A new method for mapping variability in vertical seepage flux in streambeds

Xunhong Chen · Jinxi Song · Cheng Cheng · Deming Wang · Susan O. Lackey

Abstract A two-step approach was used to measure the flux across the water-sediment interface in river channels. A hollow tube was pressed into the streambed and an in situ sediment column of the streambed was created inside the tube. The hydraulic gradient between the two ends of the sediment column was measured. The vertical hydraulic conductivity of the sediment column was determined using a falling-head permeameter test in the river. Given the availability of the hydraulic gradient and vertical hydraulic conductivity of the streambed, Darcy's law was used to calculate the specific discharge. This approach was applied to the Elkhorn River and one tributary in northeastern Nebraska, USA. The results suggest that the magnitude of the vertical flux varied greatly within a short distance. Furthermore, the flux can change direction from downward to upward between two locations only several meters apart. This spatial pattern of variation probably represents the inflow and outflow within the hyporheic zone, not the regional ambient flow systems. In this study, a thermal infrared camera was also used to detect the discharge locations of groundwater in the streambed. After the hydraulic gradient and the vertical hydraulic conductivity were estimated from the groundwater spring, the discharge rate was calculated.

Keywords Groundwater/surface-water relations · Equipment/field techniques · Hyporheic zone · Infrared camera · USA

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X. Chen (✉) · J. Song · C. Cheng · D. Wang · S. O. Lackey
School of Natural Resources,
University of Nebraska-Lincoln,
3310 Holdrege, Lincoln, NE 68583-0996, USA
e-mail: xchen2@unl.edu
Tel.: +1-402-4720772

J. Song
College of Water Sciences,
Beijing Normal University,
9 Xinjiekouwai St, Beijing, 100875, China

Introduction

Water often exchanges across the stream-sediment interface. The upward or downward direction and magnitude of this vertical flux can affect the chemical, biological, and hydrological processes occurring near the interface of stream-sediments. Seepage meters have been commonly used by researchers to estimate flux across the stream-sediment interfaces in river channels, in lakes, and in submarine environments. Murdoch and Kelly (2003) summarized numerous applications of seepage meters to various water-sediment environments. Landon et al. (2001) combined seepage meter and hydraulic potentiomanometer to determine vertical streambed hydraulic conductivity.

Standard seepage meters consist of a drum and a plastic bag (Lee 1977). The plastic bag is partially filled with water and is connected to the drum with a hose. Some seepage meters also have a gas venting tube, which is attached to the most elevated top part of the drum (Bureau of Rural Sciences 2007). The drum, open at the bottom, must be slowly pressed into the sediment. The plastic bag, often containing a known amount of water, collects water if water flows from the streambed into the stream. Otherwise, the water in the bag will be reduced as stream water enters into the streambed. The flux rate is calculated from the change in volume of water in the bag for the given test period and the surface area of the sediment contained inside the drum during the test. Lee (1977) gave a detailed description for operation of a seepage meter in the field.

Seepage meters are considered cost-effective apparatuses for the determination of flux. However, researchers have identified some problems in the operation of seepage meters and identified possible errors that can be potentially introduced during the field operation (Belanger and Montgomery 1992; Libelo and McIntyre 1994; Isiorho and Meyer 1999; Schincariol and McNeil 2002; Murdoch and Kelly 2003). Improvements to such drum-bag seepage meters have been made by researchers to enhance the quality of flux measurements (Murdoch and Kelly 2003; Taniguchi et al. 2003). In fluvial settings, the hydraulic disturbance presented by the seepage drum and the velocity-head gradient at the flexible surface of the seepage bag can corrupt the seepage measurement

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(Rosenberry 2007). Cey et al. (1998) reported that seepage meters installed in a stream of Canada failed to provide any measurements of water flux into or out of the stream while other methods detected groundwater seepage into the stream.

Mini-piezometers have been installed by researchers in the top part of submerged sediments for determination of the vertical hydraulic gradient in lakebeds (Lee and Cherry 1978; Woessner and Sullivan 1984) and streambeds (Cey et al. 1998; Woessner 2000; Baxter et al. 2003; Conant et al. 2004). This technique provided supplementary information for the analysis of the flux direction but is not able to yield direct measurement of the vertical flux. Kennedy et al. (2007) used piezomanometer to measure the hydraulic head difference between streambed and stream water. Vertical hydraulic conductivity of streambed is determined from a nearby point as well. Water flux was then calculated using Darcy's law.

