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Hydrological Model for the Tude Å Catchment

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A numerical hydrological model has been developed for a 450 km^2 Danish catchment using comprehensive field data. The model integrates a simple evapotranspiration model, a lumped flow model for a phreatic aquifer found in till, and a traditional two-dimensional groundwater model for a confined fluvio-glacial aquifer. A minimal but adequate number of model parameters were calibrated by trial and error to make the model fit 29-year time series of hydraulic head and stream runoff data.

By simulating a "semi-natural" hydrological situation unaffected by withdrawals it is demonstrated that groundwater development can change the water balance considerably. In the actual case withdrawals induce a 25% increase in leakage from the phreatic to the confined aquifer, and reduce stream base flow by up to 30% in normal years, and up to 35% in dry years. On the other hand the reduction in base flow is considerably smaller for the upper stream catchments.

Introduction

Within the last decade groundwater pollution has been recognized as a serious problem threatening many Danish water works which depend entirely on groundwater. Several of the country's major water supplies have been forced to give up wells or even well fields because of groundwater pollution. This is particularly the case in the vicinity of urban areas, where many well fields are still situated. It is

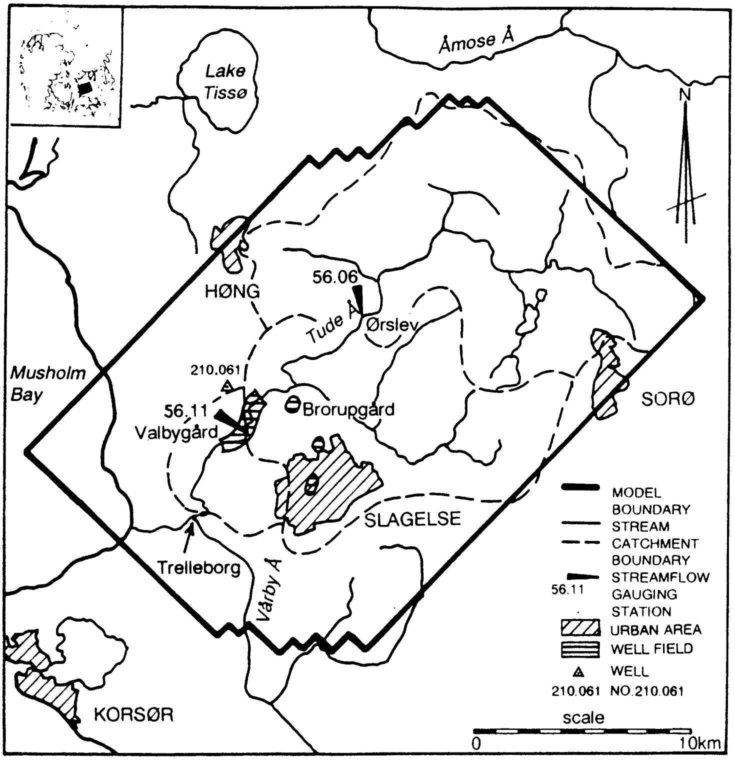

Fig. 1. Map of the investigated area.

therefore becoming the accepted policy for major water works to move potentially threatened well fields to unthreatened areas. In order to effect this in a rational manner the traditional hydrogeological investigations must be supplemented by hydrological modelling. Only thereby can one make realistic estimates of the recoverable groundwater resources, and weigh the pros and cons of alternative development plans, taking into account all the hydrological consequences.

During the period from 1988 to 1991 the Water Supply of Slagelse carried out an ambitious groundwater investigation. For decades one third of the required groundwater (1.3 million m^3 in 1989) was developed in the urban area, while two thirds (2.3 million m^3 in 1989) were developed from well fields along the stream Tude å, see Fig. 1. In 1987 the Water Supply recognized that the well fields in the urban area were threatened by pollution spreading from factory sites, such as the city's derelict gasworks. Hydrogeological investigations were initiated in order to find potential new well fields outside the urban area, and to predict the hydrological consequences of such a move of well fields, *i.e.* the effects on groundwater flow and hydraulic head in the confined aquifer and on stream flow in Tude å.

The investigations included the mapping of geology, groundwater catchments and topographical catchments in a 600 km^2 area and the recording of hydrological time series. This information was used to develop a hydrological model for a 450 km^2 sub-catchment, the catchment of Tude å. The present paper gives a description of the hydrogeology of the catchment and of the developed model. Model calibration is tested against time series of potential heads and stream flow. In Kemp & Lauritzen (1991) the model is used to point out alternative ways of replacing the threatened well fields in the urban area of Slagelse and to predict the consequences of each alternative. In contrast to these specific and local aspects, the present paper concentrates on the general application of the model to a problem which is of common interest to hydrogeologists, namely the consequences of groundwater withdrawals on the water balance in general and on stream flow in particular. This is demonstrated by repeating the dynamic hydrological simulations respectively with and without the current withdrawals within the area.

Hydrogeology

The description of the hydrogeology of the investigated area is mainly based on lithological data from hundreds of wells registered in the database of the Geological Survey of Denmark. However, in an area 5-10 km north of Slagelse, where wells are sparse, the lithological information has been supplemented with indirect measurements of lithology through geoelectric soundings, Kemp & Lauritzen (1991).

The Danish landscape was formed by glaciers and therefore the Quarternary geology is dominated by glacial till and fluvioglacial deposits. In the investigated area the total thickness of these Quarternary deposits varies between 40 and 80 m, see Fig. 2. The underlying Tertiary deposits generally consits of clay covering glauconitic sandstone and Danian chalk. However, in some areas the Tertiary clay is missing, so that the Quarternary is in direct contact with the sandstone.

Till is the dominant Quarternary deposit in the eastern, north-western and south-western areas, Figs. 2 and 3. Thin heterogeneous layers of fluvioglacial deposits, from which groundwater is hardly recoverable, are embedded in the till. However, in some areas, such as in the urban area of Slagelse, such embedded coarse deposits are more substantial and are important for water supplies. South of Slagelse and north-east of Høng the glaciers apparently mixed substantial portions of Tertiary deposits into the glacial till.

In the north-eastern, central and south-eastern part of the area the Tertiary deposits are overlain by a 10-30 m thick and apparently continuous layer of fluvioglacial sand and gravel. This, in turn, is overlain by till from later glacier advances.

From the ground surface downwards one can distinguish five hydrogeological units: a phreatic aquifer; an aquitard of glacial till; a mainly confined fluvioglacial aquifer; an aquitard of Tertiary clay; and a confined aquifer in Tertiary deposits.

