A 3-dimensional, time-variant, numerical groundwater flow model of the Øvre Romerike aquifer, southern Norway

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Using the U.S.G.S MODFLOW code, coupled with a Penman-Grindley type recharge model, it has been possible to produce a transient, 3-dimensional groundwater flow model of the Øvre Romerike aquifer. The MODFLOW code was modified to allow the water table to rise across the boundary of aquifer layers, and the recharge model was improved to allow for snow storage effects during winter.

183 groundwater level observation data from autumn 1975 and flows in groundwater fed springs and streams have been used for calibrating a steady state model. Allowing for the limited resolution attainable using 500 x 500 m grid blocks in areas of high water table gradient, the fit achieved was satisfactory. The calibrated distribution of hydraulic conductivity was then used to simulate time-variations in groundwater levels over a period in excess of 30 years. Long-term calibration data at three observation wells showed satisfactory fits with modelled data, again allowing for the limited spatial resolution of the model, and the uncertain elevations of the real data points. A seven-month running average filter was able to simulate the damping effects of the unsaturated zone on recharge maxima and minima, resulting in an improved fit.

The modelling work has indicated that hydraulic conductivity values obtained from grain size distributions tend to lead to underestimates of aquifer transmissivity. The model provides a framework for further modelling work on contaminant transport at Trandum landfill, but it is debatable whether the model can be used, unmodified, for contaminant modelling in its present form.

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Background

The Øvre Romerike aquifer (Fig. 1) is the largest discrete aquifer in Norway; in fact, it is one of the few areally extensive Quaternary aquifers in the country, covering an area of approximately 105 km^2. Situated some 40 km north of Oslo, the proximity of the Øvre Romerike aquifer to Norway's most populated and industrialised region has led to increased interest in the aquifer as a potential source of water. At present, however, the aquifer is utilised to only a few percent of its potential (Bryn 1992), supplying several military bases situated on the aquifer, and some small local communities. Nevertheless, there has already been considerable controversy arising from the conflicting interests of industry/the military and those who wish to preserve groundwater quality.

Authorities have placed emphasis on protecting the quality of the groundwater in the aquifer from 'potentially polluting activities', as it represents a possible future water resource for municipalities in the Romerike area (Østlandskonsult et al. 1991) with a potential exploitable capacity of up to 570 l/s (Snekkerbakken 1992).

Most of the area is either forest or farmland. A limited degree of 'urbanisation' (villages, airport) is concentrated on the southern part of the aquifer. Substantial areas of the aquifer are also occupied by military bases and training grounds.

Known sources of pollution to date on the Øvre Romerike aquifer include various military activities, a military and civil airport at Gardermoen (Davidsen 1991), two known leakages from oil storage tanks (Storrø & Banks 1992) and several landfill sites (Sæther et al. 1992, Omejer et al. 1992).

NORTHERN ROMERIKE
QUATERNARY MAP

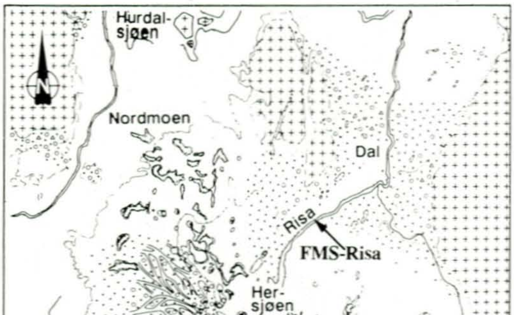


Fig.1. Øvre Romerike area showing Hurdalsjøen, Hersjøen, Transjøen, Dagsjøen, airport, Trandum, the rivers Risa, Leira, Vikka and flow gauging stations on the Risa and Vikka.

Perhaps the greatest controversy surrounds the relocation of Oslo's main international airport to Gardermoen, on the aquifer (Englund & Moseid 1992, Solnørdal 1992).

Geology & hydrology

The geology of the area is described in detail by Longva (1987). The Romerike aquifer consists of an approximately 105 km^2 expanse of Quaternary ice-marginal delta sediments built up to, and in some locations above, the marine limit. The upper part of the aquifer consists of dominantly glaciofluvial sand and gravel deposits, with areas of aeolian sand and glaciolacustrine sands and silts. These are underlain by glaciomarine/marine silts and clays. The upper, coarser part of the deposit exceeds 30-40 m thickness in some areas, while the total depth to bedrock (including marine silts and clays) may be as much as 100 m (Østmo 1976, Jørgensen & Østmo 1990). The Romerike deposit is believed to contain 150-200 million m^3 of good quality sand and gravel (Wolden & Erichsen 1990) and a number of pits have been excavated for the exploitation of this resource throughout the area.

The aquifer is bounded below by gneisses of diverse composition and origin and to the south and west by marine clay sediments which have very low hydraulic conductivities. The delta top is relatively flat, lying at an altitude of around 200 m above sea level and is surrounded by hills of outcropping basement rocks, except to the south and south west where the marine sediments form a lower lying plain at an altitude of around 150 m.

There has been intense hydrogeological and hydrochemical investigation of the aquifer due to its selection as a study area for the International Hydrological Decade (Falkenmark 1972, Norwegian National Committee for IHD 1973, 1975). This has resulted in the publication of a hydrogeological map (Østmo 1976), and descriptions of the hydrogeology (Jørgensen & Østmo 1990) and hydrochemistry (Jørgensen et al. 1991).

The aquifer is entirely fed by recharge from precipitation. The main surface water drainage of the aquifer consists of the northwards-flowing River Risa and Hersjøen Lake. The river and lake are almost entirely groundwater-fed (Jørgensen & Østmo 1990, Hongve 1992). Østmo's (1976) map indicates that the major central part of the aquifer drains towards Hersjøen and the River Risa. The marginal parts of the aquifer drain outwards towards springs in the periphery of the delta. Hydraulic and hydrochemical balances for the aquifer have been calculated.

Objective of the modelling work

Groundwater flow modelling provides a quantitative tool that, when sufficient effort is invested in model calibration and verification, can be used to predict the short and long term effects of varying recharge and groundwater abstraction. It can also provide a basis for further modelling of groundwater pollution incidents and their remediation, and can help in planning preventive action to maintain groundwater quality.

As competence in such modelling is at a rather low level in Norway, part of the aim of the project reported here has been to help build modelling competence at the two participating institutions (the IBM Bergen Environmental Sciences and Solutions Centre [IBM/BSC] and the Geological Survey of Norway [NGU]). The more concrete objective of the project has been to develop a three-dimensional, time-dependent model of groundwater flow in the Øvre Romerike aquifer. This model has been constructed as a basis for a more detailed modelling study of the impact of a landfill site at Trandum, near the centre of the aquifer, on groundwater quality. The Trandum site lies in relatively close proximity to the Military's abstraction wells, has been investigated in detail by NGU (Misund & Sæther 1991, Sæther et al. 1992).