This paper describes an alternative method for measuring vertical flux in a streambed and presents application examples. This method consists of two steps: determination of vertical hydraulic gradient across the given thickness of a streambed, and the determination of the vertical hydraulic conductivity of this streambed. The hydraulic gradient and hydraulic conductivity are measured at the exactly same point. Darcy's law is then used to calculate the specific discharge at each test location. The method was used at a number of sites in the Elkhorn River and its tributaries of Nebraska, USA to determine vertical flux in the hyporheic zone and in a spring. A thermal infrared camera was used to identify groundwater discharge spring occurring in streambed. This paper presents the results from two test sites: one site is in the Elkhorn River, and the other is in the tributary.

Methods

Determination of vertical hydraulic gradient

If stream water and water within the channel sediments (subsurface flow) exchange, a vertical hydraulic gradient must exist across the depth of the streambed. In this study, plastic tubes were used for determination of vertical hydraulic gradient in the streambed. The tubes were 1.5 m in length and 5 cm in inner diameter. The thickness of the tube wall was about 1 mm, and the tube material is transparent, but durable. At a test point, the tube is erected vertically and is pressed into the submerged streambed sediments to a desired depth. A thin-wall tube can keep the disturbance of the sediments to a minimum during passage.

When the tube passes through the streambed sediments, a column of channel sediment is formed inside the tube (See Fig. 1). If vertical flux exists in the streambed, initially a hydraulic head difference between the top and bottom end of the sediment column appears. However, the head difference will gradually disappear, and the hydraulic heads at both ends will eventually equilibrate. Thus, the water level inside the tube will reflect the hydraulic head at the bottom end of the sediment column.

Since both ends of the tube are open, some surface water remains inside the tube when the tube passes through the stream water. During the passage of the tube through the streambed, there exists friction between sediment and the wall of the tube. This can result in slight compaction of the sediment in some cases. Because of these factors, after the sediment column is formed inside the tube, the tube must stay in the channel for an appropriate length of time to allow the water in the tube to equilibrate. During this period, the hydraulic head inside the tube is frequently checked. For sandy sediment, a period of 1.5 h is needed to reach equilibrium.

The transparent tube materials allow a reading of the hydraulic head difference (Δh) inside and outside the tube. The water level outside of the tube is in the stream stage which can be considered as the hydraulic head of the streambed surface or the top of the sediment column. The length of the sediment column (L_v) is the difference between the total length of the tube and the depth from the top of the tube to the sediment surface inside the tube. The hydraulic gradient (i) can thus be calculated from

$$i = \Delta h / L_\mathrm{v} \tag{1}$$

The mini piezometer techniques (Lee and Cherry 1978; Woessner and Sullivan 1984; Cey et al. 1998) push away streambed sediments during the passage. In contrast, sediments are retained within the tube for the method proposed in this study. The vertical hydraulic conductivity of this sediment column can then be determined. Thus, the hydraulic gradient and vertical hydraulic conductivity of the streambed are evaluated at the same test point.

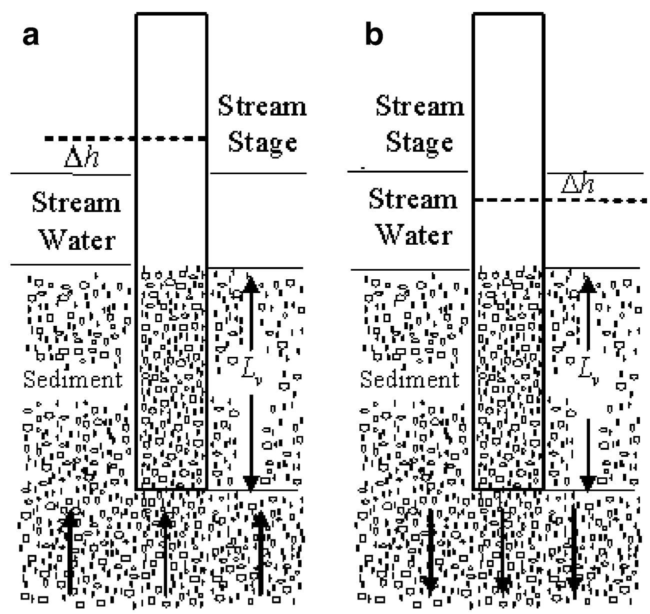

Fig. 1 A sediment column formed inside a hollow tube. After a given time, hydraulic head inside the tube reaches hydraulic equilibrium. The hydraulic head inside the tube is higher than the stream stage **a** upward arrows indicate an outflow, and the hydraulic head inside the tube is lower than the stream stage **b** downward arrows indicate an inflow