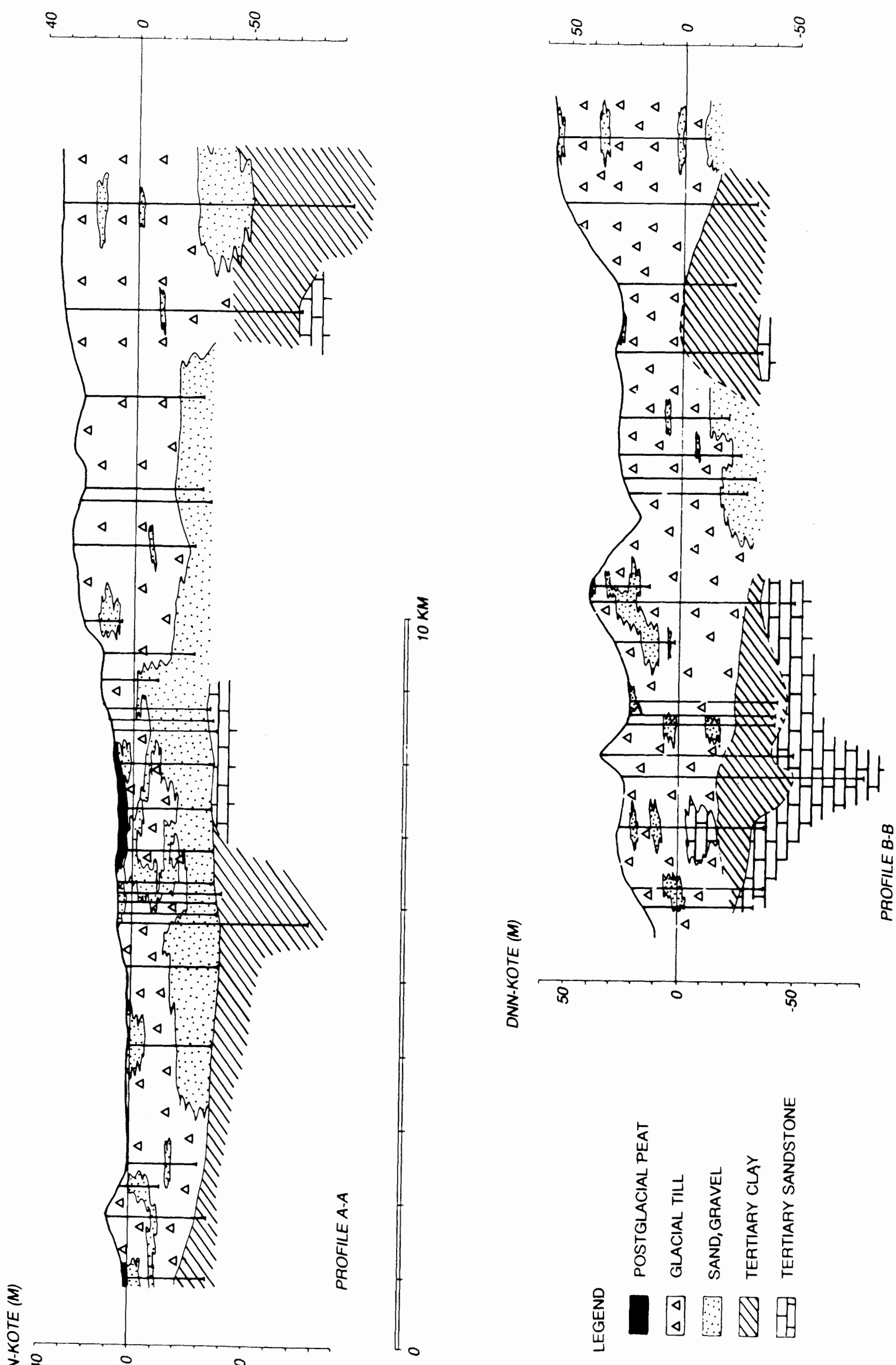

Fig. 2. Geological profiles (location shown in Fig. 3).

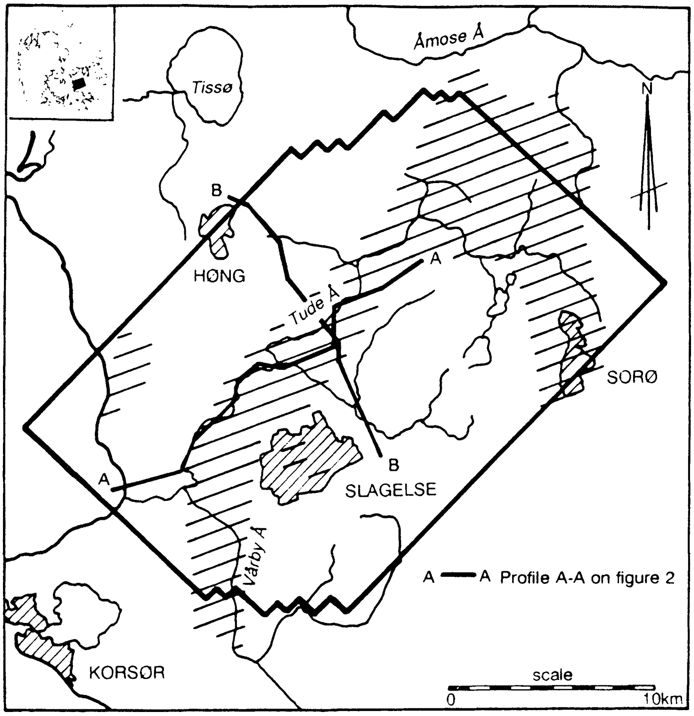

Fig. 3. Areas with a thick layer of fluvioglacial deposits.

The phreatic aquifer is found in the more permeable upper zone of the glacial till. The water table is located a few metres below the ground surface. As in the Suså catchment (Refsgaard and Stang 1981) it is thought that because of a small but strongly varying hydraulic conductivity the flow is dominated by near-surface horizontal flow. This flow is determined by artificial and natural drainage systems directing the water flows to the streams. Natural drainage systems are assumed to be concentrated in deposits with relatively high hydraulic conductivities.

The till represents the upper aquitard. Even though horizontal flow in the upper zone is dominant, vertical flow is significant and of major importance to the overall water balance of both the till and the underlying fluvioglacial aquifer. In areas where terrain elevations are high the water table elevation in the till is higher than the hydraulic head in the fluvioglacial aquifer and leakage is directed downwards, whereas it is directed upwards in areas with low elevations such as stream valleys and coastal areas. A few pumping tests give leakage coefficient values between 10^{-10} and 10^{-9} s^{-1} corresponding to a vertical hydraulic conductivity of the till of the order 10^{-9} to 10^{-8} m/s, Kemp & Lauritzen (1991).

