Characterization of spatial heterogeneity of groundwater-stream water interactions using multiple depth streambed temperature measurements at the reach scale

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Abstract

Streambed temperatures can be easily, accurately and inexpensively measured at many locations. We obtained 140 vertical streambed temperature profiles along a 220 m section of a small artificial stream to characterize patterns of groundwater-stream water interaction with a high spatial resolution. Groundwater temperature at a sufficient depth remains nearly constant while stream water temperatures vary seasonally and diurnally. In summer, streambed temperatures of groundwater discharge zones are relatively colder than downwelling zones of stream water. Assuming vertical flow in the streambed, the observed temperatures can be correlated to the magnitude of water fluxes. The water fluxes can then be estimated by applying a simple analytical solution of the heat diffusion-advection equation to the observed vertical temperature profiles. The calculated water fluxes through the streambed ranged between $10.0 \, \mathrm{L m^{-2} \, d^{-1}}$ of stream water entering the streambed and $455.0 \, \mathrm{L m^{-2} \, d^{-1}}$ of groundwater discharging to the stream. The investigated reach was dominated by groundwater discharge with two distinct high discharge locations accounting for 50% of the total flux on 20% of the reach length.

1 Introduction

Understanding and quantifying physical processes and ecological implications of groundwater surface water interaction is becoming an important subject in hydrogeological and river ecological studies. Stream water and groundwater can interact in different ways on a wide variety of scales down to heterogeneities within meters to centimetres (Brunke and Gonser, 1997; Woessner, 2000).

Investigation of groundwater-stream water interactions (water fluxes through the streambed, hyporheic flowpaths, subsurface flow velocities and travel times) can be classified according to "where-you-stand" as viewing interactions from the stream or the subsurface (Packman and Bencala, 2000).

In studies where the point of view is from the stream, the hyporheic exchange is often the focus. Hyporheic exchange is the downwelling of stream water into shallow sediments and the return to the stream after a certain distance. These flow systems transport oxygenated stream water, nutrients and dissolved organic carbon into the hyporheic zone. This leads to increased mibrobial activity and significantly influences the nutrient and carbon cycling in stream systems. Nonetheless, the continuous hyporheic exchange also affects the downstream transport and fate of contaminants.

Various studies incorporating different methods have analyzed hyporheic exchange. Deterministic approaches have shown that stream morphologic features can induce advective flow from the surface to the subsurface. Theory, laboratory experiments and field studies have investigated the influence of scale (cm to tens of m) and shape of bedforms and stream morphology on flowpaths, pore flow velocities and residence times of surface water in the hyporheic zone (Thibodeaux and Boyle, 1987; Elliott and Brooks, 1997a, b; Cardenas et al., 2003; Storey et al. 2003; Salehin et al., 2004; Saenger et al., 2005; Anderson et al., 2005; Gooseff, 2005). In general, an increased bed form wavelength and amplitude leads to increased depths and lengths of hyporheic flow paths for vertical features like pool and riffle sequences. The presence of meanders, secondary streams and streamsplits induce lateral near stream flow paths (Harvey and Bencala, 1993; Wroblicky et al., 1998; Kasahara and Wondzell, 2003).

Another scope is considered when viewing the interactions from the subsurface. As Storey et al. (2003) suggested, groundwater discharge can have a significant impact on the extent of the hyporheic zone and can affect the distribution of benthic end hyporheic fauna (Brunke and Gonser, 1997). Temporal changes of hydraulic gradients between an aquifer and a stream can alter the near stream groundwater flow field and the magnitude of both downwelling streamwater and upwelling groundwater (Wroblicky et al., 1998). Furthermore, it becomes essential to consider the spatial patterns and magnitude of groundwater discharge when the transport and the fate of contaminants from the aquifer to the stream has to be assessed (Conant et al., 2004; Conant, 2004). Independently from the point of view of the investigation, whether from the stream or

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the subsurface, it is crucial to consider the spatial distribution and the magnitude of groundwater discharge to a stream.

In general, a variety of factors from the catchment scale to single bedforms are controlling the interactions of groundwater and stream water. As a result of different mechanisms, flow patterns within the streambed can vary on small spatial scales. Investigations at the stream reach scale which consider small-scale patterns of flow require a high density monitoring network. Due to instrumentation and measurement effort, such studies are often limited to a relatively small spatial extent (Baxter and Hauer, 2000). Thus there is a need for an inexpensive, quantitative method that has the capability to characterize the spatial heterogeneity of groundwater-stream water interactions.

In this study, we show that streambed temperatures can be used to delineate patterns of groundwater-stream water interactions in fine detail on the scale of stream reaches. On the basis of the observed streambed temperature profiles, the vertical water fluxes through the streambed were quantified by applying a simple one dimensional analytical model of the heat advection-diffusion equation.

2 The temperature concept

The horizontal and vertical temperature distribution in the streambed is a result of heat transport by the flowing water (advective heat flow) and by heat conduction through the solid and fluid phase (conductive heat flow) of the sediments. While groundwater temperature remains nearly constant at the mean annual air temperature at a sufficient depth, stream water temperatures vary seasonally and diurnally. For example, in summer, streambed temperatures in groundwater discharge zones should be relatively colder than in stream water downwelling zones. The temperature measurements can be used as a surrogate for head and hydraulic conductivity measurements (Anderson, 2005). Analytical solutions to solve the heat transport equation for water flux were developed in the 1960s (Suzuki, 1960; Stallman, 1963; Bredehoeft and Papadolus, 1965). In recent years there have been several applications of temperature profiles

for estimating magnitude and direction of water flow at the groundwater surface water interface (e.g., Bartolino and Niswonger, 1999; Constantz et al., 2003; Lapham, 1989; Silliman, 1995; Stonestrom and Constantz, 2003).

