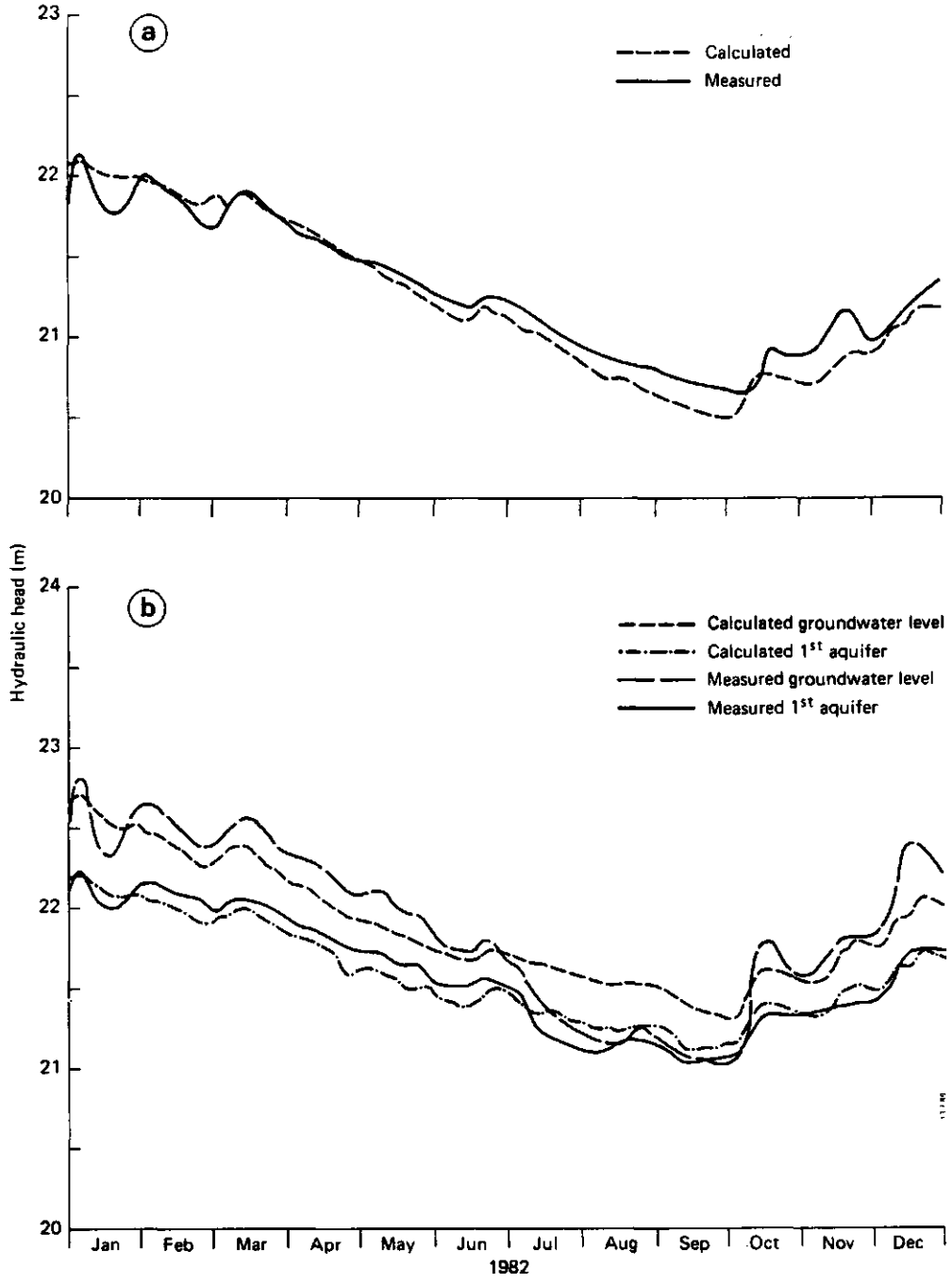

Figure D3. Measured and calculated hydraulic heads, h_m respectively h_i for nodes 38 (a) and 105 (b) (for location of nodes see Figure D1)

Reference

REES VELLINGA, E. and J.P. BROERTJES, 1984. Some results of the geohydrological research carried out in the Southern Peel region. Nota 1590. Institute for Land and Water Management Research (ICW), Wageningen, The Netherlands, 17 pp. (in Dutch).

