Nonuniform Groundwater Discharge across a Streambed: Heat as a Tracer

Time series analysis of continuous streamed temperature during a period of 47 d revealed that discharge to a stream is nonuniform, with strongly increasing vertical fluxes throughout the top 20 cm of the streambed–aquifer interface. An analytical solution to the transient heat transport equation was used to analyze several pairs of observed amplitude damping with depth. A nonuniform pattern in discharge across the stream width was also observed, which could have been caused by lateral or horizontal flow. Head measurements in a meadow area and below the stream showed strong convergence of flow near the streambed. Seepage meter measurements in the middle of the stream often resulted in highly variable flux estimates, which could have been caused by hyporheic flow due to the presence of a gravel layer. Discharge and recharge to the stream at the bank near the meadow was relatively steady throughout the period. On the other hand, discharge to the stream at the opposite bank near a steep hillslope decreased significantly toward the end of the period (early June), which was attributed to a drop in the water table on this side of the stream. The results from the time series analysis were compared with seepage meter measurements and the results from a steady-state analytical solution to the heat transport equation. The different methods agreed on the pattern of discharge across the stream width, and the mean values during the studied period generally agreed well but with different ranges.

The European Union (EU) Water Framework Directive requires EU member countries to administer their water resources in an integrated fashion, where the ecological statuses of different water bodies are treated together. In this respect, groundwater–surface water interactions play a central role (Dahl et al., 2007). In the Skjern River Catchment (SRC) in Denmark, groundwater–stream and groundwater–lake interactions have been investigated at various scales. For example, Kidmose et al. (2011) modeled lake-scale seepage patterns at a flow-through lake in the upper parts of the SRC to see how seepage is affected by lake bathymetry and the depositional character of the lake bed sediments. On-land and offshore geophysical explorations have helped to improve the inclusion of the lake in the regional groundwater flow model, improving the model’s ability to simulate the observed seepage patterns.

Heat has been used in many studies as a natural tracer to assess groundwater–surface water exchange at various scales. The seasonal changes in temperature contrasts between the surface water and deeper groundwater make it possible to quantify the direction and magnitude of the exchange (Constantz, 2008). At the kilometer scale, distributed temperature sensing has been used to study groundwater–lake and groundwater–stream interactions (Selker et al., 2006a,b). At the meter scale, point or profile measurements of temperature have been popular for some time (Constantz, 2008). A number of investigations have relied on synoptic measurements, where point measurements of temperature either at a given depth below the streambed (Conant, 2004; Schmidt et al., 2007) or at several depths (Schmidt et al., 2006; Anibas et al., 2009) were used to estimate streambed water exchange. The measurements were usually obtained by pushing a temperature probe into the bed sediments and recording the temperatures after sufficient equilibration time (minutes) or measuring temperature directly in wells (Duque et al., 2010). An advantage of this approach is that several hundreds of meters of a stream can be mapped in a relatively short time to reveal spatial heterogeneity in water exchange (Conant, 2004; Schmidt et al., 2007). Other investigations have relied on time series data at a single depth (Silliman and Booth, 1993; Constantz et al., 2001) or several depths below the streambed (Constantz et al., 2002; Keery et al., 2007). The advantage of this approach
is that temporal changes in water exchange can be quantified along a given stream reach (e.g., Keery et al., 2007) or used to determine stream flow frequency and duration for ephemeral streams (Constantz et al., 2001; Duque et al., 2010).

Analytical one-dimensional models solving the heat transport equation have been popular to interpret temperature data at the scale of a streambed because of their simplicity and because the heat flow parameters vary much less across a range of sediment textures than, for example, hydraulic conductivity (Constantz and Stonestrom, 2003). In some cases, more sophisticated one- or two-dimensional numerical models have been used (Constantz, 2008), but the use of numerical models generally requires more information to be operational and therefore can be difficult to apply. Analytical solutions, however, are limited to strictly steady, uniform vertical flow. The analytical solutions are furthermore often based on an assumption of steady-state heat transport (Schmidt et al., 2006; Anibas et al., 2009), where fixed surface water and groundwater temperatures are specified at the two boundaries, using the solution derived by Bredehoef and Papaopoulos (1965). Schornberg et al. (2010) demonstrated that the one-dimensional analytical solutions give reasonable results for discharge (in the following, we use the term discharge to indicate a positive and upward flow of groundwater to the stream and recharge to indicate a negative and downward flow of water from the stream) rates >0.1 m d\(^{-1}\) but only during periods with the greatest contrast in stream water and groundwater temperatures, i.e., typically during winter and summer. Hatch et al. (2006) developed a time series analysis based on an analytical transient solution previously given by Stallman (1965) and Goto et al. (2005). They showed that it is possible to relate exchange fluxes to amplitude damping and phase changes as a function of depth below the streambed. Their analysis was still based on the assumption of one-dimensional steady uniform vertical flow but had the advantage that amplitude damping and phase changes could be computed from sensor pairs located at different depths. Therefore it was not necessary to know the stream temperature, which often is affected by frequencies different from the diurnal signal, e.g., annual frequencies, but also high-frequency noise due to shading by clouds, the effects of vegetation, etc. (Keery et al., 2007). Keery et al. (2007) extended this method and developed a dynamic harmonic regression technique, where the damping and phase change of the temperature signal is directly computed.