The characterization of spatial patterns of flow at the groundwater surface water interface requires a measurement concept that allows many measurements with high spatial resolution during a relatively short period of time. Conant (2004) was the first who showed that streambed temperatures measured in a short period of time at many locations can be related to spatial variations of groundwater discharge. In contrast to the work of Conant (2004) who used temperatures observed at a single depth, we used temperature measurements obtained simultaneously at five depths.

3 Study site

The temperature measurements were carried out along a 220 m long reach of the Schachtgraben near the town of Wolfen (Fig. 1). For the past hundred years, Wolfen has been a major chemical industry site in Germany. In the second half of the 20th century the spectrum of products was extended to 5000 substances, including chlorinated solvents, pesticides and plastics (Walkow, 1996; Chemie AG Bitterfeld-Wolfen, 1993). The deposition of contaminated waste products in abandoned lignite pit mines nearby the production sites as well as inappropriate handling and transport of chemicals and war damages led to a large scale contamination ($25 \, km^2$) of groundwater, soils, surface water and floodplain sediments (Heidrich et al., 2004). For decades, untreated process waste waters were discharged via the Schachtgraben and the Spittelwasser into the river Mulde which is a tributary of the river Elbe. The Schachtgraben is an artificial channel located in the river Mulde floodplain system with a regular width between 2.5 and 3 m. The channel cuts the floodplain sediments and is located in the sediments of the shallow Quaternary aquifer. The channel bed itself is constructed of a homogeneous coarse gravel layer of 0.4 m thickness. Groundwater levels in the adjacent unconfined aquifer are generally 0.1 to 0.2 m higher than the water level in

the stream. The shallow aquifer is composed of Weichselian glacio-fluvial sandy gravels. Today streambed sediments and the groundwater in the adjacent aquifer and in the streambed sediments are contaminated with a wide range of substances but mainly with chlorinated benzenes and hexachlorocyclohexanes. Further downstream in the Spittelwasser floodplain, sediments were found to be contaminated with polychlorinated naphthalenes and dioxins (Brack et al., 2003; Bunge et al., 2003; Walkow, 2000). The investigated reach of the Schachtgraben and the river Mulde floodplain are the subject of additional studies concerning water flow as well as transport and fate of heavy metals and organic contaminants at the interface between groundwater and surface water.

4 Field methods

4.1 Temperature measurements

The streambed temperatures were measured along two longitudinal transects in a four day measuring campaign from 30 August until 2 September 2005. The campaign consisted of 140 measurements with 70 for each transect. The measurements were generally taken with 3 m spacing but were refined between locations with high temperature differences. The longitudinal transects were located at one third and two thirds of the total river width. A multilevel stainless steel temperature probe (TP 62, Umwelt Elektronik GmbH; Geislingen, Germany) was used to measure the temperatures simultaneously at 0.1 m, 0.15 m, 0.2 m 0.3 m and 0.5 m depths in the streambed at each location. During the study, stream temperatures were measured hourly using a self containing Stowaway TidbiT –5 to 37°C range temperature logger (Onset Computer Coporation, Pocasset, Massachusetts). Groundwater temperature was observed hourly with temperature and pressure transducers placed directly into the aquifer with a vertical spacing of 1 m between depths below ground surface of 1 m to 5 m.

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4.2 Piezometer installation and slug testing

To confirm the fluxes obtained from the streambed temperature profiles with an independent method, streambed piezometers were installed to gain information on hydraulic gradients and hydraulic conductivity. Locations were chosen according to high and low groundwater discharge zones indicated by the observed temperatures (high discharge locations: P2, P4, P5 P7; low discharge locations: P1, P3, P6). One pair of piezometers (P4, P5) was installed at a distinct groundwater discharge location at Transect A with 1 m spacing to obtain the small scale heterogeneities of streambed hydraulic properties and fluxes.

The piezometers consist of 1.6 m long HDPE (high density polyethylene) pipes with 0.04 m outside diameter. The 0.2 m screened section of each piezometer was installed between 0.3 and 0.5 m below the streambed surface. The hydraulic head differences between the stream surface and the piezometers were estimated following the method of Baxter et al. (2003). To obtain the hydraulic head differences, an additional open pipe was attached outside the piezometer ("stilling well") to minimize the influence of turbulence on stream water elevation. The hydraulic head difference was measured using parallel chalked wires connected at the top. The chalked wires were inserted into a piezometer and the attached stilling well and after removal the distance between the water marks was measured.

Each piezometer was tested twice with a falling and rising head slug test. Rising head slug tests were performed by removing the water from the piezometer using an Eijkelkamp 12 V peristaltic pump (Eijkelkamp, Giesbeek, The Netherlands). Falling head slug tests were carried out by releasing water from an attached reservoir at the top of the piezometers. The rise and fall of the water level in the piezometers was observed with an "HT 575 Kompakt" pressure transducer (Hydrotechnik GmbH, Obergünzburg, Germany).

