Groundwater recharge of fluvial deposits at Haslemoen, Solør, southeastern Norway

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(1) Modelling of the water balance of the unsaturated zone and (2) calculating the Darcian flux have been applied for determining groundwater recharge. Modelling was based on field and laboratory studies, including soil moisture measurements, infiltration experiments, textural analysis and retention measurements. Method (1) is not particularly sensitive to the uncertainty in the characterization of the soil properties and is dependent on reliable estimates of surface fluxes. Method (2) on the other hand is very sensitive to changes in soil physical parameters. Although it seems straightforward it gives more uncertain estimates of groundwater recharge than a simulation model of the soil water balance.

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Groundwater recharge is the process whereby the surplus of infiltration over evapotranspiration drains from the root zone and continues to flow downwards through the unsaturated zone towards the groundwater table. A distinction has to be made between two modes of recharge (Gee & Hillel 1988): (i) continuous, spatially distributed (diffuse) recharge resulting from widespread percolation through the entire unsaturated zone; (ii) transient (occasional), concentrated recharge resulting from the short-term penetration of water through distinct pathways that cut through the unsaturated zone and bypass the greater part of its volume.

There are a number of approaches for dealing with the problem of groundwater recharge as reviewed by Lang (1977) and Gee & Hillel (1988). These include areal water balance calculations, simulation models of water balance, simplified water balance models, water balance measurements, lysimeter measurements, Darcian flux calculations and waterborne tracers. The only direct measure of water flux is provided by lysimeters, which unfortunately have the drawback of being very costly to construct and maintain, they disturb natural conditions and modify bottom boundary conditions, and it is difficult to evaluate the representativity of the data obtained. In all other methods the water flow is deduced from the measurements of other flow and storage components of the hydrological cycle and/or from physical properties of soil and vegetation.

One aim of the present study was to analyse the effect of variability in soil properties on groundwater recharge. Two of the approaches enumerated above, namely simulation models of water balance and Darcian flux calculations, allow a deduction of deeper percolation from the known physical properties of the soil. These two methods have been applied in this study.

Geological setting and description of the sediments

The study was carried out in an area with fluvial deposits at Haslemoen in Solør, southeastern Norway. Similar deposits are found along other main Norwegian rivers. Large parts of these areas are cultivated land, and in some places pollution of the aquifers by fertilizers may become a future problem (Englund & Haldorsen 1986) and models for transport of pollutants will gain increasing interest. The study indicates the minimum variability one can expect for the unsaturated zone in Norway, since the Norwegian sediments are usually much more variable than the fluvial sediments in the investigated area.

The area along the river Glåma in Solør is characterized by thick fluvial deposits (Bjørlykke

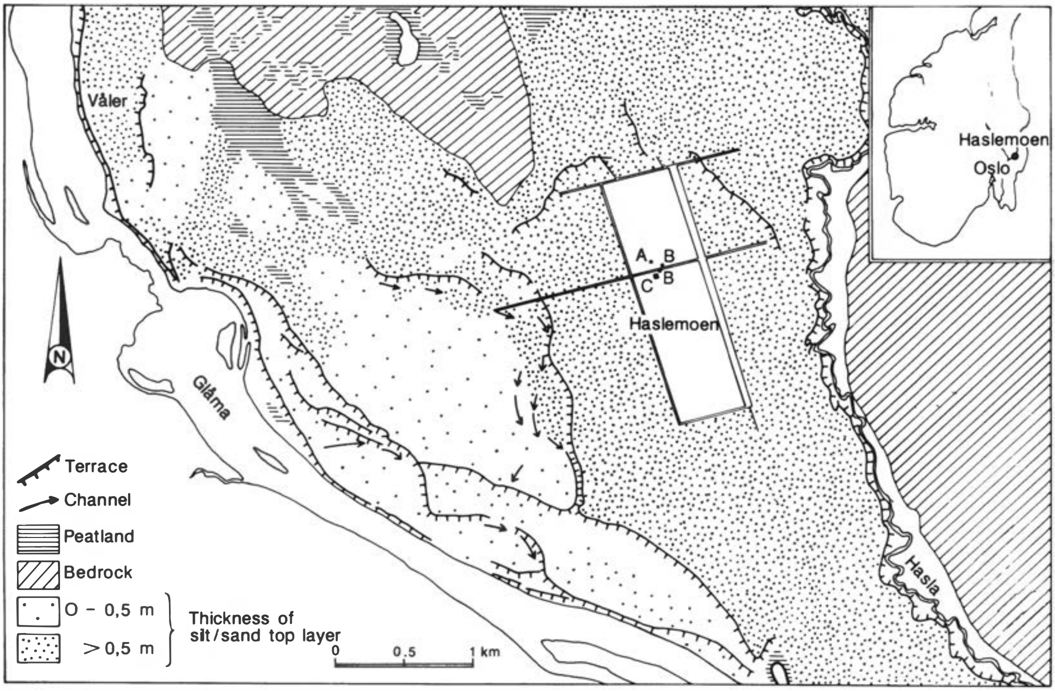

Fig. 1. Key map of Haslemoen. The white area shows the cultivated land where the studies were carried out. A: infiltration experiment; B: neutron meter gauge; C: samples for spatial variability.