The Quarternary fluvioglacial deposits form the aquifers of primary importance within the area. The above-mentioned layer of continuous deposits forms a regional confined aquifer which is exploited by most of the major well fields, *e.g.* at Valbygård and Brorupgård. In the remaining area the fluvioglacial layers are heterogeneous. However, there are local aquifers that can be exploited by minor well fields, *e.g.* in the vicinity of Slagelse. These local aquifers are mostly confined too, but due to large drawdowns in hydraulic head the aquifer in the urban area of Slagelse is now unconfined.

Groundwater flow in the fluvioglacial aquifers is dominated by horizontal flow. This is even true in discharge areas because the discharge rates are low compared to the horizontal flow rates (typically 1:1000). Numerous pumping tests have shown that the transmissivity in the regional aquifer varies between 10^{-3} and 10^{-2} m^2/s whereas transmissivities in local aquifers are lower, 10^{-4} to 10^{-3} m^2/s (Kemp & Lauritzen 1991).

Only little is known about the hydrogeology of the Tertiary aquitard and aquifer. The aquitard consists of a clay layer whose thickness varies and which can even be missing in some minor areas, Fig. 2. The leakage coefficient of the aquitard is unknown but thought to be low.

The Tertiary aquifer is in glauconitic sandstone. A few pumping tests indicate transmissivities in the order of 10^{-4} m^2/s near and west of Slagelse while values are one order of magnitude higher at Høng, Geological Survey of Denmark (1979). It is thought that the flow rates in the Tertiary aquifers are low and that the interchange of groundwater with the Quarternary aquifers is of minor importance to the overall waterbalance of the Quarternary aquifers. Only a few wells exploit groundwater from the glauconitic sandstone, *e.g.* in the western part of Slagelse and east of Høng.

Hydrological Data

The available hydrological time series of precipitation, potential evaporation, temperature, stream runoff, groundwater withdrawal and hydraulic head cover the period from 1961 to the autumn of 1989.

Daily precipitation data is available from the Danish Meteorological Institute's 5 gauging stations covering the model area. The measured precipitation data are corrected for wind effect, Allerup and Madsen (1980). The average annual precipitation for the 29-year period varies between 610 mm at the coast to 680 mm in the central/eastern part of the area. The annual variation of mean precipitation is considerable: the maximum deviation from the average is about 40%. Precipitation data are distributed according to Thiessen polygons around the gauging stations.

Daily potential evaporation is estimated according to Makkink (1957) by the Department of Agrometeorology, Danish Institute of Plant and Soil Science. The estimation is based on measured temperature and radiation at meteorological stations in the vicinity of the model area. For the period 1961 to 1989 the annual average potential evaporation is estimated at 570 mm with a maximum deviation of 15%.

Daily stream runoff data are taken from the Danish Land Development Service's two gauging stations 56.06 and 56.11 along the Tude å, see Fig. 1. At station 56.11 data only covers the period from 1977 to 1989. The catchment of the gauging stations are respectively 146 km^2 for 56.06 and 259 km^2 for 56.11 (including the catchment for 56.06).

In Tude å the low flow in the late summer is critical for the stream fauna. The median of the annual daily minimum runoff (AM50, in the following termed "minimum runoff") is approximately 80 l/s at 56.06 and 250 l/s at 56.11. In dry summers runoff sometimes falls to extremely low values (20 to 30 l/s at station 56.06) for a few weeks and then returns to "normal" minimum runoff levels again. The falls and rises are so abrupt, that they are almost certainly due to direct intake for irrigation purposes, though this can not be proven.

Sewage and waste water from the town of Slagelse is discharged to the Tude å between the two gauging stations. Approximately 120 l/s or 50% of the minimum runoff at station 56.11 is estimated to be waste water (Kemp & Lauritzen 1991).

Data for the annual groundwater withdrawal have been supplied by the water works of the area. The total withdrawal was almost constant during the period of investigation at a rate between 7.5 and 8 million m^3 per year.

Time series of monthly or bimonthly hydraulic head in the confined aquifer are available from 15 wells within the model area. Furthermore the hydraulic head in the confined aquifer was measured in 180 wells during november 1988 at a time when the change in groundwater storage was small. These measurements were used to draw a regional map of hydraulic head, Kemp & Lauritzen (1991).

Hydrological Model

The Tude å catchment lies west of the Suså catchment. The hydrogeologies of the two catchments are quite similar and therefore it is obvious to choose an integrated hydrological model of a type similar to that used in the earlier "Suså investigation" (see Dyhr-Nielsen (1981) for an overview and Refsgaard and Stang (1981) or Refsgaard and Hansen (1982) for details of the model). The hydrological model for the Tude å catchment outlined in Fig. 4 contains the three main components of the Suså model in modified (*i.e.* simplified) versions:

The First Component is an evapotranspiration model containing a simple degree-day snow model and an interception-root-zone model. The Suså model is a four layer model representing interception and root zone storages. Some of the model parameters are functions of a time-variable leaf area index. The Tude å model has only one layer representing both the interception and root zone storage and the

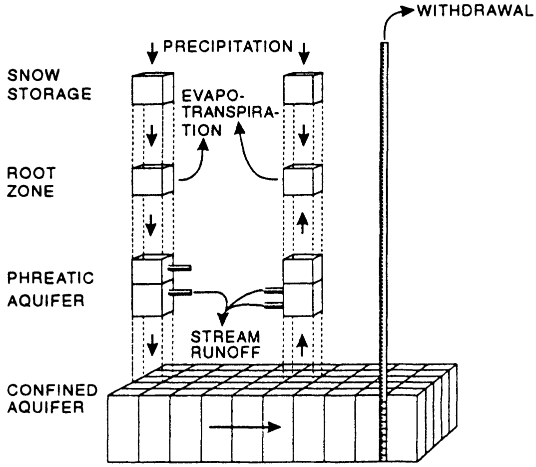

Fig. 4. Structure of the Tude å model.

only model parameters are the moisture capacity and a simple relation between moisture content and evapotranspiration.

The moisture capacity, which includes both the interception and root zone capacity, is assumed to be time-invariable, although it is known to be high from spring to harvest and low during the winter. The simplification is reasonable because during the winter the actual evapotranspiration is practically equal to the small potential evaporation, and this is simulated whether the model moisture capacity is reduced or not.

