

# Heat tracing to examine seasonal groundwater flow beneath a low-gradient stream in rural central Illinois, USA

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**Abstract** The thermal profile of a streambed is affected by a number of factors including: temperatures of stream water and groundwater, hydraulic conductivity, thermal conductivity, heat capacity of the streambed, and the geometry of hyporheic flow paths. Changes in these parameters over time cause changes in thermal profiles. In this study, temperature data were collected at depths of 30, 60, 90 and 150 cm at six streambed wells 5 m apart along the thalweg of Little Kickapoo Creek, in rural central Illinois, USA. This is a third-order low-gradient baseflow-fed stream. A positive temperature gradient with inflection at 90-cm depth was observed during the summer period. A negative temperature gradient with inflection at 30 cm was observed during the winter period, which suggests greater influence of stream-water temperatures in the substrate during the summer. Thermal models of the streambed were built using VS2DHI to simulate the thermal profiles observed in the field. Comparison of the parameters along with analysis of temperature envelopes and Peclet numbers suggested greater upwelling and stability in temperatures during the winter than during the summer. Upwelling was more pronounced in the downstream reach of the pool in the riffle and pool sequence.

**Keywords** Groundwater/surface-water relations · Numerical modeling · Thermal conditions · VS2DH · USA

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## Introduction

The use of streambed temperatures to study groundwater flow has advanced in the last five decades. One of the earliest studies by Stallman (1963) utilized temperature measurements to solve the inverse problem of groundwater velocity and hydraulic conductivity. After interest in thermal transport revived in the late 1980s, streambed temperatures were used to identify gaining and losing portions of small creeks in northern Indiana, USA (Silliman and Booth 1993). Time-series measurements of sediment temperature and water temperature were compared to identify regions of groundwater inflow and outflow relative to surface waters. Further work was done to quantify the downward flux through the streambed (Silliman et al. 1995). Groundwater and surface-water interaction has since gained increasing attention with several authors using different techniques to demonstrate groundwater and surface-water interactions (Kenoyer and Anderson 1989; Winter 1986). Since heat is an important nonreactive, naturally occurring, robust tracer (Constantz and Stonestrom 2003), new observation and modeling techniques employing heat as a tracer have led to cost-effective and accurate methods of analyzing groundwater–stream interactions (i.e. Briggs et al. 2014; Caissie et al. 2014; Constantz 2008; Swanson and Cardenas 2010).

Groundwater temperature data and associated tools have been used in a number of applications to investigate stream losses in ephemeral streams (Constantz et al. 2002, 2001; Constantz and Thomas 1996), to identify and quantify seepage (Bartolino and Niswonger 1999; Briggs et al. 2012; Conlon et al. 2003; Hoffmann et al. 2003; Prudic et al. 2003; Silliman and Booth 1993; Silliman et al. 1995; Suzuki 1960; Swanson and Cardenas 2010), to delineate groundwater flow patterns (Fanelli and Lautz 2008; Peterson and Sickbert 2003), to calculate groundwater-flow velocity (Beach and

Peterson 2013; Gordon et al. 2012; Hatch et al. 2006; Swanson and Cardenas 2011), and to measure hydraulic conductivity of bed sediment (Bartolino and Niswonger 1999; Constantz et al. 2003; Lapham 1989; Su et al. 2004). Changes in temperature profile within a streambed as a result of seasonal changes are not well studied. Inherent change in thermal and hydraulic properties due to seasonal changes in temperature (Genereux et al. 2008; Hatch et al. 2010) can be expected to cause changes in the temperature profile of the streambed. A method to quantify surface-water/groundwater interaction was developed using time series analysis of streambed thermal records from known depths (Hatch et al. 2006). With growing confidence in the methods developed, use of sediment thermal data to determine reasonable groundwater flux rates is likely to be an effective field method that is quicker and more cost effective than setting up piezometers or seepage meters (Schmidt et al. 2007).

