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Scale dependent hydraulic variability of a stream bed on an outwash plain

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Abstract The stream bed leakage coefficient of a small alluvial stream was estimated using three different methods based on data representing different length scales: (a) small-scale observations of seepage and hydraulic head (~1 m); (b) mixed-scale observations (~1 m hydraulic head combined with ~500 m streamflow gain); and (c) large scale-observations of hydraulic head and streamflow gain (~500 m). Thus, the mixed-scale and large-scale methods estimate the efficient leakage coefficient along a reach. The estimated $\log_{10}$ leakage coefficients were similar for all three estimation methods. The mixed-scale method is less time consuming then the other methods. However, when applying the mixed-scale or the large-scale method in large streams the stream reach has to be long in order to measure the streamflow gain with sufficient accuracy. The large-scale method should be preferred if the stream-bed leakage coefficient is to be used in a basin scale groundwater model.

Key words leakage coefficient; mini-piezometer; seepage meter; stream bed

INTRODUCTION

In groundwater modelling it is essential to specify the proper boundary conditions. When studying, for instance, the interaction between a stream and an underlying aquifer, the leakage coefficient of the stream bed (λ) must be estimated in order to calculate the flux between the stream and the underlying aquifer. In the present study λ was estimated from data representing three different length scales. First, data from seepage meters and mini-piezometers installed at locations regularly placed along a 480 m stream reach were used to obtain a small-scale value (~1 m). Second, data from the mini-piezometers were combined with measurements of the groundwater seeping to the stream reach. These "mixed-scale" data represent local observations of the head below the stream bed combined with large-scale observations of the seepage. Third, a large-scale (~100 m) value of λ was obtained from calibration of a stationary groundwater model. The calibration was based on regional observations of the hydraulic head and groundwater seepage for five sub-sections of the stream (Nyholm & Christensen, 2000). The purpose of using the different methods is both to compare the methods and to quantify the value and the variance of λ at different scales. This is important for evaluating how to obtain useful data needed for the design of a model of groundwater/stream interaction and for the estimation of parameter values that are decisive for the interaction.

FIELD SITE

The field site is situated on an alluvial plain a short distance in front of the main Weichselian ice border through Denmark. The deposits between the surface and 30 m depth are primarily sand and gravel. Below that, more than 60 m of predominantly semi- or low permeable beds of clay and silt (Nyholm, 2000) is found. The Haller stream becomes perennial about 2 km west of the ice border where it runs in a small valley which is incised in the alluvial plain (Fig. 1). The slope of the stream is about 0.4% (Fig. 2) and the contribution from baseflow is about 40 l s^{-1} km^{-1} (Nyholm & Rasmussen, 2000). Originally, the stream meandered across the entire valley, but in the 1940s, an artificial course was cut through the peat covering the valley floor. The stream bed is composed of predominantly sand, gravel, pebbles, and cobbles, but some areas are dominated by fine organic material.

Nyholm (2000) found several layers of supposedly low permeability at various depths within approximately 1.5 m below the stream bed.

During summer the streamflow is dominated by groundwater flowing to the stream from the southeast while overland flow and interflow is restricted to short periods after intense precipitation (Nyholm & Rasmussen, 2000). Therefore, during a dry summer spell two sets of synchronous streamflow measurements (Fig. 2) were recorded between Q2 and Q2a (Fig. 1) in order to investigate the groundwater seepage along the reach. Since the differences between the two sets of streamflow are also small (Fig. 2), the data indicate that groundwater seepage is almost uniform along the reach. There is no indication in the geology that any part of the reach is losing between Q2a and Q2, so the apparent downstream decrease in discharge between 275 m and 325 m is probably due to random errors. About 90% of the groundwater seepage enters the stream either through the banks of the channel or through the bed. The remaining seepage flows to swamps, peat-hags and small depressions in the riparian zone (Fig. 1) from where it flows superficially to the stream. This part of the flow also appears to be uniformly distributed along the reach.

