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ORIGINAL ARTICLE

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Influence of streambed hydraulic conductivity on solute exchange with the hyporheic zone

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Abstract A conservative solute tracer experiment was conducted in Indian Creek, a small urban stream in Philadelphia, Pennsylvania to investigate the role of subsurface properties on the exchange between streamwater and the hyporheic zone (subsurface surrounding the stream). Sodium Bromide (NaBr) was used as a conservative tracer, and it was monitored in the surface water at two stations and in the upper bed sediments (shallow hyporheic zone extending from 7.5 to 10 cm below the streambed). The hydraulic conductivity (K) of the upper bed sediments and the lower bed sediments (10–12.5 cm below the streambed) was measured in situ. High tracer concentrations were observed in the upper layer at locations where the hydraulic conductivity of the upper layer was larger than that of the lower layer. Low concentrations in the upper layer were observed in the converse case. A statistically significant relationship between the mass retained in the upper layer and the difference of K values between layers was observed.

Keywords Hyporheic exchange · Hydraulic conductivity · Streambed heterogeneity · Indian Creek · Philadelphia · Pennsylvania · USA

Introduction

The exchange between stream water and the subsurface surrounding the stream (the hyporheic zone) has been shown to affect the transport and fate of solutes in streams. The exchange flow is affected by variation in the hydraulic conductivity, K, of the hyporheic zone and in the head gradient along the stream. While most studies focused on topography (or bathymetry) as the major factors generating the local variation in head gradient (and subsequently exchange flow), more recent studies have pointed out the role of heterogeneity (Cardenas et al. 2004; Salehin et al. 2004). Cardenas et al. (2004) examined hyporheic exchange in a homogeneous streambed and in a streambed with $\sigma^2_{\ln K} = 0.86$ and concluded that hyporheic exchange is dominated by heterogeneity when pressure head variation (e.g. due to bed forms) is small. Salehin et al. (2004) examined hyporheic exchange in heterogeneous sand beds with $\sigma^2_{\ln K} = 1.0$ and 2.0 and found that heterogeneity resulted in a shallower hyporheic zone with a higher exchange rate. These recent studies were in the laboratory or numerical. This study is at the field scale, reporting results from a tracer study in an 80 m portion of Indian Creek, an urban stream in Philadelphia, Pennsylvania, USA. Measurements of the conservative tracer NaBr were made in the stream and the hyporheic zone. The tracer results were then analyzed in conjunction with the K values to evaluate potential trends between the two.

Site description

Indian Creek is within the Piedmont physiographic region and the streambed is primarily gravel and cobble sediment whose pore spaces are filled with significant amounts of sand and silt. The stream begins in the urbanized southwest part of Montgomery County in southeast Pennsylvania, USA. It flows in a generally southerly direction and enters the western edge of the City of Philadelphia, where it is protected as part of the city's Fairmount Park system. A wide riparian corridor (150–200 m) consisting of deciduous forest on steep valley sides (20–25% slope) exists within the study reach. It is expected that this corridor allows some infiltration of stormwater runoff with a concomitant reduction in many pollutants (Hachmöller et al. 1991; Paul and Meyer 2001; Pinay et al. 1992). However, this protection is mitigated by the presence of multiple Combined Sewer Outfalls (CSO) which discharge a mixture of stormwater runoff and raw sewage into Indian Creek during large rain events. More information on the site can be found in Ge and Boufadel (2006) and Ryan and Boufadel (2006).

Two surface water monitoring stations were established (Fig. 1). The first station (Station 1) was at a riffle 138 m downstream of the injection point. The second station (Station 2) was 246 m downstream of the injection point. Hence, the length of the reach between stations was 108 m. Between these two stations, the stream consists of three pools separated by two riffles. The first pool was 15 m long, 4 m wide, and as much as 0.6 m deep. It was the narrowest of the pools. A large gravel bar (15 m long by 13 m wide) was to the right of this pool. The second pool was 25 m long, 8 m wide, and 0.8 m deep at its deepest point. A CSO that was dry during this study was on the right bank at the upstream end of this pool. The third pool was 19 m long, 6 m wide, and 0.3 m deep at its deepest point. Downstream of the third pool was a riffle-pool/step-pool transition zone. Measurements of hyporheic samples and K were obtained within the pools between the two surface water monitoring stations.

Methods

The stream channel and portions of the floodplains were surveyed using a Philadelphia Rod and a Spectra-Physics Autolevel 280 Laserline transit level. Points were surveyed on an approximate 1 m × 1 m grid resulting in approximately 3,000 points in the study reach. The survey occurred over a period of approximately 8 weeks during the winter and early spring of 2005. There were no large storms or extreme flow events that would have significantly altered the channel morphometry between the survey and the stream tracer study.