Focusing on identifying the seasonal changes in hydraulic and thermal parameters and in determining the dominant parameters for respective temperature profiles, a two-dimensional (2D), energy-transport model, VS2DH, was used to simulate streambed temperatures for weeklong nonstorm time periods in spring, summer and winter. The simulations provided data for the comparison of thermal and hydraulic parameter values between those time periods. Specifically, the study explored the following two questions: (1) What is the nature of seasonal change in streambed thermal profile? and (2) How do the hydraulic parameters of the streambed and the thermal transport mechanisms change seasonally?

## Materials and method

### Study area

Model development was based upon a stretch of the Little Kickapoo Creek (LKC), a third-order low-gradient perennial stream, in rural central Illinois (Fig. 1). The site has been well described in previous work (Beach and Peterson 2013; Peterson and Benning 2013; Peterson and Sickbert 2006; Peterson et al. 2008; Sickbert and Peterson 2014; Van der Hoven et al. 2008), but relevant details are provided here. Situated in the extent of Wisconsinan Glaciation, the study site is south of the Bloomington moraine and overlies valley-train outwash sand and gravel of the Mackinaw Member of the Henry Formation. In the area, the Henry Formation is 5–7 m thick and is overlain by about 2 m of silt and clay with sand of the Cahokia Alluvium. Clay-rich glacial till of the Tiskilwa Formation of the Wedron Group underlies the Henry Formation.

A 30-m segment was selected where meandering was minimal in order to reduce any effects from hyporheic flow through meander necks (Peterson and Sickbert 2006; Van der Hoven et al. 2008). The stream cuts a narrow channel with

steep banks, 2–3 m from bank tops to the streambed, through the Cahokia Alluvium. The stream resides along the top of Henry Formation. The streambed substrate is composed of mostly coarse sand and gravel with interstitial finer sediments and some cobbles. The unsorted sand and gravel of the Henry Formation hosts an unconfined aquifer that is in direct communication with the stream. The underlying Tiskilwa Formation serves as the basal confining unit for the Henry Formation.

### Climate and precipitation

Central Illinois has hot, wet summers and cold, dry winters. Based on average monthly temperatures and precipitation between 1971 and 2000, July and August are the two hottest months with average temperatures over 22.5 °C. January and February are the two coldest months with average temperatures less than –2.5 °C (Table 1). Spring and fall have moderate temperatures. Late spring–early summer on average receives more precipitation than other periods. Average winter precipitation is the least among the seasons. Average monthly precipitation between 1971 and 2000 was highest for May and June with more than 100 mm of precipitation and lowest for January and February with less than 45 mm of precipitation. In 2009, April received over 150 mm of precipitation. Precipitation for July was over 90 mm and that for August was over 120 mm. Monthly precipitation for January of 2010 was less than 40 mm and made for one of the driest months during the study period.

### Streambed wells and piezometers

Six streambed wells made up of 3.81-cm PVC pipes were installed in the thalweg at an interval of 5 m (Fig. 2a). In each streambed well, temperature loggers were mounted at depths of 30, 60, 90 and 150 cm (Fig. 2b). Holes drilled at corresponding depths allowed for thermal equilibrium with the surrounding sediment. The loggers were separated using foam sealant. Two temperature loggers were installed within the stream, at wells 1 and 6, to measure surface-water temperatures; an additional temperature logger was emplaced in a nearby groundwater well (Fig. 1), screened in the Henry Outwash at a depth of 10 m to measure groundwater temperature. A stilling well was installed to measure stream stage. StoyAway TidbiT Temperature Loggers (accuracy:  $\pm 0.2$  °C and resolution: 0.16 °C at 25 °C) and HOBO® Pendant Temperature and Light Data Loggers (accuracy:  $\pm 0.47$  °C and resolution: 0.10 °C at 25 °C) recorded temperature at 15-min intervals. Data were collected from February 2009 to March 2010. The first set of wells and stream temperature loggers was installed in February of 2009 and retrieved in August 2009 and replaced with a second set of wells and

loggers. The second set of wells and loggers were retrieved in early March of 2010.

Data were analyzed to identify any problems in data collection and to select appropriate time periods for thermal modeling. Stage data, collected using Solinst Levellogger® (temperature accuracy  $\pm 0.05$  °C and a resolution 0.01 °C at 20 °C, and water level resolution of 0.002 m) installed in the stilling well, were used to select week-long periods in spring, winter, and summer that did not have storm events. Hyporheic zone and the thermal regime of the streambed behave differently between storm events and standard flow (Owari 2010), making it necessary to avoid the effects of storm events in the streambed to successfully study seasonal changes in the thermal profile of the streambed.