In many systems with significant aquifer–stream interaction, the exchange of water between the two systems cannot always be regarded as strictly vertical. A lateral or horizontal flow component will often be present, making the interpretation with the analytical solutions more difficult. Lateral or horizontal flow can occur on various scales: exchange with shallow groundwater systems at the stream banks (Modica et al., 1997, 1998), from hyporheic flow with stream water driven into the streambeds and re-entering the stream farther downstream (Hannah et al., 2009), or from geologic heterogeneities in the streambed or underlying aquifer (Schornberg et al., 2010).

The objectives of this work were (i) to evaluate if discharge to a small stream reach can be regarded as uniform across the streambed by performing a time series analysis of streambed temperature data, and (ii) to evaluate the error introduced by simulating temperature profiles assuming uniform steady-state discharge. This work builds on the time series analysis presented by Hatch et al. (2006) and Keery et al. (2007) but extends their investigations by using multiple sensor pairs below a streambed. Hatch et al. (2006) collected temperature time series at depths of 0.1 and 0.4 m and found temporal changes in recharge fluxes of −3.2 to −0.8 m d\(^{-1}\) based on one pair of sensors. They were not able to use temperature data from the deep sensor and, thus, only had recharge fluxes at one depth. Keery et al. (2007) installed sensors at depths of 0.1 and 0.2 m, but only the shallowest sensor at 0.1 m was used together with the stream temperature. In both of these studies it was therefore not possible to investigate changes in flux with depth. The purpose of our work has therefore been to obtain high-resolution time series data on temperature mainly in the upper 0 to 20 cm of a streambed, allowing us to quantify the change in exchange fluxes across a streambed. Three temperature stations were installed in the stream to depths of approximately 2.5 m below the sediment–water interface (SWI), with a total of 14 sensors at each station and with three to six sensor pairs in the upper 20 cm near the SWI. Two stations were located near the banks and one in the middle of the stream, which furthermore allowed us to investigate the differences in exchange fluxes across the stream width.

**Field Site**

The Holtum Stream field site is located in Jutland, Denmark, on a glacial floodplain valley west of the last glacial maximum of the Weichsel ice period. The western part of Jutland was the only ice-free part of Scandinavia during the last glacial period. The present landscape is therefore dominated by floodplain sediments on top of deposits from the previous Saale ice period. The contributing catchment is around 126 km\(^2\), forming the upper part of the Skjern River basin (Fig. 1). The mean annual precipitation is 888 mm yr\(^{-1}\) (1970–2007) and mean annual air temperature is around 9°C (2008).

Stream flow is approximately east to west at the field site, with an annual mean of 1.3 m\(^3\) s\(^{-1}\) in 2005 to 2007. Stream temperatures typically range from 16°C in the summer to 2°C in the winter. A low-lying meadow area is found to the north of the stream. A man-made hill with steep slopes and a tree plantation (cut down in the spring of 2008) borders the southern stream bank. Streambed sediments are dominated by sand, sometimes with a layer of coarse boulders in the middle of the stream. Groundwater levels in the meadow area are higher than the stream stage and groundwater flow is toward the stream. The meadow area is always wet, with free-standing water. A
small pond is found in the middle and often a small canal connects the pond directly to the stream (seen as a dark shadow from the pond past Piezometer 16 and to the stream, Fig. 1).

Field Methods

Synoptic measurements of hydraulic heads and temperature profiles were conducted during a period of about 9 mo. In addition to this, continuous temperature distributions in the streambed were monitored every 20 min during a 53-d period in the spring of 2008, with a significant rate of change in temperature of the stream water (from 8 to 14°C). Seepage meter measurements were conducted 26 d after this period to get an independent estimate of discharge to the stream.

Hydraulic Characterization of Discharge to Stream

Piezometer Network

Twenty-one piezometers were installed in the early spring of 2008 (Fig. 1). The piezometers are constructed of metal pipes (2-cm i.d.) equipped with a 9-cm-long metal screen. The piezometers were pushed 1 to 3.7 m into the subsurface using a pneumatic hammer. Piezometers P1, P2, P3, and P13 were installed in the streambed, the rest in the meadow area. All piezometers were leveled using a standard total station (CTS-1, Topcon Positioning Systems, Livermore, CA) relative to a fixed point above sea level (measured using a Trimble R8 global positioning system, Trimble Navigation Ltd, Sunnyvale, CA). Right after installation, each piezometer was pumped clean for finer sediments that may have been pushed through the screen. The hydraulic heads were measured during a period of about 9 mo.