1901; Sortdal 1921; Holmsen 1954). Terraces are found in different levels above the present river and the sediments are dominated by sand with a cover of silt which represents overbank deposits (Fig. 1). The fluvial deposits are in many places used for groundwater supply because they form important aquifers. In a Norwegian context the sediments are very homogeneous. The study was carried out at Haslemoen in the central part of the Solør area at a fluvial terrace at 172–175 m a.s.l. (Figs. 1, 2). The terrace is covered by a layer of silt or very fine sand from 0.5 to 1 m thick. The underlying sand sediments extend down to the groundwater level (Fig. 2A), sometimes interrupted by thin horizons of coarse silt. The whole terrace is very uniform stratigraphically.

An area of cultivated land (Fig. 1) was chosen for detailed studies. Here the thickness of the fine-grained top layer is about 0.5 m and depth to the groundwater level is 3.5 m (Fig. 2A). The top layer of silt and fine sand is influenced by soil weathering and cultivation and no primary structures are visible. Transition to the underlying fine sand is sharp, due to the abrupt change in the fluvial sedimentation. The sand sediments show an alternation between fine-grained rippled sand and medium to coarse sand with trough cross bedding (Fig. 2A). Two fine-grained layers are observed under the surface layer at this locality.

Two profiles were dug to a depth of about 120 cm (Fig. 2B) with a distance between the two profiles of about 16.5 m. As shown in Fig. 2B there is a difference between the two profiles, especially in the thickness of the horizons. The horizons can be clearly separated from each other by sharp boundaries. In some cases the depth down to the boundary varies over distance. In Fig. 2B this is indicated as a sloping horizon boundary.

Above a depth of 50 cm, there are few or no mottles (chromaspots). In profile 1 some mottles were found between 45 and 90 cm, which shows that saturated conditions may occur in periods of the year.

The roots were found mainly in the upper 30 cm, where the root diameter is less than 2 mm. Lower down the number of roots is very small. The diameter of some of the roots below 30 cm depth is up to 34 mm; it is possible that some of them are old roots which are no longer active.

The content of macro-pores (i.e. pores with a diameter greater than 0.06 mm) was 0.5 to 2% in

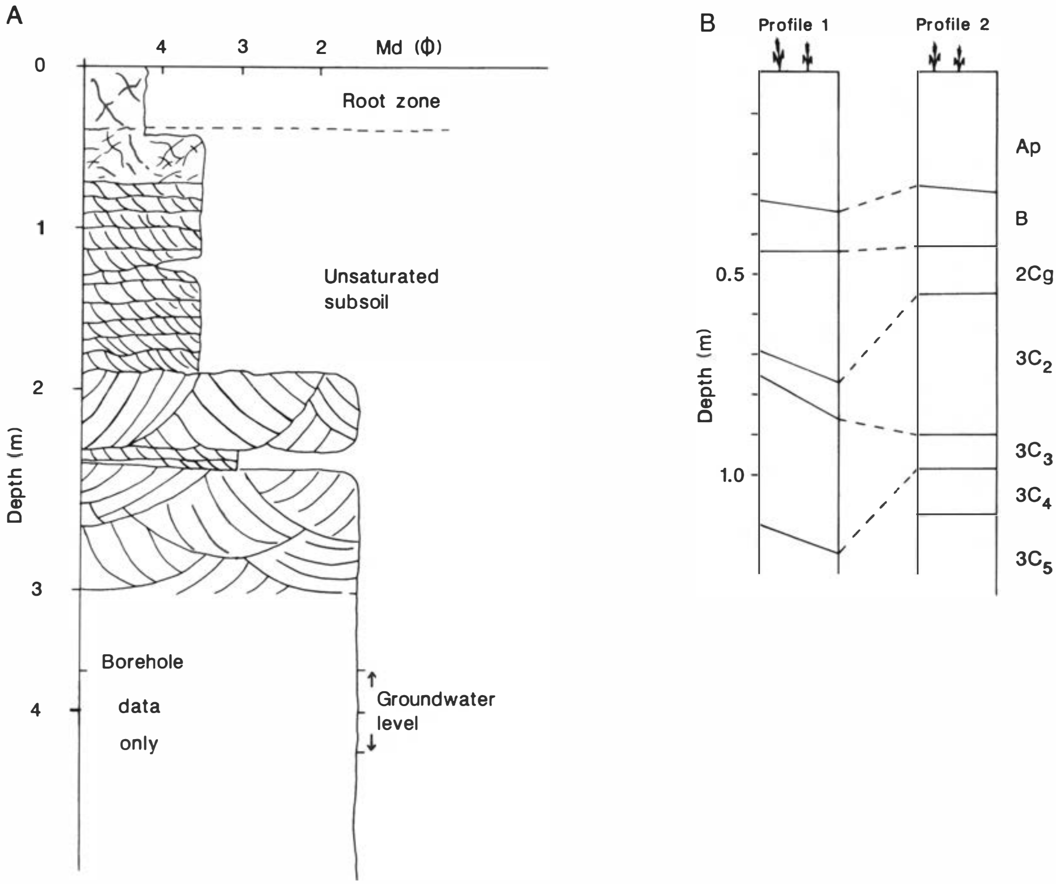

Fig. 2. A: Profile through the sediments down to the groundwater level, showing the top layer of silt and fine sand, the upper fine rippled sand and the lower medium to coarse sand with dunes. Borehole data from the lower part of the profile give information only about the grain-size distribution. B: Two soil profiles from the same locality showing the thickness of the different parts of the A, B and C horizons. For location, see Fig. 1.

the upper 30 cm. Below 30 cm the content of macro-pores was less than 0.5%. Most of the macro-pores at this depth seemed to be old root passages.