5 Analytic procedure

Streambed temperatures have a highly transient character due to seasonal and diurnal changes of stream water temperatures. It is essential for the concept of streambed temperature mapping that differences of temperature can be attributed to spatial differ-
ences of water fluxes and are not a result of temporal variations. Streambed tempera-
tures measured at a sufficient depth below the influence of diurnal variations represent the quasi-steady-state conditions of streambed temperatures for the finite time of the mapping campaign (Fig. 2).

With the assumption that water flow in the streambed is essentially vertical, the gov-
erning equation for one-dimensional conductive and advective heat transport is:

$$\frac{K_{fs}}{\rho c} \nabla^2 T_z - \frac{\rho_f c_f}{\rho c} \nabla \cdot (T_z q_z) = \frac{\partial T}{\partial t} \tag{1}$$

where T_z [°C] is the streambed temperature at depth z; t is time [s]; q_z is the vertical Darcy velocity [ms^{-1}]; ρc is the volumetric heat capacity of the solid – fluid system which can be written as $\rho c = n\rho_f c_f + (1-n)\rho_s c_s$ where $\rho_f c_f$ is the volumetric heat ca-
pacity of the fluid, $\rho_s c_s$ is the volumetric heat capacity of the solids [Jm^{-3}K^{-1}] and n is the porosity [–]. K_{fs} is the thermal conductivity of the solid – fluid system [Js^{-1} m^{-1}K^{-1}].

With boundary conditions $T = T_0$ for $z=0$, and a fixed temperature T_L for $z=L$, where L [m] is the vertical extent of the domain, the solution of Eq. (1) can be obtained as (Bredehoeft and Papadopolus, 1965):

$$\frac{T(z) - T_0}{T_L - T_0} = \frac{\exp\left(\frac{q_z \rho_f c_f}{K_{fs}} z\right) - 1}{\exp\left(\frac{q_z \rho_f c_f}{K_{fs}} L\right) - 1} \tag{2}$$

Equation (2) can be solved for q_z for a given L. It is assumed that the vertical tem-
perature distribution at different locations is only a function of q_z, i.e. other parameters

on the right-hand side of Eq. (2) are considered to be homogeneous for all observed temperature profiles. The objective function for obtaining q_z is given with:

$$\text{Error}_k(L) = \sum_{j=1}^{5} \left[T_{jk} - \left(\frac{\exp\left(\frac{q_{zk}\rho_f c_f}{K_{fs}} z_j\right) - 1}{\exp\left(\frac{q_{z_k}\rho_f c_f}{K_{fs}} L\right) - 1}(T_L - T_0) + T_0 \right) \right]^2 \tag{3}$$

where q_{zk} is the value of q_z that minimizes $\text{Error}_k(L)$ for a given L at each temperature profile consisting of $j=5$ temperature observations.

It was tested if a change of L has an influence on the estimated q_z and the quality of the fit. The objective function to find one optimal L for all observed temperature profiles implies the optimization of $\text{Error}_k(L)$. We computed an optimal q_{zk} at each profile k for the overall L ranging between 0.6 and 10 m. For $k=140$ observed temperature profiles, the objective function is given with:

$$f(L) = \sum_{k=1}^{140} [\text{Error}_k(L)] \tag{4}$$

Once the optimal q_z for a chosen L is obtained from Eq. (3), q_z can be substituted into Eq. (2) to obtain a simulated streambed temperature distribution. To test the quality of fit between observed and simulated temperatures, the difference of temperatures ΔT [K] can be obtained from Eq. (5):

$$\Delta T = T(z) - \frac{\exp\left(\frac{q_z\rho_f c_f}{K_{fs}} z\right) - 1}{\exp\left(\frac{q_z\rho_f c_f}{K_{fs}} L\right) - 1}(T_L - T_0) + T_0 \tag{5}$$

6 Results and discussion

6.1 Stream water temperatures

During the field campaign, the stream water temperatures showed variations with a low of 15.8 and a high of 23.0°C (Fig. 3). The dotted line in Fig. 3 illustrates the 24 h moving average of stream water temperatures. It varies only between 17.6°C and 18.6°C around the overall average of 18.4°C during the field campaign. This indicates that the temperature oscillations are of diurnal character. The temperature regime is characterized by anthropogenic influences which become apparent in temperature peaks in the early morning (Fig. 3).

6.2 Streambed temperatures

The observed streambed temperatures varied spatially between 11.5 and 17.5°C at a depth of 0.5 m in the streambed. At the shallow depth of 0.1 m, the temperatures showed a wider range and a higher minimum and maximum of 12.2 and 19.9°C. In summer, groundwater discharge is indicated by relatively low streambed temperatures. Along the observed 220 m reach, two major groundwater discharge zones were identified. The first discharge zone is located between 20 and 50 m and the second between 125 and 170 m (Fig. 4).

The discharge zones are characterized by streambed temperatures at 0.5 m being less than 15°C. Within the second discharge zone, there are distinct locations showing temperatures less than 13°C at 0.5 m depth and even at 0.1 m depth, temperatures are less than 15°C (Fig. 4). These distinct locations of very low temperatures are restricted to a length of 3 to 5 m. Both major discharge zones have a very similar spatial extent.

Along both longitudinal transects, very similar patterns of streambed temperature are visible. Variations of streambed temperatures are occuring primarily and along the reach while the differences between the eastern and western bank are of minor significance.

