TRACER TEST IN HOBØL CREEK, NORWAY,
UNDER DIFFERENT FLOW CONDITIONS

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Abstract

The current paper describes a repetition of a tracer test using tritiated water in Hobøl Creek, Norway, under two markedly different flow conditions in August and in October 2002. By carefully fitting a one-dimensional model of the solute transport in the creek, various mixing parameters could be evaluated and compared between the two occasions. The change in residence time in and exchange rate with the hyporheic zone caused by the difference in flow could be well accounted for in terms of the “pumping” theory. The pumping theory represents the hydro-mechanics of the exchange flow. Further, we found that at least 70% mass recovery is needed to provide an accurate evaluation of model parameters and this can only be achieved up to a limited stream distance.

Key words: tracer test, hyporheic zone, solute contaminants.

1. INTRODUCTION

Transport of solute contaminants in streams is affected markedly by exchange with the hyporheic zone and retention of, particularly, sorbing solutes (Jackman et al., 1984; Jonsson et al., 2004) or colloids (Packman et al., 2000). Mathematical models representing both in-stream processes and the hyporheic exchange are often evaluated versus tracer injections with inert and reactive solutes and the model parameters are assessed by interpretation of observed breakthrough curves at different stations downstream of the section of injection.
This paper aims to describe a tracer test performed in a small stream under different flow conditions. The two experiments aimed to shed light on the problem of generalisation of exchange relationships for the hyporheic zone to other conditions than evaluated experimentally. Hence, the experiment was evaluated versus existing theories for the hyporheic exchange based on the so-called pumping mechanism (Wörman et al., 2002).

Formal optimisation of parameters using a direct fit versus the observed breakthrough curve gives a linear weight to all parts of the breakthrough curves (Bencala and Walters, 1983; Hayes et al., 1966), which may not be the optimal approach for the determination of all parameters. Since various parts of the breakthrough curve are controlled by various mechanisms (Harvey and Wagner, 2000), we may also apply fitting-by-eye combined with a log-transforms of the data to provide for a higher weight of the tails of the breakthrough curves (Jonsson et al., 2004). Log-transforms of the tracer data and a complete emphasis on the tail permits a more detailed evaluation of mechanisms responsible for exchange processes in the hyporheic zone, but ignores in-stream mixing (Hart, 1995; Haggerty et al., 2002).

The methods and limitations are demonstrated based on results from a stream tracer test in Hobøl River, Norway. A tritium injection was undertaken and 5 breakthrough curves were determined along 14 km of the river.

2. REPEATED TRITIUM INJECTIONS IN HOBØL RIVER, NORWAY

Two tracer injections of tritiated water were performed in Hobøl River, Norway, in August and October 2002 and the results are used as a reference in this paper. The tracer tests were performed under markedly different flow conditions and they are partly of different qualities, which makes them a valuable reference for analyses of the evaluation methodology and generalisation problems.

On 26 August 2002 at 13:10 about 222 GBq of H-3 was injected instantaneously about 10 km downstream of Lake Mjaer in Hobøl River at the Mill at Brekka. The discharge was 387 l/s at a distance from the injection site \( x = 471 \text{ m} \) and 644 l/s at \( x = 3531 \text{ m} \) (Fig. 1).

A repeated injection was undertaken with about 200 GBq of H-3 at the same location on 11 October 2002 at 17:35 to 18:25. The discharge was 226 l/s at a distance from the injection site \( x = 471 \text{ m} \). Consequently, the discharge was markedly lower at this second occasion.