Climate and land use

The meteorological data used in the models are taken from the nearest meteorological station at Flisa. The area is a typical inland area in south-eastern Norway with a maximum precipitation in the summer. The main part of the snow melt occurs in April. The average annual precipitation for the period 1931 to 1960 is 623 mm with about two-thirds occurring during the months of May to October (DNMI 1986). The average potential evaporation during these months is calculated at 365 mm.

The studied area is situated on cultivated land which is mainly used for barley and oat growing. The retention capacity of the top silt layer is so great that irrigation is not needed, even in the driest summer periods (Haldorsen et al. 1986).

Field and laboratory methods

The field studies were carried out during the period 1985–86. The field installations are shown in Fig. 1. The pF cylinders and samples for grain-size analysis were collected close to the described soil profiles shown in Fig. 2B.

The water content in the unsaturated zone was

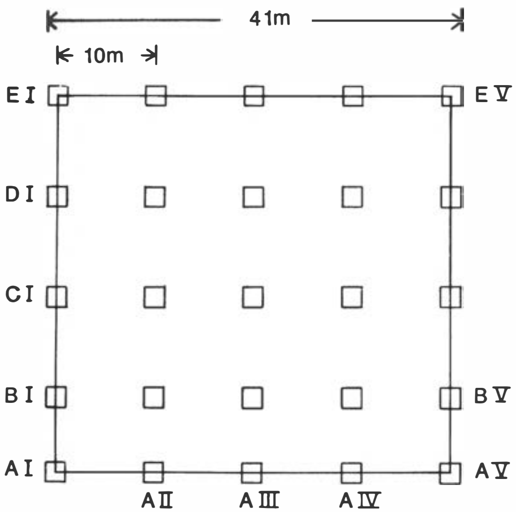

Fig. 3. Sample net for spatial variability studies of the top layer. For location, see Fig. 1.

measured in two 50 mm tubes using a neutron moisture meter (Troxler Depth Moisture Gauge 1255 S/N 729). An additional tube of the same type was used in the infiltration studies described below. Counting was carried out from 40 cm below the ground at 20 cm intervals down towards the groundwater level. Calibration was in accordance with the calibration curve from the manufacturer. This was regarded to be satisfactory for the actual sediments which are completely dominated by quartz, feldspars and muscovite.

An area of $41 \times 41\ m^2$ was used for taking measurements of the spatial variability (Fig. 3). The area was placed so far inside a field that any influences from the edges were avoided. Samples were taken from 25 sites inside the quadratic area along straight lines with a distance between the sites of 10 m. Sample sets were taken at each site from depths of 10–14, 35–39 and 50–54 cm. These sampling depths were chosen in order to study the variability due to a combination of the primary sediment variability and the variability due to the cultivation methods. The upper samples were taken above the ploughing depth and in the fine-grained silt layer, the intermediate samples in a zone of significant root activity. The deepest samples are believed to be below the zone influenced by the cultivation and were mainly collected from the fine sand. Three parallel samples were collected from each depth, each consisting of three subsamples for water retention studies and for grain-size analysis.

The suction, or negative pressure head, was measured using tensiometers (Soil moisture 2710, 24, 36 and 60 cm and Quick Draw Soil moisture probe 2900F). The grain-size analyses were carried out by sieving and the hydrometer method. For the samples collected above 0.5 m depth the organic material was removed with a 10% H_2O_2 solution.

Samples for pF measurements were collected by hammering steel cylinders (100 cm^3 soil moisture pF rings) vertically down into the sediment. The samples thus represent undisturbed sediments. In the laboratory the water content was calculated at suctions of 0.02, 0.1, 1.0 and 15 bars.

Results of field and laboratory studies

Moisture variation during 1985 and 1986

The variation in moisture content during the late summer and fall 1985 and the summer and fall 1986 is shown in Fig. 4 for the northernmost tube 4 (Fig. 1). The measurements in the other tube gave similar results. The water content was low in June 1985 and the profile was probably close to an equilibrium situation. A period with high precipitation gave a high water content in the upper part of the profile early in September. In November the water content was low in the upper half of the unsaturated zone, while the percolation of the excess water from the early fall was still draining through the lowest part of the zone. During 1986 the water content was decreasing during the growing season and fall, with the lowest water contents in August. The autumn rain is shown by a higher water content in the upper part of the profile in November compared with August. The shape of the moisture curves is the same for all the measurements, which again means that there is an even drainage throughout the whole profile. There is no indication of a water front moving downwards through the profile. The top silt layer is clearly seen by the high water content above 50 cm depth. In the middle of May the water content is also high just below the silt layer, which is explained by the percolation of the snow melt water from middle and late April. There are water maxima between 1 and 1.5 m depth and at 2 m depth. These are due to the position of two beds with fine sand and coarse silt (see Fig. 2A) with a higher water retention