The geological model and surface features

NGU possesses large amounts of both geological and geophysical data from the Øvre Romerike area which have been used in the study. The main sources of data used to construct the geological model are:

i) Borehole data collected during the International Hydrological Decade (IHD). These data have been published by Misund & Banks (1993).

ii) Borehole data from groundwater contamination investigations at Trandum and Sessvollmoen (Storrø 1991, Banks 1991, Misund & Sæther 1991, Storrø & Banks 1992 and Sæther et al. 1992).

iii) Data collected in connection with the new Oslo International Airport at Gardermoen (NGI 1991).

iv) Seismic profiles run during the IHD; summarised on Østmo's (1976) map and by Longva (1987).

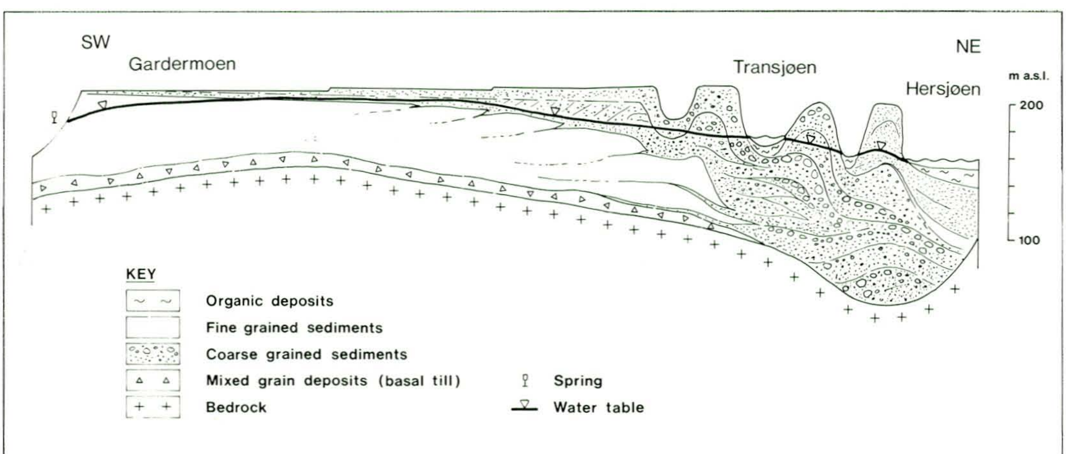


Fig.2. An interpretation of the structure of the Romerike delta, after VIAK (1990).

It is widely accepted that, although complex, the Romerike deposit is a generally fining-downwards glacio-fluvial and glacio-marine sequence (Longva 1987). In some boreholes, and most seismic profiles, it is possible to distinguish at least two distinct subdivisions or 'layers'; an upper layer of medium-to-coarse sands and gravels and a lower layer of fine sand and silt grading down into clay. It must be pointed out that this is a considerable oversimplification; in the upper coarse section, wedges of fine sands and silt occur. In the silty layer recent investigations have indicated that several discrete sand horizons can be detected in the Gardermoen area, possibly associated with marginal spring horizons (Sønsterudbråten 1992). VIAK (1990) have also constructed block diagrams (Fig.2) which, even though they are based on a mixture of 'enlightened guesswork' and real data, do illustrate the aquifer's considerable complexity. However, because of the limited available data and the resolution possible in a numerical model, the aquifer structure chosen for modelling is simplified to two layers; a coarse-grained (high conductivity) upper layer and a finer (lower conductivity) lower layer.

The topography, rivers, roads and other prominent features were derived from 1:50,000 topographical maps of the Romerike area. The topography is taken to represent the top of the upper aquifer layer. Contour maps of the interface between the two layers and the aquifer base (top of bedrock or, where present, very low permeability marine clays/till) have been constructed, primarily from the seismic profiles and some borehole data, and thereafter digitised (Fig. 3,a,b,c; Fig. 4).

The map of the aquifer base shows a deep channel in the bedrock lying under Hersjøen and the bed of the Risa river, and a deep basin in the south created by a channel in the bedrocks against which the marine sediments abut. The sediments of the aquifer essentially fill in this topography to produce the flat-lying topography of the delta top. The geological model shows that both the upper sediment layer and the total sediment pack are thickest in the region of Hersjøen.

Hydraulic parameters

Using samples from the boreholes from sources i) to iii) above, grain size distribut-

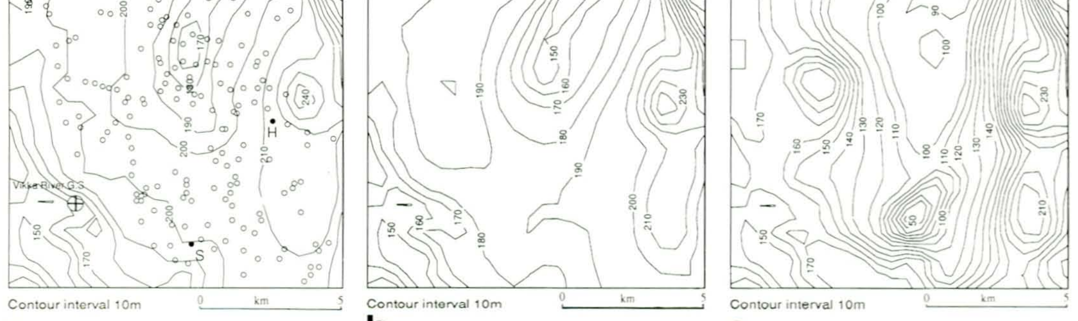


Fig.3 (a) Contour map of topography, showing observation boreholes (circles) and stream gauging stations (crosses) used for model calibration. Filled circles show Nordmoen (N), Sand (S) and Hauerseter (H) observation boreholes. (b) Contour map of top of fine layer. (c) Contour map of base of aquifer.

ions have been analyzed at NGU. These distributions have been used to estimate hydraulic conductivity using 10% and 60% (d_{10} and d_{60}) grain size fractions by the Bayer method (Langguth and Voigt 1980).

In using these estimates of hydraulic conductivity, the limitations of the method and the quality of the data must be considered. The samples from the boreholes are of varying quality. The data from i) above are largely from sediment samples which have been rinsed up with drilling water, or pumped up through a slotted pipe, and are hence likely to be depleted in both fines and the coarsest fraction. Samples from ii) and iii) above are likely to be more representative. Boreholes under investigations ii) were sampled using a 'throughflow' sampling device. Many of these considerations indicate that the estimates of hydraulic conductivity using this method are probably incorrect. However, they provide a useful starting point for modelling work, under the expectation that they will need to be adjusted during calibration. All hydraulic conductivity data were plotted on a histogram, and four main maxima were identified corresponding to clay, silt, fine sand and course sand/gravel (Fig. 5b).

Very few pumping tests have been carried out in the Øvre Romerike aquifer which have yielded reliable values of storage coefficients. The following values have thus been used, as indicated by Fetter (1988)

Coarse layer $S_Y = 0.25$, $S_S = 0.0003\ m^{-1}$
Fine layer $S_Y = 0.20$ $S_S = 0.0003\ m^{-1}$

Where S_Y is specific yield (i.e. unconfined) and S_S = specific storage (confined).