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Determination of vertical hydraulic conductivity

The second step is to determine the vertical hydraulic conductivity of the sediment column within the tube. The principles of the falling-head permeameter test are followed. As shown in Fig. 1, the upper part of the tube is empty. Water is continuously poured into the tube until it is full. Then, the permeameter test begins immediately. A series of water level readings at given times are recorded. The vertical hydraulic conductivity K_v is calculated from (Hvorslev 1951)

$$K_v = \frac{\frac{\pi D}{11m} + L_v}{(t_2 - t_1)} \ln(h_1/h_2). \quad (2)$$

where L_v is the length of sediment core in the tube; h_1 and h_2 are hydraulic head inside the tube measured at times t_1 and t_2, respectively, D is the interior diameter of the tube, and $m = \sqrt{K_h/K_v}$. K_h is the horizontal hydraulic conductivity of the streambed sediment around the base of the sediment core. In this equation, h is the height of the added water column inside the test tube. Chen (2000) reported that K_h values can be several times larger than K_v values in the upper part of the streambed sediments.

In Eq. 2, K_h is an unknown variable. Theoretically, an inverse approach can be developed to simultaneously calculate K_h and K_v from a series of hydraulic head readings and their associated times. However, h is very insensitive to K_h in this equation, and its sensitivity coefficient with respect to K_h is small. Inverse computations often lead to non-convergence. To minimize the uncertainty resulting from K_h, a longer sediment column, compared to the tube diameter, can be used. For example, if L_v=10D, a choice of m=1 or m=10 in Eq. 2 gives a very small difference in the resultant K_v.

Any two pairs of h~t readings can be used to calculate a K_v value. However, a slight difference in K_v can occur from using different pairs of data. This can potentially be caused by a measurement error in the hydraulic reading. Thus, a regression procedure is used to calculate K_v from more than two pairs of h~t readings.

Estimation of specific discharge

Given the values of hydraulic gradient i, and vertical hydraulic conductivity K_v at each test location, the specific discharge is calculated using Darcy's law

$$q = -i \times K_v \quad (3)$$

Multiplication of q by the surface area of the streambed gives the volumetric rate of subsurface flow discharge or the volumetric rate of stream water infiltration.

Results

The method was applied in the Elkhorn River, northeastern Nebraska. Results from two sites are reported; the Neligh site is on the main stem of the Elkhorn River, and the Hadar site is on a tributary of this river. Figure 2 shows the study locations in the Elkhorn River.

Vertical flux at the Neligh site

The measurements at the Neligh site were conducted on 5 and 6 August 2007. The US Geological Survey (USGS) gauge near Ewing, about 30 km upstream, indicated a flow of 0.85 m^3/s. The average daily discharge rate for 6 August during the period from 1947–2006 was 2.1 m^3/s at this gauge. The USGS gauge near Norfolk, about 50 km downstream was 3.3 m^3/s on 6 August, but the average daily flow rate was 8.8 m^3/s for the period 1896–2006. This region had experienced a dry period for at least a month in 2007; so during testing, the river stage was below the daily historical average discharge rate. Because of this low flow period, the effect of return flow from bank storage (Chen and Chen 2003) was expected to be minimal during the tests. Figure 3 shows the orientation of 12 measurement locations within the river channel. During the test, only the northern half of the channel was submerged. The southern part was exposed because of the low stream stage.

