Nordic Hydrology, 23, 1992, 73-88

Modelling Streamflow Recession in Two Danish Streams

Bente Clausen

Department of Earth Sciences,
Aarhus University, Denmark

The streamflow recession for two Danish streams is found not to follow the traditionally used single exponential expression $Q = Q_0 K^t$, where K is the recession constant. Three alternative equations, based on a simple hydrogeologic model, are applied. A good fit is obtained by use of the equation $Q = B + C K^t$, where K is a constant for each catchment and B and C are constants within each recession. Recessions with constant evapotranspiration are used to determine some of the hydraulic parameters for one of the two catchments.

Introduction

This paper is part of a larger baseflow study. Previous papers have dealt with synchronous streamflow measurements in the Danish River Alling (Clausen and Rasmussen 1988, Clausen 1989). The present paper deals with time series of streamflow recessions in the same river. Another Danish stream, the Tude, is used for comparison.

Generally, streamflow recessions in the two Danish streams are concave upwards on a semilogarithmic plot as shown in Fig. 1. It follows that these recessions are not well described by the widely used single exponential equation (*e.g.* Tschinkel 1963, Weisman 1977, Grip 1977, Gottschalk and Perzyna 1989, Demuth 1989, Tallaksen 1989, *etc.*)

$$Q = Q_0 \exp(-at) = Q_0 K^t \qquad (1)$$

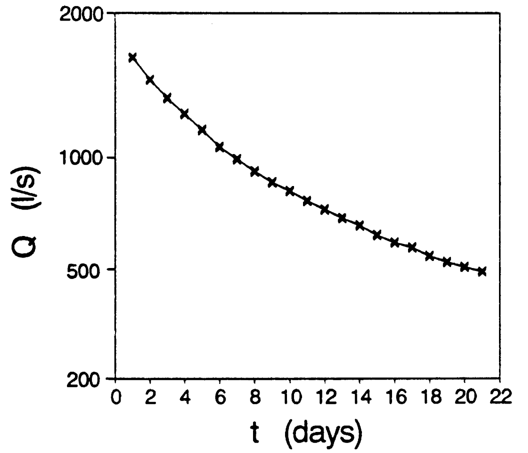

Fig. 1. A typical recession in the Tude observed in May 1983.

where Q is the flow t days after the start of the recession, Q_0 is the initial streamflow, and a and K are recession constants. Eq. (1) is a solution of the linearized basic partial differential equation of unsteady flow from an unconfined, isotropic and homogeneous strip aquifer with shallow stream incision, which is not influenced by evapotranspiration, leakage, irrigation, *etc.* (Singh 1968). In this case the recession constant a is proportional to $T S^{-1} m^{-2}$, where T is the transmissivity, S is the storage coefficient, and m is the distance from the stream to the divide.

Many other theoretical and empirical recession equations are given in the literature, for example

$$Q = B + C K^t \tag{2}$$

$$Q = C K^t + D M^t \tag{3}$$

$$Q = B + C K^t + D M^t \tag{4}$$

where K and M are recession constants ($K>M$) and B, C and D are constants within each recession period.

Eq. (2) was applied by Toebes and Strang (1964) and referred to as the 'Ice melt exponential' because the discharge asymptotically approaches the constant value B and therefore may typify baseflow recessions in areas with permanent snow and ice. This equation was also applied to recessions in a Polish river system by Radczuk and Szarska (1989), who interpreted B as a constant baseflow and K as exp($-v$ $1/A$ S). Here v is the infiltration velocity, 1 the total length of the stream and A the catchment area.

Eq. (3) was applied by James and Thompson (1970), who made a least-square estimation of the recessions of some Kentucky streams. They considered the two exponential terms as interflow and baseflow. Nutbrown (1975) showed theoretically that, even for the simplest aquifer structure, Q takes the form

$$Q = \sum_{i=1}^{\infty} B_i \, K_i^t \qquad (5)$$

where the individual flow components $B_i K_i$ for $i=1,2..$ dominate the recession curve at different times. Nutbrown and Downing (1976) pointed out that the recession curve depends on the initial distribution of the pressure head, and that Eq. (5) therefore depends simply on the dynamics of groundwater flow.

In this paper the recession Eqs. (2), (3) and (4), which are all special versions of Eq. (5), are considered as results of a simple lumped riparian groundwater model with two aquifers. The groundwater model presented in the first part of the paper is based on the geology of the two Danish catchments. In the second part the four recession Eqs. (1), (2), (3) and (4) are tested on data from the Danish streams by use of least-squares estimation in order to find the best equation. Finally, after testing the reliability of the groundwater model some of the physical parameters are estimated.