During both experiments, samples were taken by the use of auto-samplers at the distances \( x = 471 \text{ m}, 3531 \text{ m}, 10138 \text{ m}, 12274 \text{ m} \) and \( 14048 \text{ m} \). The samples were collected and analysed in a beta counter. The injection activity was also determined by samples from the injection containers and analyses in the beta counter. Breakthrough curves from the two experiments and model curves are shown in Fig. 2.
Fig. 1. Injection site (left-hand side) and parts of the first reach of Hobøl River in October 2002. The left-hand side figure shows the insertion of the tritium isotope into the injection container that shields the surrounding from radioactivity. The right-hand side figure shows the third reach, 3531 m < x < 10138 m. The stream changes its character from the upper reach with more boulders and coarser bed material to the lower reaches where the stream cuts through clayey sediment deposits. Discharge at a distance of 471 m from the injection site is 644 l/s in August and 387 l/s in October. In-situ measurements indicate that the depth and width of the August flow are only slightly larger compared to the October flow conditions. At x = 471 m the mean depth decreased from 0.33 to 0.31 m and width from ~6.5 to ~6.0 m, whereas the mean flow velocity decreased significantly from 0.265 to 0.139 m/s.

Fig. 2. Breakthrough curves at different distances downstream of the injection site (Fig. 1) from the August and the October experiments, respectively. Model results (solid curves) as obtained from “fitting-by-eye” are included for the October experiment. The fit gets worse with distance, which is partly due to the change of the influence of hyporheic time scales with distance. The duration of the breakthrough is markedly different as a result of the different flow conditions.
Flow velocities and stream depth were surveyed in a limited number of cross-sections along the investigated reach. The cross-section was divided in intervals of 10 cm and the width was between 3 and 6 meters.

3. MATHEMATICAL FORMULATION OF THE PROBLEM AND TRACER TEST EVALUATION

General problem statement

The interpretation of breakthrough curves of inert solutes in streams is commonly based on a statement of mass conservation in the form of

$$\frac{\partial c_d}{\partial t} + \frac{1}{A_T} \frac{\partial (AU_d)}{\partial x} - E \frac{\partial^2 c_d}{\partial x^2} = J_s,$$  

(1)

where \(c_d\) is the solute concentration [kg/m³], \(A_T\) (in m²) is the cross-sectional area of the main stream including side pockets, \(A\) is the cross-sectional area of the main stream excluding side pockets, \(U\) is the flow velocity in the main stream [m/s] (\(Q = UA\)), \(Q\) is the discharge [m³/s], \(E\) is the main stream dispersion coefficient [m²/s] and \(J_s\) is the mass exchange rate with the hyporheic zone. The effective flow velocity in the main stream channel corrected for side pockets with stagnant water is given by \(U_e = Q/A_T\) (Wörman, 1998).

The exchange with the hyporheic zone leads to a typical skewed breakthrough following a tracer injection. The skewness indicates a prolonged transport process and a retention of solutes that can be very significant, especially for solutes that sorb onto solid surfaces in the hyporheic zone (e.g., soil particles) (Jonsson et al., 2004). The uptake has been expressed as a first-order mass transfer (Hays et al., 1966), diffusive (Jackman et al., 1984) and based on a general residence time distribution (Wörman et al., 2002; Haggerty et al., 2002). However, despite apparent differences between the physical rationale behind the model formulation, their mathematical characteristics, notably those defined in terms of the central temporal moments of the breakthrough curve, are similar (Wörman, 2000; Wörman et al., 2002). Therefore, this paper makes use of the general formulation of residence times in the hyporheic zone without preconceptions about what mechanisms are responsible for the exchange.

Using this framework, the net solute mass flux [kg m⁻³ s⁻¹] in the dissolved phase in the stream water can be found by integrating over the distribution of transport pathways (Wörman et al., 2002):

$$J_s = \frac{1}{2} \int_0^\tau f(T) \frac{P}{A} \left[-V_z(\tau,T) \right]_{r=0} \cdot c_d + \left(V_z(\tau,T) g_d \right)_{r=0} \right] dT,$$  

(2)
where $g$ is the solute mass per unit volume of water in the hyporheic zone [kg/m$^3$], $f(T)$ is the probability density function (PDF) of the residence time in the hyporheic zone $T$ weighted by the velocity component normal to the bed surface $V_n$, $V_z$ is absolute value of the flow velocity, $P$ is the wetted perimeter, $A$ is the cross-sectional area of the stream, and $\xi$ is an area reduction factor equal to $V_n/V_z$ that accounts for the fact that the streamlines are not necessarily always perpendicular to the bed surface.