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6.3 Groundwater temperatures

Groundwater temperatures were observed in hourly intervals at depths between 1 and 5 m below the streambed surface, adjacent to the stream (Fig. 1). At depths of 4 and 5 m the groundwater temperatures are 11°C. Temperatures increase to 15°C at a depth of 1 m below the streambed surface. The groundwater temperatures were measured at a location close to a zone of relatively high streambed temperatures. Thus the shallow groundwater temperatures correspond well with the streambed temperatures being at 16.8°C at a depth of 0.5 m. The coldest streambed temperatures are nearly identical to groundwater temperatures observed at a depth of 4 m.

6.4 Fluxes obtained from temperature profiles

As temperature can be used as a surrogate for head measurements and can be easily obtained at hundreds of locations, the water fluxes in the streambed can be estimated with a high spatial resolution. The water fluxes were obtained at each location from Eq. (2) by minimizing the differences between observed and modelled temperature profiles (Eq. 3). At each temperature profile, q_z was estimated for L ranging from 0.6 to 10 m. It was found that q_z for L larger than 1.0 was essentially independent from L (Fig. 5). The resulting fluxes are not influenced by the depth at which T_L is obtained as long as T_L remains constant with the increasing depth. This is basically the case when upward flow from groundwater to surface water is present. The observed groundwater temperature at a depth of 4 m below the streambed surface was 11.0°C and was constant during the measuring campaign. Hence, the lower boundary condition T_L was set to 11.0°C. The upper boundary condition T_0 was set to 18.4°C which is the average stream water temperature of the four-day mapping period. Equation (2) requires the thermal conductivity K_{fs} as an input parameter which was not measured within this study. However, the range of thermal conductivities of water saturated sediments is small thus K_{fs} can be reliably estimated and was set to $2 \, \mathrm{Js^{-1}m^{-1}K^{-1}}$ (Stonestrom and Blasch, 2003). The parameter set used for estimating q_z from the observed tempera-

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ture profiles is summarized in Fig. 2.

The resulting water fluxes ranged between -10.0 and $455.0\,\mathrm{L\,m^{-2}\,d^{-1}}$ (Figs. 4a and c). The average groundwater discharge is $58.2\,\mathrm{L\,m^{-2}\,d^{-1}}$ and the average recharge is $2.3\,\mathrm{L\,m^{-2}\,d^{-1}}$. Figures 4a and c illustrate the spatial distribution of fluxes in relation to the length of the observed reach. Analogous to the temperatures, the flux distribution is very similar in the two longitudinal transects. Recharge occurs only along less than 1% of the reach. The zones with discharges higher than $100\,\mathrm{L\,m^{-2}\,d^{-1}}$ are present on 16% of the total length of Transect A and on 19% of Transect B. Approximately 20% of the total length contributes 50% of the total discharge (Fig. 7). Around 85% of the total discharge occurred at 50% of the total length (Fig. 7) Only four profiles were observed to have discharges higher than $200\,\mathrm{L\,m^{-2}\,d^{-1}}$ which contribute about 10% to the total discharge (Figs. 4a and c). These relations as well as the order of magnitude of fluxes are comparable to those observed by Conant (2004). Spatially distinct high discharge zones were also observed in other studies but with higher maximum discharges (Baxter and Hauer, 2000; Conant, 2004). Yet the maximum discharges are more than 4 times higher than the average discharge. The reduced spread of fluxes compared to natural rivers can be explained with a reduced streambed heterogeneity in terms of morphologic features and hydraulic properties. According to the maximum fluxes the homogeneous streambed might lead to less significant preferential flowpaths and thus to lower maximum fluxes.

Recharge occurs only at few locations and at low flow rates (up to $10\,\mathrm{L\,m^{-2}\,d^{-1}}$). Vertical hydraulic gradients observed in the streambed piezometers and the water table elevation adjacent to the Schachtgraben indicate a gaining reach. Nonetheless, downward flow can be topographically induced at pool and riffle structures and at smaller spatial scales at streambed ripples and obstructions (Thibodeaux and Boyle, 1987). Because of the artificial origin of the homogeneous gravel streambed, natural pool and riffle sequences are assumed not to be present at the Schachtgraben. Storey et al. (2003) reported that a streambed hydraulic conductivity below $10^{-4}\,\mathrm{m\,s^{-1}}$ will result in a restricted downwelling of water at a pool and riffle structure. However, the

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combination of a streambed with no apparent geomorphological heterogeneity and low streambed hydraulic conductivities can result in the observed low recharge fluxes and restricted hyporheic exchange. It is likely that downwelling of stream water occurs mainly due to streambed roughness induced by the single gravel grains. Since hyporheic flowpaths are related to the vertical extent of the streambed morphologic features, hyporheic flow in the Schachtgraben will only occur in the upper few centimetres of the streambed.

6.5 Differences between observed and simulated temperatures

The mean difference between all observed and simulated temperatures is 0.023 K at Transect A and 0.028 K at Transect B. The highest calculated difference at Transect A was 2.1°C, located at a depth of 0.1 m and at was −1.6°C Transect B also at a depth of 0.1 m (Figs. 4b and d).