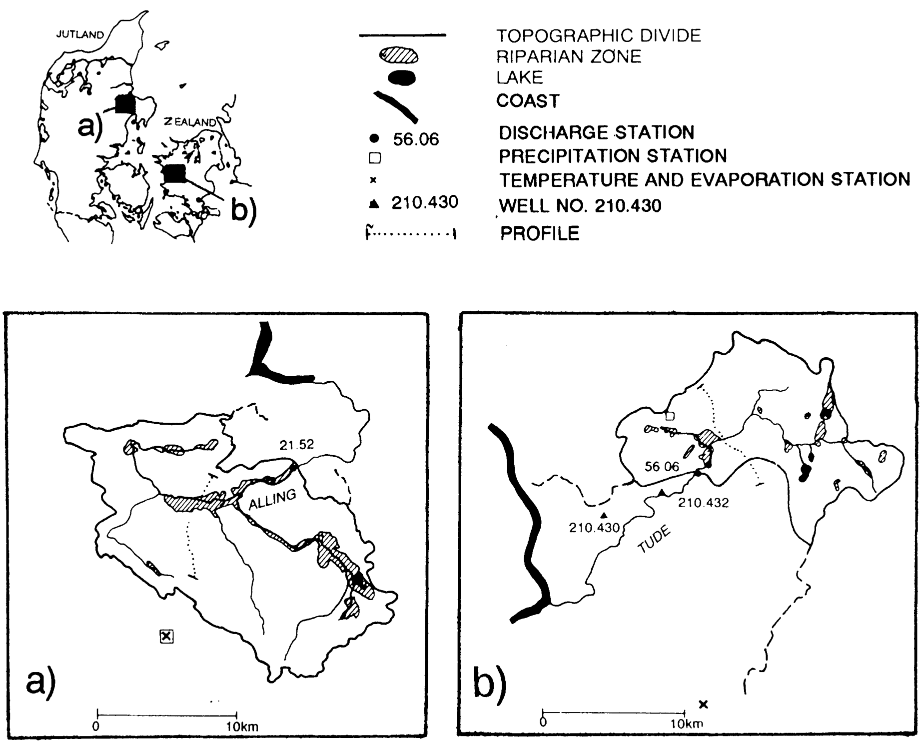

Fig. 2. The catchments of the two small rivers: a) the Alling in Jylland (Jutland) and b) the Tude in Sjælland (Zealand).

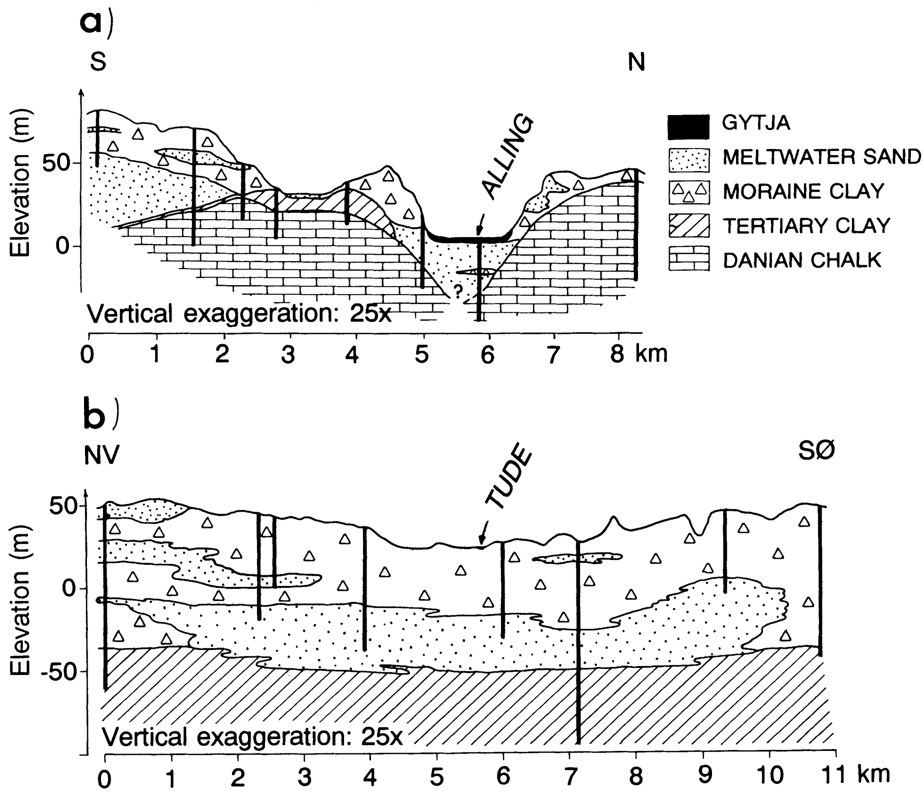

Fig. 3. Geological profiles of the two catchments. The locations can be seen in Fig. 2.

Field Areas

The locations of the two streams, the Alling and the Tude, are shown in Fig. 2. The elevation ranges from near sea level to 120 m in the Alling catchment and 63 m in the Tude catchment. The land is mostly cultivated; only minor areas are forested, and there is pasture in most of the low-lying riparian areas. The extent of the riparian zone is taken from a soil classification map (Danish Research Service for Plant and Soil Science 1978) and is about 5 % of the total area for both catchments (see Fig. 2). Lakes occupy only a minor part of the catchment area (0.1 % for the Alling and 0.5 % for the Tude).