In the model the relation between moisture content and evapotranspiration stipulates, that if the moisture content exceeds 50% of the moisture capacity actual evapotranspiration equals potential evaporation. For lower moisture contents the actual evapotranspiration is reduced proportionately. Similar simple evapotranspiration models have been used with satisfying results in other hydrological studies, *e.g.* Tage Sørensen (1986).

The Second Component is the model for the phreatic aquifer in the till. Because of heterogeneity of this aquifer the flow model has to be strongly simplified. Therefore the horizontal flows in the phreatic aquifer towards the streams are modelled as a lumped flow from a linear reservoir through two outlets. The elevation of the phreatic surface relative to the outlets determines whether there is flow or not, and the specific yield gives the amount of water stored in the reservoir. Outflow from the upper outlet respond quickly and may be interpreted as flow through natural and man-made drains, whereas the more slowly responding outflow from the lower outlet may be interpreted as groundwater flow (including base flow) from near-surface reservoirs. Additional elements of the water balance for each of the reservoirs are the recharge from the root zone and the leakage through the aquitard between the phreatic aquifer and the underlying confined aquifer.

The water balance for each of the reservoirs is given by

$$S_y \frac{\partial h_p}{\partial t} = R - (Q_{dr} + Q_{gf} + L) \tag{1}$$

where S_y is the specific yield of the aquifer, h_p is the water level in the phreatic aquifer, t is the time, R is the recharge rate from the root zone (input from the evapotranspiration model), Q_{dr} is the flow rate through drains to the stream, Q_{gf} is the base flow or groundwater flow rate from the aquifer to the stream, and L is the leakage rate of groundwater between the phreatic aquifer and the underlying confined aquifer through the till aquitard.

Drainage and groundwater flow are calculated as

$$Q_{dr} = (h_p - h_{dr}) \frac{S_y}{K_{dr}} \tag{2}$$

$$Q_{gf} = (h_p - h_{gf}) \frac{S_y}{K_{gf}} \tag{3}$$

where h_{dr} and h_{gf} are reservoir outlet elevations while K_{dr} and K_{gf} are time constants for drainage and groundwater flow.

Leakage is calculated from

$$L = (h_p - h) \frac{p'}{m'} \tag{4}$$

where p' is the vertical hydraulic conductivity and m' the thickness of the aquitard.

The Third Component is a 2-dimensional horizontal flow model for the confined aquifer described by the equation (*e.g.* Konikow and Bredehoeft 1978; Trescott *et al.* 1976)

$$\frac{\partial}{\partial x}\left(T_{xx} \frac{\partial h}{\partial x}\right) + \frac{\partial}{\partial y}\left(T_{yy} \frac{\partial h}{\partial y}\right) + L - W = S\frac{\partial h}{\partial t} \tag{5}$$

where x and y are the horizontal Cartesian coordinates, T_{xx} and T_{yy} are the transmissivities in the x- and y-directions respectively, h is the hydraulic head in the confined aquifer, W is the withdrawal from the confined aquifer, S is the storage coefficient and t is the time. Strictly speaking Eq. (5) is only valid for flow in confined aquifers but it can also be used to approximate flow in phreatic aquifers when drawndowns are small compared with the saturated aquifer thickness.

Eqs. (1) and (5) are linked through Eq. (4). This is a simplification relative to the Suså model in the sense that the storage term of the aquitard is neglected. To smooth the flow between the phreatic and the confined aquifers it is thus excepted that the calibration of the present model will yield a higher specific yield than a similar model which includes an aquitard storage term.

The numerical scheme is, that the flow Eq. (5) is solved by a finite-difference approximation and the system of equations is solved by a modified version of the IADI-solver of Konikow and Bredehoeft (1978) which allows the grid dimensions to vary in space. The water balance Eq. (1) is solved sequentially by a finite difference approximation to give h_p for the next time step. The sequential scheme is accepted due to the short time step length at one day which is used in all the dynamic simulations.

Model Area and Boundary Conditions

To fulfil the objective (see Introduction) the model area includes the recharge area of the part of the aquifer exploited by the Water Supply of Slagelse as well as the catchment of Tude å. The regional map of hydraulic head in the confined aquifer is used to delimit the boundary so that a no-flux condition is fulfilled.

In most of the boundary areas till is the dominant Quarternary deposit, see Fig. 3, and the transmissivity of the confined aquifer is low. In these areas the no-flux condition will therefore be fulfilled even if the well fields are extended or the rate of withdrawal is increased in the central model area.

Along parts of the north-eastern and south-western boundaries the confined aquifer extends across the boundary, see Fig. 3. Despite the high transmissivity the potential map indicates that there is no significant boundary flux in these areas. Furthermore, because of the distances involved, a moderate change in groundwater withdrawal in the central model area will not provoke any major flux across these parts of the boundary.

The no-flux condition in the confined aquifer is also applied to the parts of the boundary which are at sea. It is assumed that within the model boundaries groundwater leaks from the confined aquifer to the sea.

For the phreatic aquifer model the water table may vary in all model elements except those situated seaward of the coastline. In such elements the water table is kept constant at sea level.

The 450 km^2 model area is divided into rectangular elements with a side length of 500 m. This distance is reduced to 250 m in riparian zones so as to define the stream locations more accurately.

Model Calibration

The hydrological model is calibrated to fit both hydraulic head measurements in the confined aquifer and the stream runoff. The calibrated parameters are transmissivity and storage coefficient for the confined aquifer, vertical hydraulic conductivity for the confining till layer, specific yield, outlet elevations and time con-

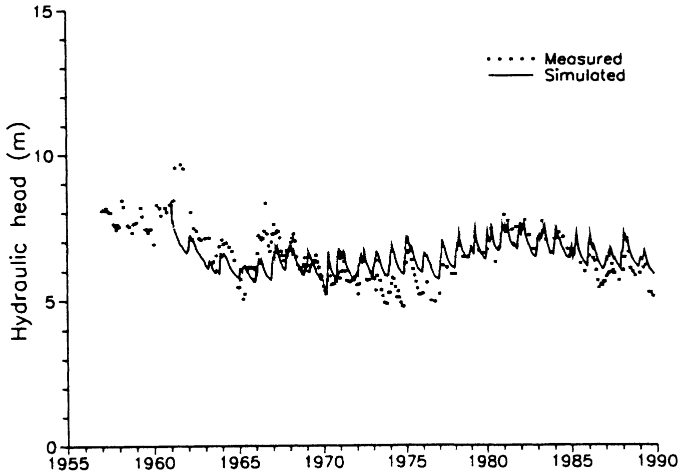

Fig. 5. Hydraulic head at well 210.061.

stants for the phreatic aquifer and moisture capacity for the evapotranspiration zone (including interception and root zone capacity). Each of these parameters can be distributed freely between the model elements, but the actual case uses parameter zonation. As described below the division into parameter zones is based on prior knowledge of the hydrogeology.