Recharge model

Recharge has been estimated from available daily rainfall and potential evapotranspiration data from Gardermoen meteorological station. Surface runoff is assumed to be negligible, due to the porous sediment material, the flat delta surface and the observed lack of surface watercourses on the aquifer. A 'soil moisture model' of daily Penman/Grindley type (see e.g. Rushton &

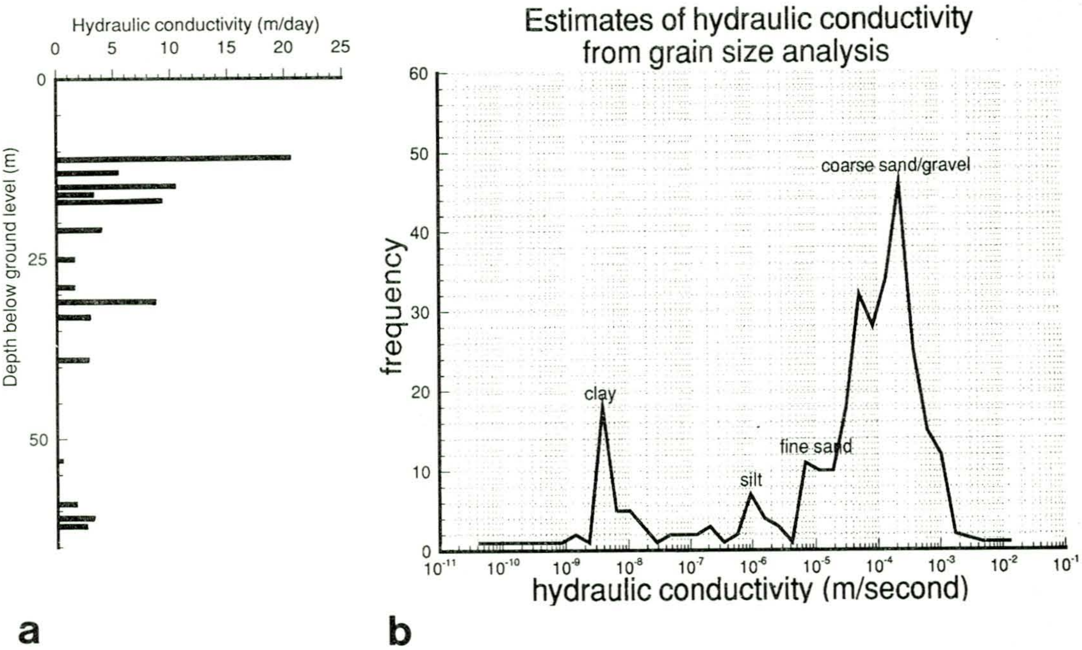


Fig. 5. (a) Hydraulic concictivity estimates from grain size distributions for borehole 75 at Nordmoen (UTM coordinates 167816). Note the rapid variation in hydraulic condictivity values with depth. (b) Histogram of all hydraulic condictivity estimates (using the Bayer method - from d_{10} and d_{60} grain sizes).

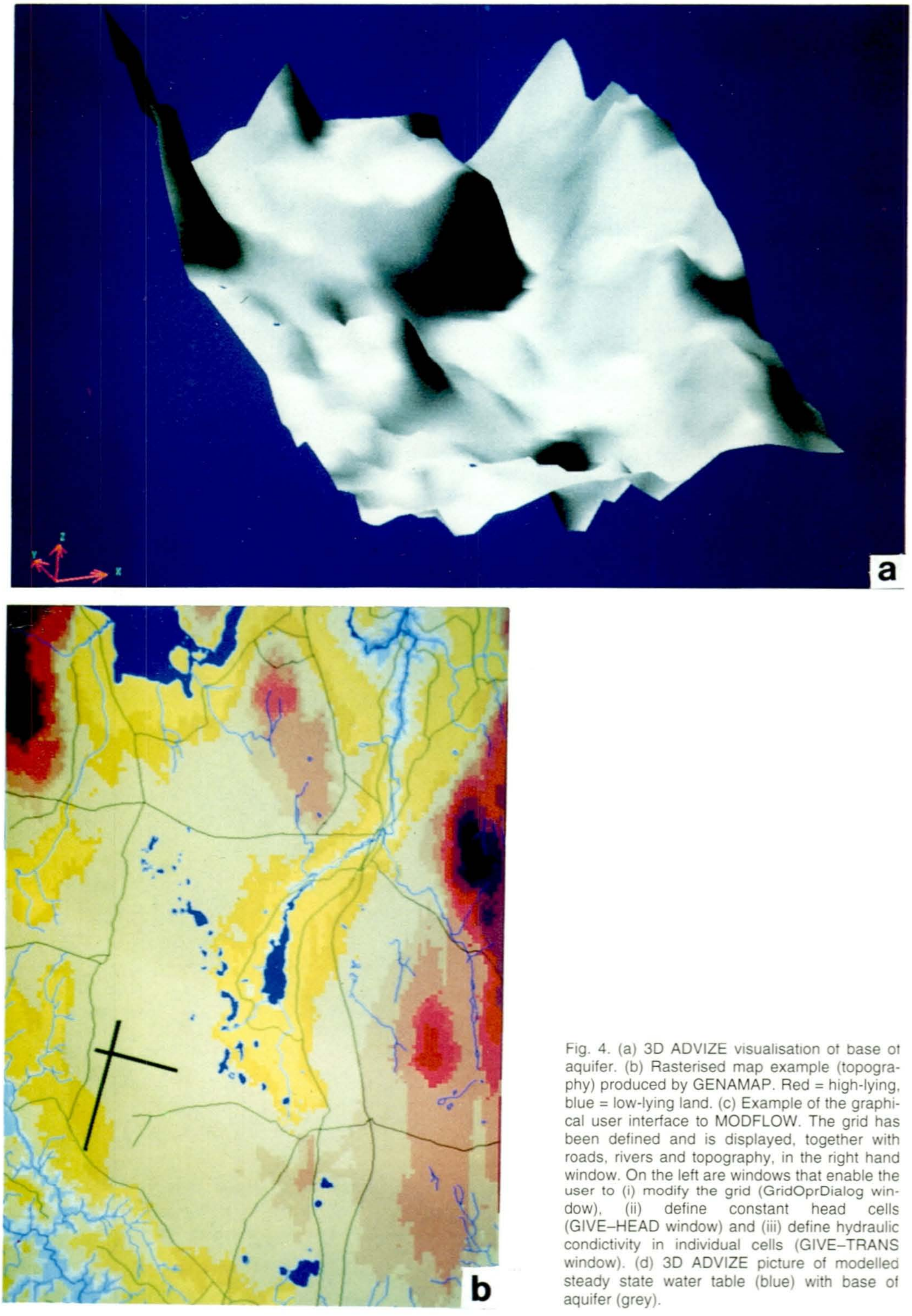
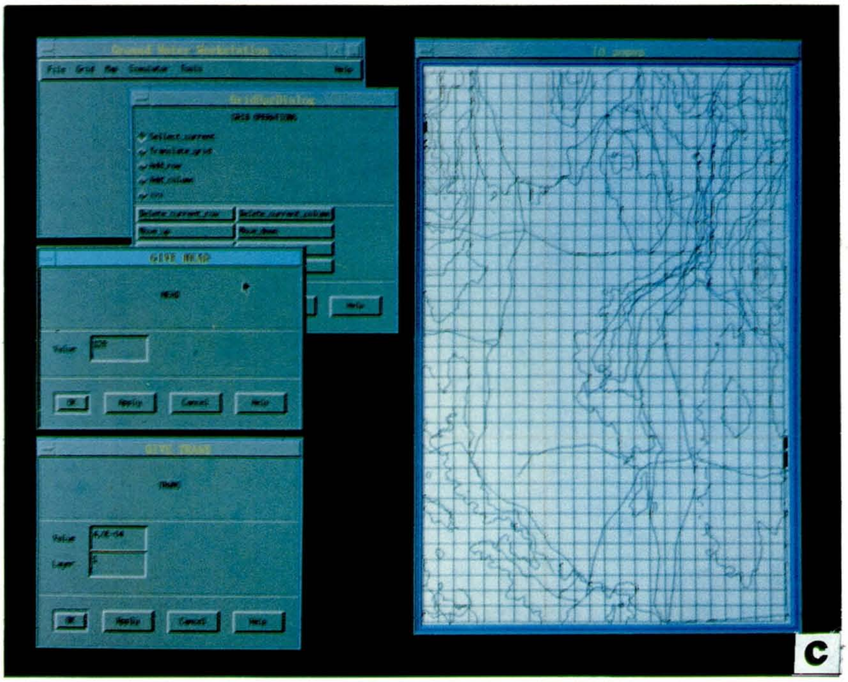