Seepage Meter

Discharge to the stream was measured with seepage meters at three locations in the stream (upstream of temperature stations T1, T2 and T3; see below) on 2 July 2008 (Fig. 2). Each seepage meter was constructed from the bottom part of a steel drum, with a diameter of 57 cm (Lee, 1977). The seepage meters were assembled with fitting valves, 1.5-cm i.d., designed for reducing the effect of friction (Rosenberry, 2005). A 4-L plastic collection bag was attached to the valve. The collection bag was covered with a rigid plastic casing to minimize the effect of stream-flow-induced seepage from movement of the bag (Murdoch and Kelly, 2003; Rosenberry, 2008). The seepage meters were installed as near as possible to the individual temperature stations (and thus called S1, S2, and S3; see below) without influencing the temperature measurement itself. The seepage meters were left to equilibrate for 24 h before any measurements were taken. In total, 12, 22, and 34 measurements were performed at S1, S2, and S3, respectively.

Temperature Characterization of Discharge to Stream

Several synoptic temperature measurements were taken with a Model TP 62 temperature probe similar to that used by Schmidt et al. (2006). The probe has a length of 151 cm, with six temperature sensors distributed unevenly across the lower 50 cm. The sensors have an equilibration time of about 10 min. Some conduction of heat through the metal probe is to be expected and the measurements are therefore best at demonstrating qualitatively how the temperature pattern changes below a stream during a season.

Three permanent temperature stations (T1, T2, and T3) were established in April 2008, Fig. 2. Each temperature station consisted of two separate pipes: a 4.2-m-long metal pipe, which was constructed and installed in the same way as the piezometers, and a plastic pipe carefully pushed 0.65 m into the streambed. A second plastic pipe with four temperature sensors in the lower 1.6 m of the pipe was installed inside the metal pipe. The metal pipe would also lead to some conduction of heat, but the temperature data from these depths were nearly constant and therefore not used in the time series analysis (see below). The temperature data from the deepest sensor (2.95 m under the SW1), however, was used to specify the ambient groundwater temperature as a lower boundary condition in the steady-state heat transport solution (see below).
There were 10 temperature sensors in the plastic pipe positioned directly into the streambed. The first and most shallow sensor measured the stream temperature and the rest measured streambed temperatures from right under the streambed surface and down to a depth of approximately 0.65 m (depending on scouring and sedimentation, however). All 14 sensors in a station were connected to a multiplexer located at the top of the plastic pipe positioned directly into the streambed (Fig. 2). All data from the three multiplexers were sent to a datalogger located in a box in the meadow (Fig. 2).

The temperature sensors were made of thermocouples having an absolute accuracy of about ±1°C. All the sensors were made from the same coil of cable and it was assumed that the accuracy of all the sensors was the same. Thermocouples have the advantage of being easy to deploy because they can be fabricated as needed from a thermocouple cable. Each thermocouple end installed in the plastic pipes was welded to a copper ring, which was placed in a groove carved on the outside of the plastic pipe and then covered by a thin layer of plastic insulation to resist water. This approach was used in an attempt to ensure good contact between the individual temperature sensors and the streambed sediments. Thermocouples operate on the principle that dissimilar metals in a circuit develop a voltage proportional to the temperature difference between their connections (Stonestrom and Constantz, 2003).

A disadvantage of using thermocouples is that a known reference temperature inside the box with the multiplexer has to be measured. The reference temperature was measured with a thermistor with a calibrated accuracy of 0.2°C. This can lead to small errors, which will influence the temperature measurements for the entire streambed profile (J. Constantz, personal communication, 2008).

To reduce this error, the multiplexer box and the reference temperature measurement point were insulated with rubber foam. The temperature records were still corrected, however, for the influence of any small fluctuations inside the multiplexer box, which could be the result of small temperature differences between the reference point and the individual connections of the thermocouple cable ends in the multiplexer box (J. Constantz and R. Niswonger, personal communication, 2008). Temperatures were measured at each sensor every minute and the mean temperature during a 20-min period was stored in the datalogger. The period of temperature measurements was from 21 April to 6 June 2008.

**Analytical Methods**

**Theory**

Conduction and convection of heat in the upper part of the streambed is assumed to be governed by the one-dimensional conduction–convection equation (Stallman, 1965; Goto et al., 2005; Hatch et al., 2006):

\[
\frac{\partial T}{\partial t} = \alpha_e \frac{\partial^2 T}{\partial z^2} - \frac{\nu v_e}{\gamma} \frac{\partial T}{\partial z} \tag{1}
\]

where \(T(z,t)\) is temperature (°C), which varies with time \(t\) (s) and depth \(z\) (m), \(\alpha_e\) is the effective thermal diffusivity (m² s⁻¹), \(\gamma\) is the ratio of the volumetric heat capacity of the streambed to the fluid (= \(\rho c/\rho c_f\)), \(\rho c\) is the volumetric heat capacity of the saturated sediment–fluid system \(J \text{ m}^{-3} \text{ °C}^{-1}\), \(\rho\) is the density of the sediment–fluid system \(\text{kg m}^{-3}\), \(c\) is the specific heat capacity of the sediment–fluid system \(J \text{ kg}^{-1} \text{ °C}^{-1}\), \(\rho c_f\) is the volumetric heat capacity of the fluid \(J \text{ m}^{-3} \text{ °C}^{-1}\), \(\rho_f\) is the density of the fluid \(\text{kg m}^{-3}\), and \(\omega\) is the specific heat capacity of the fluid \(J \text{ kg}^{-1} \text{ °C}^{-1}\), \(n\) is porosity, and \(v_e\) is the pore water velocity (m s⁻¹).