The moisture capacity of the evapotranspiration zone determines the amount of precipitation that infiltrates the groundwater zone and is therefore calibrated to give an overall realistic infiltration as shown by the observed stream runoff and the hydraulic heads.

The trial and error calibration of aquifer parameters compares simulated and observed time series of hydraulic head and runoff for the period from 1978 to 1989. Furthermore, the calculated hydraulic heads are compared with simultaneously measured heads in 160 wells within the model area from November 1988. The model is then tested by comparing simulation results with observations for the entire 29-year period of investigation. At this step transmissivities and storage coefficient in the urban area of Slagelse needed some recalibration. The main results of the calibration are summarized below. For a detailed discussion the reader is referred to Kemp & Lauritzen (1991).

The calibrated model gives a good overall simulation of the observed temporal and spatial variations in hydraulic head in the central parts of the confined aquifer, whereas the simulation results are only fair in the boundary areas and in the urban area of Slagelse. The agreement between observations and simulations is particularly good in areas where the aquifer is thick and homogeneous, *e.g.* along Tude å and Vårby å. As an example Fig. 5 shows the simulated and measured hydraulic head 1.5 km to the northwest of the Valbygård well field.

In the urban area of Slagelse there is a good agreement (better than 1 m deviation) between simulated and observed heads from 1973 and onward, whereas the

Table 1 – Observed and simulated runoff accumulation (1980-1988) at stations 56.06 and 56.11 (million m^3)

	Station 56.06	Station 56.11
Observed	323	636
Simulated	337	622
(Sim – Obs)	14	–14
(Sim – Obs)/Obs	4%	–2%

simulated head is approximately 5 m too low before 1973. A possible explanation of the discrepancy is that the model fails when the groundwater withdrawal in the urban area exceeds a certain rate. This is suggested by the fact that after 1973, when fifty per cent of the pumping was moved from the urban area, simulations and observations agree. The possible failure of the model at high pumping rates may be due to an inadequate description of the heterogenous geology in the urban area and also to ignoring the effects of drawdown on the transmissivity of this phreatic part of the aquifer.

Table 1 shows the observed and simulated runoff accumulation in Tude å for the period 1980-1988. At runoff gauging station 56.06 the simulated runoff accumulation exceeds the observed by approximately 4%. At gauging station 56.11 the simulated runoff accumulation is approximately 2% less than the observed runoff. Notice that in Table 1 the actual observed runoff accumulation at station 56.11 is reduced by 45 million m^3 to allow for the estimated contribution of sewage and waste water from Slagelse.

The model simulates the temporal variations in stream runoff quite well, Fig. 6. Simulations and observations agree during the summers but the model overestimates runoff peaks slightly during some of the winters.

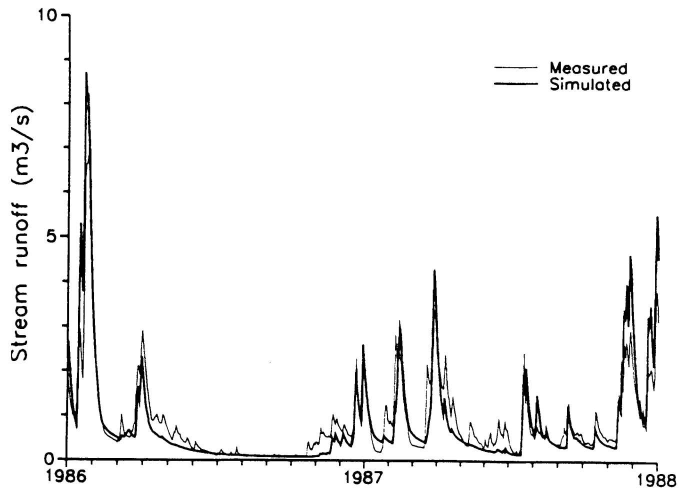

Fig. 6. Stream runoff at station 56.06.

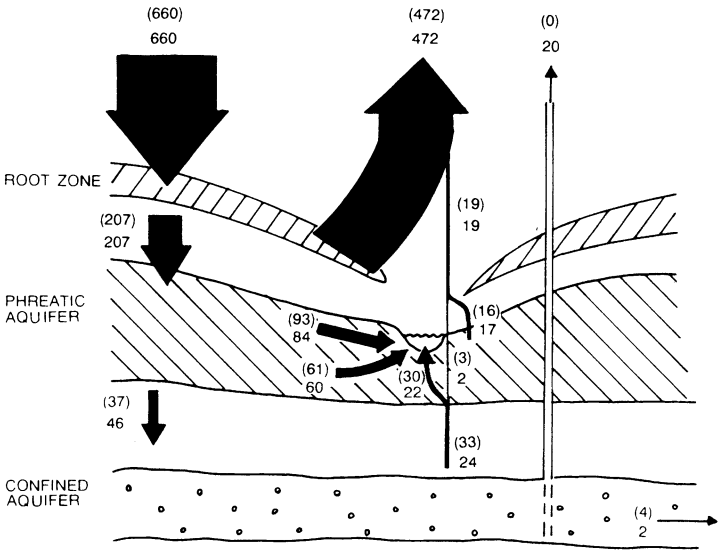

Fig. 7. Overall water balance in millimetres per year – including and in brackets excluding current withdrawals from the simulation.

The simulated overall water balance for the entire model area during the period of investigation is shown in Fig. 7. 72% of the precipitation evaporates or transpires, 22% is directed to the streams as drain flow or as groundwater flow through the phreatic aquifer, while 7% leaks to the confined aquifer in recharge areas. Just under half of this recharge is withdrawn from the aquifer for consumption and irrigation purposes while the remaining is either discharged to the phreatic aquifer or leaks to the sea. Most of the discharge from the confined aquifer to the phreatic aquifer is discharged to the streams as base flow while a minor part is evapotranspirated in the riparian zones. Base flow is approximately the same as the minimum runoff rate in the streams, *i.e.* in the late summer periods all runoff is groundwater discharged from the confined aquifer.

When the simulated components of the water balance are related to the precipitation the water balance of the Tude å catchment is comparable with that of the neighbouring Suså catchment, see Dyhr-Nielsen (1981).