Introduction of the T-PDF is essential to facilitate the treatment of different exchange mechanisms that occur on various geometrical scales of the stream and the landscape. For instance, surface roughness of the bed surface is known to cause hyporheic flow paths that extend over distances of decimetres (Thibodeaux and Boyle, 1987; Elliott and Brooks, 1997). On the other hand, larger obstacles in streaming water may cause a minor fraction of the flow paths to extend over longer distances. In addition, landscape topography leads to alternatingly dis- and recharge areas over the stream bottom. All these mechanisms added together, may produce a complicated T-PDF that can be difficult to observe in all details relevant for generalisation over larger scales.

**PARAMETER OPTIMISATION BY DIRECT FITTING OF MODEL SOLUTION VS. BREAKTHROUGH CURVE**

Evaluation of the model parameters is commonly done by fitting the model solution $c_d(t_i)_{\text{model}}$ to observations $c_d(t_i)_{\text{obs}}$ based on the points $i$ in which the breakthrough curve is observed. One should have in mind, however, that independent observations of model parameters are a valuable approach both to decrease the difficulty of the fitting process and as a test of the model validity. This problem is not addressed in this paper, but rather it is assumed that the theoretical basis (eqs. 1 and 2) generally applies.

From eqs. (1) and (2), we can find four independent parameters, namely $U_e$, $<T>$, $E$ and $W<T>/R$, where the exchange velocity $W = \xi 2 V_z$, the hydraulic radius $R = A/P$, and $<...>$ denotes arithmetic average. In terms of the transient storage parameters (e.g. Harvey and Wagner, 2000), we have the first-order exchange rate $\alpha = 1/T$ and the ratio of cross-sectional areas of the storage zone to the stream $A_s/A = W<T>/R$.

One reason to refrain from using the formal optimisation method described above is that the least-square estimates of the parameters do not automatically consider a “sophisticated” weight to various parts of the breakthrough curve. Log-transforms increase the weight of the tail (Haggerty et al., 2002). Another possibility is to optimise the fit-by-eye utilising the fact that the various parameters control different parts of the breakthrough (Harvey and Wagner, 2000). Figure 2 shows optimised model fit to the log-normal breakthrough curves from the October experiment using those fit-by-eye criteria.
4. COMPARISON RESULTS UNDER DIFFERENT FLOW CONDITIONS

From a visual comparison of the breakthrough curves in Fig. 2, we can see a marked difference in the slope of the tails from the two experiments. The deviating flow conditions between the August and October periods caused a significant differences in both the residence time of the hyporheic zone, $<T>$, and $W_{<T>/R}$. Table 1 summarises the results obtained using the three evaluation methods on the results from the August and October experiments. In this section, we discuss how the change in transport parameters (Table 1) can be explained from physical principles and the change in flow conditions.

<table>
<thead>
<tr>
<th>Reach [m]</th>
<th>Tracer test in August 2002</th>
<th>Tracer test in October 2002</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>$&lt;T&gt;$ [s]</td>
<td>$\xi^2/V_{&lt;T&gt;/R}$ [m/s]</td>
</tr>
<tr>
<td>0–3,531</td>
<td>15,000</td>
<td>0.0154</td>
</tr>
<tr>
<td>3,531–10,138</td>
<td>45,000</td>
<td>0.0320</td>
</tr>
<tr>
<td>10,138–12,274</td>
<td>500,000</td>
<td>0.071</td>
</tr>
<tr>
<td>12,274–14,048</td>
<td>500,000</td>
<td>0.114</td>
</tr>
</tbody>
</table>

The dispersion coefficients included in Table 1 are determined with a relatively low accuracy (±50%). Dispersion controls the width of the breakthrough curves, which is also somewhat controlled by the mixing with the hyporheic zone (Harvey and Wagner, 2000). The tails, however, are not affected so much by the dispersion, which means that fitting of the tail provide relatively good estimates of $<T>$, and $W_{<T>/R}$. Sensitivity analyses indicated that $W_{<T>/R}$ was determined with an accuracy of ±20% and $<T>$ within ±30% – ±70%.