The geology of the two catchments is typical of Eastern Denmark. A deep primary aquifer is overlain by Quaternary sand and clay. In the Alling catchment the lower aquifer consists of Danian chalk (see Fig. 3a). Above this in the peripheral area there is Tertiary clay with a very low permeability, while semipermeable moraine clays and permeable meltwater sands make up the upper layer. The valleys are mostly dominated by sand (see Fig. 3a), but in some areas moraine clay prevails. In the Tude catchment Quaternary sand makes up the primary aquifer (see Fig. 3b). Above this is a 30-50 m thick layer of moraine clay, which also occupies the valley. Small lenses of sand are found in the moraine clay.

Mean values of precipitation *P*, potential evapotranspiration *E*, and runoff *Q* for both catchments are presented in Table 1. The specific discharge is greater in the Alling (7.24 l/s km^2) than in the Tude (6.24 l/s km^2) because of the larger annual precipitation. The 2-year minimum flow Q_2 (see Table 1) is much higher in the Alling (2.20 l/s km^2) than in the Tude (0.50 l/s km^2) due to differences in climate and geology.

A Simple Lumped Riparian Groundwater Model

The groundwater model is a lumped, conceptual model applied to the riparian zone and based on the hydrogeologic system with two aquifers as illustrated in Fig. 4. The upper (sand) aquifer, which is either phreatic or confined, is separated from the lower (chalk or sand) confined aquifer by an aquitard (moraine clay). The concepts are simplified from the subsurface component in the distributed model (Hansen and Dyhr-Nielsen 1982) applied to the catchment of Suså.

The upper aquifer is assumed to behave as a linear reservoir with one outlet (the stream). The outlet level corresponds to the stream surface, and this is the reference level. Thus the water discharge *Q* through the outlet is proportional to the pressure head *h* of the upper aquifer. The input from the atmosphere is referred to as the recharge *R* (see later definition Eq. (16)), which is negative during recession periods. Leakage to or from the lower aquifer Q_1 is proportional to the difference in hydraulic head between the upper and the lower aquifer h-ϕ, while the change in the water content ΔW in the upper aquifer depends on the change in hydraulic

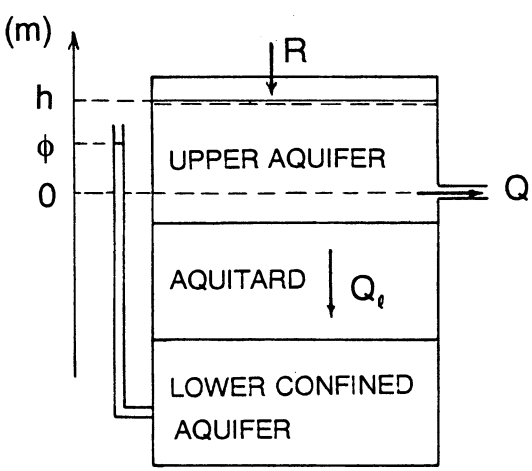

Fig. 4. The structure of the lumped groundwater model.

head dh/dt, as well as the storage coefficient S. If A is the area of the riparian zone, the equations become

$$Q \equiv A \frac{S}{\tau} h \equiv A\sigma h \tag{6}$$

$$Q_1 \equiv A \frac{K'}{B'} (h-\phi) \equiv A\lambda (h-\phi) \tag{7}$$

$$\Delta W \equiv AS \frac{dh}{dt} \tag{8}$$

where σ is a constant defined as S/τ, and τ is a time constant for the upper aquifer. The leakage coefficient λ is defined as K'/B', where K' is the hydraulic conductivity of the aquitard and B' its thickness. Applying the principles of continuity to the upper aquifer gives

$$R A \equiv Q + Q_1 + \Delta W \tag{9}$$

Substituting Eqs. (6), (7) and (8) into Eq. (9) gives

$$R \equiv (\sigma+\lambda) h - \lambda\phi + S \frac{dh}{dt} \tag{10}$$

In order to solve this differential equation we will assume that R is constant and ϕ decreases exponentially, $\phi = \phi_0 \exp(-\omega t)$. The equation becomes

$$\frac{dh}{dt} = -\alpha h + \beta + \gamma \exp(-\omega t) \tag{11}$$

where

$$\alpha = \frac{\sigma+\lambda}{S}, \quad \beta = \frac{R}{S}, \quad \gamma = \frac{\lambda\phi_0}{S}$$

The solution of Eq. (11) is

$$h \equiv c_1 + c_2 \exp(-\alpha t) + c_3 \exp(-\omega t) \tag{12}$$

where

$$c_1 \equiv \frac{\beta}{\alpha}, \quad c_2 = h_0 - \frac{\beta}{\alpha} + \frac{\gamma}{\omega-\alpha}, \quad c_3 = -\frac{\gamma}{\omega-\alpha}$$

Substituting Eq. (12) into Eq. (6) gives the solution for Q

$$Q \equiv d_1 + d_2 \exp(-\alpha t) + d_3 \exp(-\omega t) \tag{13}$$

where

$$d_1 = A\sigma c_1, \quad d_2 = A\sigma c_2, \quad d_3 = A\sigma c_3$$

This solution has the same form as Eq. (4).