At both transects, although the observed patterns of temperatures were very similar, the highest differences between simulated and observed temperatures occured at different locations. The differences are clearly related to certain depths but seemed to be randomly distributed along the transects (Fig. 4). At 82 out of 140 temperature profiles (58.6%), the maximum difference between observed and simulated temperatures occurs at 0.1 m depth. At the other depths of 0.15 m, 0.2 m, 0.3 m and 0.5 m the maximum differences are similarly distributed, respectively 13.6%, 5.9%, 11.4% and 11.4%. This distribution of differences indicates that there is an influence of diurnal stream water temperature oscillations at the shallow depth of 0.1 m, disturbing the quasi-steady-state profile. As well it is possible that shallow, non-vertical hyporheic flow paths could have influenced the upper 0.1 m of the streambed. Figure 6 illustrates examples of simulated temperature profiles after q_z was obtained from observed temperatures using Eq. (3) and the related observed temperatures at $T(z)$. A recalculation of fluxes excluding the temperature measurements at a depth of 0.1 m showed that there is no significant influence for low and medium fluxes. At high flux locations the resulting fluxes decrease when the most shallow measurement is excluded. For exam-

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ple the calculated maximum flux is reduced from 455 to 325 $\mathrm{L\,m^{-2}\,d^{-1}}$ without the temperature at 0.1 m depth. Although there are indications that temperatures within 0.1 m are influenced by diurnal temperature oscillations in the surface water, there is no evidence for an increased uncertainty in the resulting fluxes. In particular for groundwater
5   discharge, where streambed temperatures change from groundwater temperature to stream water temperature in the upper few centimetres of the streambed, it is essential to have an observation point at a shallow depth.

6.6   Verification of flux calculations with piezometer data

A total number of 7 streambed piezometers was installed and tested to confirm the
10   magnitude of water fluxes obtained from the temperature profiles (Figs. 4a and c). The observed head differences Δh between the streambed and the aquifer indicated an upward flow direction at all streambed piezometers. The maximum Δh occurred at piezometer P2 with 0.118 m, the minimum Δh at piezometer P7 with 0.017 m. The vertical hydraulic gradient was obtained by dividing Δh with Δl which is the length between
15   the centre of the piezometer screen and the top of the streambed. All piezometers were installed at the same depth in the streambed and thus Δl is 0.4 m at all piezometer locations. The resulting vertical hydraulic gradients are between 0.043 and 0.295.

    Streambed hydraulic conductivities were estimated from rising and falling head slug tests using the Hvorslev (1951) case G, basic time lag equation. Horizontal hydraulic
20   conductivities K_h varied within a relatively small range of one order of magnitude between 1.39×10^{-4} and $1.26\times10^{-5}\,\mathrm{m\,s^{-1}}$. As the cobbely streambed makes it impossible to install permeameters to obtain the vertical hydraulic conductivity K_v in the field, the anisotropy ratio has to be estimated. Freeze and Cherry (1979) gave an anisotropy ratio of core samples K_h/K_v between 3 and 10. The resulting hydraulic conductivi-
25   ties are within the range given by Calver (2001) and lower than the vertical hydraulic conductivities observed by Chen (2004)

    Employing both an anisotropy ratio of 3 and 10, the resulting fluxes based on the

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piezometer data are within one order of magnitude of the fluxes obtained from the temperature data (Table 1 and Fig. 5). In general, the fluxes obtained from Eq. (2) correspond reasonably well with the fluxes obtained from Darcy's law calculations. At piezometer locations P3 and P5, fluxes calculated with an anisotropy ratio of 3 over-

5 estimate the temperature based fluxes while fluxes based on an anisotropy ratio of 10 underestimate them.

6.7   Applicability and limitations

Using streambed temperatures to quantify groundwater-stream water interactions is limited to locations and time periods were groundwater and stream water have suffi-

10 cient temperature differences. For a precise quantification of fluxes, we suggest that the average temperature difference during a measuring campaign should be between 5 and 10˚C. In our approach the conceptualization of water fluxes in the streambed is based on the assumption of vertical flow. In streams with intense non-vertical hyporheic flow in the streambed, the presented approach may not be valid. At locations

15 with a very high groundwater discharge, streambed temperatures can be nearly equal to groundwater temperature. If the flux is doubled or tripled, the temperature will remain the same (Conant, 2004). A similar behaviour occurs for high downward fluxes. In these cases, the observed streambed temperatures can be essentially equal to stream water temperatures. At these particular locations the calculated fluxes would represent

20 the minimum flux but the true fluxes could be higher. Streambed temperatures cannot be used for a reliable quantification of the water fluxes at these locations. The presented method focuses on spatial patterns of groundwater-stream water interactions. Temporal changes of flow conditions in the streambed are beyond the scope of this approach.

7   Conclusions

We measured streambed temperatures at depths of 0.1, 0.15, 0.2, 0.3 and 0.5 m along a 220 m long reach of an artificial stream. Based on the observed temperatures, the vertical water fluxes were estimated applying a one-dimensional analytical solution of the heat-advection-diffusion-equation. As temperature can be inexpensively and easily measured, hundreds of measurements can be taken to draw a high resolution picture of groundwater-stream water interactions on the scale of stream reaches. The simple concept of relating streambed temperatures to spatial differences of vertical water flux might be subject to several limitations and uncertainties but provides a reasonable agreement between simulated and observed temperatures. Furthermore, the independent results of Darcy's law calculations based on streambed piezometer data confirmed the fluxes derived from the temperature profiles.

Although the artificial streambed at our study site appears to be relatively homogeneous in comparison to natural streams, a high spatial heterogeneity of groundwater-stream water interactions was observed. Only 20% of the total length contributes to 50% of the total groundwater discharge to the stream. A significant downwelling of streamwater was not observed.