Insufficient duration of stream water sampling or tail concentrations that sink beneath the detection limit of the tracer too early, may lead to an incomplete representation of the tail of the breakthrough curve. In both experiments described here, we had a nearly 100% mass recovery at $x = 3531$ m, but mass recovery dropped after this section due to both problems mentioned above. The consequence is that the exchange parameters of the August experiment could not be determined with an acceptable accuracy after 10,138 m (Table 1). Sensitivity analyses revealed a too wide range of possible interpretations. Also the exactness of, particularly, the residence time $T$ decays with distance in the October experiment.
Wörman et al. (2002) and Salehin et al. (2003) showed that the variation of hyporheic exchange between various stream reaches could be physically represented by the pumping mechanism caused by pressure variability along an uneven bed surface (Elliott and Brooks, 1997). This representation is believed to be a reasonable approximation also in this study, because results of Table 1 indicate that the length of hyporheic flow paths is 1-2 cm. This length interval is obtained as the product \( W \times <T> \) (mean depth at both occasions was about 30-40 cm at the uppermost 10 km). Thus, since the predominant exchange mechanism should be due to surface roughness of the stream, such as moving bed-forms or boulders, we express the hyporheic residence time as (Wörman et al., 2002)

\[
<T> = 3.8 \times 0.34 \left( \frac{h}{H} \right) \frac{d_b \lambda g h}{K^3 U^2},
\]

in which \( d_b \) is the depth limitation of the hyporheic zone, \( \lambda \) is the wavelength of the surface roughness and \( K \) is the hydraulic conductivity of the hyporheic zone. Fehlman (1985) found from hydraulic flume experiments that the pressure variability on the bed surface is described by \( r \approx 3/8 \) for sizes of the roughness elements on the bed surface, \( H \), less than 34% of the flow depth. Here, we tested both \( r = 3/8 \) and \( r = 0 \) to demonstrate the sensitivity of the results.

Consequently, we can expect a change of hyporheic residence time with flow conditions according to \( h \propto U^{-2} \). According to Manning formula for friction losses, \( h \propto U^{3/2} \) and this leads to \( <T> \propto U^{(3r/2-2)} \). Further, we can show from the same physical principles that the retention factor for solute elements, \( W <T>/R \), follow \( W <T>/R \propto d_b/h \propto d_b U^{-3/2} \).

Manning formula is introduced because of the difficulties to measure flow depths at exactly the same section between the occasions. Flow velocities were determined from stream gauge measurements as well as the tracer evaluation (Table 1). The difference between the flow velocity in the main-stream channel, \( U \), and the effective flow velocity, \( U_e \), was essentially a factor of 2 at \( x = 471 \) m, both in August and October. At \( x = 471 \) m, we got \( U/U_e \approx 0.273/0.137 = 1.99 \) in August and \( 0.139/0.073 = 1.90 \) in October.

The steam gauge measurements gave a flow velocity ratio between the two experiments of \( U_{\text{August}}/U_{\text{October}} = 1.96 \) at \( x = 471 \) m and \( U_{\text{August}}/U_{\text{October}} = 1.75 \) after about 20 km. Hence, if we use the proportionality \( <T> \propto U^{(3r/2-2)} \) and \( r = 3/8 \), at the first reach we can estimate \( T_{\text{August}}/T_{\text{October}} = 0.38 \). This number is 0.26 if \( r = 0 \). The corresponding ratio based on fitting-by-eye to the tracer data is \( T_{\text{August}}/T_{\text{October}} = 0.31 \). Accordingly, the results indicate fairly well that the expression for pumping exchange, (3), can be used to generalise hyporheic exchange rates to various flow conditions. This conclusion is further confirmed by studying the exchange velocity \( W \).
If we use the proportionality $W<T>/R \propto d_b U^{-3/2}$ and assume that the depth of the hyporheic zone is constant between the experiments, we get an expected change $(W<T>/R)_{\text{August}}/(W<T>/R)_{\text{October}} = 0.36$. The corresponding ratio evaluated from the tracer data (Table 1) is about 0.5 (as an average for all three methods).