If $R = 0$, then $d_1 = 0$, and the solution corresponds to Eq. (3). If ϕ is constant ($\omega = 0$) the solution becomes

$$Q = (d_1 + d_3) + d_2 \exp(-\alpha t) \qquad (14)$$

which has the same form as Eq. (2). Incidentally, it should be mentioned that if *R* in this last case varies linearly, the solution has the form

$$Q = e_1 + e_2 t + e_3 \exp(-\alpha t) \qquad (15)$$

where e_1, e_2 and e_3 are constants.

Selection of Recession Periods

The selection of recession periods is illustrated in Fig. 5, which shows three recessions in the Tude in 1987.

The selection of recession periods is based on the discharge curve and the recharge *R* to the groundwater in the riparian zone, defined as

$$R = P - E \qquad (16)$$

The definition is based on the assumption that the high groundwater level in the riparian zone implies that actual equals potential evapotranspiration *E*. The recharge thus becomes negative during dry periods. Periods with frost and snow were left out in order to relate recession to aquifer properties. The criteria for the

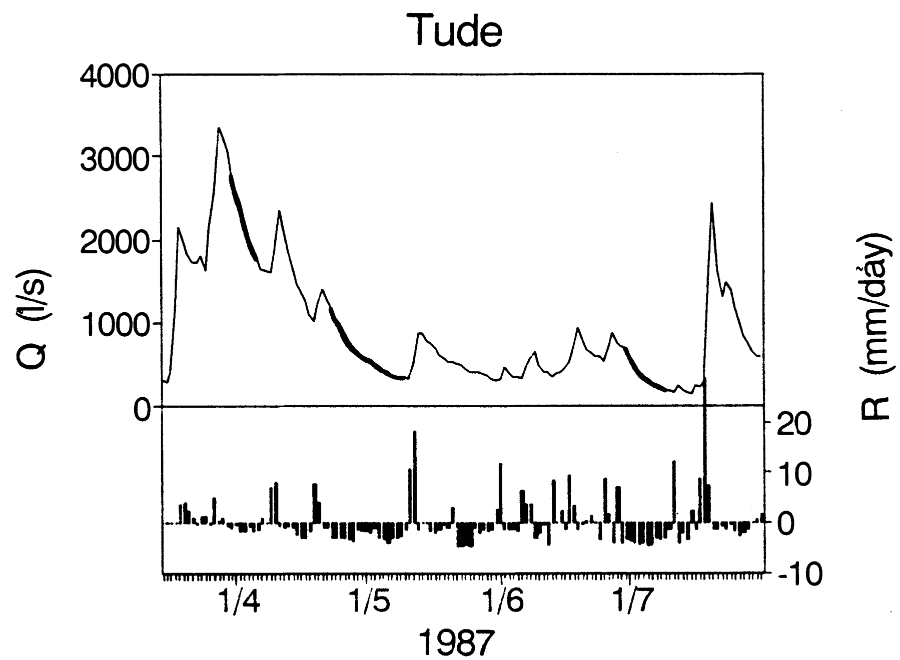

Fig. 5. Three examples of selected recessions in the Tude in 1987.

selection of recession periods are:

1) The recession period involves at least 8 consecutive days.
2) The discharge decreases during the recession period as well as the two preceding days.
3) The recharge R in the riparian zone is negative during the recession period as well as the two preceding days.

Based on these criteria, 63 recession periods were selected for the Alling and 87 periods for the Tude during the years shown in Table 1.

Table 1 – Means of precipitation $\bar{P}$, potential evapotranspiration $\bar{E}$, runoff $\bar{Q}$, and the 2-year minimum flow Q_2, for the Alling and the Tude catchments.

Discharge station	Area km^2	Period	$\bar{P}$ mm/y	$\bar{E}$ mm/y	$\bar{Q}$ l/s	Q_2 l/s
The Alling st. 21.52	242	1974-1990	713	565	1751	532
The Tude st. 56.06	146	1961-1989	682	572	916	73

Testing the Recession Equations

Eqs. (1), (2), (3) and (4) were fitted to the individual recession segments by minimizing the Weighted Sum of Squared Deviations

$$WSSD \equiv \sum_{1}^{L} \left(\frac{Q-Q'}{Q} \right)^2 \tag{17}$$

where Q is the observed discharge, Q' the calculated value and L the number of days in the recession period. The weighting was used to prevent large values of Q from dominating the results.

The goodness of fit of the four models was measured by the Weighted Mean Squared Deviation WMSD, which is the WSSD divided by the degrees of freedom. Fig. 6 shows the average WMSD for all recession periods of a given length L, while Fig. 7 shows the average WMSD for all recessions with $Q_0/\bar{Q}$ in a given interval. The bars at the bottom depict the numbers of recessions.

Figs. 6 and 7 show that WMSD is much higher for Eq. (1) than for Eqs. (2), (3) and (4). This is most pronounced in the Alling, where the difference is highest for recessions with a high Q_0. Moreover, it is seen that the WMSD is almost the same for Eqs. (2), (3) and (4). Only for recessions in the Alling with $Q_0/\bar{Q} > 2$ there is a small difference between the WMSD using Eq. (2) and the WMSD using Eq. (3) or Eq. (4). It is also worth noticing that with Eqs. (2), (3) and (4) the WMSD does not increase markedly with L. This means that the equations fit long recessions just as well as short recessions.

