

# Evaluation of streambed hydraulic conductivity heterogeneity in an urban watershed

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**Abstract** It has long been understood that streambed hydraulic conductivity plays an important role in surface-subsurface solute exchange. Using a portable falling head permeameter in situ, we estimated the horizontal hydraulic conductivity,  $K$ , of the near-surface streambed sediments at a total of 85 locations encompassing two depth intervals: 7.5–10 and 10–12.5 cm. The measurements were conducted in an 80 m reach of Indian Creek, a small urban stream in Philadelphia, PA, USA. We found that the  $\ln K$  data within each sediment layer were Gaussian, but the combined data set was not. The results indicated that while the mean hydraulic conductivity decreased with depth, horizontal heterogeneity (e.g. the variance) increased with depth. This strong contrast between layers suggests that they should be treated as separated entities in modeling studies. Variogram analyses across the stream suggested symmetry with respect to the thalweg in the upper layer and fractality in the lower layer. The variograms along the streams suggested that the  $K$  data are random.

**Keywords** Hydraulic conductivity · Streambed heterogeneity · Indian Creek

## 1 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 (Bencala and Walters 1983; Ge and Boufadel 2006; Ryan and Packman 2006). 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 morphology (e.g. Harvey and Bencala 1993; Elliot and Brooks 1997a, b; Kasahara and Wondzell 2003) as the major factor generating the local variation in head gradient (and subsequently exchange flow), more recent studies have demonstrated that when the pressure head variation is small, hyporheic exchange is driven by hydraulic conductivity heterogeneity (Cardenas et al. 2004) and that increased heterogeneity results in faster hyporheic exchange along shorter hyporheic flowpaths (Salehin et al. 2004). These recent studies were based on laboratory and numerical experiments in which streambed heterogeneity was represented by random fields developed from hydraulic conductivity values of deep bed sediments and aquifers. However, the shallow sediments are expected to play a major role in controlling hyporheic exchange (Ryan and Packman 2006). This may be especially important within pools and glides since the pressure head variation within these geomorphic features should be less than the pressure head variation within, for example, a pool/riffle sequence. In addition, heterogeneity in natural streams would not be expected to be completely random. For example, downstream fining and settling of fines along channel sides are well-known processes (Leopold et al. 1995).

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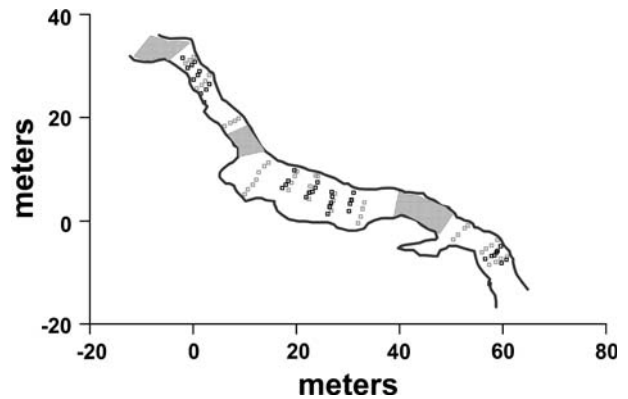
Much data have been reported for the statistical properties of aquifers (Gelhar 1993; Boufadel et al. 2000; Tennekoon et al. 2003, and references therein). Streambeds have been less studied. Calver (2001) reported on 41 studies conducted between 1967 and 2000. The studies were undertaken in the United Kingdom (19), United States (11), Europe (8), Africa (1), Arabia (1), and Australia (1). Approximately 41% (17) of these studies involved estimating  $K$  based on field measurements, however, details of the field methods were not provided. Duwelius (1996) reported 32 hydraulic conductivity measurements made at depths ranging from 12 to 198 cm below the streambed along a 13.5 km reach of the East Branch Grand Calumet River (IN, USA). Springer et al. (1999) reported 73 hydraulic conductivity values measured at depths ranging from 1.52 to 6.1 m in five reattachment bars spread over a 320 km reach of the Colorado River. Cardenas and Zlotnik (2003) appear to have reported the most spatially dense dataset to date with 456  $K$  measurements at 76 locations in a 45 m reach of Prairie Creek (NE, USA). However, their measurements were made at depths ranging from 1 to 2.2 m. We know of no field studies that documented the statistical properties of streambed hydraulic conductivity based on a spatially dense dataset collected near the surface–subsurface boundary.

This manuscript reports measurements and statistical analysis of hydraulic conductivity of the sediments within 12.5 cm of the streambed in several pools of an 80 m reach of Indian Creek, an urban stream in Philadelphia, PA, USA.