The zonation of model parameters and the results of calibration are as follows:

For the evapotranspiration model the moisture capacities are distributed within zones taking into account the actual land use and soil type. Five zone types are used: cultivated clayey soil, cultivated sandy soil, forest areas, urban areas and riparian or meadow areas.

For cultivated areas the moisture capacity is calibrated at 170 mm for clayey soils and 135 mm for sandy soils. The resulting infiltration from clayey soil areas is approximately 80% of the infiltration from sandy soil areas.

For forest areas the moisture capacity is estimated at 500 mm which reduces infiltration to about 50% of the infiltration from cultivated areas.

For urban areas it is estimated that the diversion of precipitation from streets and roofs reduces the infiltration by 20% as compared to the neighbouring cultivated areas.

For riparian or meadow areas the moisture capacity is set at 0 mm and the actual evaporation is assumed to equal potential evaporation (except during frosty weather). This means that when precipitation exceeds potential evaporation the root zone discharges water to the phreatic aquifer, and conversely the root zone is recharged with water from the aquifer when potential evaporation exceeds precipitation. This immediate exchange of water is due to the coincidence of the root zone and the free water table in these areas.

The phreatic aquifer model uses spatially constant values of specific yield, time constants and depth from average ground level to the outlets. Specific yield is estimated at 0.05. For the upper outlet the estimated time constant is 10 days and the depth to the outlet is 3 m. The corresponding values for the lower outlet are 30 days and 3.5 m. The time constants are almost identical to those in the Suså model (generally 10 days and 33 days), whereas the water content between the outlets is 25 mm in the Tude å model and between 10 and 22.5 mm in the Suså model, Refsgaard (1981). As stated above this larger value may be due to the fact that in the present model the aquitard storage has to be simulated by increasing the specific yield of the phreatic aquifer.

For the confined aquifer model the transmissivity zonation and calibrated transmissivity values are shown in Fig. 8 and Table 2. The zonation is mainly based on a transmissivity field estimated by ordinary kriging (Kemp & Lauritzen 1991) using 225 experimental transmissivity values. These values consist of direct measurements from pumping tests in 48 wells and indirect estimations from the specific capacity in 177 wells using a linear regression between transmissivity and the specific capacities according to Ahmed and De Marsily (1987), Clifton and Neuman (1982) and others. The resulting uncertainty for the estimated transmissivity field is one to two orders of magnitude in the Tude å area and two to three orders of magnitude around the town of Slagelse and in the boundary areas. Therefore the kriged field is only used for establishing the initial transmissivity zones while the transmissivity values within the zones are calibrated. During calibration zonation was slightly modified.

The calibrated transmissivities are on the whole comparable to the transmissivity field estimated by kriging except for the north-west boundary area, *i.e.* zones 1 to 5, where experimental transmissivity values are sparse and the uncertainty is high. In this area till is the dominant deposit, fluvioglacial deposits are heterogeneous, and experimental values can therefore be expected to be purely local characteristics.

As indicated by Fig. 8 the calibrated model shows that the aquifer is very

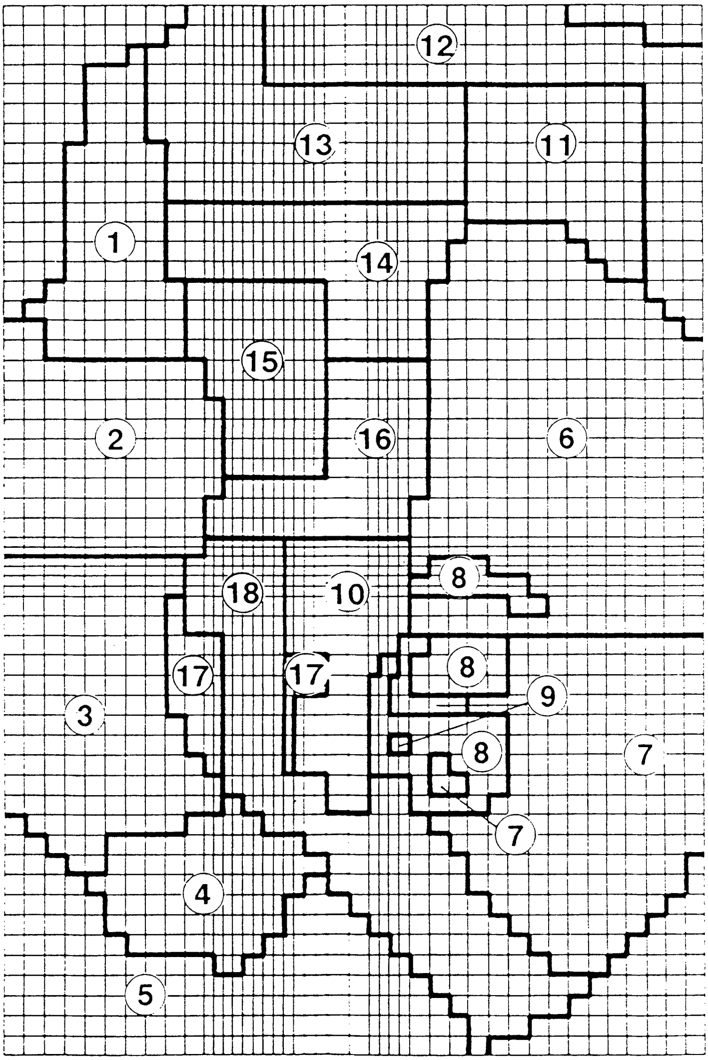

Fig. 8. Transmissivity zonation.

Table 2 – Zone transmissivities (m^2/s)

Zone	T	Zone	T	Zone	T
1	1.0×10^{-4}	7	6.4×10^{-4}	13	1.8×10^{-2}
2	2.2×10^{-4}	8	5.0×10^{-5}	14	1.7×10^{-3}
3	5.0×10^{-3}	9	1.0×10^{-3}	15	9.0×10^{-3}
4	1.0×10^{-2}	10	7.5×10^{-4}	16	1.2×10^{-3}
5	5.0×10^{-3}	11	3.5×10^{-3}	17	7.0×10^{-3}
6	1.0×10^{-3}	12	9.0×10^{-3}	18	1.0×10^{-2}

heterogeneous in the urban area of Slagelse, *i.e.* around zones 8 and 9, where the transmissivity is rather low and apparently varies with a factor of 20 within short distances. This is in agreement with the hydrogeological model and with the registered high sensitivity of hydraulic head to the distribution and rate of pumping in this area.