SMALL-SCALE MEASUREMENTS

The small-scale variation of the stream bed hydraulic conductivity was investigated by Lee (1977) and Lee & Cherry (1978) using a combination of seepage meter and piezometer data. In their set-up, the water seeping into the seepage meter is collected in a folio bag flowing in the stream and connected to the seepage meter outlet. However, experiments by for example Lee & Cherry (1978) and Cey *et al.* (1998) show that the use of folio bags may produce uncertain results. We duplicated the set-up of Lee & Cherry (1978) and observations showed a large scatter. Thus, in some cases zero flux values were observed. Therefore, a modified set-up was tested. Here, a silicon tube connected the seepage meter outlet to a small container positioned at the bank of the stream (Fig. 3). The tube was 1.15 m long with 7 mm inner diameter and negligible head loss. The container inlet was positioned at the stream stage level and the system was allowed to equilibrate before the final measurements were made. The seepage rates obtained from the modified set-up (Fig. 4) at the six locations along the reach were significantly larger than the rates obtained from the classical set-up with

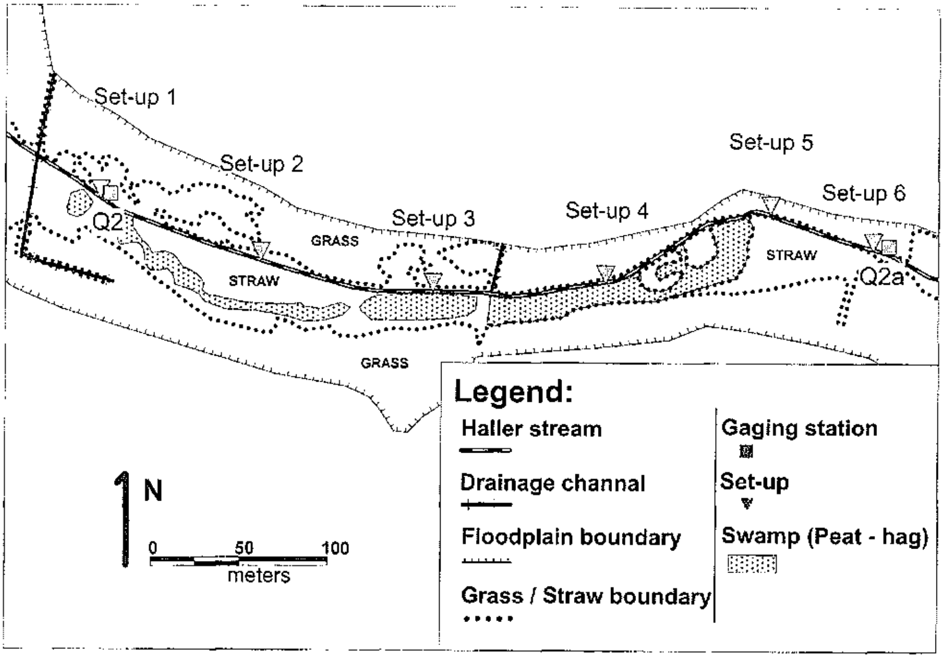

Fig. 1 The field site at the Haller stream between discharge stations Q2 and Q2a.

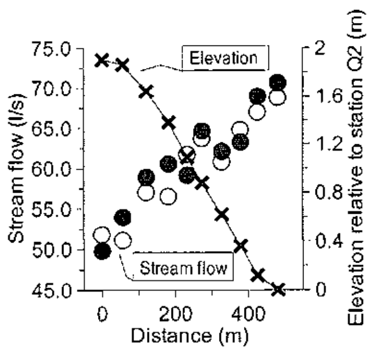

Fig. 2 Streamflow and relative elevation between the discharge stations.

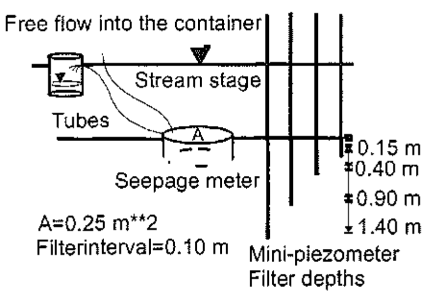

Fig. 3 Mini-piezometer and modified seepage meter set-up used in the study.