Bromide was chosen as the conservative tracer because its background concentration was low (typically <0.1 mg L^{-1}). The tracer injection began on the morning of May 7, 2005 and continued for 24 h. The tracer injectate solution was made by mixing approximately 86 kg of NaBr with stream water in a 1,200 L tank using a wooden paddle, and allowing it to sit overnight. The injectate was remixed on the morning of the injection. The injectate concentration was measured and found to be equal to 71,627 mg L^{-1} Br^{-}. The injectate solution was gravity-fed to the stream and the flow rate was monitored using a King Instruments float-type flow meter valve attached to the tank outlet. The valve was adjusted as needed in an effort to maintain a constant discharge rate. A 3.0 m long manifold was used to make the delivery of the solution more uniform across the stream. The manifold was fabricated by perforating a 3.8 cm (ID) diameter PVC pipe at 10 cm intervals. The manifold was placed approximately 25 cm above the water surface. The injection rate through the manifold system averaged about 9 mL s^{-1}.

Surface water samples were collected from May 7, 2005 to May 8, 2005. Hyporheic zone samples were collected from May 7, 2005 to May 9, 2005. The discharge was measured by integrating ten velocity measurements across the stream 15 m downstream of the injection point. These measurements were made using a SONTEK ADV acoustic Doppler velocimeter. The ADV reads velocity values at an interval of 0.004 s (i.e., 250 Hz) and reports them by averaging over the interval requested by the user, which was 1.0 s herein. The accuracy of the ADV is 1% of the reported velocity.

To estimate the plateau value (i.e., maximum) of the stream concentration at the injection point, the following equation (Kilpatrick and Cobb 1985, p 5) was used:

$$C_{\mathrm{IP}} = \frac{Q_{\mathrm{inj}} C_{\mathrm{inj}}}{Q_{\mathrm{IP}}}, \tag{1}$$

where Q_{inj} is the injection flow rate, C_{inj} is the concentration of the injectate, and Q_{IP} is the measured stream discharge at the injection point (actually 15 m downstream of the injection point).

Surface water samples at the two stations were obtained using two methods. During the rising and falling limb of the breakthrough curve, samples were obtained by hand using 120 ml polypropylene bottles that had been triple rinsed in the stream prior to collecting the sample.

During the plateau portion of the experiment, autosamplers (ISCO Model 3700) were used. The collected samples were transferred from the ISCO bottles to 120 mL polypropylene bottles for transport to the laboratory. Sampling was conducted at 3 min intervals for the hand samples and hourly for those by the autosamplers. All samples were collected from the center of the cross sections.

Groundwater elevation near the stream was determined from six bankside groundwater wells (GW, Fig. 1). Each well was hand dug with the use of a 10 cm