Fig. 4. (a) 3D ADVIZE visualisation of base of aquifer. (b) Rasterised map example (topography) produced by GENAMAP. Red = high-lying, blue = low-lying land. (c) Example of the graphical user interface to MODFLOW. The grid has been defined and is displayed, together with roads, rivers and topography, in the right hand window. On the left are windows that enable the user to (i) modify the grid (GridOprDialog window), (ii) define constant head cells (GIVE–HEAD window) and (iii) define hydraulic condictivity in individual cells (GIVE–TRANS window). (d) 3D ADVIZE picture of modelled steady state water table (blue) with base of aquifer (grey).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Redshaw 1979, for details) has been used which considers the soil zone as a reservoir containing a certain amount of water (soil moisture). This quantity is typically quoted in mm (i.e. $m^3/1000\ m^2$). The model assumes that percolation of water from the soil down to the water table only occurs if the soil is at 'field capacity'. The degree of undersaturation with respect to field capacity is called the 'soil moisture deficit' (SMD), and is expressed in mm.

The model assumes the following scenario:

For any given day:

$$\nabla SMD - R = AS = AE - P$$

Where AS = actual change in soil moisture deficit (ΔSMD - R)
P = precipitation (corrected for surface runoff)
AE = actual evapotranspiration
R = recharge to groundwater

And R = 0 if SMD > 0.

All values are quoted in mm. AS and AE are calculated from potential evapotranspiration (PE), using the relationships:

AE = PE **if** PS < 0 **or** (PS > 0 **and** SMD < C)
AS = 10% x PS **if** (PS > 0 **and** SMD > C)

where PS = potential change in soil moisture deficit (PE - P)
and C = root constant (taken as 175 mm in this case), a constant which reflects the decrease in available water for evapotranspiration by plants as the soil dries out.

Use of snow data in the recharge model

The most common versions of such a model treat snow exactly as rain, i.e. that snow infiltrates directly into the ground. In Scandinavia, this is not satisfactory, as during winter a store of precipitation is built up in a long-lasting snow cover, without sig-

Table 1. Comparison of recharge model with empirical data. Annual averages for period 1968 - 1974. (In model, run-off = 0, root constant = 175 mm). The recharge models with and without the snow package yield similar average results for the period in question, but the snow package radically improves the *distribution* of recharge within a given year. In the third column, recharge is not exactly equal to (precipitation - evapotranspiration) due to end effects of snow storage at the beginning and end of the simulation period).

	Empirical (Jørgensen & Østmo 1990)	Recharge model without snow package (snow treated as rain)	Recharge model (with snow package)
Precipitation mm	794	844	843
Evapotranspiration mm	400	445	447
Recharge mm	394	399	407

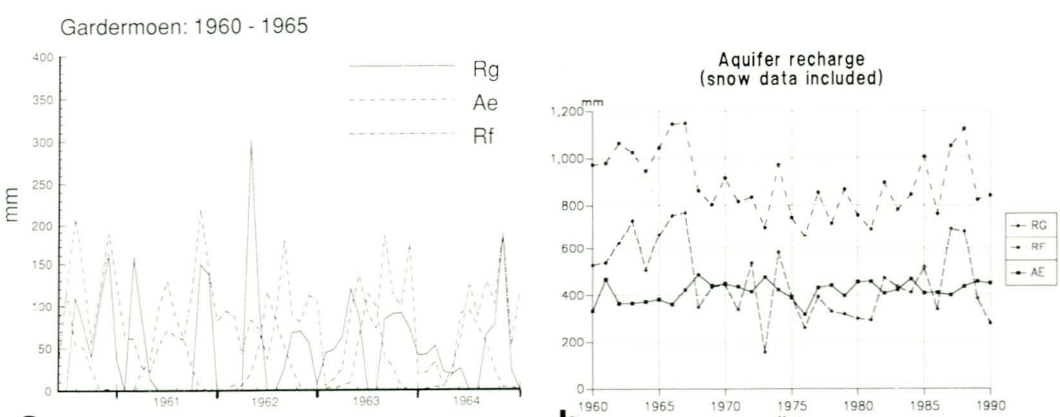


Fig. 6. Recharge model, (a), monthly data, (b) yearly data.

nificant amounts being recharged to the soil and further to the water table. In Norway, the winter is the period when groundwater levels recede to their lowest levels (Kirkhusmo & Sønsterud 1988, Nordberg 1980), in contrast to e.g. England, where the majority of recharge occurs in winter. Unfortunately there is no simple relationship between the amount of water (in mm) which is stored in the snow cover, and its thickness, as the snow becomes steadily more compacted during winter (Nor.Nat.Comm. I.H.D. 1975). One cannot therefore use changes in snow *depth* to assess whether accumulation or melting of the snow cover is occurring. In this work it has been chosen to use a simplified model. Data from the Meteorological Institute contain daily measurements of precipitation, snow depth and degree of snow coverage (S) from 0 (no cover) to 4 (100 % cover). The model considers 3 different modes of treatment of snow data:

i) When $S \leq 1$, the model functions as the normal summertime model.
ii) When $S > 1$, the model switches to 'accumulation mode'. No precipitation seeps into the soil layer; it is all stored in the snow layer. Evapotranspiration of existing soil moisture is allowed, such that the SMD steadily increases during this period, and no groundwater recharge occurs.

iii) When the snow *cover* (as opposed to depth) begins to decrease ($S_n < S_{n-1}$), the model switches again to 'melting mode'. New precipitation is allowed to enter the soil zone and all precipitation which is stored in the snow layer is allowed to melt and seep into the soil, where it is treated by the model as normal rainfall. The melting occurs as follows:

Daily amount of meltwater = (precipitation stored in snow) / (duration of melting period)

During the melting period, relatively large amounts of meltwater seep into the ground, and the SMD is quickly satisfied, allowing recharge to the water table. The melting period finishes when S reaches 0 again (and the model returns to 'normal mode'), or when S begins to rise again (and the model returns to mode ii). It should be mentioned that the model does not take into account changing surface run-off due to frozen ground conditions.