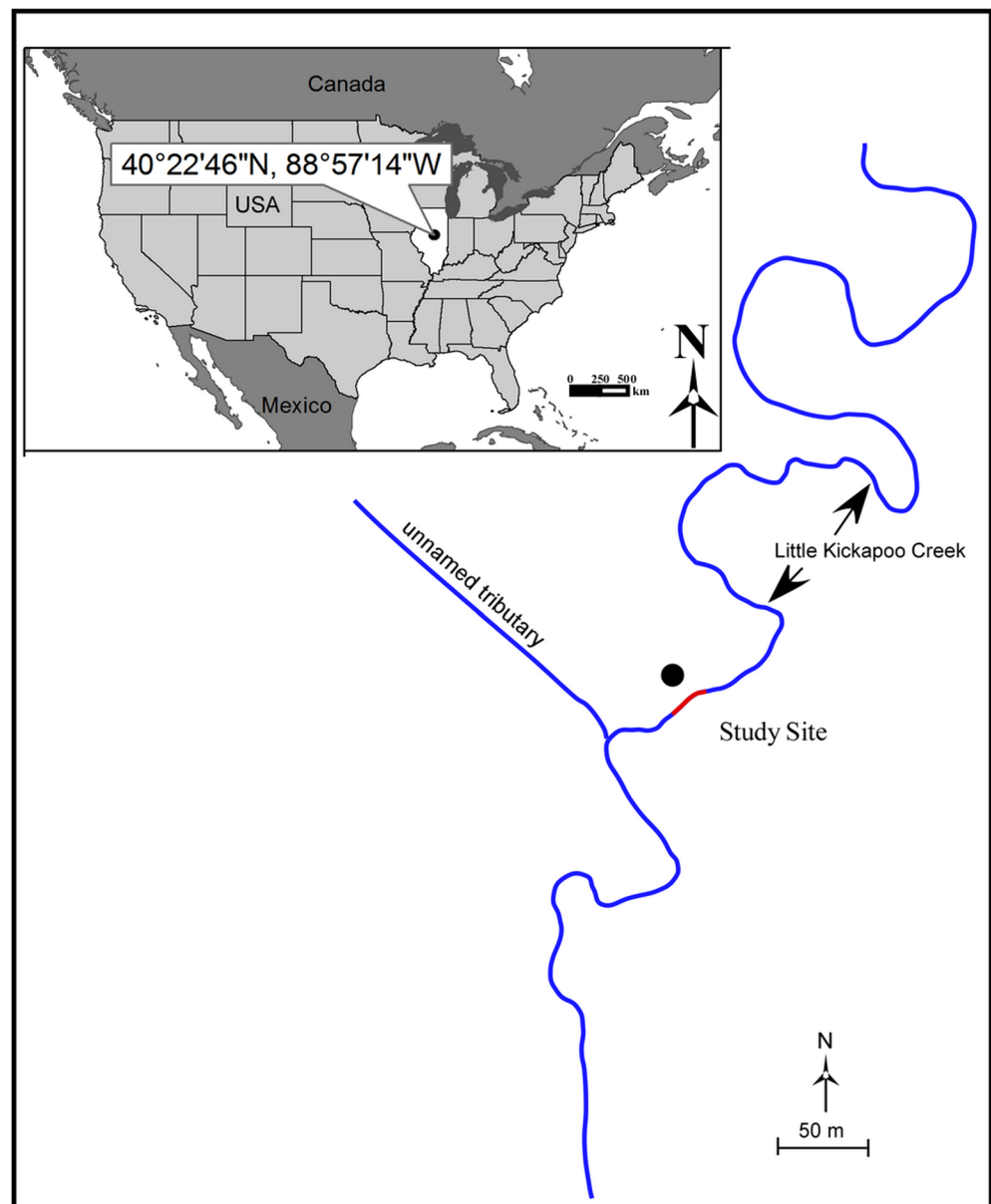
### Thermal modeling

VS2DH simulates 2D heat and groundwater flow (Healy and Ronan 1996). The model makes use of the finite difference method to solve the advection-dispersion equation (Eq. 1):