Table 1 summarizes the results of Δh, L_v, K_v, q, and water depth of the river for each test location. Among the

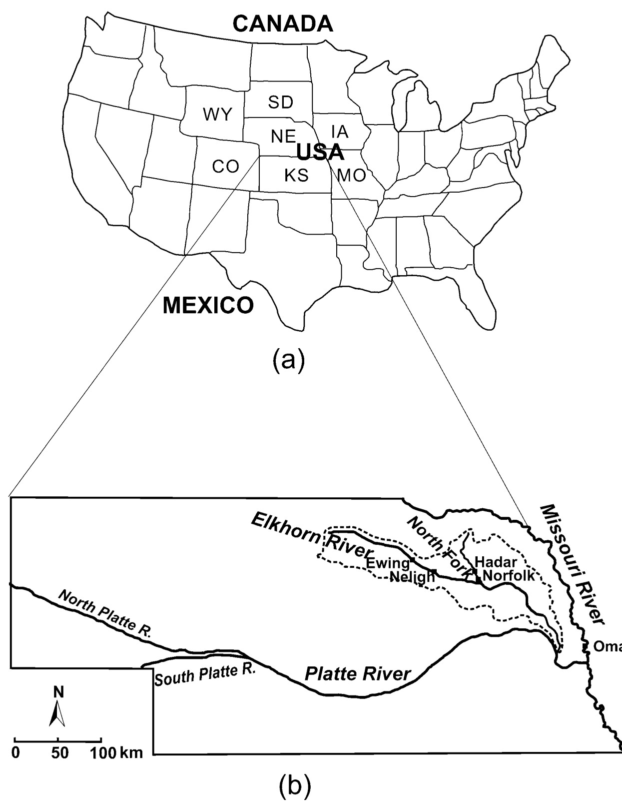

Fig. 2 Map of **a** states in USA (*CO* Colorado; *IA* Iowa; *KS* Kansas; *MO* Missouri; *NE* Nebraska; *SD* South Dakota; *WY* Wyoming), **b** Nebraska, showing the Elkhorn River and the North Fork tributary where vertical flux measurements in streambeds were conducted

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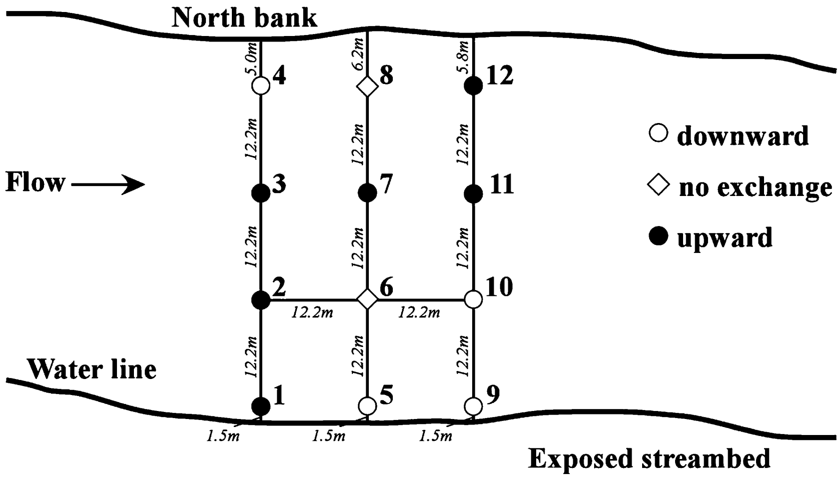

Fig. 3 Orientation and distances between the 12 measurement locations at the Neligh test site on the Elkhorn River

12 measurements, four locations had downward hydraulic gradient, indicating infiltration of river water into the streambed, six locations show upward hydraulic gradient and supply water to the river, and the remaining two locations had no hydraulic gradient. Note that neighboring two locations were only 12.2 m apart, yet the hydraulic gradient can change from upward to downward. Among the locations with upward flux, locations 1, 7, and 12 had a larger discharge rate. Among the locations with a downward flux, location 9 had a large infiltration rate.

The values of K_v among the 12 locations vary from 16 to almost 39.4 m/day, with an average value of 28.1 m/day. Additionally, 32 permeameter tests at the study site show K_v ranging from 0.96 to 42.3 m/day, with an average of 25.7 m/day. If the test with K_v=0.96 m/day is excluded, the K_v values of the other 31 tests range from 11.5 to 42.3 m/day with an average of 26.4 m/day. The K_v= 0.96 m/day result was probably affected by a small silt or very fine sand lens. These relatively large values of vertical hydraulic conductivity indicate that the exchange of stream water and subsurface flow can occur quickly if a vertical hydraulic gradient exits across the streambed. The sediments are relatively uniform and permeable. Sieving analysis of sediment samples from the streambed indicated dominant medium-grain sand.