### 1.1 Site description

Indian Creek begins approximately 1.5 km upstream of our study site in Montgomery County in southeast Pennsylvania, USA. It flows generally north to south before entering 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 Philadelphia. It is expected that this corridor allows some infiltration of stormwater runoff with a concomitant reduction in many pollutants (Hachmöller et al. 1991; Pinay et al. 1992; Paul and Meyer 2001). However, this protection is tempered by the approximately 50% of the 2.5 km<sup>2</sup> drainage area which is covered with impervious surface. More information on the site can be found in Ryan and Boufadel (in press).

The reach where  $K$  measurements were made was 80 m long (Fig. 1) and consists of three pools separated by riffles. The first pool was 15 m long, 4 m wide, and



**Fig. 1** Map of Indian Creek. Riffles are indicated by shading. The locations of  $K$  measurements are shown with *black symbols* (upper layer sediment) and *gray symbols* (lower layer sediment)

as much as 0.6 m deep. It was the narrowest of the pools. The second pool was 25 m long, 8 m wide, and 0.8 m deep at its deepest point. The third pool was 19 m long, 6 m wide, and 0.3 m deep at its deepest point. The bed sediment of this Piedmont stream consists largely of gravel- and cobble-sized particles. However, there is a significant amount of sand- and silt-sized particles that fill the pore spaces between the larger sediment particles and which likely play a major role in controlling hydraulic conductivity.

## 2 Methods

Landon et al. (2001) examined several methods of estimating hydraulic conductivity, including falling and constant head permeameters, slug tests, seepage meters and grain size distribution. The authors concluded that the variability of  $K$  was greater than the variability between the different methods and thus the choice of method may be less important than collecting enough data to describe the spatial variation in  $K$ .

In order to collect enough data to characterize the spatial variation of  $K$  in Indian Creek, we chose to use a falling head slug test conducted in situ with a small diameter portable permeameter. The permeameter was easily installed in the streambed and the test was relatively quick (< 20 min).

The permeameter design was a modification of a design by A. Wörman (personal communication) and consisted of a reservoir made from a 45 cm length of clear PVC pipe (10.2 cm diameter) attached to an iron pipe (ID = 1.5 cm, OD = 1.9 cm). The reservoir was sealed at one end and the seal was drilled and fitted with a threaded pipe nipple with a standard plumbing-type ball valve attached to it. A steel drive point was attached to the bottom of the iron pipe. The pipe was

perforated and screened over a 2.5 cm interval starting a few millimeters from the drive point. The top of the pipe was threaded to accept the standard plumbing-type ball valve. Prior to using the permeameter in situ, the reservoir and ball valve were removed and a protective cap was threaded to the top of the pipe. The pipe was then inserted or pounded into the streambed to the appropriate depth. Once the selected depth was reached, the protective cap was removed and the ball valve and reservoir were attached. The reservoir was filled with stream water while the ball valve was closed. The ball valve was then opened instantly causing a decrease in the water level in the reservoir. The water level in the reservoir was recorded at pre-determined timed intervals. The test was run until the reservoir emptied or 20 min had passed.

The hydraulic conductivity values were calculated using the measured field data and the method of Hvorslev (1951, Fig. 18-G). All  $K$  measurements were made within the pools of the reach. Initial attempts to measure  $K$  in situ indicated that the disturbance of surficial sediment (0–5 cm) during permeameter placement resulted in preferential vertical flowpaths along the permeameter casing. These flowpaths were not observed when the permeameter was installed to a depth of 7.5 cm or greater. The possibility that fluidization of the bed sediments (i.e. “flowing sands”) might be induced by measuring hydraulic conductivity at a depth of 7.5 cm was considered and rejected because the head induced by our test method would be expected to dissipate to within 5% of the background head within a vertical distance equal to only 1.25 cm (Bouwer and Rice 1976, Fig. 3). Thus  $K$  measurements were made at a depth interval of 7.5–10 cm below the streambed and 10–12.5 cm below the streambed. For simplicity of notation, these will be labeled “upper layer sediments” and “lower layer sediments,” respectively.

Measurements in the upper layer sediments were made at 38 locations on ten lateral transects, while those of the lower layer sediments were made at 47 locations on 11 lateral transects (Fig. 1). For each depth, the transects were a minimum of 5 m apart along the stream, and measurements were made at approximately 1 m interval within each transect (i.e. across the stream).