Investigations aiming at characterization of groundwater surface interaction can benefit from using multiple methods and techniques. The quantification of water fluxes through the streambed is of particular importance when mass fluxes of solutes and contaminants at the interface between groundwater and surface water are of interest. In cases of groundwater contamination, high groundwater discharge locations will contribute a great extent to the contaminant input into the stream. It is essential that these locations can be identified precisely on river segments to 1 km length to assess, for instance, the potential impact of large scale groundwater contamination on the stream. We consider streambed temperature measurements to be a useful tool to gain insight into the spatial heterogeneity of fluxes along a stream reach. Because of its proven effectiveness, this method can be applied on a field site before other methods are used

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for choosing the locations of additional instrumentation.

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Table 1. Hydraulic conductivities, hydraulic gradients and vertical fluxes obtained from slug-tests using streambed piezometers.

Name	Horizontal Hydraulic Conductivity K_h ms^{-1}	Vertical Hydraulic Gradient	Vertical Hydraulic Conductivity K_v ms^{-1} for K_h/K_v=3	Vertical Flux q_z Lm^{-2}d^{-1} for K_h/K_v=3	Vertical Hydraulic Conductivity K_v ms^{-1} for K_h/K_v=10	Vertical Flux q_z Lm^{-2}d^{-1} for K_h/K_v=10
P1	2.38 E–05	0.070	7.92 E–06	47.9	2.88 E–06	17.4
P2	5.82 E–05	0.295	1.94 E–05	494.6	7.04 E–06	179.5
P3	1.99 E–05	0.113	6.63 E–06	64.4	2.41 E–06	23.4
P4	1.20 E–05	0.113	4.01 E–06	39.0	1.46 E–06	14.1
P5	6.87 E–05	0.245	2.29 E–05	484.6	8.31 E–06	175.9
P6	1.36 E–05	0.063	4.52 E–06	24.4	1.64 E–06	8.9
P7	5.82 E–05	0.043	1.94 E–05	71.9	7.04 E–06	26.1

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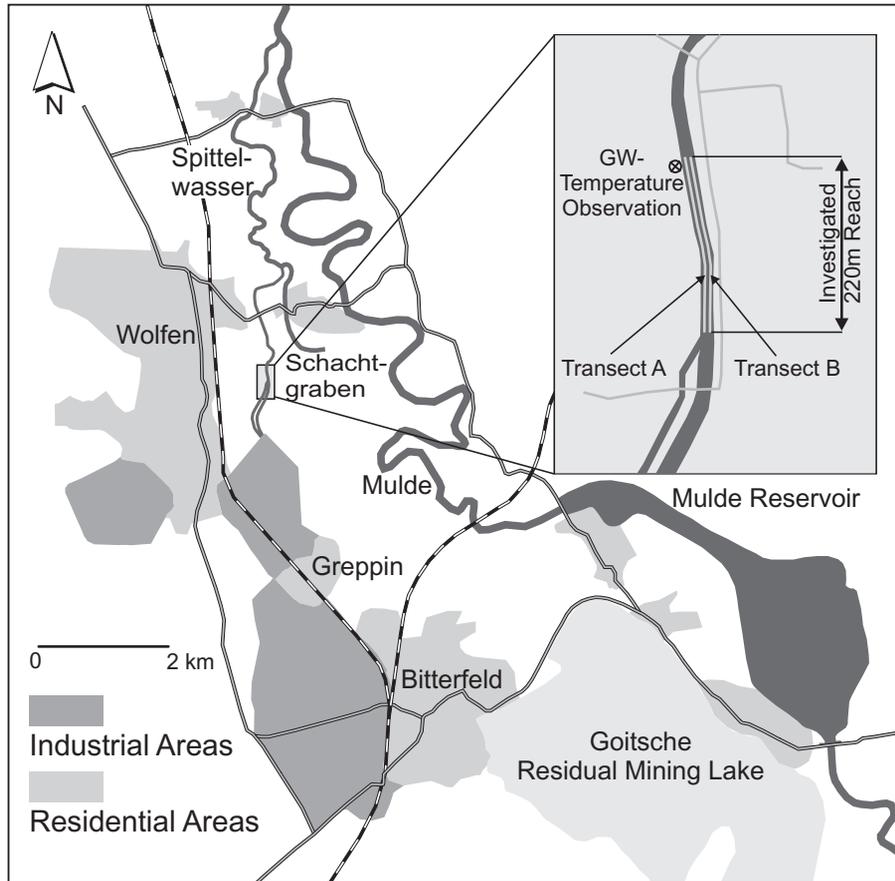
Fig. 1. Overview of the Bitterfeld-Wolfen region, showing the location of the Schachtgraben and illustrating the position of the investigated transects.