From Eq. 2, the approximate time that will allow the hydraulic head to reach a near equilibrium inside the tube can be estimated. Assume the following parameter values: K_v=25 m/day, L_v=50 cm, m=3, and D=5 cm and assume that the hydraulic heads at the two ends of the sediment column are equal. After introducing a 2-cm water column to the tube, it takes only 1.45 h for a 1.9-cm water column to infiltrate from the tube into the subsurface. At this study site, the average K_v value is greater than 25 m/day. The time of 1.5 h for the hydraulic head to reach equilibrium seemed to be sufficient during the measurement for hydraulic gradient across the streambed.

Near the edge of the southern riverbank, a very small brook flowed on the exposed streambed. The water was apparently the mixture of spring discharge and some surface water diverted from the main river channel. This part of the streambed would have been submerged if the stream stage were at the normal levels. A thermal infrared camera was used to determine the distribution of water temperature within the brook in the early morning (6:30 a.m.) of 6 August 2007. At this time, the sky was

Table 1 Hydraulic variables of the Elkhorn River at the Neligh test site, Nebraska

Test number	Hydraulic head difference Δh (cm)	Thickness of the sediment column in pipe L_v (cm)	Vertical hydraulic conductivity K_V (m/day)	Specific discharge $Q = K_V \frac{\Delta h}{L_V}$ (cm/day)	Water depth of river (cm)
1	1.0	56.9	36.76	64.6	11.0
2	0.3	62.5	30.67	14.7	15.6
3	0.6	63.8	30.18	28.4	18.8
4	–0.2	74.7	16.00	–4.3	37.0
5	–0.6	63.8	29.58	–27.8	23.8
6	0.0	66.3	31.02	0.0	14.4
7	1.1	63.8	27.90	48.1	14.2
8	0.0	69.1	16.49	0.0	23.9
9	–0.6	61.9	39.39	–38.2	24.4
10	–0.2	65.0	26.35	–8.1	8.0
11	0.5	62.5	29.21	23.4	8.6
12	1.1	59.9	23.84	43.8	44.4
GW1[*a]	1.3	45.2	16.97	48.8	3.0
GW2[*a]	1.9	27.5	13.67	94.4	1.3

[*a] Measurement at spring site

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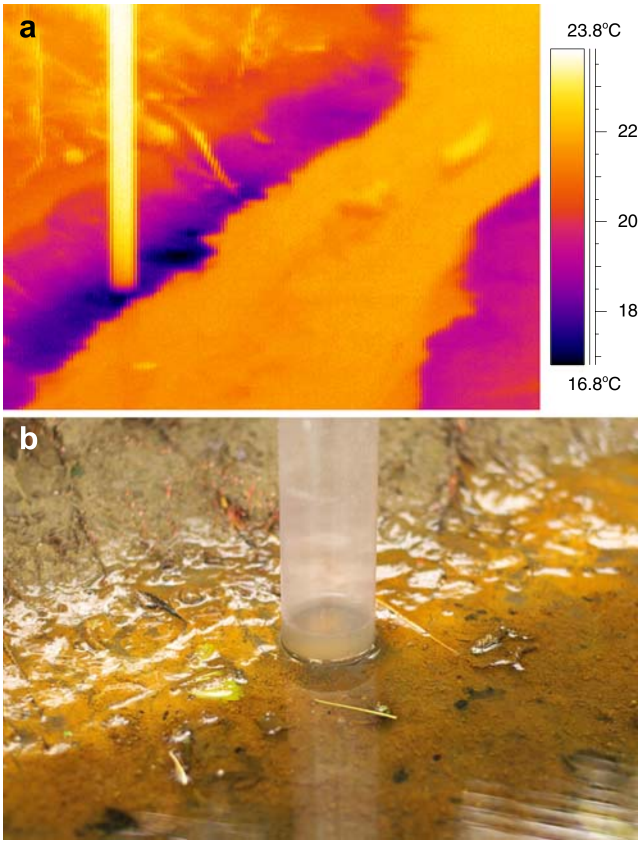

Fig. 4 Thermal infrared picture **a** showing a spring in a streambed and temperature variations of a brook where spring water mixed with river water; visible picture **b** of this spring showing the water level inside the test tube was higher than the water level in the brook. In the thermal infrared picture, the highest temperature was near the upper part of the test tube. The diameter of the tube in both pictures was 5 cm

cloudy and the sun had not risen. Thus, there was no reflection effect on the water surface. The thermal pictures identified a number of locations along the brook having lower water temperatures. Figure 4a is a thermal picture of the brook showing variations of water temperature from 16.8 to about 23°C. After the tube was pressed into the sediments of the spring, it raised the temperature of the surface water around the tube slightly. The water temperature in the main channel of the river was about 23.8°C at this time.