Sensitivity analysis shows that the calibrated transmissivity of zone 4 is uncertain and appears to be positively correlated with the leakage coefficients within the area and negatively correlated with the transmissivity of the neighbouring zones 3 and 5. This means that if the calibrated transmissivity in zone 4 is too high, the discharge to the phreatic aquifer and Tude å could be higher than estimated in this area, or there could be a high transmissivity zone in a neighbouring area.

The leakage coefficient between the phreatic and the confined aquifers is a function of vertical hydraulic conductivity and the thickness of the confining till. The thickness of the confining layer in each model element is estimated on the basis of stratigraphical information from wells (Kemp & Lauritzen 1991), whereas the hydraulic conductivity, which is assumed spatially constant within the entire model area, is estimated at 8×10^{-9} m/s by calibration. This means that the leakage coefficient varies between 10^{-9} and 10^{-10} s^{-1} which is in agreement with the results from 7 different pumping tests within the area, see Kemp & Lauritzen (1991). During the hydrological investigations of the neighbouring Suså catchment the vertical hydraulic conductivity of the till was estimated at 5×10^{-9} m/s, see Refsgaard and Stang (1981).

By calibration the storativity of the confined aquifer is estimated at 0.0005 which is the correct order of magnitude for this type of aquifer and in agreement with pumping tests within the area. Due to large drawdowns in the urban area of Slagelse the aquifer is locally phreatic with a calibrated specific yield of 0.1. This is in the lower range of what was expected and probably characterizes the heterogeneity of the aquifer, *i.e.* the mixing of fluvioglacial deposits with till, in the urban area.

Influence of Withdrawal

As an example of the application of this model the influence of groundwater withdrawal on the hydrological cycle can be demonstrated by comparing the calibration/test simulation (covering the years 1961 to 1989) with an alternative simulation in which all groundwater withdrawals are set to nil. The latter can be characterized as a simulation of a "semi-natural" hydrological situation unaffected by groundwater withdrawal but affected by man's use of land.

To simulate this the no-flux boundary condition must be fulfilled even if all groundwater withdrawals are excluded from the model. As stated above the transmissivity is low along most of the boundary and no groundwater is withdrawn in these areas. Therefore no significant changes in the directions of groundwater flow are expected along these parts of the boundary.

Along parts of the north-eastern and south-western boundaries the confined aquifer extends across the boundary. Dramatic changes in groundwater withdrawals close to these parts of the boundary can therefore provoke boundary fluxes. However, current withdrawals within these parts of the model area are balanced by withdrawals of comparable magnitude from the aquifer outside the model boundary. In the "semi-natural" hydrological situation withdrawals inside as well as outside the model area must be neglected and therefore no significant change of fluxes across these parts of the boundary are expected. It is therefore a fair assumption that the no-flux boundary condition will be fulfilled even if all groundwater withdrawals are neglected during the simulation.

The simulated overall water balances in Fig. 7 show, that the withdrawal of groundwater induces a 25% increase in the leakage of water from the phreatic to the confined aquifer (from 37 to 46 mm/year), the discharge to the phreatic aquifer in the riparian areas is reduced by 21% (from 33 to 24 mm/year), and the discharge to the sea is reduced by 50% (from 4 to 2 mm/year). The increase in leakage from the phreatic aquifer is balanced by a 10% reduction in drain flow to the streams (from 93 to 84 mm/year) while the decrease in discharge from the confined aquifer reduces the stream base flow by 27% (from 30 to 22 mm/year).

The withdrawal of groundwater lowers the hydraulic head of the confined aquifer relative to the water-table elevation in the phreatic aquifer. The area in which groundwater is leaking to the confined aquifer is thereby enlarged from 63% to 70% of the entire model area and the leakage rate per unit leakage area is increased by 12% (from 59 to 66 mm/year). Similarly, the area where groundwater is discharging from the confined aquifer is reduced from 37% to 30% of the entire model area and the discharge rate per unit discharge area is reduced by 10% (from 89 to 80 mm/year).

In Fig. 7 we distinguish between three contributions to stream runoff: drain flow, which is water flowing through the upper outlets of the phreatic aquifer model; groundwater flow in the phreatic aquifer, which is flowing through the model's lower outlets but is not base flow; and base flow, which is groundwater leaking from the confined aquifer to the riparian zones, thus generating the base flow of the streams. As shown in Fig. 7 the withdrawal of groundwater reduces both the base flow and the drain flow, whereas the influence on groundwater flow through the phreatic aquifer is insignificant.

The influence of groundwater withdrawal on stream runoff varies through the year. Fig. 9 shows, for a two year period, the simulated reduction of base flow, groundwater flow and drain flow due to the withdrawal (reductions are accumulated for the entire model area). The influence on base flow is practically constant in time, drain flow is influenced in the winter (when drain flow is present), whereas the influence on groundwater flow in the phreatic aquifer is small but varies

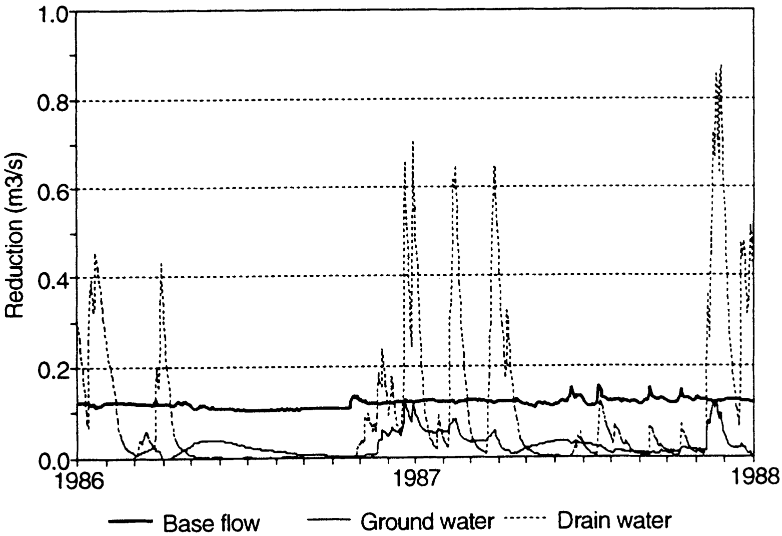

Fig. 9. Simulated stream flow reductions due to current groundwater withdrawals.

through the year. This shows that the groundwater withdrawal has the greatest influence on runoff in the winter season and the least influence in the summer season.