### 3 Results and discussion

The actual measured  $K$  values and their locations are shown in Table 1 (upper layer) and Table 2 (lower layer). The  $K$  values of the upper layer sediments

**Table 1**  $K$  values for upper sediment layer

Northing (m)	Easting (m)	$K$ (cm/s)
57.8	-6.8	1.71E-03
2.1	22.9	3.59E-03
26.9	4.7	3.63E-03
59.7	-8.1	4.52E-03
-1.1	29.6	4.57E-03
30.4	3.2	5.21E-03
18.4	7.7	5.42E-03
0.1	27.4	5.44E-03
58.4	-6.6	6.24E-03
0.9	28.2	6.35E-03
30.2	1.9	7.10E-03
23.7	6.4	7.12E-03
26.4	2.7	7.15E-03
59.6	-4.9	7.18E-03
57.4	-12.2	7.23E-03
23.1	5.6	7.32E-03
24.1	7.4	7.96E-03
22.5	5.5	8.11E-03
58.8	-6.0	8.79E-03
19.6	9.8	8.86E-03
1.5	24.6	8.90E-03
26.8	5.5	8.95E-03
0.3	30.8	8.99E-03
3.1	26.5	9.01E-03
56.6	-7.3	9.78E-03
26.1	1.4	9.82E-03
26.7	3.4	1.05E-02
58.9	-5.8	1.12E-02
1.2	29.0	1.16E-02
2.5	25.4	1.21E-02
60.7	-7.5	1.23E-02
30.7	3.9	1.23E-02
21.9	4.5	1.24E-02
18.0	7.0	1.24E-02
31.1	5.4	1.32E-02
17.3	6.3	1.33E-02
-0.3	30.2	1.51E-02
30.7	4.0	1.69E-02
-2.0	31.6	1.78E-02

ranged from  $1.71 \times 10^{-3}$  to  $1.78 \times 10^{-2}$  cm s<sup>-1</sup>. The mean  $K$  value was  $8.98 \times 10^{-3}$  cm s<sup>-1</sup> and the variance of the natural logarithm of  $K$ ,  $\sigma_{\ln K}^2$ , was equal to 0.22, a relatively small value. The  $K$  values of the lower layer sediments ranged from  $1.16 \times 10^{-4}$  to  $3.69 \times 10^{-2}$  cm s<sup>-1</sup>, with an average of  $5.76 \times 10^{-3}$  cm s<sup>-1</sup> and a  $\sigma_{\ln K}^2$  value of 2.28, an order of magnitude larger than that observed in the upper layer sediments. Analysis of variance indicated that there was no significant difference between replicate  $K$  tests ( $p = 0.01$ ).

Our results are similar to those of other stream studies. Cardenas and Zlotnik (2003) reported values in the range of  $1.7 \times 10^{-4}$  to  $8.6 \times 10^{-2}$  cm s<sup>-1</sup> for Prairie Creek, a sand bed stream in central Nebraska. Duweliuss (1996) reported  $K$  values of  $3.5 \times 10^{-6}$  to  $4.2 \times 10^{-1}$  cm s<sup>-1</sup> for the East Branch Grand Calumet River, an urban river in northern Indiana. Duweliuss

**Table 2** *K* values for the lower sediment layer

Northing (m)	Easting (m)	<i>K</i> (cm/s)
23.7	8.7	1.16E-04
32.5	0.9	1.33E-04
1.7	26.3	1.56E-04
18.5	5.9	2.04E-04
22.6	6.6	2.23E-04
19.7	8.6	2.31E-04
11.6	7.0	3.21E-04
22.4	5.3	4.70E-04
58.6	-7.9	5.36E-04
20.1	9.5	1.14E-03
2.3	27.1	1.39E-03
27.2	3.8	1.43E-03
12.7	9.3	1.67E-03
57.8	-4.7	1.80E-03
0.1	31.8	2.28E-03
60.8	-6.9	2.58E-03
56.0	-6.0	2.66E-03
50.4	-3.6	2.75E-03
33.2	3.5	2.83E-03
27.5	5.2	2.92E-03
6.1	18.3	3.64E-03
26.8	2.0	4.17E-03
7.2	18.9	4.83E-03
32.8	2.4	4.97E-03
8.1	19.3	5.27E-03
58.9	-3.7	5.31E-03
12.2	7.9	5.41E-03
19.0	7.3	6.04E-03
22.5	4.2	6.19E-03
14.6	11.2	6.74E-03
10.6	6.1	6.95E-03
9.9	5.1	7.55E-03
8.8	19.8	7.89E-03
51.3	-2.6	8.03E-03
56.7	-5.3	8.05E-03
3.1	28.2	8.55E-03
-0.6	31.1	8.62E-03
24.1	8.6	9.53E-03
26.7	3.1	9.58E-03
57.4	-8.5	1.14E-02
-1.5	30.5	1.24E-02
53.2	-0.8	1.58E-02
59.6	-7.3	1.60E-02
13.8	10.6	1.72E-02
0.7	25.7	2.41E-02
32.0	-0.4	3.32E-02
52.6	-1.4	3.69E-02