Results

The recharge model was run for the period 1968-1974 (Fig. 6). The results were found to correspond very well to empirical estimates calculated from the hydrological balance of the aquifer by Jørgensen & Østmo (1990) for the same period (Table 1). The results of the model also give the typical two recharge peaks every year, one at snow melt and one in late summer, a typical pattern for inland Norway (Nordberg 1980).

MODFLOW

To model groundwater flow, the numerical model MODFLOW, written by the USGS, was chosen (McDonald & Harbaugh 1988). This is a finite difference model which is widely used throughout America and Europe and thus has the advantage of being well tested, with clear instructions for its use. MODFLOW can only simulate flow in the saturated zone (below the water table).

MODFLOW requires a geological model defined on a grid (dimensions and size chosen by the user) as well as arrays defining the cell type, horizontal and vertical hydraulic conductivities, and storage coefficients (for transient simulations) for each grid block in each layer. In addition, files describing the properties and location of rivers, drains, wells and areas of constant head must be provided by the user. In many cases, these files of data must be compatible with each other. Much of the work involved in using MODFLOW lies in the creation of these files. At IBM/BSC, a GUI (graphical user interface) was built to automatically perform as much of this work as possible. The contour maps of the geological model were automatically digitised using the data capture software developed at IBM/BSC in

Bergen. The prepared maps were scanned and the resulting images vectorised to give the contours as a series of points. Each contour was then assigned a height value using the AUTOCAD package. These digitised contour maps were then rasterized using the commercial GIS package GENAMAP (GENYSIS). This procedure results in maps consisting of small square regions (pixels) each of which has a height associated with it. Each pixel represents a region of 50 by 50 m. These rasterised maps provided the basic input for the GUI and an example is shown in Fig. 4b. MODFLOW requires a grid that is rectilinear (i.e. squares or rectangles) when viewed from above. The GUI was used to interactively generate this grid and to assign constant head and 'no-flow' cells, river segments, and horizontal conductivities. The interface then generates the input files for MODFLOW, checking for internal consistency. Files containing information on recharge, drains and output control flags must be generated separately. The groundwater flow modelling study of the Øvre Romerike aquifer provided a useful test case for software developers at IBM/BSC to develop skills in graphical user interface programming. Although still in the prototype phase, the GUI provides the most important functions provided by commercial available interfaces to MODFLOW and to other groundwater modelling packages. The geological model and the modelling result were vizualised using a software package, ADVIZE, that has also been developed at IBM/BSC.

The grid and boundary conditions

A grid of 36 by 24 square (500 m x 500 m) blocks, covering almost the entire areal extent of the Øvre Romerike aquifer was generated using the user interface. Boundary conditions are defined so as to represent external influences on the groundwater system. As much as possible, natural features were used to determine conditions around the model perimeter. The crystalline bedrock underlying the aquifer and the clay-rich marine sediments in the southwest are assumed to be impermeable, and grid cells representing these lithologies are defined as 'no-flow' cells. The lakes Hurdalsjøen and Hersjøen were modelled as constant head cells (constant water level throughout the simulation). This seems reasonable since the level of these lakes varies

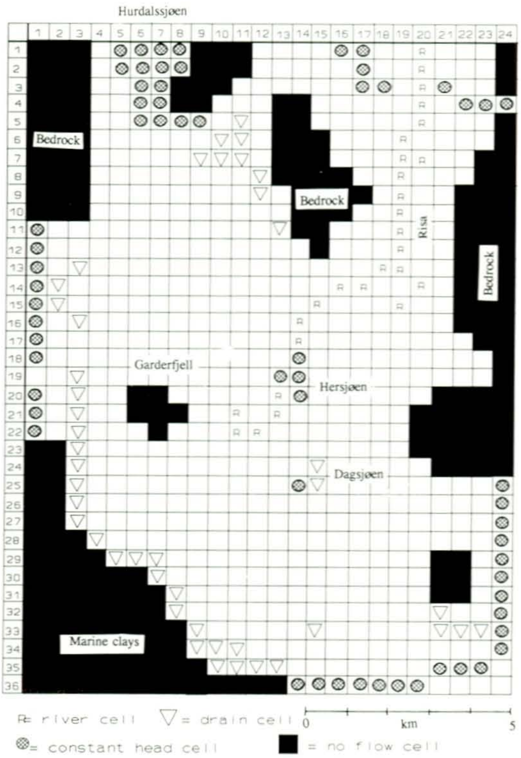


Fig. 7. MODFLOW grid for Øvre Romerike with cells defined.

very little with time and they are believed to be in some degree of hydraulic continuity with the aquifer. These no-flow and constant head cells fix conditions on a large proportion of the model's boundary.

The Risa river, the smaller stream from Transjøen to Hersjøen and the small eastern tributary of the Risa were modelled using the river package in MODFLOW. Here rivers are assumed to be continually active and are either influent or effluent with respect to the aquifer depending on the relative heights of the river bottom and the water table. The sediments in the river bed have been modelled with the same hydraulic conductivity as the aquifer at this level.

Springs which occur along the west and the southwest margins of the aquifer, near the boundary with the underlying marine sediments, are modelled using the drain package of MODFLOW. The springs are thus only active when the water table is higher than the drain level, and flow is proportional to the water table height above the spring elevation. This allows the simulation of streams or springs which can go dry, and prevents influent (recharge) conditions occurring.

The remaining small portions of the model's outer boundary, which cannot be defined as no-flow cells, drains or truly valid constant head cells, are assigned a reasonable value of constant head derived from Østmo's (1976) map. The water level in these cells should strictly be variable, but since there are only a small number of them and they are situated in regions where the aquifer is relatively thin, holding their water levels constant has only minor effects on the rest of the model.

The boundary conditions and allocation of cell types in the model are shown in Fig. 7. Throughout all simulations, it has been assumed that recharge is constant over the whole area.