The influence on annual minimum runoff can be expressed by the influence ratio on minimum runoff, which Refsgaard and Stang (1981) define as

$$I_1 = \frac{\Delta Q_{min}}{\Delta Q_{av}} \tag{6}$$

where ΔQ_{min} is the reduction of the annual minimum runoff rate and ΔQ_{av} is the reduction of the average runoff rate through the entire simulation period (in this case the average reduction through the period 1961 to 1989). Because ΔQ_{av} is a constant, I_1 only tells us how the reduction of the annual minimum runoff rate, ΔQ_{min}, varies from year to year. Therefore two alternative influence ratios, I_2 and I_3, are proposed. I_2 relates the reduction in annual minimum runoff to the average reduction in runoff for that same year ($\Delta Q_{an.av}$)

$$I_2 = \frac{\Delta Q_{min}}{\Delta Q_{an.av}} \tag{7}$$

The inclusion of the variation in annual average runoff reduction makes I_2 more useful than I_1 as a measure of the variation in minimum runoff loss relative to average loss.

I_3 relates the reduction in annual minimum runoff to the minimum runoff (when no withdrawal of groundwater occurs) for that same year (Q_{min0})

$$I_3 = \frac{\Delta Q_{min}}{Q_{min0}} \tag{8}$$

I_3 is a measure of the relative reduction in annual minimum runoff due to the withdrawal of groundwater. This ratio, rather than I_1 and I_2, should be used during any discussion of the influence of groundwater withdrawal on minimum runoff in streams.

Table 3 lists statistics for the simulated influence ratios for different sub-catchments: the average value, an approximated 95% confidence interval (two standard deviations, 2 σ), and the correlation coefficient (ϱ_{min}) between the annual influence ratio and the annual minimum runoff. The table shows that the average values of the I_1 influence ratio varies between 0.39 and 0.48. This is higher than the influence ratio of 0.33 estimated for the Suså catchment, Dyhr-Nielsen (1981). A possible explanation of the rather high influence ratios for the Tude å catchment might be that the groundwater withdrawals are mostly situated close to the streams. The standard deviations of I_1 indicate that there is some variation in the influence of withdrawals on the reduction of minimum runoff through the years. The correlation coefficients indicate a moderate positive correlation between I_1 and the annual minimum runoff. This means that the largest absolute reductions in minimum runoff are most likely to occur in years where minimum runoff is high.

According to Table 3 the average of I_2 and I_1 are practically identical but the variation of I_2 is considerably larger than that of I_1. This shows that a larger fraction of the annual runoff loss can occur in the summer than is indicated by I_1. As stated above this is because the variance of I_1 is proportional to the variance of the reduction of annual minimum runoff, whereas the variance of I_2 is also a function of the variance of the reduction of annual average runoff. Apparently there is no correlation between I_2 ratios and the annual minimum runoff.

The influence ratio I_3 is a measure of the relative reduction in annual minimum runoff due to the withdrawal of groundwater. According to Table 3 the average reduction is 0.29 for large catchments while it is smaller for the subcatchments. The

Table 3 – Statistics for the simulated influence ratios on minimum runoff from different catchments for the period 1961 to 1989

	I_1			I_2			I_3		
	Av.	2 σ	ϱ_{min}	Av.	2 σ	ϱ_{min}	Av.	2 σ	ϱ_{min}
Ørslev	0.46	0.09	0.80	0.47	0.15	-.06	0.22	0.04	-.79
Valbygård	0.39	0.08	0.78	0.41	0.16	-.08	0.24	0.04	-.66
Trelleborg	0.44	0.08	0.65	0.46	0.17	-.13	0.29	0.06	-.71
Total area	0.48	0.08	0.63	0.50	0.17	-.14	0.29	0.06	-.74

average ratio is for instance 0.29 for the Trelleborg catchment whereas the ratio is only 0.22 for the Ørslev catchment, which is a sub-catchment in the upper part of the Trelleborg catchment. The influence ratio varies from year to year and is moderately negatively correlated with the annual minimum runoff. This means that the influence of withdrawals on minimum runoff is relatively larger in years when minimum runoff is low.

Conclusions

The hydrogeology of the 450 km^2 catchment of the stream Tude å is quite similar to the hydrogeology of the neighbouring Suså catchment and therefore the developed hydrological model contains the main components of the Suså model: an evapotranspiration model; a lumped model for the phreatic aquifer in the glacial till; and a traditional two-dimensional finite-difference groundwater model for the confined, fluvioglacial aquifer.

The evapotranspiration model contains a simple degree day snow model and a one-layered model for the interception and root zone storage. Taking into account the spatial variability of land use and soil type and using measurements of daily precipitation, potential evaporation and temperature the model calculates actual evapotranspiration and infiltration from the root zone.

Because of the inhomogeneity of the till the model for the phreatic aquifer must be greatly simplified when compared with the (unknown) natural system. Therefore the streamward horizontal flows in the aquifer are modelled by a lumped approach as flow from linear reservoirs (one for each model element) with two outlets. The phreatic aquifer model is coupled with the flow model of the confined aquifer through a linear leakage term, whereas the aquitard storage is neglected.

The model calibration uses trial and error to fit a 29-year long time series of hydraulic head and stream runoff. To reduce the calibration effort most of the calibrated parameters are assumed spatially constant, *i.e.*: the linear reservoir outlet parameters; the specific yield of the phreatic aquifer; the storativity of the confined aquifer; and the vertical conductivity of the aquitard. The transmissivity of the confined aquifer is assumed constant within 18 zones.

The average annual water balance simulated by the hydrological model for the period 1961 to 1989 shows that 71% of the precipitation is evapotranspirated, 22% is diverted to the streams through drainage systems or as groundwater flow through the phreatic aquifer while only 7% leaks to the confined aquifer. Just under half of the confined aquifer recharge is developed by water supplies or for irrigation purposes while the remainder is discharged to the riparian zones of the phreatic aquifer and thereby generates the base flow of the streams. In minimum runoff situations all stream runoff is due to the discharge of groundwater from the confined aquifer.

The influence of current groundwater withdrawals on the hydrological cycle is demonstrated by comparing simulations respectively including and excluding groundwater pumping. This shows that the withdrawal of groundwater induces a 25% increase in groundwater leakage to the confined aquifer which is mainly balanced by reduced drain flow to the streams. Further, the withdrawal reduces the discharge from the confined aquifer to the riparian zones and thereby induces a reduction of the stream base flow. For the entire catchment the base flow reduction is 30% in normal years, and up to 35% in dry years. On the other hand the reduction is considerably smaller for the upper stream catchments. For instance the reduction at Ørslev is approximately 20% in normal years.

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