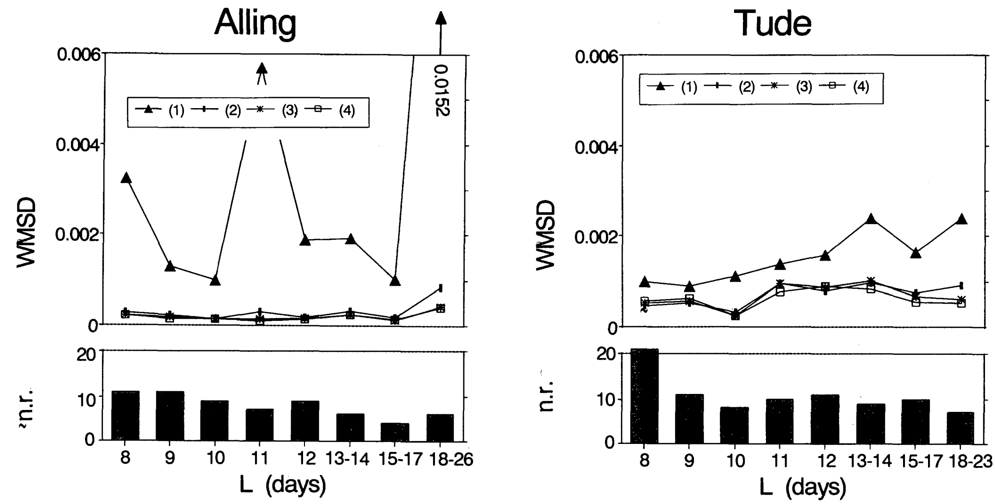

Fig. 6. The weighted mean squared deviation (WMSD) of the four recession equations for periods of a given length *L*. The bars at the bottom depict the numbers of recessions (n.r.).

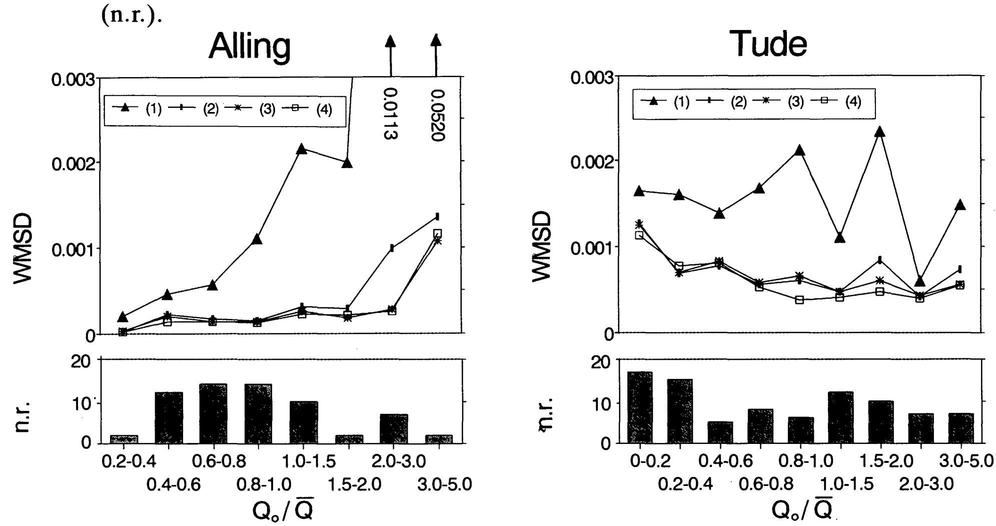

Fig. 7. The weighted mean squared deviation (WMDS) of the four recession equations for periods with a given initial discharge Q_0. The bars at the bottom depict the numbers of recessions (n.r.).

The conclusion is that the more advanced Eqs. (3) and (4) do not fit the recession better than Eq. (2) except for cases in the Alling where $Q_0/\bar{Q} > 2$. Assuming that our physical interpretation of Eqs. (2), (3) and (4) is correct, this means that the decrease of ϕ within a recession period is too small to be seen on the recession curve. It is therefore reasonable to consider ϕ as a constant within the limited time of a recession period.

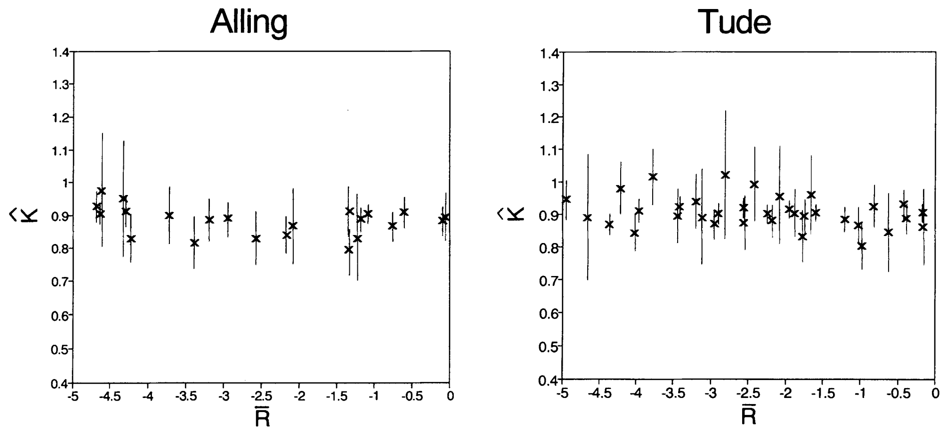

Fig. 8. Estimated values of K ($\hat{K}$) for recessions with an almost constant recharge (R). The 95 % confidence interval for the estimates of K is marked.