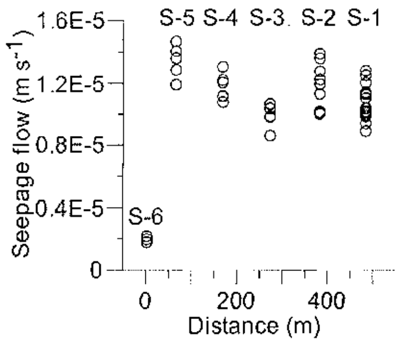

Fig. 4 Seepage rates measured at set-up 1 to set-up 6.

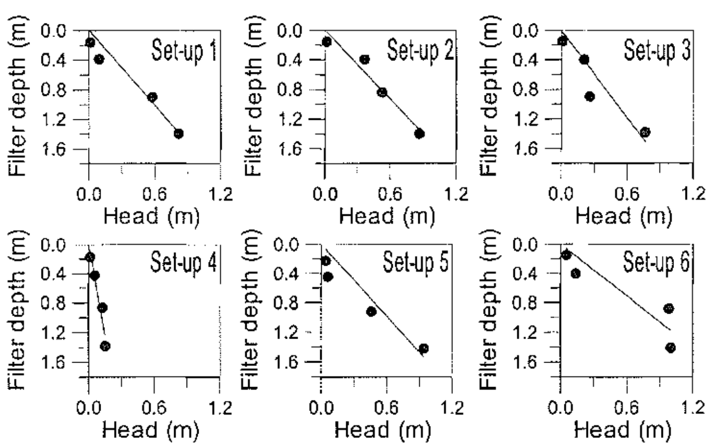

Fig. 5 The hydraulic head below the stream bed at set-up 1 to set-up 6.

folio bags, and the scatter was smaller. Therefore, the modified seepage meter set-up was used to obtain the seepage values presented below.

Mini-piezometers were installed at four depths at each seepage meter (Fig. 3). The seepage meters and the piezometers were left undisturbed for at least a day before the simultaneous seepage and hydraulic head measurements were made. Replicate rates of seepage (Q) obtained at the six locations are presented in Fig. 4. The largest rate, which was obtained at S-5, was 8–9 times larger than the smallest rate obtained at S-6. Otherwise, the seepage rates measured at S-1 to S-5 differ by less than approximately 35%. Figure 5 shows the measured heads at various depths for S1-S6.

The hydraulic head (h) was recorded at depth z below the stream bed. Using Darcy's law the $\log_{10}$-hydraulic conductivity of the stream bed ($\log(K)$) can be estimated from:

$$K = \frac{Q}{A}\frac{z}{h} = q\beta^{-1} \Rightarrow \log(K) = \log(q) - \log(\beta) \tag{1}$$

where A is the area of the seepage meter (Fig. 3). An estimate of the average $\log_{10}$-seepage rate ($\log(q)$) and its variance can be computed from the repeated seepage meter measurements. Likewise an estimate of the average $\log_{10}$-vertical hydraulic

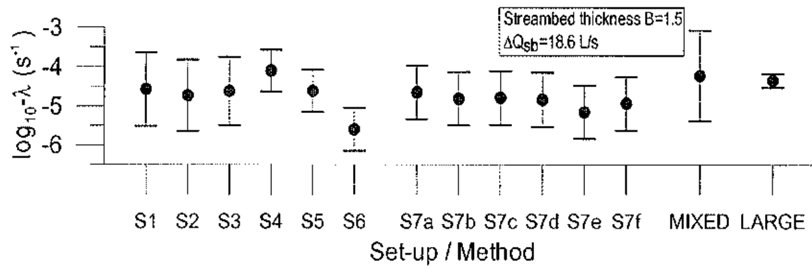

Fig. 6 The $\log_{10}$-values of the leakage coefficient (λ) calculated from measurements representing three different length scales.