(1996) describes the riverbed as being primarily sand, sandy clay and clay with a surface layer of “soft, fine-grained sediments” whose thickness ranged from 0 to 4 m. Values reported by Calver (2001) were predominantly in the range of  $1 \times 10^{-6}$  to  $1 \times 10^{-1}$  cm s<sup>-1</sup>. Data reported by Gelhar (1993, p. 291) indicate that  $\sigma_{\ln K}^2$  ranges from 0.15 to 4.8 for aquifers.

The *K* values measured 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.

A *t*-test for unpaired data with unequal variance (Haan 1977, p. 172) indicated that the lower layer sediments have a mean  $\ln K$  value that is significantly lower than that of the upper layer sediments ( $\ln K_U = -4.81$ ;  $\ln K_L = -5.76$ ;  $p < 0.0002$ ). This could be due to the trapping of fine sediments in the lower layer while they continuously get flushed in the upper layer. In other words, the finest sediments, which are the easiest to transport, appear to be driven furthest into the streambed, resulting in a lower average  $\ln K$ . The large variance of the lower layer sediments indicates that the pathways and final destinations of the fine sediments are not uniformly distributed in the lower layer.

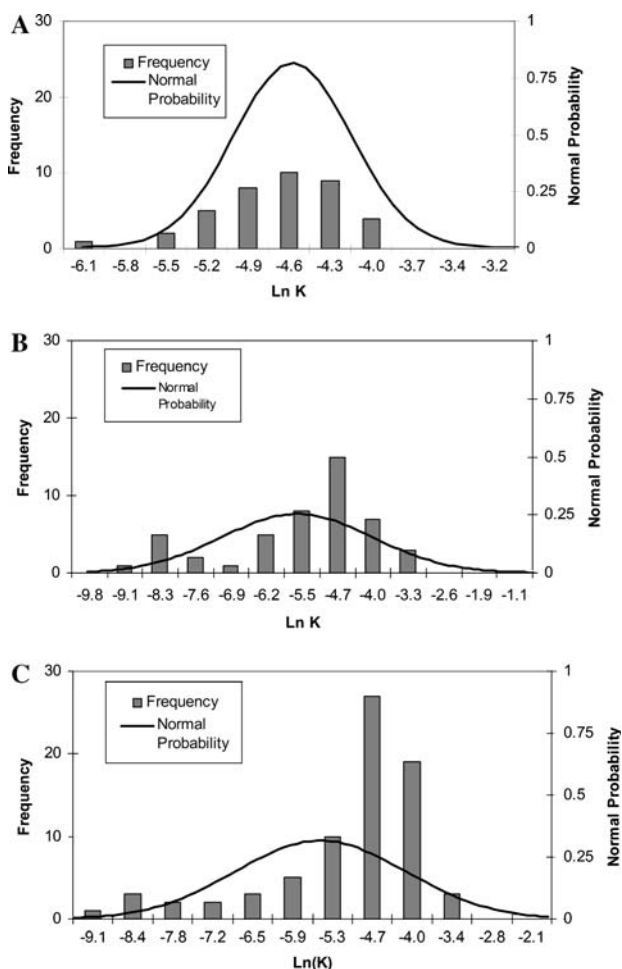
Histograms of the natural log transformed *K* values are shown in Fig. 2 along with the probability distribution corresponding to the Gaussian distribution (with mean and variances equal to those of the data). The measurements at each depth followed the Gaussian distribution fairly well (Fig. 2a, b). However the combined data set appears to deviate from the Gaussian (Fig. 2c). The cumulative distributions on normal probability plots in Fig. 3 indicate that the measurements at each depth (Fig. 3a, b) are distributed normally at the 95% confidence level, but such was not the case for the combined data (Fig. 3c). The Kolmogorov–Smirnov test of normality (Conover 1980, p. 344) further confirmed these findings.