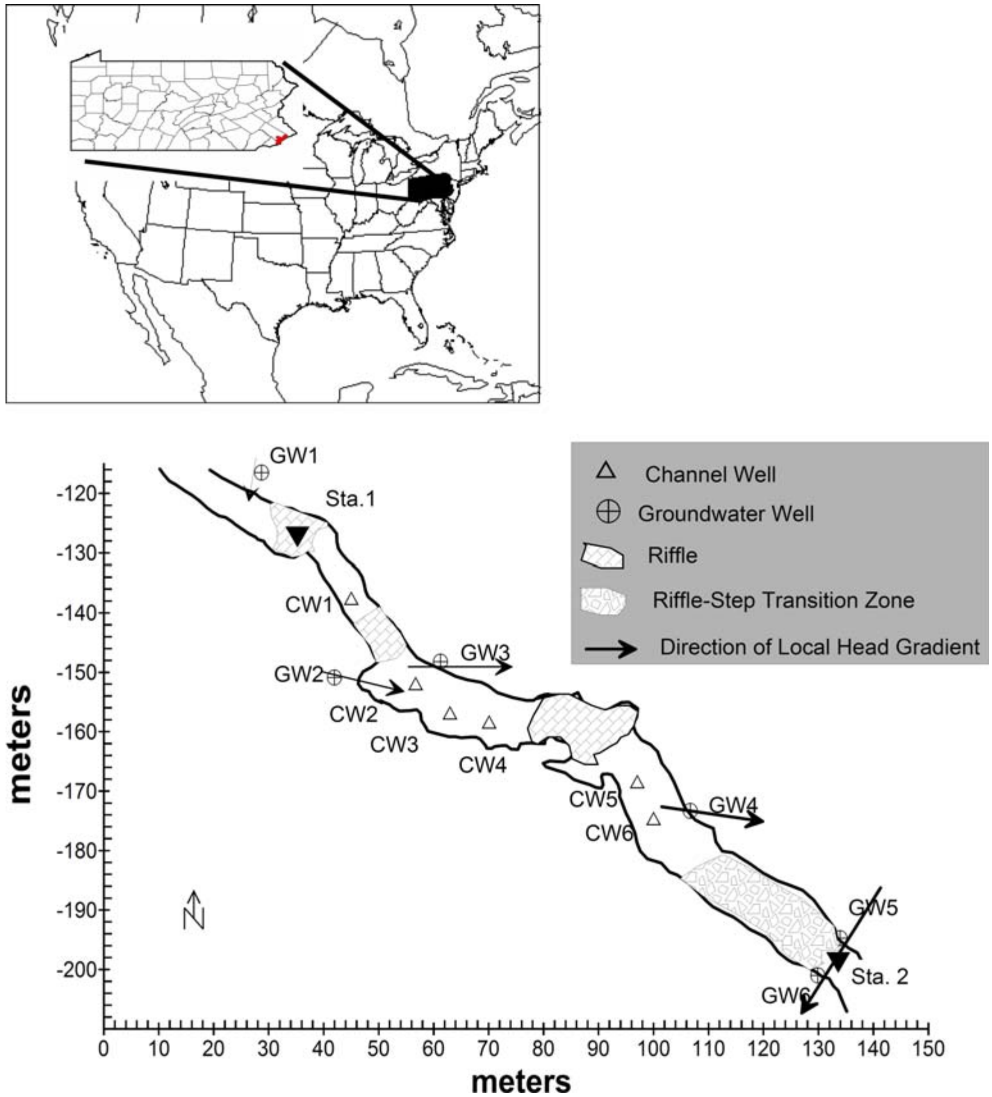

Fig. 1 Location of Indian Creek and detailed map of reach from Station 1 to Station 2

diameter hand auger and were finished with PVC pipe (5.0 cm diameter) that was capped at the bottom and perforated over an interval of approximately 10 cm. The perforated interval was screened using several layers of standard window screening which was secured with twine. Once the PVC was placed into the borehole, the borehole was backfilled by hand. The depth of the well, the depth to water in each well, and the height of the casing above the ground surface was then recorded.

Six wells were installed in the channel. They are referred to by the symbol CW in Fig. 1. The downstream distances of the CWs are reported in Table 1. The location of each well was determined through triangulation with known benchmarks set during the surveying effort. CW1 was located 7 m downstream of Station 1 in a small pool. CW2, CW3, and CW4 were located in the next pool downstream, 36, 42, and 48 m downstream of Station 1, respectively. Finally, CW5 and CW6 were located in the third pool, 76 and 82 m, respectively, downstream of Station 1. Each CW consisted of a 1.5 cm ID iron or copper pipe. The pipe was capped on one end and perforated over a 2.5 cm segment on that end. The wells were pounded into the stream bed such that the top of the perforated interval was 7.5 cm below the streambed. Thus, the water samples represent the hyporheic zone between 7.5 and 10 cm below the streambed. This depth was chosen to coincide with hydraulic conductivity measurements made in the upper sediment layer.

Samples from all CWs were collected approximately 2, 9, 21, and 48 h after the start of the injection. All well samples were collected using a 1.8 m length of 0.32 cm diameter Tygon tubing attached to a 60 mL syringe. Each hand sampling event consisted of withdrawing three 20 mL aliquots of water. The first aliquot was used to rinse the tubing (the total volume of the tubing was 15 mL). This aliquot was then disposed of on the stream bank. The second aliquot was used to rinse the 120 mL

Table 1 Estimated average plateau concentration in surface water (SW) and maximum observed concentration in channel wells (CW)

Label	X (m)	h (m)	SW conc. (mg L^{-1})	CW conc. (mg $L^{-1)}$
CW1	148	0.10	10.8	9.8
CW2	160	0.23	10.8	8.6
CW3	172	0.32	10.8	6.4
CW4	184	0.20	10.7	0.3
CW5	196	0.13	10.7	1.2
CW6	208	0.14	10.7	9.5

X (m), Alongstream distance from the injection point; h (m), Depth of stream at well location

polypropylene sample bottle, and the third aliquot was returned to the laboratory for analysis.

All surface water and hyporheic samples were analyzed using a Dionex 500 Ion Chromatograph, which was calibrated each day using a six point calibration procedure. One out of every 10 samples was run in duplicate. The coefficient of variation (CV) was 0.033. The injectate samples were diluted by a factor of 10^5 prior to analysis. The injectate samples were also analyzed (without dilution) using an Orion IonAnalyzer model 407A meter with a Mettler Toledo bromide ISE and a Cole-Parmer single junction reference electrode. The results of both methods were in agreement.