The effective thermal diffusivity is defined as (Ingebritsen and Sanford, 1998)

\[
\alpha_e = \frac{\kappa_e}{\rho c_f} \tag{2}
\]

where \(\kappa_e\) is the effective thermal conductivity (W m⁻¹ °C⁻¹).

Thermal dispersion is ignored because the thermal conductivity of different materials is relatively high and efficient in homogenizing any local temperature variations, making the influence of hydrodynamic dispersion less important in heat transport than in solute transport (Ingebritsen and Sanford, 1998).

**Steady-State Solution**

The steady-state analytical solution to Eq. [1] is (Bredehoeft and Papaoepulos, 1965)

\[
T(z) = T_s + \left(T_b - T_s\right) \frac{\exp\left[N_{pe}(z/L) - 1\right]}{\exp(N_{pe} - 1)} \tag{3}
\]

where \(T_s\) is the stream temperature (°C), \(T_b\) is the groundwater temperature (°C) at a depth \(L\) under the streambed (m), and \(N_{pe}\) is the Peclet number expressing the ratio of convection to conduction:

\[
N_{pe} = \frac{q_z \rho f c_f L}{\kappa_e} \tag{4}
\]

where \(q_z\) is the vertical Darcy flux (m s⁻¹). The steady-state solution is most suitable in periods where there is a marked difference in temperature between the groundwater and stream and during periods with stable temperatures as, for example, in winter and summer (e.g., Schmidt et al., 2006; Anibas et al., 2009).

**Transient Solution**

The solution to Eq. [1] given sinusoidal variations in stream temperature at the streamed surface is (Hatch et al., 2006)

\[
T(z,t) = A \exp\left(\frac{\nu z}{2\alpha_e} - \frac{z}{2\alpha_e} \sqrt{\frac{\varepsilon + \nu^2}{2}}\right) \times \cos\left(\frac{2\pi t}{\rho} - \frac{z}{2\alpha_e} \sqrt{\frac{\varepsilon - \nu^2}{2}}\right) \tag{5}
\]
where $A$ is the amplitude of temperature variations at the streambed–aquifer interface (°C), $P$ is the period of temperature variations (s) defined by $P = 1/f$, where $f$ is the frequency (cycles s$^{-1}$), and $\varepsilon$ is defined by

$$\varepsilon = \sqrt{v^4 + \left(\frac{8\pi \alpha_c e}{P}\right)^2}.$$  \hspace{1cm} [6]$$

The rate of penetration of the thermal front, $v$, is proportional to the pore water velocity and defined by $v = n v_f / \gamma$ (modified from Hatch et al., 2006). The exponential term on the right-hand side of Eq. [5] defines the damping of temperature amplitude variations with depth into the streambed. The cosine term defines the shift in phase with depth. As a result, the predicted temperatures at different depths are a nonlinear function of the thermal properties of the sediment and fluid, the fluid velocity, and the frequency of surface temperature variations (Hatch et al., 2006).

Hatch et al. (2006) proposed a method for quantifying exchange fluxes based on phase and amplitude variations between pairs of subsurface sensors (Fig. 3). Equation [5] is separated into components and solved for the ratio of amplitude variations between pairs of temperature measurement points at different depths:

$$v \Delta_A = \frac{2 \alpha_c}{\Delta z} \ln \frac{A_d}{A_s} + \sqrt{\frac{\varepsilon + v^2}{2}}$$  \hspace{1cm} [7]$$

where $\Delta_A$ is the ratio of the amplitude at the deep ($A_d$) and shallow ($A_s$) locations, $A_d/A_s$; or it is solved for the phase shift between the two measurement points ($\Delta \phi$):

$$v \Delta \phi = \sqrt{-2 \left(\frac{\Delta \phi \ 4\pi \alpha_c e}{P \Delta z}\right)^2}.$$  \hspace{1cm} [8]$$

where it has been assumed that the sediment properties do not change between the two points.

The thermal front velocity, $v$, appears on both sides of the equations, explicitly or embedded in the $\varepsilon$ term, requiring that Eq. [7] and [8] be solved iteratively. The Darcy flux ($q_z$) is then found from $v = n v_f / \gamma$. Equations [7] and [8] were used to quantify vertical temporal changes in discharge to the stream during the period with continuous thermal data. Before using Eq. [7] and [8], the temperature data were filtered using a band-pass filter to obtain a diurnal signal. Due to filtering edge effects (Hatch et al., 2006), the data record was shortened to 47 d.