The steady state model

The aim of the steady state modelling was to generate a hydraulic conductivity model for the two aquifer layers that produces water table levels and water fluxes in rivers and springs in acceptable agreement with observations. A total of 183 water levels measured in boreholes, largely in the period 5-7 November 1975, from Østmo's map (1976) were used in model calibration (Fig. 3a). Two parameters were used to test the fit of the model results to these borehole measurements. To test the fit of the model results to the observations on the scale of the whole model, the average discrepancy, D, was calculated:

$$D = \frac{\Sigma\,(h_m - h_o)}{n}$$

where h_m and h_o are the modelled and observed water table levels, respectively, and n is the number of grid blocks where borehole observations are available. To test the fit of model results to the observations on the scale of one grid block, the root mean square of the discrepancies, rmsD, was used:

$$rmsD = \sqrt{\frac{\Sigma\,(h_m - h_o)^2}{n}}$$

A good fit is obtained when D and rmsD are close to zero, although as individual deviations can be positive or negative, more emphasis was laid on rmsD than D itself. Since, in reality, several boreholes in a single grid block can exhibit considerably different water levels in areas of steep water table gradient, an exact fit (rmsD = 0) should not be expected. In the simulation, one grid block is allowed a single water level and only variations in water table on the scales larger than a single grid block can be modelled.

Other available methods of calibrating the model are to compare the fluxes in rivers and springs. Data are available for the water flux in the Risa river at Risabru (the NVE gauging station, at UTM Grid.Ref.^{6}203 ^{66}801 - Fig. 1, 3a). The flow in the Risa (Jørgensen & Østmo 1990, Norweg. Nat. Comm. IHD 1973, 1975) varies little throughout the year and has an average value of around 0.7 m^3/sec. In MODFLOW, the flux in rivers is not directly calculated, only the exchange rates between rivers and the aquifer. However, an estimate may be made by summing the fluxes from the aquifer to:

1) constant head cells representing Hersjøen
2) cells representing the stream from Transjøen to Hersjøen
3) the drain cells representing the springs at Dagsjøen
4) cells representing the Risa river upstream of Risabru

This sum is likely to overestimate the flow at Risabru as it does not include the effects of evaporation from Herjøen or possible losses to the aquifer between Dagsjøen and Hersjøen.

Water flowing from the springs along the southwest margin of the aquifer provides the main input to the Leira and Vikka rivers which are situated on the marine sediments and flow parallel to the aquifer's margin. Sporadic data is available on flux in the Vikka river (at UTM $^{6}154$ $^{66}722$ - Fig.1, 3a) for the period 1989 to 1991 (NVE 1991), giving an average of 0.089 m^3/sec. Approximately 25 % of the total number of springs feed the Vikka upstream of the gauging station, and it is assumied that up to 50 % of the Vikka's flow may be derived from surface run-off. The total flux from the springs along the southwestern margin is therefore estimated to lie between 2 and 4 times the gauged flow in the Vikka, i.e., between 0.36 and 0.18 m^3/sec.

The hydraulic conductivity model

For a given recharge rate, there are many possible hydraulic conductivity models that can satisfy the calibration data outlined above. It is therefore not possible to arrive at a unique solution to the hydraulic conductivity

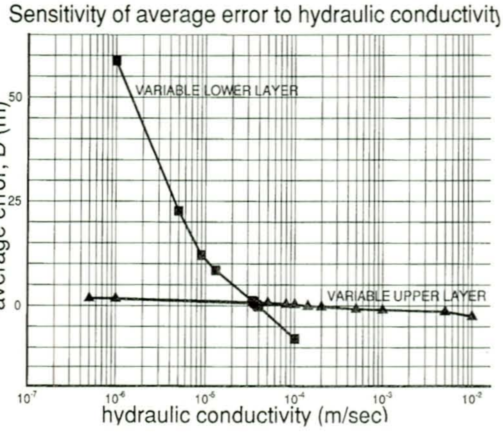


Fig. 8. MODFLOW, steady state model: homogeneous layer model - sensivity of D (average error) to variation in hydraulic conductivity of coarse and fine layers.

field, but by using the available data on hydraulic conductivity and recharge as a guide, a likely solution can be obtained. Time-dependent data indicate that the water table lies at approximately its average level for the year of 1975 during the month of November, although recharge is normally below average in this month. This discrepancy is due to the delaying action of the unsaturated zone on recharge arrival at the water table, so that November lies on the decaying limb of the previous hydrograph peak of earlier autumn. For the purposes of the steady state model, where the aim is to reproduce the water table levels in November 1975, the average modelled recharge rate for 1975 (1.27×10^{-8} m/s) was used, rather than the November value for this year (c. 1×10^{-8} m/s).

Due to the complex aquifer structure, it has not been practical to conclusively identify the boundary between the two modelled layers using the hydraulic conductivity estimates from borehole samples or to assign a given grain size sample to a specific layer. In detail, both the coarse and fine layers have a complex structure composed of discontinuous layers of widely varying grain sizes. The grain size samples reflect narrow, widely spaced intervals along the borehole, and hydraulic conductivities estimated from these samples show a wide range and rapid fluctuating trends, see Fig.5a. Thus the indentification of the effective boundary between the coarse and the fine aquifer layers and the calculation of an representive average hydraulic conductivity for each of these aquifer layers is impractical for individual boreholes. Unfortunately, little reliable pumping test data was available at the time of the study that might give an effective avarage hydraulic conductivity of the sediments. However, the grain size samples do provide a statistical sample of hydraulic conductivity estimates from the aquifer as a whole. A histogram of hydraulic conductivity estimates based on d_{60} and d_{10} data shows several peaks which are typical for coarse, fine and clay-rich sediments (Fig. 5b). In the absence of spatial data on hydraulic conductivity, this histogram was used as a guide to the likely average hydraulic conductivity values of the two modelled layers, which were thus estimated at $2.0x10^{-4}$m/s (coarse layer) and $7.0x10^{-6}$m/s (fine layer), corresponding to the 'coarse sand' and 'fine sand' peaks on the histogram. These conductivities were used as a starting point for model calibration. A model run using these values was, however, unsatisfactory, yielding water table levels which were too high regardless of the hydraulic conductivity assigned to the coarse layer.