The spatial structure of the data can be assessed using the variogram of  $\ln K$  viz:

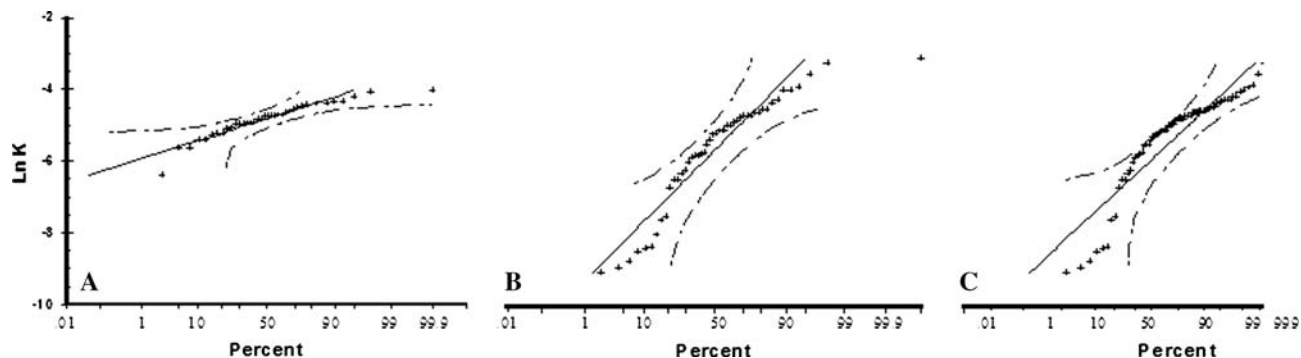
$$\gamma(s, h) = \left\langle |\ln K(s) - \ln K(s + h)|^2 \right\rangle, \quad (1)$$

where *s* indicates the position in the horizontal plane, and *h* is the lag distance between measurements.

For each data set, we applied Eq. 1 across and along the stream. Due to the spatial placements of the transects, the lags of the variograms across the stream were from 1 to 5 m, and those of the variograms along the stream were from 6 to 44 m. The lag values that fell within 15% of the lags reported in Tables 3 and 4 were averaged over the number of pairs (also reported in Tables 3, 4) to obtain the variogram for the corresponding lag. The standard deviations of the variogram



**Fig. 2** Histograms and theoretical Gaussian probability density functions of  $\ln K$ . The upper layer sediment (**a** 7.5–10 cm below the streambed) and the lower layer sediment (**b** 10–12.5 cm below the streambed) conform to the Gaussian distribution fairly well. However, when the upper and lower layer sediments are combined (**c**), they do not conform to the Gaussian distribution



**Fig. 3** Normal probability plots (with 95% confidence intervals) of  $\ln K$  in **a** upper layer sediment; **b** lower layer sediment; and **c** both upper layer and lower layer sediments. The  $\ln K$  values of the individual sediment layers are normally distributed based on

values were also computed, and are reported in Tables 3 and 4. They were generally small, less than 5% of the variogram values.

Figure 4a reports the variogram of the upper layer sediments across the stream; the variogram generally decreased with the lag, with the 3.0 m lag as a sole exception. Recalling that the transects were centered at the thalweg of the stream, the results indicate that the largest difference occurred between the thalweg and the stream banks. The smallest variability occurred between  $K$  values located at about 2.0 m from the thalweg on each side of the stream. The symmetry with respect to the thalweg was not observed in the lower layer (Fig. 4b), where the variogram increased with the lag. Such an upward facing variogram could indicate fractality, observed in aquifers in North America (Boufadel et al. 2000; Tennekoon et al. 2003). This will be investigated in future studies. But, it appears that the physical mechanisms affecting the value of  $K$  differ greatly between the upper layer and lower layer sediments.

Figure 5a shows the variogram of the upper layer sediments along the stream. It indicates that for lags greater than 6.0 m, the variogram for the upper layer sediment increases with lags. However, the possible sill value is larger than the variance of the data, which could indicate periodicity in the data. Therefore, a traditional stationary variogram (Gelhar 1993) model will not be assumed, and the data are treated as random. The same could be hypothesized regarding the lower layer sediments along the stream (Fig. 5b), where the variogram seems to be independent of the lag (i.e. random). The decrease in the variogram value at lags of 22–25 and 44 m could indicate periodicity of

the Kolmogorov–Smirnov test ( $p = 0.05$ ). However, when the  $\ln K$  of the two sediment layers are combined, the Kolmogorov–Smirnov test indicates the data are not normally distributed ( $p = 0.05$ )