Results and Discussion

Groundwater Flow and Seepage Meter

The isopotential lines in the meadow area and below the stream along the transect (Fig. 1) are shown in Fig. 4. The data are from 12 June 2008. The stream stage was 50.6 m, giving an overall horizontal gradient of 0.018 and a vertical gradient below the stream of approximately 0.67. The interpreted flow lines converge below the streambed and indicate that a significant amount of groundwater will discharge to the stream, not just from the adjacent meadow area but also from deeper parts. At the bank near the meadow area, there could be a significant horizontal component of discharge to the stream, while in the middle of the stream, discharge will be more vertical. In the northern and most shallow part of the meadow, groundwater discharges directly to the surface; this probably explains why standing water is almost always found in these areas.

The mean discharges measured by the seepage meters were very different among the three sites: 0.07 m d$^{-1}$ for S1 with a

![Fig. 3. Damping and phase attenuation of temperature with depth: (a) station with temperature sensors at selected depths; (b) temperature signals in stream and two depths (subscripts s = shallow, d = deep) showing the reduction in amplitude ($\Delta A = A_s - A_d$) and phase shift ($\Delta \phi$) (modified from Hatch et al., 2006).](https://pubs.geoscienceworld.org/vzj/article-pdf/10/1/98/2984836/98.pdf)
minimum and maximum of 0.06 and 0.08 m d−1, respectively; 1.34 m d−1 for S2 with a minimum and maximum of 0.56 and 2.05 m d−1, respectively; and 1.73 m d−1 for S3 with a minimum and maximum of 1.41 and 1.90 m d−1, respectively. Discharge was highest near the southern bank and lowest near the northern bank. The seepage meter data from S2 were the most unsteady. This might be a result of hyporheic flow caused by the imposed velocity field around the rigid plastic casing, which would have the greatest effect in the middle of the stream where the flow velocity is highest (Murdoch and Kelly, 2003; Rosenberry, 2008). Hyporheic flow can also be induced by a sudden shift in the streambed profile, which was also the case primarily near S2. The streambed was rather irregular upstream of S2, with big differences in water depth and with turbulent stream flow. The sediments in the middle of the stream were also more loose and coarse grained, which could promote a higher degree of hyporheic flow here than at the two banks, where more fine-grained sediment materials were found.

Seasonal Temperature Distributions

Three cross-sectional temperature distributions are shown in Fig. 5, spanning the period from late summer to the middle of winter (the end of August 2007 to the end of January 2008), recording the change from warm to cold streambed temperatures. In August there was an indication of high discharge in a distinct narrow zone 2 m from the northern bank (left-hand side of profile), where groundwater temperatures were observed near the streambed. The contrast in temperature at other locations was around 4°C. Data from September showed a cooling of the streambed, with almost uniform temperatures around 8°C, making it difficult to trace any discharge to the stream. The results from January 2008 show a two-layered temperature distribution, with about 5°C water in the shallow part of the streambed overlying 7 to 8°C water. Thus, there was no clear evidence of zones with distinct discharge to the stream. At this time the discharge to the stream looked very uniform. These snapshots of the temperature distribution indicate that the discharge to this small segment of a stream may be highly dynamic with time.
Steady-State Heat Transport Solutions

Discharge was estimated by fitting solutions of Eq. [3] to the observed temperature profiles at Stations T1, T2, and T3; an example from 20 May 2008 is given in Fig. 6. The values of \( T_g \) and \( T_s \) in Eq. [3] were given as 8 and 11.2°C, respectively. Data were selected at 1800 h to get the maximum temperature difference between the stream and groundwater. At this time of the day, the stream temperature on average is at its highest. The characteristic time scale for heat transport by conduction is \( \tau = l^2/\alpha_e \), where \( l \) is the transport length (Turcotte and Schubert, 2002). If it is assumed that the characteristic length is 10 cm, across which the temperature changes from that of groundwater to that of stream water, then \( \tau \) is on the order of a few hours given the parameters in Table 1. We therefore decided to use the 1800 h measurement of the stream water temperature as the boundary condition \( T_s \). This is different from Schmidt et al. (2006), who used the average stream water temperature during several days. The fitting parameter in Eq. [3] was the discharge, \( q_z \). There was high discharge at T2 (0.69 m d\(^{-1}\)) and T3 (1.17 m d\(^{-1}\)), as shown by the 8°C warm water near the streambed. The results for the other days are summarized in Fig. 7. The highest discharge was always recorded at T3 during the period of 1 May to 6 June 2008, followed by discharge at T2, then T1 (Fig. 7). There was a trend of increasing discharge at T2 (0.4 to 0.65 m d\(^{-1}\)) and T3 (0.85 to 1.2 m d\(^{-1}\)) and slightly decreasing discharge.