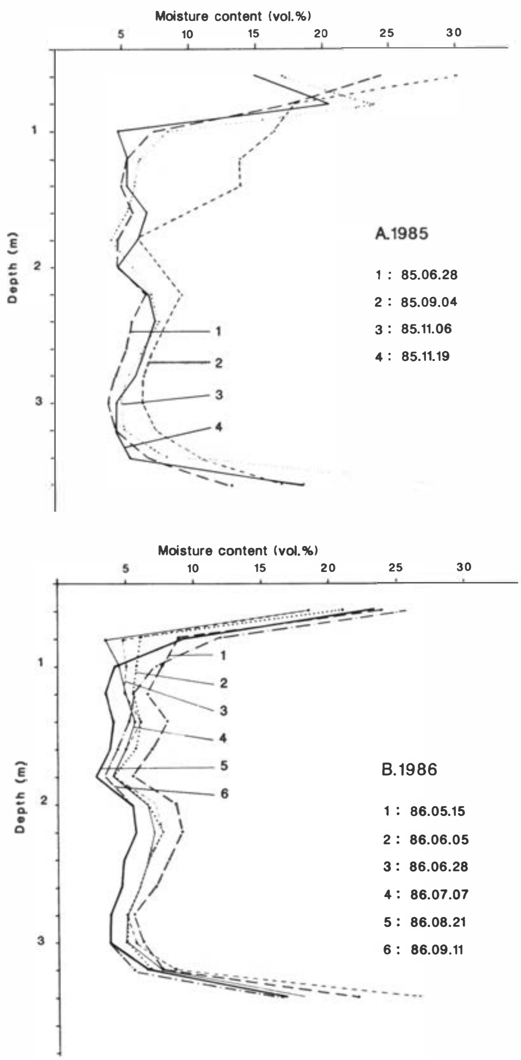

Fig. 4. Moisture content from 40 cm depth down to the groundwater level measured in the northernmost neutronmeter gauge tube shown in Fig. 1.

capacity than the overlying and underlying sand layers.

Spatial variability of the fine-grained top sediment: pF data

The spatial variation patterns of soil physical parameters have been investigated applying geostatistical methods, i.e. calculating and analysing semivariograms. Details of the theory of geostatistics can be found in a plethora of literature, see for example Journel & Huijbregts (1978) and Delhomme (1978). The semivariograms are constructed for four parameters in the area: porosity, pore size distribution (λ), saturated hydraulic conductivity (K_s) and air entry pressure (ψ_a). The semivariograms are calculated separately for each of the three sampling depths 10–14 cm, 35–39 cm and 50–54 cm. Porosity values are calculated from the measurements of air content in the pF cylinders at pF 2 (0.1 bar). The retention curves based on the results from the pF analyses are drawn on double logarithmic paper with the section in cm H_2O as the abscissa and the relative water content as ordinate. The measurement steps are saturation (0 cm H_2O), 0.02 bar (20.4 cm H_2O), 0.1 bar (100 cm H_2O), 1 bar (1000 cm H_2O) and 15 bars (15,500 cm H_2O).

To reduce the data quantity the results from the three parallels were combined. The samples from the greatest depth (50–54 cm) can be separated into two populations – a silt population and a sand population – which in some cases are treated separately.

A linear adaption to the retention curve is done by drawing a line in the flat area of the curve through about 100 and 1000 cm H_2O, depending on soil type. The rise of the curve is an expression for λ (pore-size distribution).

Air entry pressure, ψ_a, can be defined as the pressure that must be laid on the soil sample to drain the biggest pores. ψ_a is found by drawing a straight line along the retention curve as described above. The straight line intersection of this with a vertical line from the point of saturation gives a value for the air entry pressure (see Fig. 4).

The calculations of the pore-size distribution index are based on the method of Brooks & Corey (1964):

$$\frac{\theta - \theta_r}{\theta_m - \theta_r} = \left(\frac{\psi_a}{\psi}\right)^{\lambda} \quad (1)$$

θ_r = leftover water
θ_m = water content at saturation
θ = actual water content
ψ = actual pressure
ψ_a = air entry pressure
λ = pore-size distribution index

It is possible to find an expression for λ by taking the logarithm of the equation above:

$$\lambda = \frac{\ln(\theta - \theta_r)}{\ln \psi_a - \ln \psi} \quad (2)$$

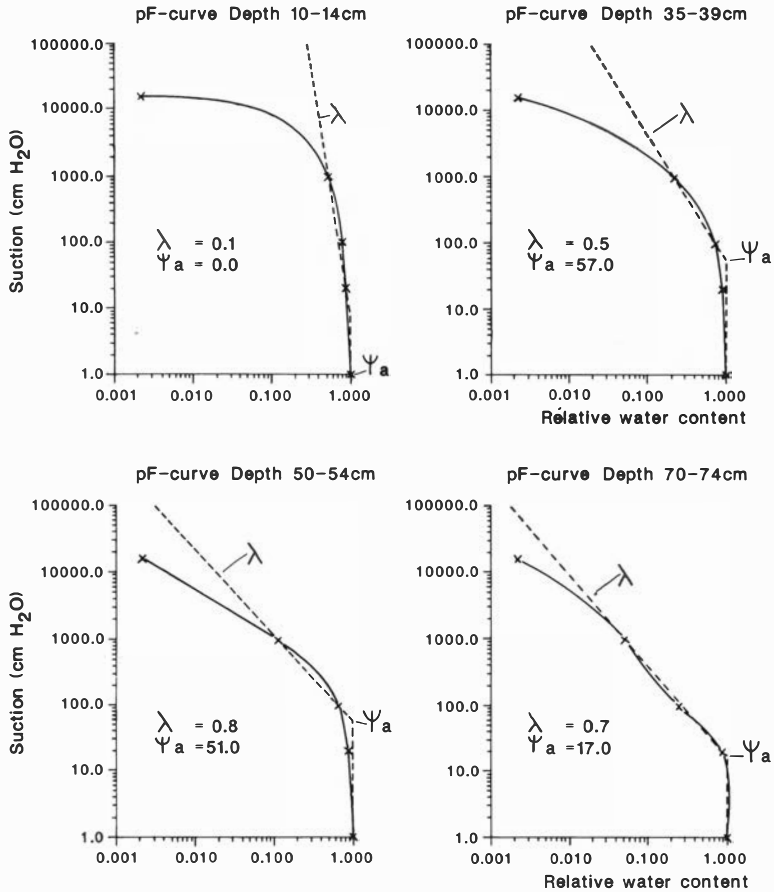

Fig. 5. Retention curves based on pF data with linear adaption. λ: pore-size distribution index, ψ_a: air entry value.

Typical examples on the retention curves with linear adaption from each of the three depths are shown in Fig. 5A–C. Fig. 5D gives the value for the subsoil sand. The values of λ and ψ_a are typical for the respective depths. Well-sorted material and uniform pore sizes are characteristic of the sand samples and the silt samples from the deepest layers. In the top samples the content of organic material is higher and the pore sizes are more varying.

In the silt zone the saturated hydraulic conductivity is estimated on the results from the 225 grain-size analyses. The Hazen's equation (Hazen 1893) for water at 5°C is used when calculating K_s:

$$K_s = 0.00116{*}d_{10}^2{*}^{2/3} \tag{3}$$

Semivariograms $\gamma(h)$ were calculated from the different sets of soil parameters in fixed distance classes:

$$R(h) = \{(i,j); h - \varepsilon < d((x_i, y_i), (x_j, y_j)) < h + \varepsilon\} \tag{4}$$

where

$$d((x_i, y_i), (x_j, y_j)) = \sqrt{(x_i - x_j)^2 + (y_j - y_j)^2},$$

i.e. the Euclidean distance between two sampling points (x_i, y_i) and (x_j, y_j), h distance and the width of the distance class is 2ε. The semivariogram is calculated from:

$$\gamma(h) = \sum_{(i,j)\in R(h)} \frac{[A(x_i, y_i) - A(x_j, y_j)]^2}{2N(h)} \tag{5}$$

where $A(x_i, y_i)$ is the respective soil parameter value observed at point (x_i, y_i). The smallest distance between two pairs of observations in this investigation is 10 m. For distances greater than

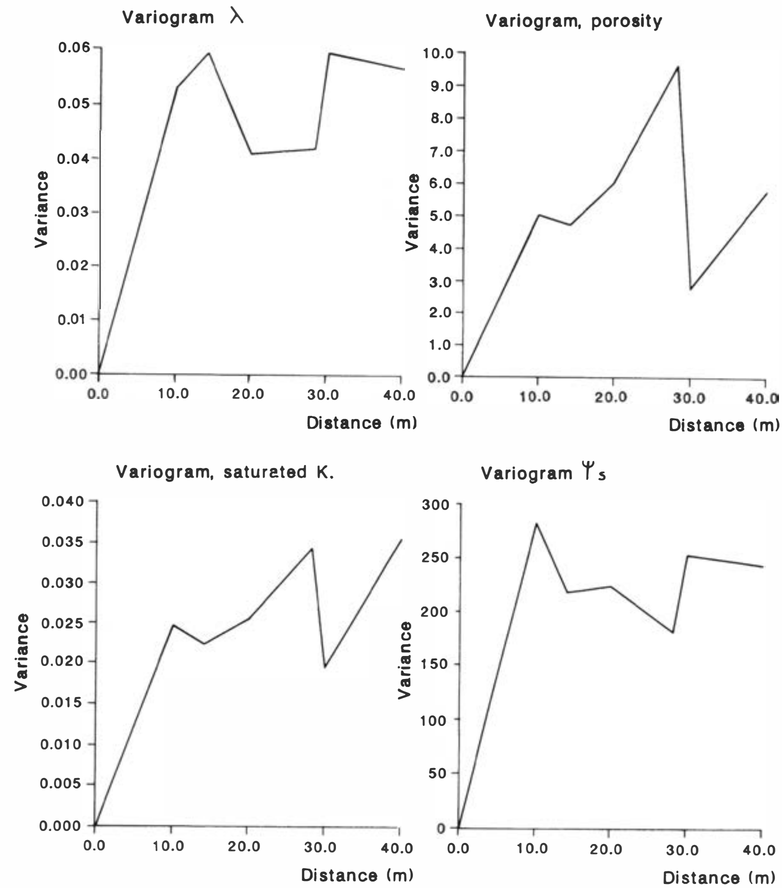

Fig. 6. Semivariograms showing the spatial variability of different soil physical parameters.

40 m there are so few pairs of observations that they are not used in the calculations.

The semivariograms (Fig. 6) are drawn on the basis of the calculations of *h*. The curves are drawn through the calculated values without any correction of the curve. The semivariograms show no trend or dependence. The variance is about constant over distance. The differences are considerable within one and the same semivariogram. If the result for one parameter and a certain depth is compared with the results for the same parameter from another depth, the semivariograms show that the variance increases with depth. Fig. 6 shows examples of semivariograms from 35–39 cm depth.