**Table 3** Data for across stream variograms

Lag (m)	Upper layer sediments			Lower layer sediments		
	$\gamma$ (h)	SD	Number of replicates	$\gamma$ (h)	SD	Number of replicates
1	0.484	0.153	11	1.96	0.049	17
2	0.430	0.099	20	2.30	0.73	11
3	0.549	0.123	21	1.99	0.37	30
4	0.379	0.061	40	3.01	0.43	50
5	0.279	0.038	56	4.03	0.57	58

**Table 4** Data for along stream variograms

Lag (m)	Upper layer sediments			Lower layer sediments		
	$\gamma$ (h)	SD	Number of replicates	$\gamma$ (h)	SD	Number of replicates
6	0.264	0.043	38	5.07	0.47	95
7	0.343	0.065	29	5.28	0.52	85
8	0.328	0.055	37	4.94	0.48	87
9	0.365	0.357	39	5.30	0.46	110
11	0.378	0.530	18	5.33	0.40	151
12	0.318	0.073	20	5.95	0.43	164
13	0.317	0.073	20	5.69	0.41	167
15	0.359	0.096	15	5.62	0.36	216
22	0.376	0.097	16	4.58	0.28	244
24	0.430	0.079	31	4.45	0.26	257
27	0.437	0.046	90	4.92	0.28	274
35	0.396	0.025	248	4.98	0.27	303
44	0.442	0.041	115	4.32	0.24	280

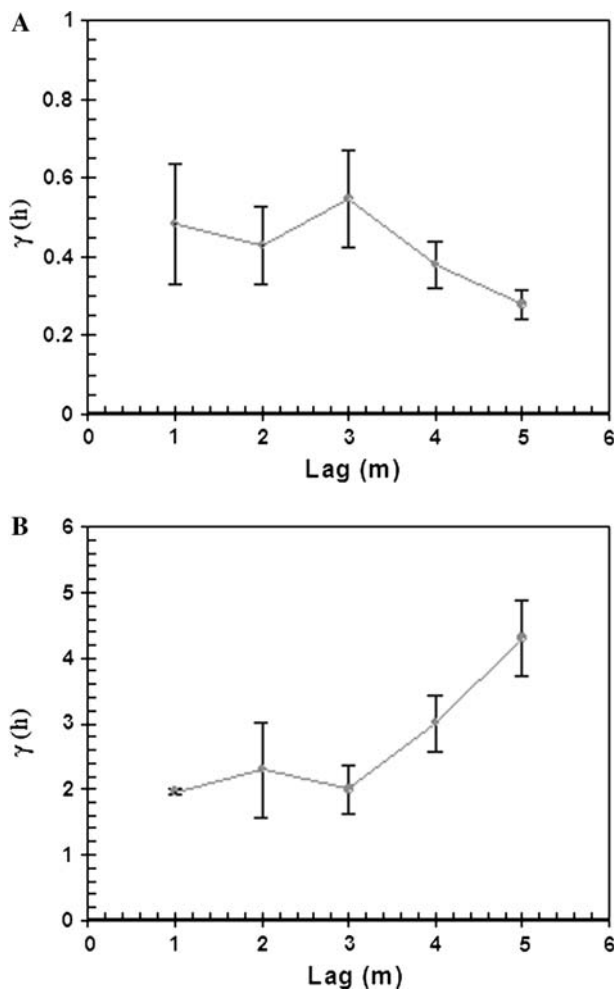
the riffle/pool sequence. It seems more plausible at this stage to assume that the variability along the stream is random, which could be the actual situation pending more data. This is an issue to investigate in a future work.

#### 4 Conclusion

These data are the first to our knowledge to report heterogeneity of near-surface (upper 12.5 cm) stream bed sediments based on data collected in situ. Using a portable falling head permeameter in situ, we estimated the horizontal hydraulic conductivity,  $K$ , of the near-surface streambed sediments at a total of 85 locations encompassing two depth intervals: 7.5–10 and 10–12.5 cm. The mean value of  $\ln K$  was significantly higher in the upper sediment layer compared to the lower sediment layer ( $\ln K_U = -4.81$ ;  $\ln K_L = -5.76$ ;  $p < 0.0002$ ). Additionally, the value of  $\sigma^2_{\ln K}$  was significantly lower in the upper sediment layer compared to the lower sediment layer ( $\sigma^2_{\ln K_U} = 0.22$ ;  $\sigma^2_{\ln K_L} = 2.28$ ;  $p < 0.001$ ). The  $\ln K$  data within each sediment layer was Gaussian. However, the combined data set was not Gaussian based on the Kolmogorov–Smirnov

Test ( $p = 0.05$ ). This suggests that different processes were controlling the bed sediment characteristics in each layer. The effects of these different processes were further highlighted by the variation in the upper and lower layer variogram statistics. We found that symmetry with respect to the thalweg was strongly present in the upper layer sediments but absent in the lower layer sediments. This strong contrast between layers suggests that they should be treated as separated entities in modeling studies.