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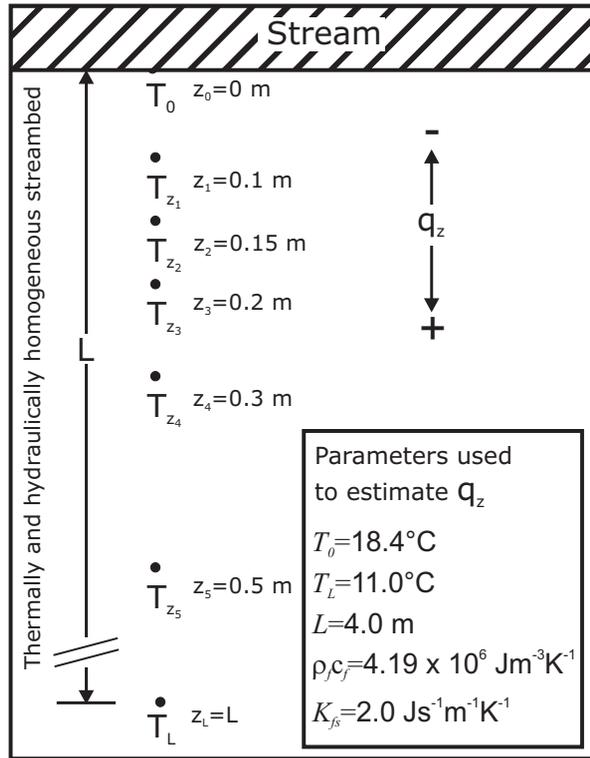
Fig. 2. Concept of vertical temperature profiles, boundary conditions and parameters used for the analytical model.

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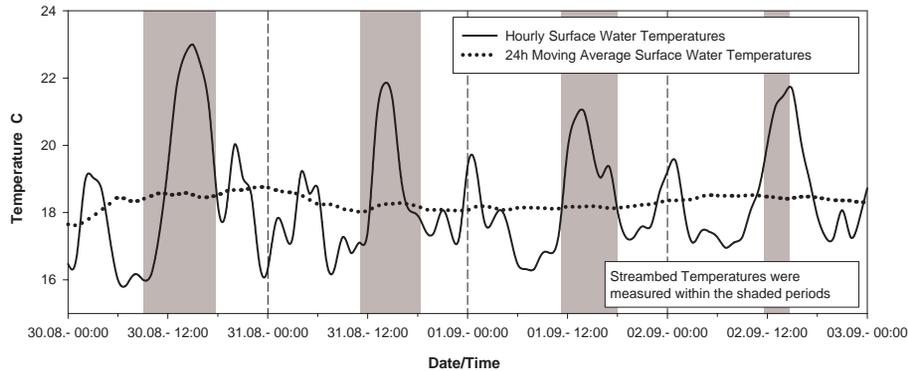
Fig. 3. Surface water temperatures during the four day measurement campaign in August/September 2005.

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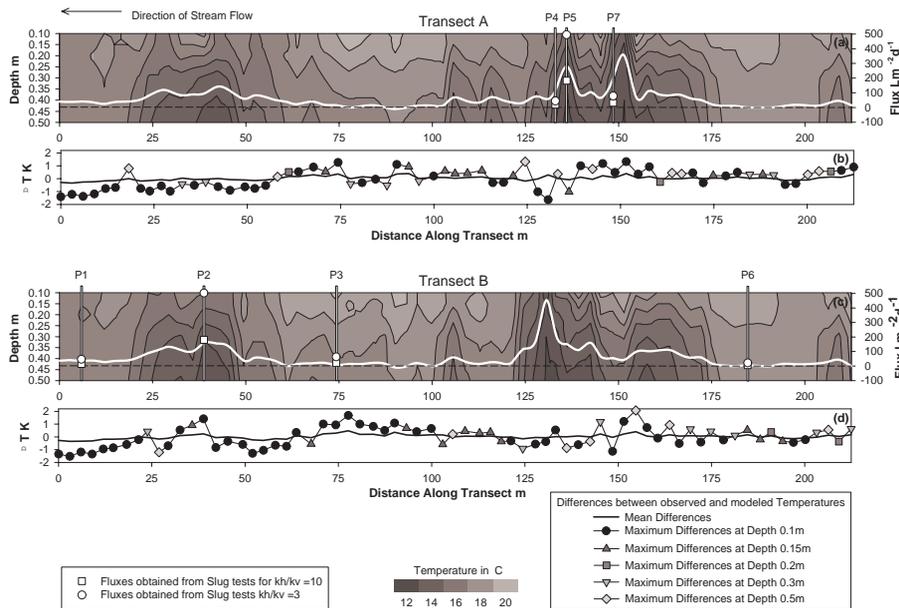
Fig. 4. Temperature distribution and temperature based fluxes and the locations of streambed piezometers and the fluxes from Darcy's law calculations at each piezometer for different anisotropy ratios along Transects A **(a)** and B **(c)**:

Mean and maximum differences between observed and simulated temperatures for each temperature profile for Transects A **(b)** and B **(d)** The maximum difference is given with the respective depth.
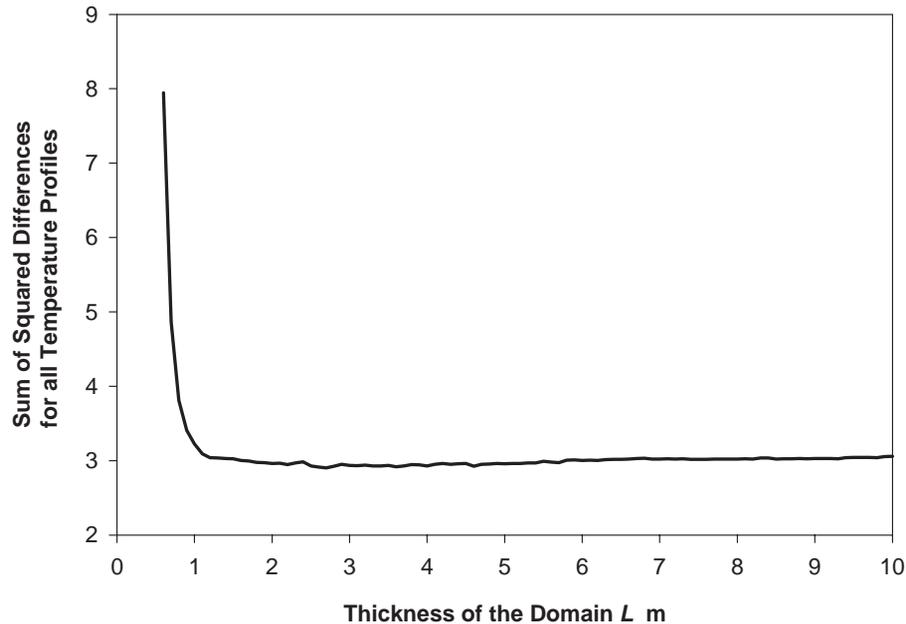
Fig. 5. Sum of squared errors of all temperature profiles vs. the length of the domain L. The results show that for the given parameter set, the quality of fit and the derived vertical fluxes are essentially constant for $L > 1.0$ m.