The hydraulic conductivity of the streambed was measured following the tracer experiment (in a period of two weeks). The measurement was in situ using a portable falling head permeameter (Landon et al. 2001; Freeze and Cherry 1979) whose results were interpreted using the method of Hvorslev (1951, p 44, Fig. 18-G). The permeameter consisted of a 10 cm diameter reservoir attached to a hollow steel pipe with inside diameter of 1.2 cm and outside diameter of 2.2 cm. A solid steel drive point was attached to the other end of the pipe and the pipe was evenly perforated over a 2.5 cm section with 12 holes each 0.9 cm in diameter. The perforated section was then wrapped with stainless steel mesh having an opening size of 178 μm. All K measurements were made within the 80 m subreach of Indian Creek, extending downstream from Station 1, in which the channel wells were sampled during the tracer test. The K measurements were obtained at the two depth intervals: 7.5–10 cm (upper layer) and 10–12.5 cm (lower layer). Attempts to measure K at depths of less than 7.5 cm resulted in short-circuiting along the permeameter casing. Measurements in the upper layer sediments were made at 38 locations on 10 lateral transects that were laid out across the pools in which the channel wells were located. Measurements in the lower layer sediments were made at 47 locations on 11 lateral transects laid out across the same pools.

Results

The flow rate was measured on the day prior to the start of the injection and on Day 2 of the injection, and was found to be approximately 53 L s^{-1}. Based on this flow rate and the injectate's mean flow rate (9 mL s^{-1}) and concentration (71,627 mg L^{-1}), the in-stream concentration at the injection point was found to be (Eq. 1) 12.3 mg L^{-1}. Equation 1 assumes that complete and instantaneous mixing occurs between the injectate and the streamwater, and that the initial background concentration is 0. As the background concentration was less than 0.1 mg L^{-1}, the error introduced by the latter assumption is negligible.

Breakthrough curves for each of the surface water monitoring stations are shown in Fig. 2. The variation in "plateau" Br^- concentration observed at each of the stations is attributable to the lack of precise injection rate control; gravity was used as the sole driving gradient for the injection flow causing the injection rate to decrease continuously over time as the head on the outlet was reduced. This head reduction was counteracted by periodically (approximately every 30 min) adjusting the metering valve on the tank outlet. The variation in the plateau Br^- concentration was smaller at Station 2, most probably because of longitudinal dispersion, which is expected to increase with the distance from the source (Fischer et al. 1979, p 131). The average plateau Br^- concentration was 10.84 and 10.64 mg L^{-1} at Stations 1 and 2, respectively. The Wilcoxon–Mann–Whitney Rank Sum Test (Thorndike and Dinnel 2001, p 389) was used to test the null hypothesis that the plateau concentration observed at Station 2 was equal to the plateau concentration observed at Station 1. This is a non-parametric test that makes no assumption regarding the underlying distribution of the data. The results of the test indicated that the null hypothesis should be rejected; the plateau concentration at Station 2 was lower than that at Station 1 with a $p < 0.001$. The reduction in the plateau value between stations could be attributed to dilution by incoming groundwater (to the stream). The bromide concentration at both surface water monitoring stations returned to near background levels (<0.2 mg L^{-1}) within 29 h of the beginning of the injection (i.e. 5 h after the injection ended).

The local head gradient between the stream and the surrounding aquifer was calculated using the observed water elevations in the groundwater wells and in the stream. As shown in Fig. 1, the gradient was generally from the groundwater to the stream in the upper section of the reach (upstream of CW2), from the stream to the left bank between CW2 and CW6, and across the stream from the left bank to the right bank at Station 2. This suggests that the slight increase in flow observed between Station 1 and Station 2 occurred upstream of CW2 or downstream of CW6.