![Simulated and modeled temperature distributions (20 May 2008) at Stations T1, T2, and T3. The fitted exchange fluxes (\( q_z \)) are shown; they are all positive, indicating upward flow (discharge).](https://pubs.geoscienceworld.org/vzj/article-pdf/10/1/98/2984836/98.pdf)

![Darcy flux estimations based on steady-state heat transport solution and on daily temperature profiles taken at 1800 h. Positive values indicate discharge to the stream.](https://pubs.geoscienceworld.org/vzj/article-pdf/10/1/98/2984836/98.pdf)

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Solution</th>
<th>T1 and T3</th>
<th>T2</th>
</tr>
</thead>
<tbody>
<tr>
<td>Density of the streambed-sediment-fluid system (( \rho )), kg m(^{-3})†</td>
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<td>2000</td>
<td>2300</td>
</tr>
<tr>
<td>Specific heat capacity of the streambed-sediment-fluid system (( c )), J kg(^{-1}) °C(^{-1})†</td>
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<td>1091.5</td>
</tr>
<tr>
<td>Porosity (( n ))†</td>
<td>transient</td>
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<td>0.3</td>
</tr>
<tr>
<td>Thermal conductivity (( k )), W m(^{-1}) °C(^{-1})†</td>
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<td>2.47</td>
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<tr>
<td>Density of the fluid (( \rho_f )), kg m(^{-3})§</td>
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<td>999.7</td>
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<td>Specific heat capacity of the fluid (( c_f )), J kg(^{-1}) °C(^{-1})§</td>
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<td>4192</td>
</tr>
<tr>
<td>Depth with steady groundwater temperature (( L )), m†</td>
<td>steady state</td>
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<td>5</td>
</tr>
<tr>
<td>Groundwater temperature at depth ( L ) (( T_g )), °C¶</td>
<td>steady state</td>
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<td>8.07</td>
</tr>
<tr>
<td>Stream temperature (( T_s )), °C¶</td>
<td>steady state</td>
<td>varies</td>
<td>varies</td>
</tr>
</tbody>
</table>

The seepage meter measurements were conducted in early July and showed slightly higher discharge at T2 (1.34 m d\(^{-1}\)) and T3 (1.73 m d\(^{-1}\)) and approximately the same discharge at T1 (0.07 m d\(^{-1}\)). In general, the pattern of discharge to the stream is correct, given the limitation of both methods. For example, the analytical solution only predicted the vertical component of discharge to the stream assuming a constant discharge rate and steady-state conditions. As discussed above, there may have been horizontal components, especially near the banks and below it is shown that the discharge rate was probably not constant below the streambed. Furthermore, the assumption of steady state can be violated, which will influence the fitted discharge (Anibas et al., 2009; Schornberg et al., 2010). The seepage meter measurements also revealed that discharge can change rapidly during a single day, especially near T2, where discharge ranged from 0.56 to 2.05 m d\(^{-1}\).

**Transient Heat Transport Solution**

Unfiltered temperature time series data from T1, T2, and T3 show clear damping and phase shift with depth (Fig. 8). The temperature data were from the shallowest sensors, except at T1, where Sensor 4 from the top was used (due to streambed scouring). There was a clear damping of the stream temperature signal at T2 and T3, with the largest damping at T3. There was much smaller damping at T1. Qualitatively, these observations match those based on seepage meter measurements and the steady-state heat transport solution, with the highest discharge (the greatest damping) at T3 and the lowest at T1.

Figure 9 shows an example of filtered data from T1 using Sensors 4 and 7 (Sensor Pair [SP] 4) that had a spacing of 15 cm. Sensors located deeper than 20 cm resulted in uncertain estimations because of too much damping. At T2 and T3, the maximum spacing used was 9 and 4 cm, respectively, and the maximum distance from the streambed was 13.5 and 9 cm, respectively. Table 2 shows the different sensors used, sensor pairs, respective spacing, and relative distance of the midpoint of a sensor pair to the streambed, as illustrated in Fig. 10 for T1. After filtering, the temperature changes showed a clear diurnal signal.

Temporal exchange rates between the stream and groundwater were estimated using Eq. [6] on a daily basis for a period of 47 d from 21 April to 6 June 2008 (Fig. 11). Six time series are available for T1 and T2, while it was only possible to obtain three time series for T3 due to significant groundwater discharge and, hence, great damping of the stream temperature signal. The mean discharge is also shown. Previous applications of Eq. [6] only solved for the flux at one location below the streambed. We were able to estimate the exchange flux at several locations, however, given the spatial and temporal resolution of the temperature signal in our data set.

At T1, it was not possible to use data from sensors located deeper than 20 cm under the streambed (Sensor 7;
Table 2. Position of each sensor pair (SP) used, spacing, and relative distance to streambed for Darcy flux estimations. Data were collected 20 May and 1 June 2008. Distance to streambed is relative because of streambed scouring but distances between individual sensor pairs are constant.