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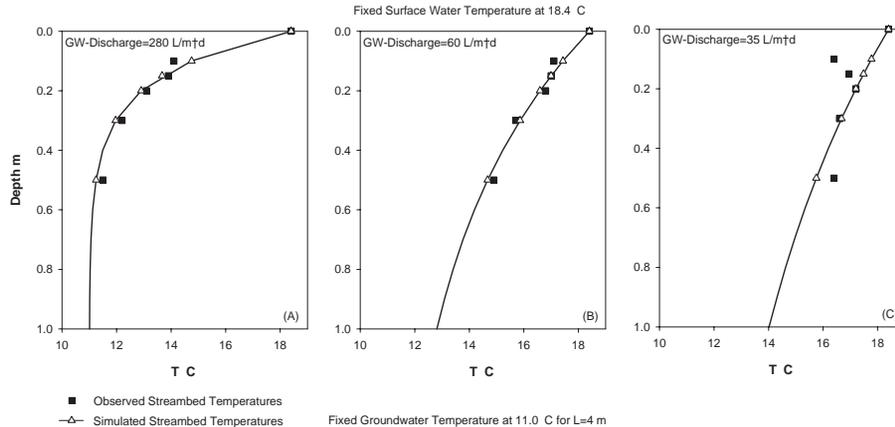
Fig. 6. Comparison of observed and simulated streambed temperatures for three example profiles. The illustrated profiles represent high **(a)**, medium **(b)** and low **(c)** groundwater discharges and are located proximal to the positions of streambed piezometers P5 (a), P4 (b) and P1 (c).
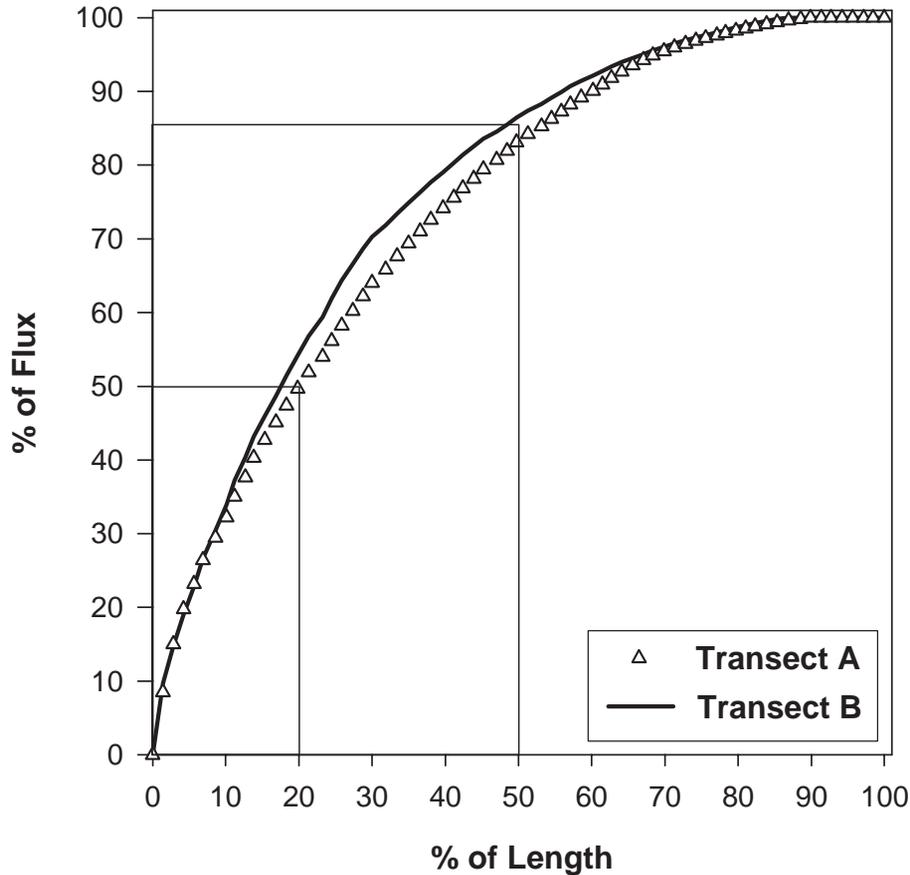
Fig. 7. Percentage of flux vs. percentage of length of the Transects A and B. Approximately 50% of the total flux occur on 20% of the total length.