The fact that the accuracy of the evaluation methods decreases with distance is a problem to generalisations over longer stream reaches and to entire watersheds. Harvey et al. (1996) and Harvey and Wagner (2000) suggested that the “experimental Damköhler number” – defined in terms of the current parameters as $D_m = x/(U_e T) + h x/(U_e W T^2)$ – should be around unity to allow a good estimate of the hyporheic exchange. Further, it is possible to show that as long as the capacity of the hyporheic zone does not limit the uptake (unlimited depth), the time scale of the process decreases with $D_m$ (Wörman, 1998), which, e.g., implies a decreasing slope of the tail of breakthrough curves with distance. Such a decrease of the slope of the tails is evident from the results in Table 1, where an increasing value of $<T>$ implies a decreasing slope of the tail. Wörman et al. (2002) found the same tendency over 30 km.

Hence, the results of this study indicate that the tracer tests does not reveal a definite limit of the time scales involved in the hyporheic zone mixing. A related reason for this is that surface-sub-surface water interaction occurs over such a wide range of geometrical scales.

7. VARIATION OF PARAMETERS WITH DISTANCE

Previous investigations have focused on providing scaling relationships for hyporheic exchange parameters with distance in streams with a particular account taken to variation in hydraulic and morphological conditions (Wörman et al., 2002; Salehin et al., 2003). The scaling relationship (3) has been found to apply reasonably well.

In this study, hydraulic conductivity of bed sediments was not measured. Qualitatively, however, the change of $<T>$ with distance in Hobøl River is plausible. As seen in Fig. 1, the stream changes its character with distance, from the upper reach with more boulders and coarser bed material to the lower reaches where the stream cuts through clayey sediment deposits. The bed material is likely to be less permeable with distance, which, according to (3), implies an increase of $<T>$ with distance. Also water depth increases slightly, which also acts to increase $<T>$ with distance. This increase is qualitatively verified by the values listed in Table 1 for both experiments.

8. DISCUSSION OF RESULTS AND CONCLUSIONS

Two tracer tests with H-3 were performed in Hobøl River, Norway, under markedly different flow conditions as a basis for evaluation of methods used to interpret tracer breakthrough curves and an analysis of problems related to generalisation of results to other stream reaches, especially larger scales. Despite the fact that the discharge and
the flow velocity decreased by a factor of 2.2 between the occasions, the ratio of mainstream velocity (from gauge meters) and “effective flow velocity” (i.e. tracer velocity) was the same, \( \sim 2 \).

The difference in hyporheic exchange parameters between the two occasions could be well explained by the pumping theory, particularly eq. (3). From hydro-mechanical principles, the change of the residence time in the hyporheic zone was found to approximately follow \( <T> \propto U^{-1.74 \pm 0.25} \). The retention factor for solutes followed \( W<\bar{T}>/R \propto U^{-3/2} \).

However, the distance of investigation, expressed in terms of a Damköhler number, is another factor that affects possibilities to generalise hyporheic exchange. As long as the hyporheic exchange is not limited with depth, more and more geometrical scales of the stream system and the surrounding landscape will influence the groundwater–surface water interaction.

There is a decreasing accuracy of results with distance. This problem is because of the increasing problem of capturing the tail with distance. If the mass recovery of the breakthrough curve is larger than about 70%, one can determine the parameters \( <T> W/h \) and \( U_e \) with an acceptable accuracy of \( \pm 20\% \).

References


Received 13 June 2005
Accepted 6 September 2005