The Physical Interpretation

The reliability of the physical interpretation of Eq. (2) is considered by studying the variation of the estimates of K ($\hat{K}$). It must be remembered that the Eq. (2) was derived from the simple lumped groundwater model for constant R and ϕ, and that $K = \exp(-\alpha)$ depends solely on the hydraulic parameters in the upper reservoir and the hydraulic connection to the lower reservoir.

The values of $\hat{K}$ for recessions during which R is almost constant are shown in Fig. 8. Recessions with $Q_0/\bar{Q} > 2$ in the Alling are omitted. It is seen, that $\hat{K}$ varies independently of $\bar{R}$ in the range 0.80-0.97 in the Alling and 0.80-1.02 in the Tude, which means that $1/\alpha$ varies between 4.5 days and a high number of days (for values of K higher than 1 α becomes meaningless. These few cases are ascribed to random variation). The 95 % confidence interval of the estimates is also marked in Fig. 8 and is seen to be considerable. When estimating a joint value of K by minimizing the sum of WSSD for all periods applied in Fig. 8, we obtain $K = 0.90$ ($1/\alpha = 9.5$ days) for the Tude and $K = 0.87$ ($1/\alpha = 7.2$ days) for the Alling.

An F-test shows that for the periods shown in Fig. 8 the value of K can be taken to be constant. Applying Eq. (2) to the joint estimate of K (but still allowing B and C to vary) the sum of WSSD increases only by about 20 %.

The variation of $\hat{K}$ for the recessions omitted from the above has also been studied. Some of the results are:

1) In some cases R varies almost linearly during the recession. Fitting Eq. (15) to these recessions reduces the variability of $\hat{K}$ considerably.
2) For the recessions in the Alling in which $Q_0/\bar{Q} > 2$, the values of $\hat{K}$ are lower than the joint estimate of K (0.87). The reason is that these recessions are better

described by Eq. (3) and (4), mainly because of a very steep curve in the first part of the recession.

3) During some recessions the slope of the discharge curve changes very abruptly, and the $\hat{K}$'s for these recessions deviate considerably from the joint estimate. These changes cannot be explained by variations of R but might originate from the sudden discharge of waste water, the influence of pumping, backwater effects induced from the growing or cutting of weeds, or simply from errors in measuring and calculating the discharge. These recessions represent about 10 % of all recessions.

Estimating some Physical Parameters

In the previous section we found an estimate of

$$\alpha = \frac{\sigma+\lambda}{S} \tag{18}$$

In order to estimate the individual physical parameters we write out Eq. (14) in the form

$$Q = d_2 \exp(-\alpha t) + A\left(\frac{\sigma}{\sigma+\lambda}\right)R + A\lambda\left(\frac{\sigma}{\sigma+\lambda}\right)\phi \tag{19}$$

To estimate λ and σ we need observed values of R and ϕ. In the Tude catchment the Slagelse Waterworks has recorded the pressure head relative to sea level $\phi_{s.l.}$ in two wells (No. 210.430 and No. 210.432) 4 to 12 times per year since 1966. Well No. 210.430 is situated close to the stream (see Fig. 2) and filtered in the primary aquifer at a depth of about 20 m below sea level. Well No. 210.432 is situated two km from the stream (see Fig. 2) and filtered in a sand aquifer from 0 to 5 m below sea level. Because the graphs of observed water level in the two wells are very similar (and not influenced by pumping) it is assumed that both graphs represent the pressure head in the primary aquifer. Fig. 9 shows the interpolated values during periods with almost constant R.

To change the reference level from sea level to the elevation H of the outlet, we write

$$\phi \equiv \phi_{s.l.} - H \tag{20}$$

Because the groundwater model is a lumped model it is not possible to specify the value of H. Substituting Eq. (20) into Eq. (19) gives

$$Q = f_1 \exp(-\alpha t) + f_2 R + f_3 \phi_{s.l.} + f_4 \tag{21}$$

where

$$f_1 = d_2, \quad f_2 = A\left(\frac{\sigma}{\sigma+\lambda}\right), \quad f_3 = A\lambda\left(\frac{\sigma}{\sigma+\lambda}\right), \quad f_4 = -A\lambda\left(\frac{\sigma}{\sigma+\lambda}\right)H$$