Grain-size distribution

The particle sizes of all samples are within the range of 0.006 and 0.2 mm, i.e. medium silt to fine sand (Fig. 7). The samples from 10–14 cm depth all consist of silt, while at the 35–39 cm depth some of the samples consist of fine sand. All the deepest samples consist of fine sand.

In situ measurements of the drainage processes

An area of about 20 m^2 was used for calculating the unsaturated hydraulic conductivity by the internal drainage method described by Hillel et al. (1972). The study was carried out in November 1986 and the area was continuously wetted for two days. The tensiometers showed that the sediments then were near saturation in the upper meter. The area was covered with a plastic sheet to prevent evaporation and the profile was allowed to drain freely. The water content during the

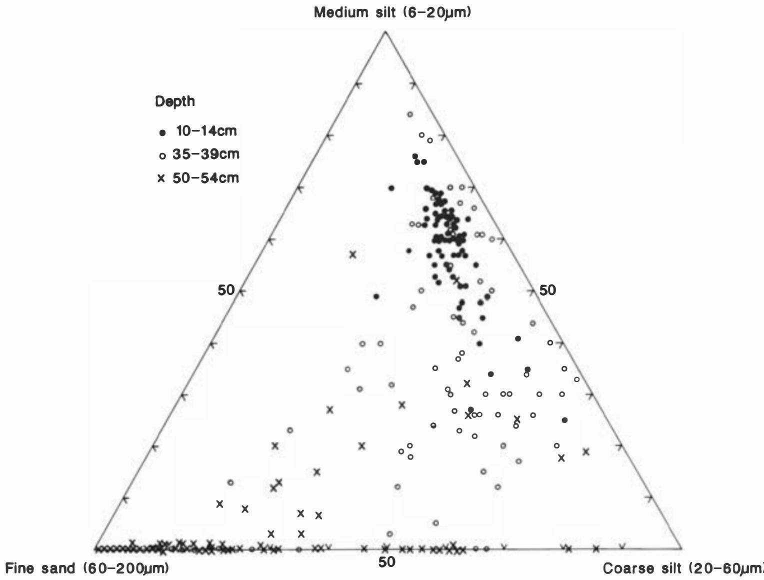

Fig. 7. Grain-size distribution curves for the samples in the net shown in Fig. 3.

following drainage event of one month was measured down to 120 cm depth using a neutron meter placed in the centre of the wetted area (Fig. 1). The results of the test are shown in Fig. 8. The upper silt layer had a rather constant water content during the whole drainage period, which reflects the great water retention capacity of the silt. The transition from the silt to the sand, where the water has to pass from small pores to greater pores (see Fig. 4), is also of importance in this connection. The loss of water during the one month is less than 5% of the total soil volume. In the underlying sand the drainage is much faster and a loss of water of about 20% was measured. No drainage front with a high water content passing down through the profile was found. There is a gradual decrease in the water content in the whole sand horizon during the whole period. The curves show, however, that there is a tendency for a higher water content at 120 cm depth in the initial part of the drainage period compared with the higher part of the sand, while at the end of the period the water content is rather constant in the whole period. The 120 cm level corresponds to the transition zone between the fine rippled sand and the underlying coarser sand layer. The high initial water content represents a full saturation of the sand where a pressure potential great enough to overcome the capillary potential from the fine to the coarser sand is built up.

Simulation model of water balance

In Norway precipitation and the resulting infiltration exceed evapotranspiration by an appreciable amount except during summer. The recharge process thus takes place almost continuously throughout the unsaturated zone. For the Haslemoen aquifer this is confirmed by the very smooth fluctuations of water content at some depth in time as well as in space (see Fig. 4). The analysis of physical soil properties from soil samples provides evidence of their large homogeneity with only a small-scale variability present. The conclusion, therefore, is that the recharge through the unsaturated zone of the fine sand–coarse silt aquifer is mainly to be considered as a uniform and slow process. No structures have been identified that can give rise to fast concentrated episodes of recharge, although the existence of such cannot be totally excluded. These conclusions, based on the observed data, form the background for mathematical model formulation.

Simulation models calculate water storage, redistribution and drainage from the soil based on specified surface boundary conditions in terms of given precipitation and evaporation data. Several detailed numerical models have been developed that calculate water flow deterministically using Richard's equation. The one used here, UNSAT2, is a programme developed by Davis & Neuman (1983).

The calculations performed are based on the following assumptions and conditions:

- The flow direction is vertical and one dimensional. The total study area is thus approximated by a set of vertical columns, each with its own set of physical parameters.
- For a certain column the layers of silt and sand are horizontal. The physical properties are the same throughout the sand layer. The silt cap has been divided into three layers with different characteristics.
- Boundary conditions are given as known fluxes of precipitation and evapotranspiration at the surface and as a free boundary with a vertical tangent at a depth of 350 cm.