Model of hydraulic conductivity for fine layer

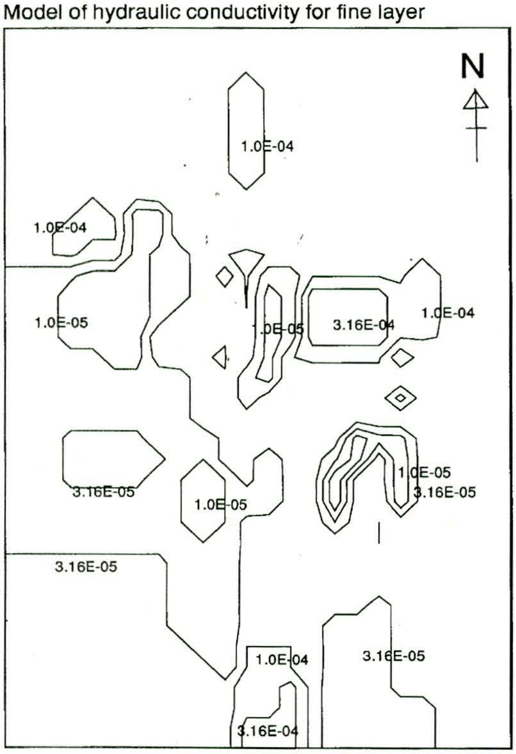
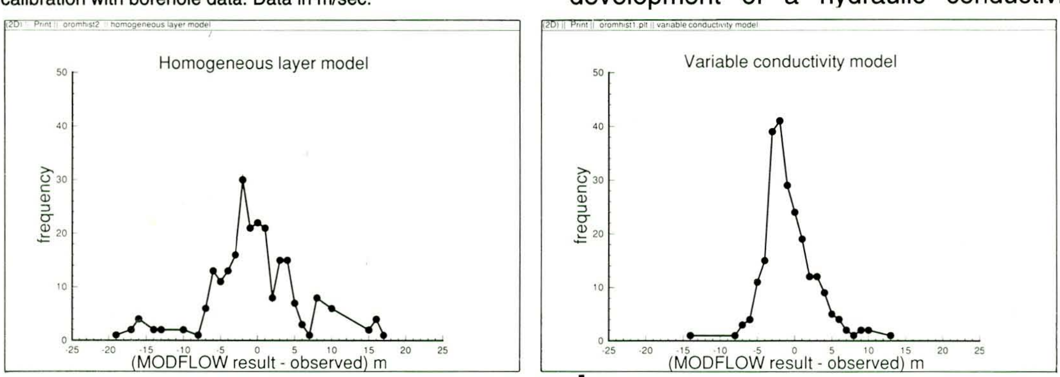


Fig. 9. Resulting hydraulic conductivity field for fine layer after calibration with borehole data. Data in m/sec.

In order to develop a strategy for modelling the hydraulic conductivity, the sensitivity of the homogeneous layer model to variations in hydraulic conductivity was tested by holding the hydraulic conductivity of one layer constant, varying the other and recording the change in the average error, D, as a measure in net change in the water table level. The results, shown in Fig. 8, show that the water table level is many times more sensitive to variation in the hydraulic conductivity of the lower fine layer than the coarse upper layer. This is consistent with the observation that the water table tends to be situated close to the interface between the two layers over large areas of the aquifer, indicating that water table levels are largely controlled by properties of the fine layer. A hydraulic conductivity of $7.2x10^{-4}$ m/sec was assigned to the coarse layer, a near maximum value. The hydraulic conductivity of the fine layer that minimized the total error, D, was then found. This resulted in a hydraulic conductivity value of $3.2x10^{-5}$m /sec for the fine layer, possibly reflecting the numerous coarser intercalations that are known to exist within that layer.

The modelled estimates of the 'average' hydraulic conductivity of each layer (i.e. 7.2×10^{-4} and 3.2×10^{-5} m/s) were considered a satisfactory intermediate point in the development of a hydraulic conductivity

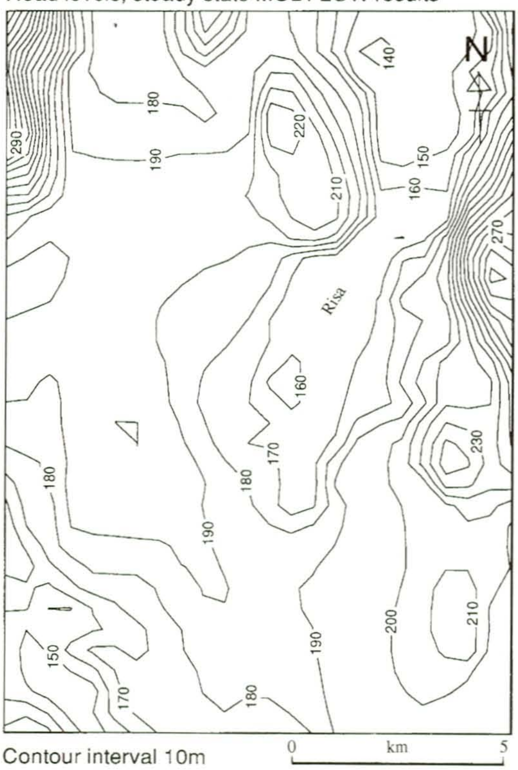


a b

Fig. 10. Histograms of discrepancies between steady state model and borehole data for: (a) homogeneous layer model. (b) calibrated hydraulic conductivity model.

model for the Øvre Romerike aquifer. The homogeneous layer model described abo-

Head levels, steady state MODFLOW results



Contour interval 10m

Fig. 11. MODFLOW steady state results for head. Contours in metres a.s.l.

ve, although miminizing the average error, D, gave a relatively high value of rmsD (5.86m). To improve the fit of modelled to observed water levels, the hydraulic conductivity field within each layer was allowed to vary horizontally. The fit of the modelled to observed water level data was improved as much as possible by modifying the spatial distribution of hydraulic conductivity within the fine layer, this layer's hydraulic characteristics being the most important for determining the water table level. Using this approach, the conductivity distribution in Fig. 9 was obtained, with rmsD values improved to 3.4m while maintaining a low value of average error, D (0.22 m) - see Fig. 10. Although the rmsD error may sound large, it must be remembered (as previously explained) that one is comparing the average water table within a 500 x 500 m grid block with the observed water level data from a single point within the block, often in areas with high water table gradients. The simulated water table level is shown in Fig. 11. A three dimensional view of the aquifer base and the modelled water table are shown in Fig. 4d. Since extensive spatial data on the hydraulic conductivity of the aquifer is not available, it is not possible to check the permeability model directly with observations. However, values of flux at Rissabru, calculated in the manner described above, and flux from the springs along the southwest margin of the aquifer were acceptably close to the observed values, and are listed in Table 2.

By comparing the groundwater head values in the upper and lower layers, one can identify vertical head gradients and areas where upward or downward flow is occurring. The 2-layer model is clearly not adequate to satisfactorily model vertical head gradients, but in general downward head differentials are observed in the area corresponding to the main watershed (i.e. recharge area), while upward differentials are observed along the River Risa (discharge area).

The transient model

Modifications to MODFLOW

The version of MODFLOW which was available free of charge from the USGS at the time of the study was designed to model situations in which the water table is, in general falling, e.g. the development of

Table 2. Outflows from springs; modelled and observed.

Location	Modelled flow m^3/s	Observed flow m^3/s
R. Risa at Risabru	0.82	0.7
Springs in SW	0.29	0.18 - 0.36

drawdown cones due to pumping. The model is so written that if a model cell goes dry, it is redefined as a 'no flow' cell and remains inactive for the remainder of the simulation. With this code it is therefore not possible to model the behaviour of a water table which is expected to rise over the boundary between two layers. Thus it was necessary to modify the MODFLOW code to allow such a rise to occur. This modification was performed by IBM/BSC and tested to check stability and convergency with steady state solutions when using constant recharge in transient mode. New versions of MODFLOW now have the capability of modelling rising water levels.

Input to the transient model

For transient simulations using time-dependent recharge data (derived from the recharge model), the length of stress period (i.e. time over which external stress conditions such as recharge are constant) was set to 1 month, the maximum time interval between successive measurements in the observed data. Two time steps of equal length per stress period were found to be sufficient to resolve movements of the water table and no significant difference in the results was obtained by increasing the number of timesteps. Recharge data was calculated daily and summed for monthly intervals from the meterological data as described above, for the time period from May 1960 to June 1991, giving a total of 374 stress periods.