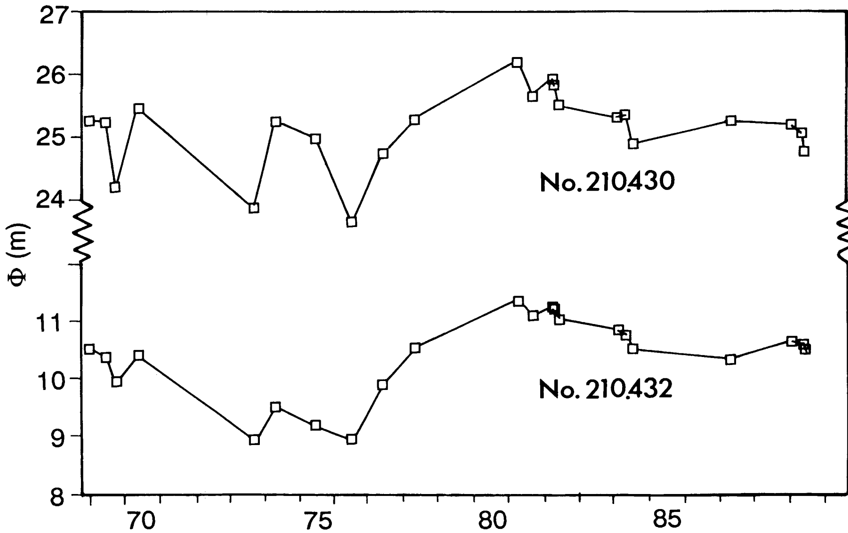

Fig. 9. Observed values of $\phi_{s.l.}$, the pressure head relative to sea level, in two wells in the Tude catchment. See the locations in Fig. 2.

On the basis of the observations shown in Fig. 9, λ and σ were estimated by fitting Eq. (21) to recessions with an almost constant *R*. Based on data from well No. 210.430 the following 95 % confidence intervals were found

$$\lambda \equiv 3.8\times10^{-8}\ s^{-1} \pm 1.9\times10^{-8}\ s^{-1}$$
$$\sigma = 7.8\times10^{-9}\ s^{-1} \pm 2.7\times10^{-9}\ s^{-1}$$

Data from well No. 210.432 give the estimates

$$\lambda = 2.5\times10^{-8}\ s^{-1} \pm 1.4\times10^{-8}\ s^{-1}$$
$$\sigma \equiv 6.4\times10^{-9}\ s^{-1} \pm 2.4\times10^{-9}\ s^{-1}$$

From these results and Eq. (18), *S* is estimated at 3.8 % from data for well No. 210.430 and 2.6 % from data for well No. 210.432.

It is not possible to estimate the physical parameters in the Alling catchment because there are no long observation series of ϕ. With the assumption that λ and *S* have approximately the same values as in the Tude catchment, we find from the estimated value of α that σ is appr. $2\times10^{-8}s^{-1}$. With the assumption that σ and *S* have the same value as in the Tude catchment, we get that λ is appr. $4.5\times10^{-8}s^{-1}$. Thus, either the upper aquifer in the Alling catchment reacts more quickly, or the leakage coefficient is higher than in the Tude catchment.

This method for estimating λ and σ relies on precise measurements of *P, E* and ϕ. The model area *A* has no influence on the estimation of λ, as λ is calculated as f_3/f_2 in Eq. (21). The estimate of σ depends on *A,* while the estimate of *S* is less sensible to changes in σ and thus in A. The definition of the model area, which is assumed to represent the discharge area, could be improved by including potential maps of the primary aquifer.

Kemp & Lauritzen (1991) set up a distributed model for the Tude catchment. This model has two outlets from the upper aquifer but is otherwise based on principles similar to the ones used in this paper. They found a value of 5 % for S and a general value for the whole catchment of the hydraulic conductivity of the leakage layer of 8×10^{-9} m/s. Thus, with thicknesses of 10 to 50 m the leakage coefficient λ varies between $10^{-9}\,s^{-s}$ and $10^{-10}\,s^{-1}$. However, when calibrating the model on observed streamflow it was necessary to adjust the leakage coefficient in the stream valley to about $2\text{-}3\times10^{-9}\,s^{-1}$ (Steen Christensen, pers. comm.), which is a factor 10 lower than the value found in this paper. The calibrated values from the distributed model were verified by using data from long pumping tests (Kemp & Lauritzen 1991).

Fredericia (1990) cites that most of the hydraulic conductivity values found from Danish field measurements lie between 10^{-8} m/s and 10^{-7} m/s. If the thickness of the moraine clay in the Tude valley is about 10 m, the distributed model gives a result at the lower end of this range, while the result from the recession model lies at the upper end.

When looking for the reasons for the difference between the two λ estimates it should be mentioned that the method presented here assumes no horizontal flow between the recharge area and the discharge area, while in the distributed model the lower outlets in the recharge area might contribute to streamflow even during dry summer periods. The conclusion is that if the physical parameters are to be estimated without using a fully distributed model it is of vital importance to include and estimate the horizontal flow in the simple model. The present method should therefore only be used in cases where it is reasonably certain that there is no horizontal flow between the recharge area and the discharge area.

Conclusions

For the two Danish streams it was found, that Eq. (2) ($Q=B+CK^t$) fits the recession much better than the single exponential Eq. (1) ($Q=Q_0K^t$). It was also found that within the period of a recession the more complicated Eqs. (3) and (4) do not fit the recession better than Eq. (2) apart from recessions in the Alling with a very high Q_0. This only means that the recessions are too short to show the decrease of the slow-reacting component.