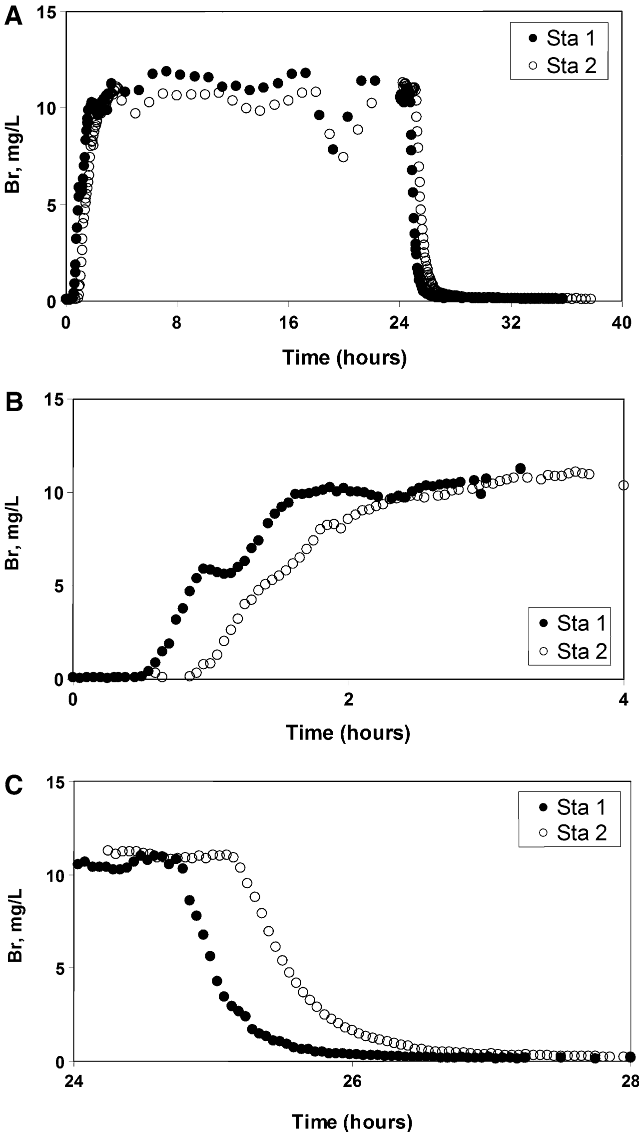

Fig. 2 Observed breakthrough curves for surface water monitoring Stations 1 and 2 : **a** complete breakthrough curves; **b** rising limb; **c** falling limb. Time is from start of injection

The breakthrough curves for each of the six channel wells are shown in Fig. 3. The maximum hyporheic tracer concentration in the wells and the estimated maximum surface water concentration near the wells are shown in Table 1.

CW1 appears to have responded the quickest of all channel wells. The concentration in CW1 was 7.9 mg L^{-1} (73% surface water) 85 min after the start of the injection. This was approximately the same response time observed in the surface water at Station 1. The maximum concentration observed at CW1 was 9.8 mg L^{-1} or 91% surface water. The maximum observed concentrations in the channel wells decreased to 8.6 mg L^{-1} (79% surface water) at CW2 and 6.4 mg L^{-1} (59% surface water) at CW3. Surface-subsurface exchange was significantly reduced at CW4 and CW5. The maximum observed bromide concentration at CW4 was just 0.3 mg L^{-1} (3% surface water) and at CW5 the maximum bromide concentration was just 1.2 mg L^{-1} (11% surface water). At CW6 (Fig. 3), the maximum observed bromide concentration was 9.5 mg L^{-1} (90% stream water), which occurred at 9 h after the start of the tracer injection. However the second highest value (8.9 mg L^{-1}), which was practically equal to the maximum, occurred at $t = 22$ h, which was the time at which the peak concentrations were observed in the other CWs.

At $t = 48$ h (i.e., 24 h after the end of injection) the concentration in all channel wells except CW3 returned to the background value. The observed concentration in CW3 was still 78% of the maximum. It is speculated that this elevated concentration may be the result of the channel well screen becoming clogged near the end of the injection, leaving water with a high tracer concentration trapped inside.

The K values measured in the upper sediment layer ranged from 1.71×10^{-3} to 1.78×10^{-2} cm s^{-1}, with a mean value of 8.98×10^{-3} cm s^{-1}. This layer was relatively homogeneous with a variance of the logarithmically transformed data ($\sigma^2_{\ln K}$) equal to 0.22. The K values measured in the lower sediment layer ranged from 1.16×10^{-4} to 3.32×10^{-2} cm s^{-1} with an average value of 5.97×10^{-3} cm s^{-1}. This layer was more heterogeneous than the upper layer with a $\sigma^2_{\ln K} = 2.26$. The mean K value of the upper layer was significantly higher than the mean K value of the lower layer ($p < 0.0002$) based on the student t test (R.J. Ryan and M.C. Boufadel, submitted). The $\sigma^2_{\ln K}$ of the upper layer was significantly lower than the $\sigma^2_{\ln K}$ of the lower layer ($p < 0.0001$) based on the F test.