gradient (log(β)) and its variance can be computed from the head observations in the four piezometers. The estimate of log(K) is computed from equation (1) and its variance is computed by summing the variances of the estimated log(q) and log(β). (It is reasonable to assume that the measurement errors on log(q) are independent of the measurement errors on log(β).) The $\log_{10}$-value of the leakage coefficient (log(λ)) for the 1.5 m deep layer where piezometers readings were taken is calculated from:

$$\lambda = K/B \Leftrightarrow \log(\lambda) = \log(K) - \log(B) \quad (2)$$

Figure 6 shows that the estimated values of log(λ) at the six positions vary from –4.1 to –5.6 (λ has the dimension s^{-1} which is used in the following). This corresponds to a 1.5 order of magnitude variation of the leakage coefficient. The 95% confidence intervals for log(λ) in Fig. 6 indicate that the log(λ) values at S-1 to S-5 may not differ significantly from each other, whereas the estimated value at S-6 is significantly lower than most of the other values.

To study the variation of the seepage and leakage coefficient on a much smaller spatial scale we installed six seepage meters in a cluster 15 m downstream of position S-6 (Fig. 1). The cluster is outlined in Fig. 7(a). The measured seepage rates vary from 2.6×10^{-6}–9.4×10^{-6} m s^{-1} (Fig. 7(b)). The piezometric level relative to the local surface of the stream was measured along a profile upstream of the seepage meter cluster (Fig. 7(c)). Iso-potential lines were drawn under the assumption of isotropy and homogeneity. They indicate that there might be a horizontal flow component, but this is still under investigation. The vertical hydraulic conductivity at each seepage meter was estimated using an average hydraulic gradient based on the three sets of piezometers placed in the stream. The estimated log(λ)-values vary from –4.65 to –5.15 s^{-1} (Fig. 6) corresponding to 0.5 order of magnitude of the leakage coefficient. The 95% confidence intervals of the log(λ)-values indicate that the leakage coefficients may not differ significantly from each other at set-up 7, or from the values obtained at set-ups 1–6.

The variance of the small-scale estimates of log(λ) is dominated by the variance of the hydraulic gradient.

MIXED-SCALE MEASUREMENTS

The total amount of groundwater seepage reaching the stream between Q2 and Q2a is calculated as the difference between synchronous streamflow measurements at the two stations during rainless periods. Previous analysis shows that at Q2 and Q2a the uncertainty

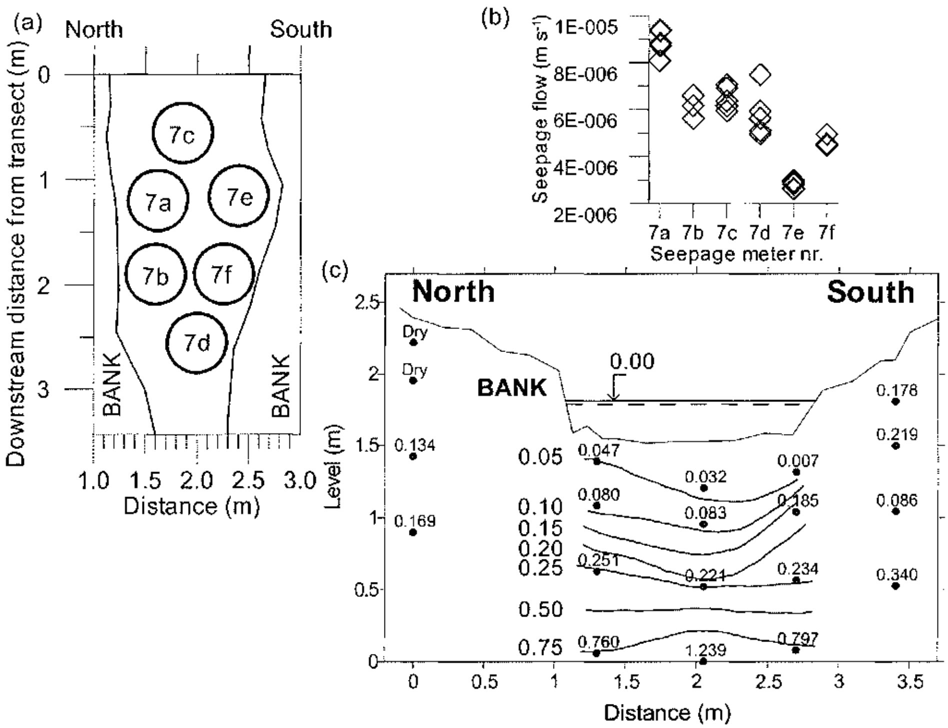

Fig. 7 (a) Position of the mini-piezometer transect and cluster of seepage meters (7a–f) immediately downstream of Q2a. (b) The hydraulic head at the transect. (c) Seepage rates.