<table>
<thead>
<tr>
<th>Temperature Station 1</th>
<th>Temperature Station 2</th>
<th>Temperature Station 3</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sensor pair</td>
<td>Spacing (cm)</td>
<td>Distance to streambed (cm)</td>
</tr>
<tr>
<td>T1_1_4 vs. T1_1_5 (SP1)</td>
<td>5</td>
<td>7.5</td>
</tr>
<tr>
<td>T1_1_4 vs. T1_1_6 (SP2)</td>
<td>10</td>
<td>10.0</td>
</tr>
<tr>
<td>T1_1_5 vs. T1_1_6 (SP3)</td>
<td>5</td>
<td>12.5</td>
</tr>
<tr>
<td>T1_1_4 vs. T1_1_7 (SP4)</td>
<td>15</td>
<td>12.5</td>
</tr>
<tr>
<td>T1_1_5 vs. T1_1_7 (SP5)</td>
<td>10</td>
<td>15.0</td>
</tr>
<tr>
<td>T1_1_6 vs. T1_1_7 (SP6)</td>
<td>5</td>
<td>17.5</td>
</tr>
</tbody>
</table>

Fig. 10. Positions of uppermost sensors at Station T1 used in the time series analysis and sensor pairs (SP) with spacings (in cm) and relative distance to the streambed for Darcy flux estimations.

Fig. 11. Exchange rates between aquifer and stream using transient heat transport solution for sensor pairs (SP) at Stations T1, T2, and T3. Negative and positive values are recharge (losing stream) and discharge (gaining stream), respectively.
Th e exchange rates were always positive and represent discharge. Th us, the time series data suggest recharge of the aquifer at the northern bank with a mean recharge of \(-0.15 \text{ m d}^{-1}\) during the whole period—quite different from the results obtained by seepage meter and the steady-state heat transport solution. The three upper sensor pairs (SP1, SP2, and SP3), however, showed upward flow (discharge) at certain times, e.g., around 10 to 20 May, when discharge increased upward. Also, these sensor pairs always showed less positive or negative exchange compared with the three other (deeper) pairs of sensors (SP4, SP5, and SP6). The location of the midpoint of the two sensor pairs SP3 and SP4 were the same. It is therefore possible to evaluate the order of magnitude of the error in the estimated discharge because the two fluxes should be approximately the same assuming that the change in discharge with depth did not change too dramatically. The flux at SP3, however, was consistently 0.1 m d\(^{-1}\) higher than at SP4. It therefore seems that the data quality and the time series analysis make it impossible to resolve fluxes with greater precision than about 0.1 m d\(^{-1}\). This is similar to the error estimated in a numerical analysis using the one-dimensional analytical solution on temperature profiles extracted from the results of a two-dimensional numerical heat transport model (Schornberg et al., 2010). This may also explain why it is possible to see upward flow (<0.1 m d\(^{-1}\)) at the three upper sensor pairs at the same time as downward flow at the three lower pairs during the period of 10 to 20 May. From the other sensor pairs, the estimated discharge clearly showed a steady pattern but with a noticeable decreasing trend from 20 May to the end of the measurement period.

At T2, the exchange rate was estimated for all sensor pairs between Sensors 2 and 5 (Table 2) (<13.5 cm below the streambed). Data from sensors located deeper than this had significant damping. The exchange rates were always positive and represent discharge to the stream. Discharge was very dynamic during the first two-thirds of the period until about 22 May, after which the fluxes stabilized. The discharge in the top 13.5 cm varied from 0 to 1.8 m d\(^{-1}\). Before 22 May, discharge was very irregular, with no apparent pattern; sometimes the flux was highest in SP6 and lowest in SP1, sometimes highest in SP3 and lowest in SP6. This is similar to the seepage meter measurements, where great variations (0.56–2.05 m d\(^{-1}\)) were recorded during a single day. Again, the reason for this might be hyporheic flow or a complex flow field within the more stony sediments found at T2. During the last third of the period (from 22 May), the fluxes showed less fluctuation, as at T1. The three and two sets of sensor pairs SP1 to SP3 and SP4 to SP5 predicted almost the same discharge (within the precision of 0.1 m d\(^{-1}\) found from the analysis of the responses in T1). The discharge therefore increased upward in a consistent way, suggesting that flow from the banks and the deeper parts converged below the middle part of the stream. From these results, it is clear that there was not a constant discharge rate below the stream (as assumed for the steady-state solution). The mean Darcy flux for the whole period was 0.80 m d\(^{-1}\); thus the stream was gaining water at T2 in good agreement with the results of both seepage meter measurements and the steady-state heat transport solution results. The estimated discharge 7.5 cm below the stream bottom (SPI) was approximately 1.2 m d\(^{-1}\), close to the mean seepage meter flux of 1.34 m d\(^{-1}\) (measured \(\sim1\) mo later, however). The discharge fluxes at T2 did not show the same decreasing trend with time as observed at T1.