Fluidization of the bed sediments (i.e. 'flowing sands') has been cited in other work as preventing accurate in situ K measurements in upper bed sediments (e.g. Cardenas and Zlotnik 2003). However, this phenomenon was not observed in Indian Creek. Further, based on data presented by Bouwer and Rice (1976), the head induced by the test method used herein would be expected to dissipate to within 5% of the background head

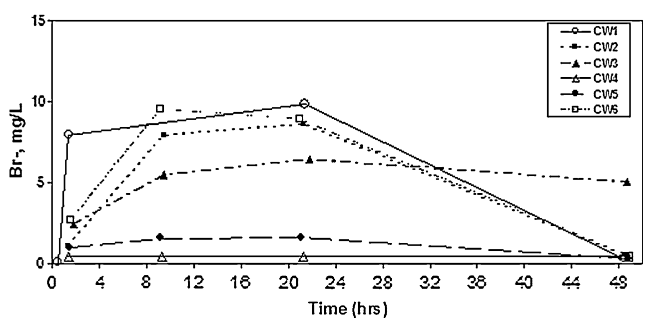

Fig. 3 Observed breakthrough curves for channel wells. Time is from start of injection

within a vertical distance equal to only 1.25 cm. For this reason, it is unlikely that the 'flowing sands' phenomenon occurred in Indian Creek. In addition, replicate hydraulic conductivity tests were conducted and analysis of variance indicated that there was no significant difference between replicate *K* tests ($p = 0.01$).

Some error in the absolute value of the *K* estimates may have been introduced by the use of the methodology of Hvorslev (1951) as this method assumed that flow will exit the permeameter through the bottom as well as horizontally. Bouwer and Rice (1976) present a method which assumes only horizontal flow out of the permeameter. A comparison of the two methods indicated that estimates of *K* using Bouwer and Rice (1976) would be approximately 55% lower than those estimated using Hvorslev (1951) and reported herein. However, the error would be constant throughout the data set, and so the statistical variation within each sediment layer and between the two sediment layers would not have changed. For the purposes of this study, the equation presented by Hvorlsev (1951) is acceptable.

The spatial structure of the *K* data of each sediment layer was analyzed and an exponential variogram (Gelhar 1993) was fit to the data and used as input in the software package SURFER (http://www.GoldenSoftware.com) to interpolate the *K* data by kriging. The resulting contours of *K* are plotted in Figs. 4 (upper sediment layer) and 5 (lower sediment layer), along with the actual location points of the *K* measurements.

The *K* values in Indian Creek were low and consistent with a silt and sand stream bed (Freeze and Cherry 1979, p 29). However, the streambed in the study reach consisted primarily of gravel- and cobble-sized particles, which have higher hydraulic conductivity (Freeze and Cherry 1979, p 29). The low *K* values are most likely due to a large amount of small-size particles (sand, silt, and clay) that filled the pore space of the coarser sediments. This was observed by Ryan and Packman (2006) in their study of a nearby urban watershed.

Figure 6 shows the *K* value of the upper and lower sediment layers (K_u and K_l, respectively) at each well and the maximum Br^- concentration observed in the upper sediment layer. K_u was larger than K_l at four channel wells (CW1, CW2, CW3, and CW6) and smaller at two channel wells (CW4 and CW5). The concentration was highest (6.5–9.8 mg L^{-1}) at wells where $K_u > K_l$ and smallest (0.4–1.2 mg L^{-1}) otherwise. This suggests that hyporheic flowpaths (i.e., those flowpaths that emerge and return to stream) remained in the upper layer sediments whenever $K_u > K_l$.

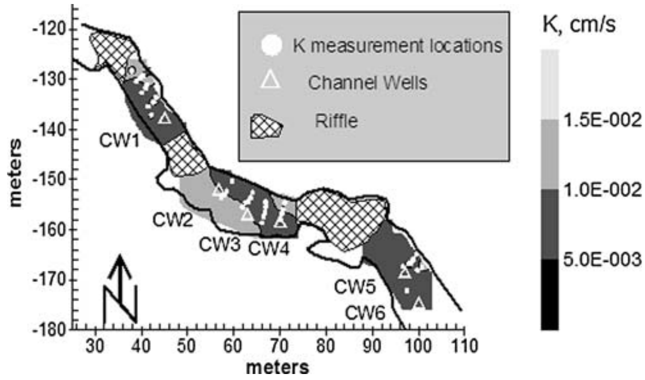

Fig. 4 Hydraulic conductivity contours of upper layer bed sediments (7.5–10 cm below the streambed)