$$\frac{\partial}{\partial t} [\theta C_w + (1-\varphi)C_s]T = \nabla \cdot K_T(\theta)\nabla T + \nabla \cdot \theta C_w D_H \nabla T - \nabla \theta C_w v T + q C_w T^* \quad (1)$$

where  $\theta$  is volumetric moisture content,  $\varphi$  is porosity,  $C_w$  is heat capacity of water,  $C_s$  is heat capacity of dry solid,  $T$  is temperature,  $K_T$  is thermal conductivity of water and solid matrix,  $D_H$  is coefficient of hydrodynamic dispersion,  $q$  is rate

**Fig. 1** Site location,  $40^\circ 22' 46''N$ ,  $88^\circ 57' 14''W$  in central, Illinois, USA, and local site layout. Flow in Little Kickapoo Creek is from north to south. The red segment of the stream highlights the section in the map above. The black circle represents the groundwater well



**Table 1** Monthly precipitation and temperatures for the study area site. 1971–2000 represents average precipitation values; values for 2009–2010 are measured

Year/s	Month											
	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec
Average precipitation (mm)												
1971–2000	38.6	42.4	76.5	90.9	108.5	101.3	95.8	93.0	84.6	66.3	81.5	73.4
2009	43.3	41.2	117.9	155.2	134.4	193.3	90.2	121.7	45.5	263.7	78.7	93.0
2010	38.6											
Average temperature (°C)												
1971–2000	-5.3	-2.6	3.7	10.3	16.6	22.1	24.0	22.9	18.7	12.1	4.3	-2.2

Data adapted from Water and Atmospheric Resources Monitoring Program (2010) and NOAA (2010)

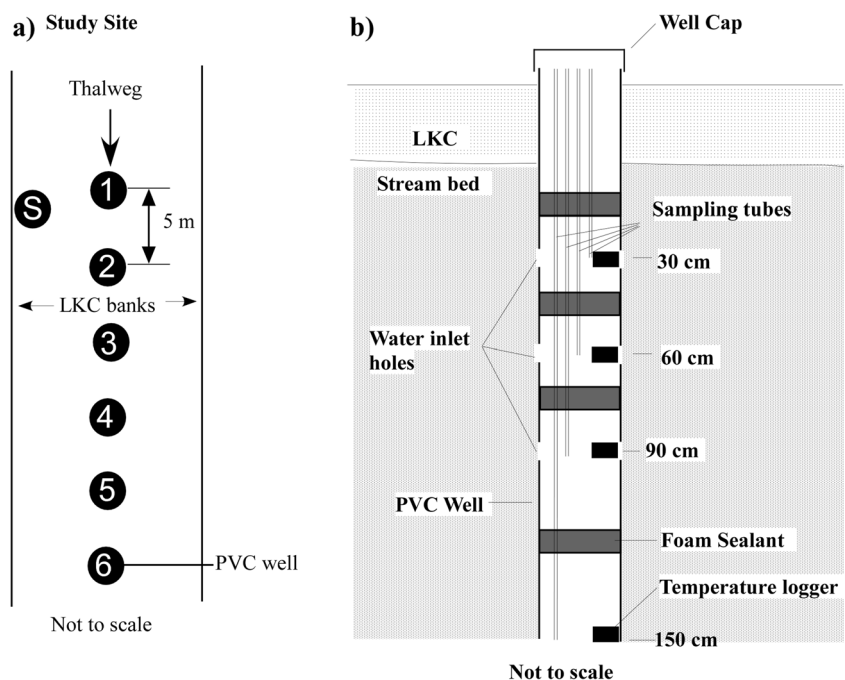
of fluid source and  $T^*$  is temperature of fluid source or sink. The terms on the right side of the Eq. (1) represent heat change in the system due to conduction, dispersion, advection and sink or source respectively. VS2DH has been effectively employed to describe heat transport and to determine hydraulic parameters for numerous study sites near streams (Constantz et al. 2002; Essaid et al. 2008; Su et al. 2004)