The soil physics is characterized by:

- porosity
- saturated hydraulic conductivity
- relation between unsaturated hydraulic conductivity and water content
- relation between soil suction and water content (pF curve)

Field calibration

The model performance was tested against the field drainage experiment described above. Some slight modifications of the soil hydraulic parameterization were necessary to allow the model to perform in accordance with the observed decay of soil moisture content. The unsaturated hydraulic conductivity when calculated from the theoretical expressions very quickly approaches zero for water content less than about 0.35. It was not possible to get an agreement between the observed and modelled water content with this relationship. Therefore a modification was made according to which the relative hydraulic conductivity decreased linearly from a value of 0.08 at a water content of 0.35 and to zero at zero water content. The model results after this change in parameterization are shown in Fig. 8. The impression is that the distinction between layers is exaggerated by the model. With this exception there is good agreement.

Simulations

Simulations with the model were performed for the period 20/3/1985 to 20/3/1986. Precipitation, temperature and potential evaporation data from the climatic station at Flisa were utilized. One snow curvey was performed on 20/3/1985. Snow melt was calculated using the degree-day method. Four parameters describing the soil physical properties, namely porosity n, saturated hydraulic conductivity k_s, air entry pressure ψ_a and the pore size distribution index λ, represent the necessary input data to the model. Each layer in the model is characterized by a parameter set. The mean and variance of these parameter values are shown in Table 1, as determined from the laboratory tests of soil samples. The parameter values are not independent of each other and their mutual covariances are also given in the Table. The theoretical distribution of the parameters was determined using simple plotting on probability paper. It was found that n and λ are well approximated by the normal distribution while k_s and ψ_a belong to the lognormal one.

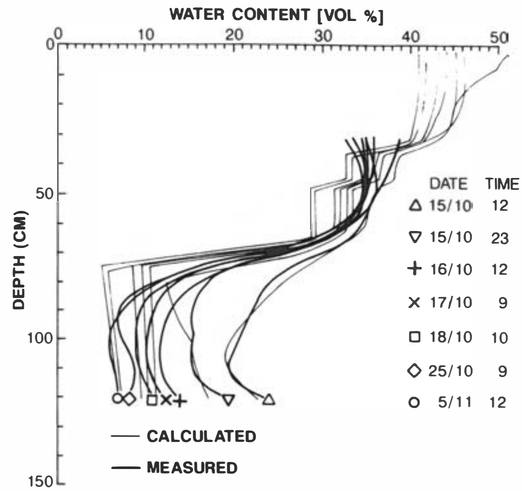

Fig. 8. Moisture variation measured during the drainage experiment.

Fifty simulation runs were performed. Precipitation, snow melt and potential evapotranspiration data were kept constant while the soil physical parameters for each layer were allowed to vary in accordance with their theoretical distributions and internal dependencies. This was achieved using a Monte Carlo technique generating random numbers. The dependencies were maintained by successive regression relations between parameters. From the sets of generated parameters pF curves as well as relations between unsaturated hydraulic conductivity and water content were constructed. The same type of modi-

Table 1. Mean values and variance-covariances of soil physical parameters, soil suction and vertical Darcian flux.

	n	$\ln(k_s)$	$\ln(\psi_a)$	λ	$\ln(\psi)$ 28/6/85	$\ln(\psi)$ 11/9/86	$\ln(q_z)$ 28/6/85	$\ln(q_z)$ 11/9/86
Mean	0.5	2.66	2.94	0.80	4.58	4.84	−3.88	−4.92
Variance		0.060	8.4E-4	0.010	0.0078	0.096	0.39	1.86
Covariance		0.0019						

fications of the relation between the unsaturated hydraulic conductivity and water content as performed for the field calibration was also applied this time.

Darcian flux calculations

In an aquifer of poorly graded soil of sand or silt with the groundwater table at a considerable depth, the deep percolation is a gradually varying process. In the upper part of the unsaturated zone the flow of water is controlled by both diffusion and gravity with relatively rapid changes. The simulation model of water balance applied here considers both of these controlling principles.

At a greater depth in the profile the moisture content settles to a uniform gradually varying level, such that the rate of .deeper percolation can be accommodated by the vertical unsaturated hydraulic conductivity having only the gravitational component of the potential gradient. In the general case the vertical flow q_z in the unsaturated zone is expressed by:

$$q_z = -k(\psi)\frac{\partial \psi}{\partial z} + k(\psi) \tag{6}$$

where $k(\psi)$ is the unsaturated hydraulic conductivity, ψ the soil suction and z a vertical length coordinate. The simplifying assumption of only a gravitational flow component implies that $\partial\psi/\partial z = 0$, i.e. $q_z = k(\psi)$. Accepting the Mualem (1976) approach for the parameterization of the unsaturated flow, the percolation is assumed to be:

$$q_z = k_s(\psi_a/\psi)^{2+(2+n)\lambda} \tag{7}$$

If observations of the soil suction or alternatively the soil moisture content are available in the zone of only gravitational flow they can be used to estimate the groundwater recharge. Because of spatial variability several observation sites are necessary if average values and variances of all parameters and variables involved are to be estimated.