of the discharge is 1–2 l s^{-1} during low flow (Nyholm & Rasmussen, 2000). The seepage reaching the stream superficially via the riparian zone is subtracted from the total seepage in order to find the component that flows through the bed (ΔQ_{sb}). The area of the stream bed (A_{sb}) was estimated from field observations of its length (480 m) and width (1.5 m). The average hydraulic gradient below the stream between S2 and S2a was calculated from the average of the local gradients at the piezometer locations. (The data are the same as used in the study of the small-scale variation.) Finally the $\log_{10}$-leakage coefficient was estimated from equations (1) and (2) using $\log(\Delta Q_{sb}/A_{sb})$ instead of $\log(q)$. The $\log_{10}$-value of the leakage coefficient was estimated to $\log(\lambda) = -4.22 \pm 1.15$ (Fig. 6). The variance is dominated by the uncertainty on the value of the hydraulic gradient.

LARGE-SCALE MEASUREMENTS

Hydraulic head measurements within the entire drainage basin and measurements of baseflow at the permanent discharge stations were used to calibrate the parameters of a steady-state numerical groundwater model (Nyholm & Christensen, 2000). The parameters included the horizontal and vertical conductivities of the aquifer and the stream bed leakage. Estimation was done using nonlinear regression, which produced a nice fit to the hydraulic head and baseflow data. The $\log_{10}$-leakage coefficient ($\log(\lambda)$) was estimated as $\log(\lambda) = -4.34 \pm 0.17$ (Fig. 6).

CONCLUSIONS

Both streamflow and the cumulative seepage increase approximately linearly with distance along the stream. However, when total seepage through the stream bed is estimated from the seepage meter measurements it amounts to only 38% of ΔQ_{sb}. There may be several reasons for the discrepancy: (a) there may be unknown bias in the seepage meter measurements, (b) seepage may be unevenly distributed with areas of high rates where measurements are lacking, (c) seepage may reach the stream through a larger area than the visible stream bed so that flow through the banks of the stream is important. However, the modified seepage meter method used here does produce results of the same order of magnitude as the streamflow measurements.

The estimated values of $\log(\lambda)$ are similar for all the three estimation methods. However, the small-scale estimate of $\log(\lambda)$ at S6 deviates significantly from all other estimates. The uncertainty of a small-scale or mixed-scale estimate is significant which is mainly due to the uncertainty of the measured hydraulic gradient beneath the stream. The uncertainty of the large-scale estimate obtained by calibration of a groundwater model is significantly smaller. However, this estimate was obtained from an entirely different dataset covering the entire drainage basin but with less data on the hydraulic gradient beneath the stream.

The small-scale method can be used to measure small-scale groundwater seepage and stream bed leakage coefficient in the studied stream. A number of such measurements along a reach can be integrated to estimate larger scale seepage and efficient leakage coefficients for the reach. However, the method is more time consuming than the mixed-scale method. The small-scale method is being tested in other larger streams.

The mixed-scale method is the fastest method of obtaining an estimate of the efficient leakage coefficient for a reach. However, in order to use the method in large streams the reach may have to be long in order to measure the streamflow gain with sufficient accuracy.

The large-scale method can be used to estimate (among other parameters) the efficient stream bed leakage coefficient along reaches by using hydraulic head and streamflow gain data from the entire (sub) drainage basin. This method should be preferred if the stream bed leakage coefficient is to be used as input to a basin-scale groundwater model. However, as for the mixed-scale method, the reaches for which the efficient leakage coefficient can be estimated accurately may have to be long in large streams.

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