Especially for prediction over long-time periods it seems important to use Eq. (2) instead of Eq. (1). The price for obtaining this more reliable prediction is that the value of the extra parameter B must somehow be found.

The physical interpretation of the recession Eqs. (2), (3), and (4) was based on the geology with two aquifers. In accordance with the groundwater model the estimated values of K were found to be the same for periods during which the evapotranspiration was almost constant. Based on this constant value of K, and observed values of ϕ, it was possible to estimate the physical parameters λ, σ and S for the Tude valley. The estimate of λ was a factor 10 higher than the value found from a distributed model and long pumping tests. This could be due to the neglect of horizontal flow in the lumped model. However, both values lie within the range mentioned by Fredericia (1990).

Acknowledgements

Daily mean values of discharge were supplied by the Danish Land Development Services, who are in charge of the database. They also maintain the gauging station in the Tude, whereas the gauging station in the Alling is maintained by the County of Aarhus. Daily values of precipitation, temperature and potential evapotranspiration were supplied by the Danish Meteorological Institute, and the Department of Agrometeorology of the Danish Research Service for Plant and Soil Science. The county of Vestsjælland supplied the well data.

I am grateful to Jens Ledet Jensen, Keld Rømer Rasmussen, Steen Christensen and Jørgen Lundager Jensen, who have contributed to the work through inspiring discussions and many ideas. I also want to thank the two anonymous referees for useful comments. Finally, I would like to thank C. Aub-Robinson and Allan Gustard for useful discussions and for improving the English text, and Ruth Nielsen for making the drawings.

References

Clausen, B., and Rasmussen, K. R. (1988) Fysisk hydrologi. Medianminimum, in Danish (Physical hydrology. Median minimum). Geokompendier 28, Department of Earth Sciences, Aarhus University, 60 pp.

Clausen, B. (1989) Alling å-systemet. Grundvandsdannelse samt udveksling mellem grundvand og vandløb, in Danish (The River Alling. Recharge and the interaction between groundwater and streamflow). M. Sc. thesis, Department of Earth Sciences, Aarhus University, 122 pp.

Danish Research Service for Plant and Soil Science (1978) Soil classification map series 1:50.000, Department of Land Data and Mapping.

Demuth, S. (1989) The application of the West German IHP recommendations for the analysis of data from small research basins, IAHS Publication No. 187, pp. 47-60.

Fredericia, J. (1990) Saturated hydraulic conductivity of clayey tills and the role of fractures, *Nordic Hydrology, Vol. 21(2),* pp. 119-132.

Gottschalk, L., and Perzyna, G. (1989) A physically based distribution function for low flow, *Hydrol. Sci. J., Vol. 34(5),* pp. 559-573.

Grip, H. (1977) Recession – en tidsinvariant process? in Norwegian (Recession – a time-invariant process?), *Vannet i Norden, Nr. 10(2),* pp. 50-53.

Hansen, E., and Dyhr-Nielsen, M. (1982) The hydrological Susaa-study, *Nordic Hydrology, Vol. 13(5),* pp. 257-262.

James, L. D., and Thompson, W. O. (1970) Least squares estimation of constants in a linear recession model, *Water Resour. Res., Vol. 6(4),* pp. 1062-1069.

Kemp & Lauritzen (1991) Grundvandsundersøgelser i Tude Å-oplandet. Hydrologiske modeller, in Danish (Groundwater investigations in the Tude catchment. Hydrological models). Report to Slagelse Waterworks.

Nutbrown, D. A. (1975) Normal mode analysis of the linear equation of groundwater flow, *Water Resour. Res., Vol. 11(6),* pp. 979-987.

Nutbrown, D. A., and Downing, R. A. (1976) Normal-mode analysis of the structure of baseflow-recession curves, *J. Hydrol., Vol. 30,* pp. 327-340.

Radczuk, L., and Szarska, O. (1989) Use of the flow recession curve for the estimation of conditions of river supply by underground water, IAHS Publ. No. 187, pp. 67-74.

Singh, K. P. (1968) Some factors affecting baseflow, *Water Resour. Res., Vol. 4(5),* pp. 985-999.

Tallaksen, L. (1989) Analysis of time variability in recessions, IAHS Publ. No. 187, pp. 85-96.

Toebes, C., and Strang, D. D. (1964) On recession curves, *J. Hydrol., Vol. 3*, pp. 2-15.

Tschinkel, H. M. (1963) Short-term fluctuation in streamflow as related to evaporation and transpiration, *J. Geophys. Res., Vol. 68(24),* pp. 6459-6469.

Weisman, R. N. (1977) The effect of evapotranspiration on streamflow recession, *Hydrol. Sci. Bull, Vol. 22(3)*, pp. 371-377.

First received: 5 April, 1991
Revised version received: 19 September, 1991
Accepted: 18 February, 1992

Address:
Department of Earth Sciences,
Aarhus University,
Ny Munkegade, Bldg. 520,
DK-8000 Aarhus C,
Denmark