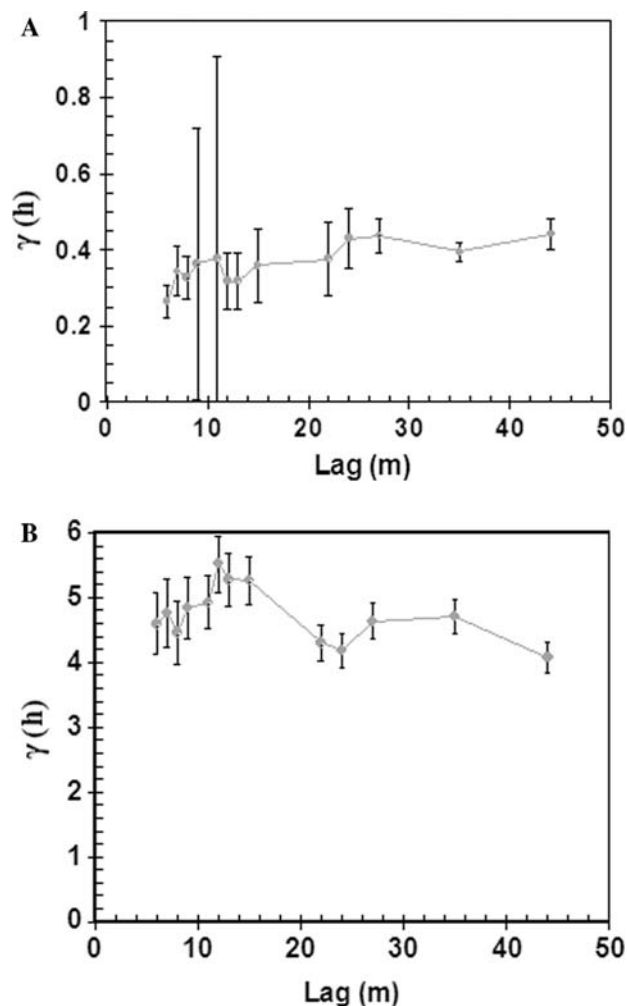
The data agree with the findings of Ryan and Packman (2006) that clay- and silt-sized sediment particles become temporarily trapped in bed sediments and disproportionately reduce the streambed hydraulic conductivity. The across stream variogram of  $\ln K$  of the upper layer sediments suggests that along the stream edge where the velocity and pressure head are low, fine sediments were trapped in the upper layer sediments. In the thalweg, where the velocity and pressure head are usually higher, the fine sediment would have been driven into the lower sediment layer. This may partly explain the lower mean  $K$  and increased heterogeneity of the lower layer sediment. Other factors also likely contributed to the lower mean  $K$  and higher heterogeneity of the lower layer sediment



**Fig. 4** Across stream variograms of  $\ln K$  for **a** upper layer sediment and **b** lower layer sediment. These variograms suggest that the upper layer sediments are symmetrical with respect to the thalweg while the lower layer sediment exhibits fractality

as well, including compression due to burial. Thus, the hydraulic conductivity field of this streambed may be best viewed as multiple thin layers each with its own statistical properties. This layering of the hydraulic conductivity field is expected to play an important role in controlling hyporheic exchange by directing subsurface flowpaths through sediment layers with relatively high values of  $K$  and limiting subsurface flux through sediments with relatively low values of  $K$ . This layering and its potential impact on hyporheic exchange should be accounted for when planning studies (whether field, laboratory or modeling) that involve flowpaths through a three dimensional matrix.

**Acknowledgments** We gratefully acknowledge the assistance of Sandeep Argarwal, Ryan Bechtel and Carl Fenderson for their help with the field work. Funding for this work was provided by the US Department of Agriculture, Grant PENR-2003-01280. However, no official endorsement should be inferred.