The location with the lowest water temperature was treated as the center part of a groundwater spring. Two measurements were conducted at two springs, about 3 m apart. The results of Δh, L_v, K_v, and q are shown in Table 1 (GW1 and GW2). The discharge rates were 94.4 and 48.8 cm/day, respectively. Figure 4b shows a picture of the spring and the test tube. The water level inside the tube is higher than that in the brook. The yellowish deposits are probably iron compounds brought up to the sediment surface by the spring water.

Vertical flux at the Hadar site

Measurements at the Hadar site were conducted on 20 August 2007. There were nine measurements taken in total. Figure 5 shows the orientation of the grid. The results are presented in Table 2. Among the nine closely spaced test locations, both upward and downward flux was encountered. The upward flux tended to concentrate on the west side of the channel. Among the four locations with downward flux, the rates for two locations were as high as 120.7 and 104.4 cm/day. The water depth for the two locations was 42.8 and 16.1 cm, respectively.

The average thickness of the sediment columns for the seepage tests was 64.2 cm at the Neligh site and 56.2 cm at the Hadar site. These depths might just be partial thicknesses of the hyporheic zone at the study sites because both downward and upward flows were observed for a short distance apart. The actual thicknesses of the hyporheic zone at these sites are unknown, but should be greater than these thicknesses of the sediment columns.

Discussion

No single definition for the hyporheic zone has been adopted by researchers from different disciplines (Woessner 2000); however, it is clear that stream water readily exchanges with fluid flow in the hyporheic zone, which includes the sediments below the channel surface. At both test sites, a strong spatial variation of inflow and outflow in the hyporheic zone was observed. Computer simulation results by Cardenas et al. (2004) suggested that heterogeneous streambed deposits (horizontal streambed hydraulic conductivity) affect the magnitude of inflow and outflow in the hyporheic zone. The vertical hydraulic conductivity measurements from the Neligh and Hadar sites indicate that K_v values of the streambed vary from one to another test location. However, the spatial variation in K_v for the streambed at the Neligh site is not as great as the variation

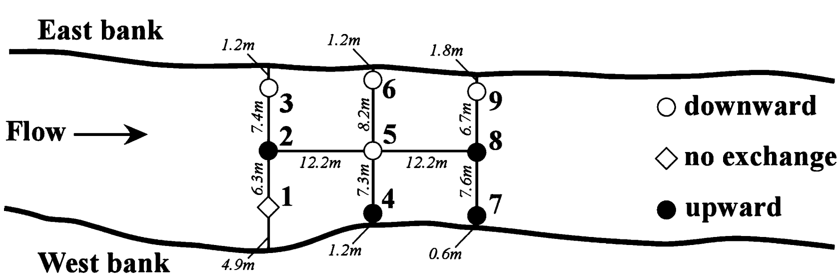

Fig. 5 Orientation and distances between the nine measurement locations at the Hadar test site in the North Fork of the Elkhorn River

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Table 2 Hydraulic variables at the Hadar test site, Nebraska

Test number	Hydraulic head difference Δh (cm)	Thickness of the sediment column in the pipe L_v (cm)	Vertical hydraulic conductivity K_V (m/day)	Specific discharge $Q = K_V \frac{\Delta h}{L_V}$ (cm/day)	Water depth of river (cm)
1	0.0	53.8	2.64	0.0	16.1
2	0.6	56.3	28.85	30.7	23.0
3	–3.1	53.8	20.94	–120.7	42.8
4	0.6	60.3	14.21	14.1	18.5
5	–3.3	60.0	18.98	–104.4	16.1
6	–0.6	54.7	0.07	–0.1	55.6
7	0.8	55.3	18.33	26.5	11.3
8	0.3	56.9	19.80	10.4	26.3
9	–4.4	54.4	0.06	–0.5	44.4

at the Hadar site. At both sites, no clear pattern between flux direction and K_v values was observed. Cey et al. (1998) determined the horizontal hydraulic conductivity of streambed sediments positioned 1 m below the water-sediment interface for the calculation of the vertical flux of water in the streambed. According to Chen (2000), the horizontal hydraulic conductivity of streambeds was larger than the vertical hydraulic conductivity. Determination of the vertical hydraulic conductivity of the streambed sediments is needed for the calculation of vertical flux in streambeds.