Discharge at T3 was estimated for three pairs of sensors, using Sensors 2, 3, and 4 (<9 cm below the streambed; Table 2). Data from sensors located lower than this showed a very high damping and were not useful in the analysis. The mean discharge for the whole period was 0.90 m d\(^{-1}\), indicating that the stream gained water at T3, in good agreement with the results from the seepage meter measurements and the steady-state heat transport solutions. The discharge was, as at T1, uniform and consistent more or less throughout the entire time series (28 April–31 May), with discharge increasing upward toward the stream bottom. After 3 June, \(A_x\) for SP3 was >1 and no solution was found. At this time, discharge was close to zero at this depth, indicating a negative flow, which does not seem likely. The mean seepage meter flux of 1.73 m d\(^{-1}\) measured in early July was much higher than predicted at the stream bottom (\(\sim0.5\) m d\(^{-1}\) on 6 June, SP1). There was a significant decrease in discharge toward the end of the monitoring period, which may be explained by a falling water table on this side of the stream with a steep hillslope.

**Summary of Exchange Rates**

Figure 12 summarizes the exchange fluxes estimated by the different methods. The temporal mean values plus the minimum and maximum values are shown. The mean values of the different temperature spacings and the results from the sensor pairs nearest to the streambed (SPI) were used in the case of the transient heat transport results. Overall, there is good agreement between the different methods in that all methods predicted a low exchange flux (recharge and discharge) at T1 and high exchange fluxes (only discharge) at T2 and T3. The mean flux measured by the seepage meters was higher than the estimates based on temperature for T2 and T3. Recall, however, that seepage meter results were all measured on a single day about 1 mo after the temperature data, whereas the results based on temperature are mean values for a 47-d period. The results from SPI give a better agreement with the seepage meter results than using mean values of all the sensor pairs. The different seepage meter results measured at T2 can be explained by hyporheic flow caused by the presence of a gravel layer and, potentially, an added pressure-driven flow through this layer due to the installation of the seepage meter. The two temperature-based methods are in reasonable agreement; however, at T1 the transient solution predicted a low recharge flux, while the
steady-state solution predicted a low discharge flux. At the moment it is not possible to explain why the transient solution gave a negative flux except that there was probably a significant horizontal flow component that made the use of both temperature-based methods problematic. The seepage meter results were somewhere in between, estimating a near-neutral exchange at T1.

The steady-state solution generally gave very good fits to individual temperature profiles (Fig. 6). The mean discharge rates compared well with the mean values obtained from the transient solutions, despite the fact that the observations from using the transient solution (Fig. 11) suggest that it is not possible to assume strictly uniform flow. This is also reflected in the larger error bars on the estimates based on the transient solution. The results from using the transient solution also demonstrate that during long periods there was a consistent trend with increase in the discharge flux (T2 and T3) as the SWI was approached. A possible explanation is that flow was not strictly vertical but rather converged near T2 and T3. Similar observations have been found from the analysis of temperature profiles in deep bore holes (Reiter, 2003).

**Conclusions**

Time series analysis of continuous streambed temperature data from three stations—two near the banks and one in the middle of the stream—shows that discharge in the upper 20 cm of the streambed was not uniform within a station or between stations. This occurred during a 47-d period in April to June of 2008, when temperature gradients between the stream and groundwater reversed. Across the stream, the exchange fluxes varied dramatically, from very little discharge (and sometimes recharge) to 0.80 to 0.94 m d⁻¹ (on average) at the two other stations. The accuracy of the equipment and limitation of the time series analysis (e.g., neglecting heterogeneity in heat parameters in the model development, data filtering etc.) indicate that only fluxes >0.1 m d⁻¹ could be resolved. The station with the smallest fluxes also has the most steady fluxes with time (but close to the precision of the method), while the station in the middle had the most unsteady fluxes, which we suspect to be caused by hyporheic flow due to the presence of a gravel layer in the middle of the stream. The unstable discharges were confirmed by seepage meter measurements. The third station displayed a significant decrease in discharge toward the end of the monitoring period, which may be explained by a falling water table on this side of the stream with a steep hillslope. The same phenomenon was not seen at the other bank due to the meadow area, which has a constant water table near the surface year-round. At the two stations with high discharge, fluxes varied as much as 1.6 m d⁻¹ across just 5.5 cm. During long periods with stable discharge, the fluxes increased upward, suggesting that there is significant lateral and horizontal flow that can account for the increase in flow upward. This was confirmed by inspecting the measured flow field below the streambed and in an adjacent meadow area. Flow along the stream and hyporheic flow are other explanations that will cause nonuniform upward flow.

The results from the time series analysis were compared with seepage meter measurements and simulation results using a steady-state heat transport solution. Generally, the pattern was the same among the different methods but with noticeable differences in mean values and ranges. The order of magnitude in exchange fluxes we believe were estimated correctly by the simple analytical methods; however, numerical methods are needed to more accurately predict the spatial and temporal discharges to the stream. Previous applications of time series analysis have been limited to only one sensor pair (one depth below the streambed); however, we used several sensor pairs (several depths below the streambed) to improve our understanding of the flow pattern below a streambed.
Acknowledgments

GEOCENTER DENMARK and the Villum Kann Rasmussen Centre of Excellence HOBE provided funding. We would like to acknowledge Peer Jørgensen for help with both development and installation of the temperature equipment as well as assistance with field equipment. Thanks to J. Constantz, R. Niwongser, and A. Fisher for their help and fast email correspondence on data filtering of the initial temperature data.

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