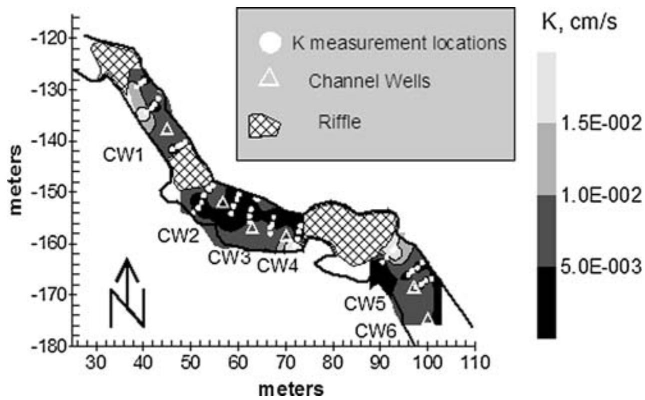

Fig. 5 Hydraulic conductivity contours of lower layer bed sediments (10–12.5 cm below the streambed)

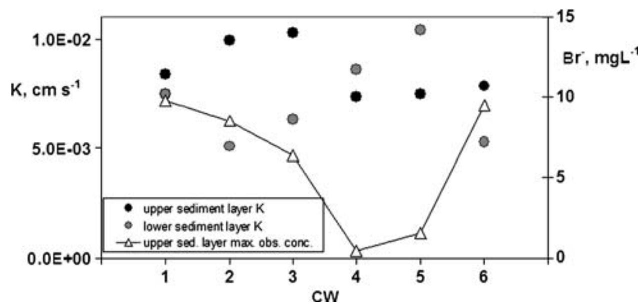

Fig. 6 Comparison of maximum observed Br^- concentration with *K* values estimated for upper layer sediments and lower layer sediments at each in-stream channel well location. Hyporheic exchange with the upper layer is maximized when the value of *K* decreases with depth and is minimized when the value of *K* increases with depth

Discussion

A complete understanding of the transport mechanisms intervening in Fig. 6 requires solving the governing equations for water flow and solute transport in the stream-hyporheic system. This is beyond the scope of this investigation. The authors seek, however, semi-empirical arguments, as discussed next.

Three major factors affect the transport of solutes in the sediments: groundwater flow, bathymetry, and heterogeneity. In the context of the results in Fig. 6, dilution by incoming groundwater to the stream could not be the reason for the small concentration values at CW4 and CW5, because the head gradient at these locations was directed from the stream channel toward the bank. To assess the role of bathymetry on hyporheic exchange, the standard deviation in the vicinity of the channel wells was computed. In the first pool, the area within 3 m of CW1 had a mean value of −1.20 m (relative to an arbitrary datum) and a standard deviation of 0.12 m. In the second pool, the area 3 m upstream of CW2 to 3 m downstream of CW4 (which includes CW3) had a mean streambed elevation of −1.42 m and a standard deviation of 0.18 m. In the third pool the area extending from 3 m upstream of CW5 to 3 m downstream of CW6 had a mean streambed elevation of −1.58 m and a standard deviation of 0.11 m. The small standard deviations suggest that bathymetry was not strongly varying in the channel wells' vicinity, which is not surprising considering that these wells are in pools. This does not rule out bathymetry completely, but suggests that bathymetry might not have played a big role in the results of Fig. 6.

In regard to heterogeneity, one notes that transport into the relatively high permeability streambed by convection is dominant with respect to transport by molecular diffusion. Thus, an understanding of water flow into the sediment provides an explanation of that of solute transport.

One can envision that the total horizontal water flow (q) through a heterogeneous streambed could be divided into flow through an upper sediment layer (q_u) and flow through a lower sediment layer (q_l), as shown illustratively in Fig. 7. Then,

$$q_u = q - q_l \tag{2}$$

and applying Darcy's Law on the lower layer:

$$q_l = K_l i, \tag{3}$$

where i is the horizontal head gradient. Thus,

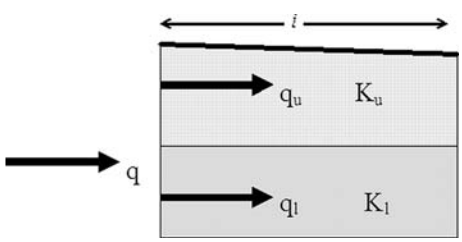

Fig. 7 Conceptual schematic of hyporheic flowpath. q is specific discharge, K is hydraulic conductivity, i is the head gradient, and subscripts u and l indicate upper and lower sediment layer, respectively. In a heterogeneous sediment with a constant head gradient, flow is divided between incremental layers based on variation in K, such that q is equal to q_u plus q_l

$$q_u \propto (-K_l i), \tag{4}$$

where the symbol $\propto$ indicates proportionality. Application of Darcy's law in the upper layer gives:

$$q_u = K_u i. \tag{5}$$

Using Eqs. 4 and 5, one may stipulate that:

$$q_u \propto (K_u i - K_l i). \tag{6}$$

Thus,

$$q_u \propto (K_u - K_l). \tag{7}$$

The transport of solute into the hyporheic zone could be inferred by considering the first 22 h of the breakthrough curves in Fig. 3; if one computes the area under each curve and divides it by the corresponding time, then one obtains the average concentration, C_{ave}, that passed the port 7.5 cm below the streambed surface. If one assumes that that zone is completely mixed (essentially ignoring the very early times), then the quantity