Vertical hydraulic gradient is another variable that is heterogeneous in space. Both heterogeneous hydraulic conductivity and hydraulic gradient lead to a complex pattern of inflow and outflow in the hyporheic zone because the two variables affect the calculation of the vertical flux. Cey et al. (1998) reported that the vertical hydraulic gradient for a reach of 450 m varied from zero to 5.5. Conant (2004) showed the variations of both vertical hydraulic conductivity and vertical hydraulic gradient of the streambeds in the Pine River in Angus, Canada. As a result, the discharge rate varied from one to another measurement within a 55-m river reach. While most of the measurements showed outflow, some of the measurements indicated inflow in the study by Conant (2004). Kennedy et al. (2007) presented another example of spatial variations in vertical flux in a stream reach of 262.5 m in North Carolina, USA.

Current-bedform in streams can induce water exchange in the subsurface sediments or the hyporheic zone. According to numerical simulation by Cardenas and Wilson (2006), groundwater discharge reduces the spatial extent of the water exchange zone below the stream. Furthermore, a high groundwater discharge may completely prevent the development of the current-bedform induced flow. The inflow and outflow measurements in the streambeds at the Neligh and Hadar sites suggest that water exchange was active at the time of this investigation. Thus, this could imply that the groundwater discharge rate to the study reach was probably not sufficiently high to prevent from the inflow into the streambeds. Actually, August is within the irrigation season in Nebraska. In the Elkhorn River watershed, a very large number of irrigation wells are constructed to pump groundwater for irrigation of crops. Pumping wells adjacent to rivers can intercept baseflow that otherwise would discharge to the rivers and induce infiltration of streamflow as well (Chen and Shu 2002).

With the method proposed in this study, it can take more than 1–2 h for the water inside the tube to reach an equilibrium even in permeable streambeds. For a lower permeability streambed, an even much longer time is required to reach equilibrium. Fortunately, these tubes are inexpensive and there is no need to assemble them. For a given study site, a large quantity of tubes can be inserted into the streambed at numerous locations in a relatively short time. Thus, the hydraulic head difference can be recorded from a large number of tests for a given time.

When the hollow tube is pressed into the streambed, it can cause compaction of the streambed column inside the tube. This compaction can be reduced by using tubes with thin walls (the wall of the tube is 1 mm in this study). If the leading end of the tube is beveled, disturbance and compaction of the sediments inside the tube (Landon et al. 2001) can be minimized. Compaction of the streambed can be readily determined by measuring the elevation of streambeds inside and outside the tube. If a large compaction of the streambed indeed occurred, the test was performed at another nearby location.

Conclusions

Compared to the traditional one used by many researchers, a better alternative method can be used to determine the vertical flux across the water-sediment interface where the depth of the water is relatively small and allows for wading. Seepage meters directly measure the volume of flux but do not produce the information about the hydraulic gradient and vertical hydraulic conductivity. The proposed method avoids the use of the flexible bag used in conventional seepage meters and produces information about hydraulic gradients and vertical hydraulic conductivities of streambeds at the same testing point. Vertical hydraulic conductivity K_v of a streambed is a key parameter in modeling the inflow and outflow and contaminant exchanges across the water-sediment interfacial zone. This study indicated that both K_v and the vertical hydraulic gradient in the streambed vary from one to another location. Thus, measurements of the two

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variables from the same location will improve the estimation of vertical flux in the streambed. The hydraulic conductivities of streambeds are generally anisotropic (Chen 2000). Determination of the vertical hydraulic conductivity is needed for the calculation of vertical flux.

Flux across the stream-water interface varies spatially within a short distance, scales of only a few meters apart. This difference not only exists in the flux magnitude, but the direction can also vary within a short distance. Downward flux can co-exist with upward flux along the channel, as well as across the channel. This up- and down-seepage may reflect in part the effect of the flow dynamics in the hyporheic zone, and not purely the regional groundwater system.

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