ANNEX E

RESULTS OF SENSITIVITY ANALYSIS

For the sensitivity analysis parameters for the saturated zone as well as the unsaturated zone have been varied. The parameters which were selected are given in Table E1. The geohydrological parameters of the saturated zone concern the runs 1 - 9 and the parameters of the unsaturated zone concern the runs 10 - 14. The runs 13 and 14 with a typical soil physical unit of 2 and 7 reflect situations with a low and high capillary rise and equilibrium moisture content respectively.

The results of the sensitivity analysis in the form of the average standard deviation and the maximum and minimum difference in hydraulic head are given in Table E2. The average standard deviation reflects the eight measuring stations as shown in Figure D1. For all the calculations the conversion factor between measured levels and average hydraulic heads calculated for a nodal point, have been taken zero (see Eq. D1). The variation in standard deviation, for the runs concerning the geohydrological parameters is in general small, except when using a higher class for the drainage resistance. The remarkable smaller deviation when using a lower drainage class, comes from the assumption of one drainage class per subregion. In the vicinity of the measuring point the drainage class can vary a bit from the average selected drainage class. Therefore it seems favourable to select a lower drainage resistance for the location of the measuring point, but this may not be the case for the entire subregion.

The increase or decrease in capillary rise has a very small effect on the standard deviation. The selection of one typical soil physical unit or the unit with extreme hydrological conditions does not show a remarkable difference from the reference run.

The variation in sprinkling, evapotranspiration, and capillary rise for the entire region is given in Table E3. These results are for the summer half year of 1982. They show clearly that variation of the geohydrological parameters has no significant effect on the overall water balance terms, except the variation of the drainage resistance on the amount of sprinkling. The variation of the parameters for the unsaturated zone (capillary rise

and soil physical units) has a more pronounced effect on these water balance terms. The evapotranspiration for instance varies from +12% to -7% (related to reference run).

Table E1. Description of the parameters and their variation for the sensitivity analysis

Run	Variation	Description
1	none	reference run
2	c - 50%	hydraulic vertical resistance
3	c - 150%	of top-layer
4	kD - 75%	transmissivity of 2nd layer
5	kD - 125%	(1st aquifer)
6	T - lower	drainage resistance class lower
7	T - higher	drainage resistance class higher
8	S - 50%	specific storage
9	S - 150%	specific storage
10	vz - 75%	capillary rise
11	vz - 125%	capillary rise
12	s - 5	one typical soil physical unit
13	s - 2	extreme soil physical unit (Fig.B1 and B2)
14	s - 7	extreme soil physical unit (Fig.B1 and B2)

Table E2. Variation of standard deviation and differences in calculated and measured hydraulic heads (m)

Variation	Standard deviation (m)	$h_m - h_i$ (m)		
		min	max	average
none	0.273	-0.49	0.11	0.20
c - 50%	0.269	-0.46	0.12	0.20
c - 150%	0.289	-0.51	0.11	0.22
kD - 75%	0.286	-0.50	0.09	0.22
kD - 125%	0.263	-0.48	0.12	0.19
T - lower	0.233	-0.43	0.19	0.15
T - higher	0.366	-0.65	0.02	0.32
S - 50%	0.257	-0.46	0.12	0.19
S - 150%	0.287	-0.51	0.10	0.22
vz - 75%	0.274	-0.49	0.11	0.20
vz - 125%	0.289	-0.50	0.07	0.22
s - 5	0.286	-0.52	0.12	0.21
s - 2	0.310	-0.56	0.07	0.24
s - 7	0.231	-0.42	0.26	0.12

Table E3. Variation of water balance terms (entire region) for the unsaturated zone (summer half year 1982; sprinkling from groundwater and/or surface water)

Variation	Sprinkling (mm)	Actual evapotranspiration (mm)	Capillary rise (mm)
none	51	436	87
c - 50%	52	432	84
c - 150%	51	437	88
kD - 75%	50	437	88
kD - 125%	51	434	86
γ - lower	56	432	84
γ - higher	42	427	84
S - 50%	53	433	82
S - 150%	49	437	91
vz - 75%	52	440	89
vz - 125%	44	416	84
s - 5	56	424	72
s - 2	55	406	67
s - 7	23	488	141