We have earlier found that k_s and ψ_a follow a lognormal distribution while λ is normally distributed. This would indicate that q_z is lognormally distributed. Applying a first-order approximation, the variance of $\ln(q_z)$ can be calculated from the expression (Benjamin & Cornell 1970):

$$\mathrm{Var}\{\ln(q_z)\} = \sum_{i=1}^{4}\left(\frac{\partial \ln(q_z)}{\partial x_i}\right)^2 \mathrm{Var}\{x_i\} + 2\sum_{i<j}\left(\frac{\partial \ln(q_z)}{\partial x_i\,\partial x_j}\right)\mathrm{Cov}\{x_i, x_j\} \tag{8}$$

where x_i; $i = 1, \ldots, 4$ represent the soil suction ψ and the parameters $\ln(\psi_a)$, $\ln(k_s)$ and λ, and $\mathrm{Cov}(x_i, x_j)$ is the covariance between the parameters x_i and x_j. Inserting the equation (7) in (8) yields:

$$\begin{aligned}\mathrm{Var}\{\ln(q_z)\} &= [-2-(2+n)\lambda]^2\mathrm{Var}\{\ln(\psi)\} \\ &+ [2+(2+n)\lambda]^2\mathrm{Var}\{\ln(\psi_a)\} \\ &+ 2(2+n)\mathrm{Cov}\{\ln(\psi_a), \lambda\} + \mathrm{Var}\{\ln(k_s)\} \\ &+ (2+n)^2[\ln(\psi_a) - \ln(\psi)]^2\mathrm{Var}\{\lambda\}\end{aligned} \tag{9}$$

It has been assumed that the suction ψ is independent of the parameter values. In contradiction to the case of applying a simulation model for water balance the variability can be explicitly deduced. No Monte Carlo experiments are needed. The variance covariances in equation (9) can be directly calculated from the standard deviations and correlation coefficients in Table 1, utilizing standard properties of the normal and lognormal distributions.

Results and discussion

In an earlier study for the same area a simplified water balance model was used (Haldorsen et al. 1986). The results of this later model only give

Table 2. Calculated water balance with simulation model for spring and summer 1985.

	Snow melt season	May	June	July	Aug	Total
Precipitation and snow melt (mm)	203	24	129	107	69	532
Evapotranspiration (mm)	13	86	88	80	47	314
Groundwater recharge (mm)	139	43	0	20	17	219
(S.D.)	(10.0)	(6.1)	(0)	(2.6)	(1.4)	(2.0)
Change in storage (mm)	51	−105	41	7	5	−1

one lumped value for the total area. For this type of simplified model it is difficult to evaluate how these simplifications affect the estimated recharges. Furthermore, the spatial variability is neglected.

The results of the calculated water balance with the simulation model for the spring and summer 1985 are shown in Table 2. Although the potential rate of evapotranspiration is controlled by atmospheric conditions, the actual evapotranspiration is also dependent on the moisture content in the soil, i.e. the actual rate of evapotranspiration may be limited by the ability of the soil to transmit water upward from below. A priori prediction of the exact boundary condition to specify under the above conditions is not possible. The UNSAT2 programme obtains a solution by maximizing the absolute value of the flux subject to requirements of it being less than the potential rate of evapotranspiration.

In the present simulations there has been no limitation in the availability of water for evapotranspiration, i.e. the evapotranspiration is equal to the potential rate. The variability in soil physical properties then only has an effect on the relation between water stored in the root zone and the deeper percolation, when the boundary conditions at the surface are fixed fluxes of precipitation/snow melt and potential evapotranspiration. The deterministic model itself produces results with high precision and good mass balance but depends entirely on reliable estimates of the surface fluxes and soil physical properties, especially hydraulic conductivity. The calculations here indicate that the uncertainty in the characterization of the soil physical parameters does not have a drastic influence on the accuracy with which the groundwater recharge can be calculated. The method stands and falls, however, with reliable estimates of surface fluxes, especially the actual evapotranspiration. In the present study this estimate has been given little attention and further studies are needed.

The Darcian flux calculations can be performed for situations when there exist simultaneous registrations of actual soil suction below the active root zone at several sites. This allows an estimation of the spatial variability in the suction values. At Haslemoen there have been no direct measurements of soil suction, but only indirectly through observations of soil water content. The corresponding suction values are determined from pF curves at the sites of observation. On two occasions, the 28/6/1985 and 11/9/1986, the observations were frequent enough to allow an estimate of groundwater recharge from Darcian flux calculation including its spatial variability. The observed mean values and variance covariances are shown in Table 1.

Applying equations (7) and (9) the mean and variance of $\ln(q_z)$ are calculated. These values are also included in Table 1. The expected values of the groundwater recharge for the two occasions are thus 4.9 [1.5, 16.8] and 1.7 [0.1, 25] mm/day, respectively. The values in brackets are 95% confidence intervals.

Although some further uncertainties have been introduced by calculating suction values from pF curves and by assuming the water content independent of the soil physical properties that can violate the results, it is obvious that this method is very sensitive to variability in soil physical parameters but also to the possible errors in the determination of these parameters as well the soil suction. The condition that the flux equals the hydraulic conductivity seems simple, but we come up against the greatest difficulty in determining the hydraulic conductivity accurately. The range of possible values as induced by the variability in physical parameters and soil suction is large. It should be noted that the range is largest for the driest situation on 11/09/1986. The most important source of variability is that of soil suction, especially in the latter case. It is possible that the achieved values reflect the natural variability in vertical fluxes and that these can vary within an order of magnitude in an aquifer like the one studied. The unsaturated hydraulic conductivity is dependent on the soil texture and structure but also on the hysteresis affected state and content of soil moisture. Small changes in water content give rise to significant changes in unsaturated hydraulic conductivity. To this we need to add the possible effect of for instance hysteresis, which is neglected in the present parameterization. Although this method seems very straightforward a conclusion is that at present it gives more uncertain estimates than for instance applying a simulation model for water balance.

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