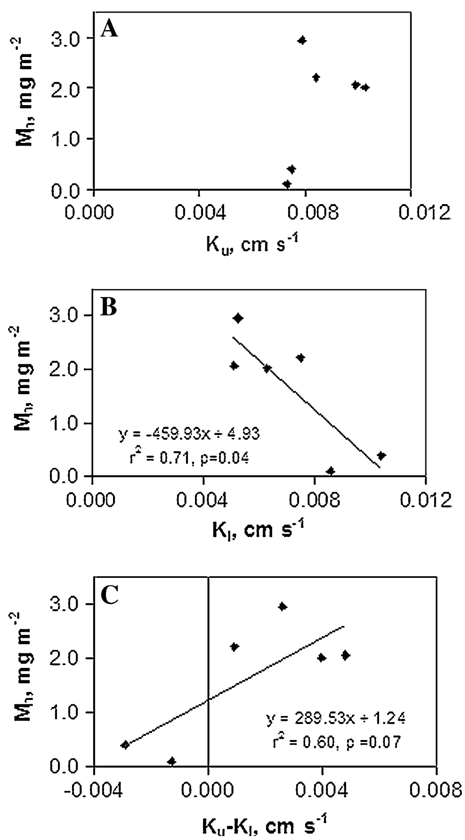

Fig. 8 Plot of M_h (the mass of tracer that entered the upper sediment layer per unit bed area) versus **a** the value of K in the upper sediments, **b** the value of K in the lower sediments, and **c** the difference in K values. The influence of vertical heterogeneity is evidenced by the significant relationship observed in (**b**) and (**c**), which suggests that the lower layer K value played an important role in controlling exchange through the upper layer

$$M_h = (C_{ave})\,(0.075)\,(1000) \quad (8)$$

represents the mass per unit area (mg/m^2) that entered the stream within 22 h at the specific location where C_{ave} is obtained. The quantity (1,000) in Eq. 8 is used to convert from liter to m^3.

Figure 8 shows plots of M_h versus K_u, K_l, and (K_u–K_l). The linear correlation was highest between M_h and K_l (r^2=0.71, p=0.04) followed by M_h versus (K_u–K_l), (r^2=0.6, p=0.07). The regression of M_h versus K_u was very poor, because the plot represents Eq. 5, which was obtained by application of Darcy's law on the upper layer without taking into account the overall mass balance in the hyporheic zone in the vicinity of wells (represented by Eq. 2).

The goodness of fit (in Fig. 8b, c) that accompanied the theoretical discussion above indicates that heterogeneity was the major mechanism behind the variation of concentration at various CW in Fig. 6. Although Fig. 8b (Eq. 4) gave the best fit, the authors favor using the regression of Fig. 8c (Eq. 7), as it seems intuitive to include the hydraulic conductivity of the upper layer in estimating hyporheic exchange.

These field-based results validated, in broad terms, the results of modeling and flume experiments reported by others (Cardenas et al. 2004; Salehin et al. 2004). The mean values of K for Indian Creek were similar to the mean values used by Cardenas et al. (2004) and Salehin et al. (2004) to model sand-bed streams. Cardenas et al. (2004) concluded that hyporheic exchange is dominated by heterogeneity when pressure head variation (e.g. due to bed forms) is small.

Salehin et al. (2004) found that heterogeneity resulted in a shallower hyporheic zone and a higher exchange rate. Both of these studies utilized sand-bed streams with relatively high heterogeneity ($\sigma^2_{\ln K}$ = 0.86–2.0). In Indian Creek, very thin sediment layers with distinct statistical properties were observed. The upper layer sediments were less heterogeneous with a higher mean K than the lower layer sediments. This upper sediment layer provided a pathway of relatively low resistance in comparison to the lower sediment layer. Thus, hyporheic exchange with deeper sediments was likely limited to those locations where the hydraulic conductivity of the lower sediment layer was less than that of the upper sediment layer (Fig. 6). Using Darcy's Law, the specific discharge through the hyporheic zone was shown to be a function of vertical heterogeneity (K_u – K_l) and using field data collected from Indian Creek, a statistically significant linear relationship was developed between (K_u – K_l) and the area under the breakthrough curve (Fig. 8, r_2 = 0.6, P = 0.07).

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