The model runs commenced in 1960, whereas observation of the water table only commenced in 1965. Two initial water table levels were used in the model to investigate the effect of initial water table levels on the model results. These two initial conditions were generated by using the calibrated, spatially variable, hydraulic conductivity model and running MODFLOW in steady state mode with recharge values corresponding to:

(1) the average value for 1975 (1.27×10^{-8}m/sec - as used in the calibration of the hydraulic conductivity model).

(2) 8.293×10^{-9}m/sec representing approximately the minimum annual recharge for the time period 1965 to 1990.

The model results are compared with observed time series data from the three observation wells at Hauerseter, Sand and Nordmoen (Kirkhusmo & Sønsterud 1988) for the period from May 1967 to September 1991 in Fig. 12. It was found that the lower initial (May 1960) water table, generated from a recharge of 8.293×10^{-9} m/sec, gave

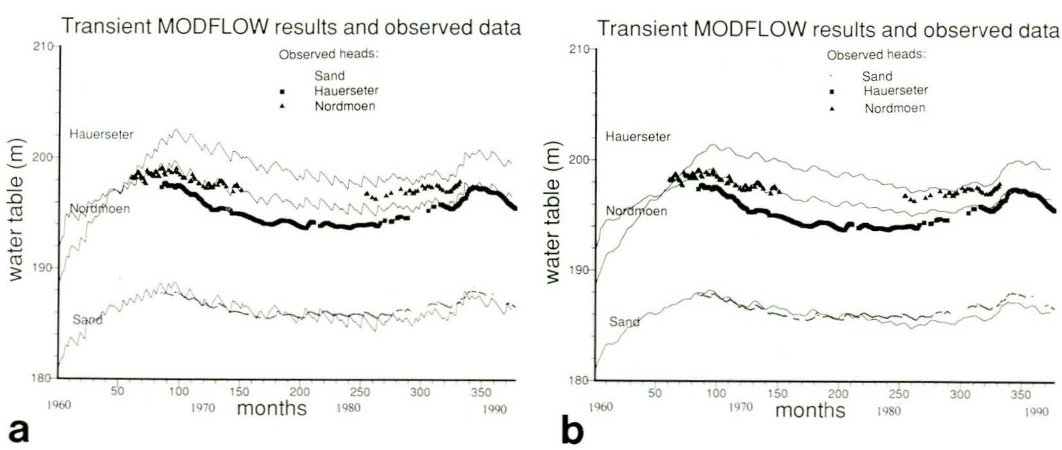


Fig. 12. Transient MODFLOW results with steady state heads generated using calibrated hydraulic conductivity field (7Fig. 16) and minimum recharge of 8.239×10^{-9} m/sec: (a) using monthly recharge. (b) using running averages of monthly recharge over 7 month intervals.

a slightly better fit to real data than the higher initial water table. The model shows similar trends to the real data, but with the absolute elevation of the model results too high for Hauerseter by some 3-4m, and an overall good fit for Nordmoen and Sand. For all wells, however, there is a slight long term 'decline' in the modelled water table relative to the real data. This leads to the modelled results for Sand and Nordmoen falling below observed water levels for the latter part of the data series. All three model trends show much greater seasonal variations than the observed data (Fig. 12a). This is due to the 'damping' effect of the unsaturated zone (i.e. the 'smearing out' of major recharge peaks) which is not included in the model. It was found that this 'smearing' effect could be simulated by taking running averages of the recharge data over 7-monthly periods, and the results are shown in Fig. 12b. This modified model produces results which show very similar variations to the observed data, but with the modelled results lagging some months behind the observed data. The magnitude of this lag varies from 0 to 2 months for Nordmoen, 2-4 months for Sand and 3-4 months for Hauerseter and is thus roughly correlated with the thickness of the unsaturated zone (1-3 m at Nordmoen, 10-20 m at Hauerseter and Sand - Bjor & Huse 1988). This lag presumably represents the time taken for the recharge water to travel through the unsaturated zone, an effect which is not included in the model.

The modelled trends all show a similar type of long term deviation from the observed data in that they fall less sharply in the time period from 1967 to 1976 and rise quicker in the time period 1980 to 1990.

Conclusions

Using the U.S.G.S MODFLOW code, coupled with a Penman-Grindley type recharge model, it has been possible to produce a transient, 3-dimensional groundwater flow model of the Øvre Romerike aquifer. It was, however, found necessary to modify the MODFLOW code to allow the water table to rise across the boundary of aquifer layers, and to modify the recharge model to take into account snow effects during winter.

A steady state model has been calibrated against 183 regional water level observation data from autumn 1975 and against the flows in groundwater fed springs and streams. Allowing for the limited resolution attainable using 500 x 500 m grid blocks in areas of high water table gradient, the fit achieved was satisfactory. The distribution of hydraulic conductivity calibrated using the steady state model was then used to simulate water table variations over a period in excess of 30 years. Calibration data at 3 observation wells showed satisfactory fits with modelled data, again allowing for the limited spatial resolution of the model, and the uncertain elevations of the real data points. A seven-month running average filter has been applied to the data to simulate the damping effects of the unsaturated zone on recharge maxima and minima, resulting in an even better fit. The seasonal variations in water table level appear to lag the modelled water level data by 0 to 4 months, indicating a delay period in the unsaturated zone.The delay seems to vary with the thickness of the unsaturated zone.

The modelling work has indicated that hydraulic conductivity values obtained from grain-size distributions tend to lead to underestimates of aquifer transmissivity. The model provides a framework for further modelling work on contaminant transport at Trandum landfill. It is doubtful, however, whether the model can be used for contaminant modelling in its present form. Average hydraulic conductivities are sufficient to reproduce the bulk flow of groundwater but contaminant flow is very sensitive to small-scale heterogeneities in the aquifer (Luckner & Schestakow 1991 and others). The rapid interchange of layers of different grain size observed in the aquifer are on the scale of metres or less. This is much smaller then the grid block dimension (500 m), the scale below which the model assumes the aquifer to be homogeneous. Coarse grain layers will allow for more rapid transport of contaminants, and if interconnection of such layers exists they will provide local rapid transport routes. The effect of these heterogeneities will therefore be to greatly increase the rapidity of propagation of the contaminant as compared to the model.

Another reason for caution with respect to contaminated transport modelling lies in the hydraulic conductivity model itself. This model is calibrated against observed heads and limited spring and stream flows only. Since contaminant transport is very sensitive to variation in hydraulic conductivity, additional checks on the validity of the model (from e.g. pumping tests) would be desirable, before it was used to model transport phenomena. Much detail data is, however, available at Trandum which could be used to generate a more detailed geological model of the area local to the landfill. Prospects appear very hopeful for a fruitful attempt at contaminant and hydrogeochemical modelling at that locality in the future, using the present modelling results as a fundament on which to build.

A full report of the modelling results can be found in Odling et al. (1993).

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