**Fig. 5** Along stream variograms of  $\ln K$  for **a** upper layer sediment and **b** lower layer sediment

## References

- Bencala KE, Walters RA (1983) Simulation of solute transport in a mountain pool-and-riffle stream: a transient storage model. *Water Resour Res* 19:718–724
- Boufadel MC, Lu S, Molz FJ, Lavallee D (2000) Multifractal scaling of intrinsic permeability. *Water Resour Res* 36:3211–3222
- Bouwer H, Rice RC (1976) A slug test for determining hydraulic conductivity of unconfined aquifers with completely or partially penetrating wells. *Water Resour Res* 12:423–428
- Calver A (2001) Riverbed permeabilities: information from pooled data. *Ground Water* 39:146–153
- Cardenas MB, Wilson JL, Zlotnik VA (2004) Impact of heterogeneity, bed forms, and stream curvature on subchannel hyporheic exchange. *Water Resour Res* 40:W08307. DOI:10.1029/2004WR003008
- Cardenas MB, Zlotnik VA (2003) Three-dimensional model of modern channel bend deposits. *Water Resour Res* 39:1141. DOI:10.1029/2002WR001383
- Conover WJ (1980) *Practical nonparametric statistics*, 2nd edn. Wiley, New York

- Duwelius RF (1996) Hydraulic conductivity of the streambed, east branch grand calumet river, northern lake county, Indiana. US Geologic Survey WRIR 96-4218
- Elliot AH, Brooks NH (1997a) Transfer of nonsorbing solutes to a streambed with bed forms: theory. *Water Resour Res* 33:123-136
- Elliot AH, Brooks NH (1997b) Transfer of nonsorbing solutes to a streambed with bed forms: laboratory experiments. *Water Resour Res* 33:137-151
- Freeze RA, Cherry JA (1979) *Groundwater*. Prentice-Hall, Englewood Cliffs, NJ
- Ge Y, Boufadel MC (2006) Estimation of solute transport parameters from field studies. *J Hydrol* 323:106-119. DOI:10.1016/j.jhydrol.2005.08.021
- Gelhar LW (1993) *Stochastic subsurface hydrology*. Prentice-Hall, Englewood Cliffs, NJ
- Haan CT (1977) *Statistical methods in hydrology*. The Iowa University Press, Ames, IA
- Hachmöller B, Mathews RA, Brakke DF (1991) Effects of riparian community structure, sediment size, and water quality on the macroinvertebrate communities in a small, suburban stream. *Northwest Sci* 65:125-132
- Harvey JW, Bencala KE (1993) The effect of streambed topography on surface-subsurface water exchange in mountain catchments. *Water Resour Res* 29:89-98
- Hvorslev MJ (1951) Time lag and soil permeability in groundwater observations, Bulletin No. 36. US Army Corps. of Engineers, Waterways Experiment Station, Vicksburg, MS
- Kasahara T, Wondzell SM (2003) Geomorphic controls on hyporheic exchange flow in mountain streams. *Water Resour Res* 39:1005. DOI:10.1029/2002WR001386
- Landon MK, Rus DL, Harvey FE (2001) Comparison of in-stream methods for measuring hydraulic conductivity in sandy streambeds. *Ground Water* 39:870-885
- Leopold LB, Wolman MG, Miller JP (1995) *Fluvial processes in geomorphology*. Dover Publications, New York
- Paul MJ, Meyer JL (2001) Streams in the urban landscape. *Annu Rev Ecol Syst* 32:333-365
- Pinay G, Fabre A, Vervier P, Gazelle F (1992) Control of C, N, P distribution in soils of riparian forests. *Landsc Ecol* 6:121-132
- Ryan RJ, Boufadel MC (in press) Incomplete mixing in a small, urban stream. *J Environ Manage* DOI:10.1016/j.jenvman.2005.09.016
- Ryan RJ, Packman AI (2006) Changes in streambed sediment characteristics and solute transport in the headwaters of valley creek, an urbanizing watershed. *J Hydrol* 323:74-91. DOI: 10.1016/j.jhydrol.2005.06.042
- Salehin M, Packman AI, Paradis M (2004) Hyporheic exchange with heterogeneous streambeds: laboratory experiments and modeling. *Water Resour Res* 40:W11504. DOI:10.1029/2003WR002567
- Springer AE, Petrouson W, Semmens BA (1999) Spatial and temporal variability of hydraulic conductivity in active reattachment bars of the Colorado River, Grand Canyon. *Ground Water* 37:338-344
- Tennekoon L, Boufadel MC, Lavallee D, Weaver J (2003) Multifractal anisotropic scaling of the hydraulic conductivity. *Water Resour Res* 39:1193. DOI:10.1029/2002WR001645