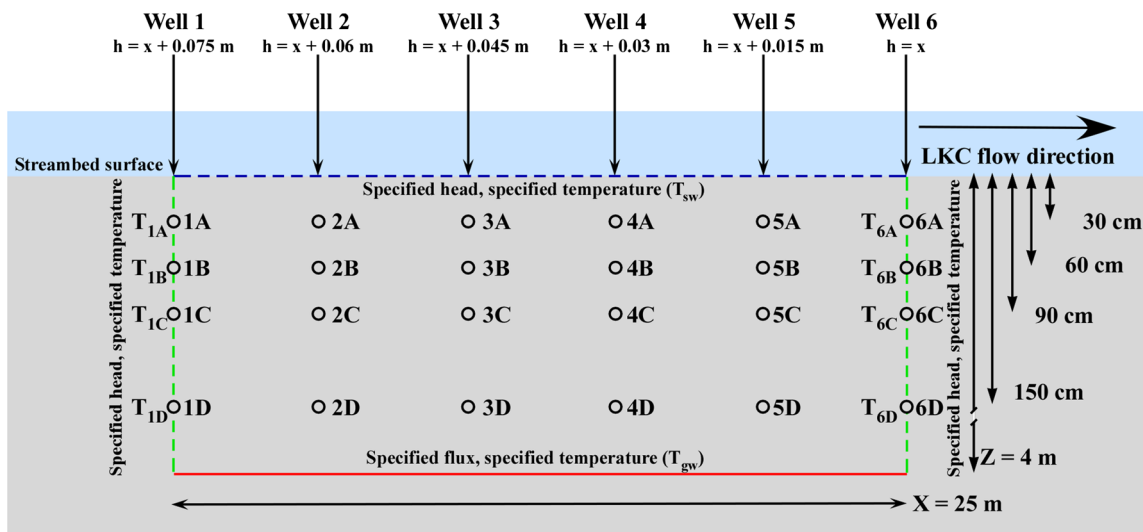
Thermal modeling was based on the conceptual model that assumed a homogeneous medium with a stream gradient of 0.003 (Fig. 3). Boundary conditions were updated using a time step of 4 h. Temperature data from wells 1 and 6 provided lateral boundary conditions for the thermal models. Temperatures from wells 2–5 were used to specify the initial temperature contours in the model and were used for calibration. Streambed topography changes represented by the scour and fill data (Bastola 2010) were used to define the top of the

domain. Total head values decreased at 0.015 m for every 5-m change in distance in the downstream direction. Thermal boundary conditions of the upper and lower boundary of the domain were defined by the temperature of the stream and groundwater respectively. Also, the thermal boundary conditions for the left and right boundaries were defined by the temperature recorded by the loggers at various depths. Hydraulic properties of medium sand provided by VS2DH were initially used in the models. Parameter values available in the literature were used to specify the thermal parameter values (Table 2). The model was set up to give output values every 15 min, thus enabling comparison with the observed values at the same scale.

Model optimization was based on reducing the mean absolute error (MAE), which provided the best-defined minimum errors for the parameters, by making proper adjustments to

**Fig. 2** **a** Schematic setup of hyporheic wells (1–6) and stilling well (S) in a map view. Well 1 is the upgradient well. **b** Design of individual hyporheic wells. LKC Little Kickapoo Creek





**Fig. 3** Conceptual model of the streambed showing location of the temperature loggers. Solid (red) and dashed (blue and green) lines represent the extent of the domain used in the numerical model. The dashed blue line represents boundary conditions of specified head and specified temperature defined by the surface water conditions. Dashed

green lines represent boundary conditions of specified head and specified temperature defined by the temperatures recorded in wells 1 and 6. The solid red line represents boundary conditions of specified flux and specified temperature defined by the groundwater temperature

parameters on a trial and error basis. MAEs were calculated using a program written in MATLAB. MAE graphs were constructed for each parameter and the parameter value associated with the vertex of the error curve was chosen within the range of plausible values. The models were adjusted for base flux, followed by adjustments to horizontal hydraulic conductivity,

saturated thermal conductivity, dry heat capacity of the medium, vertical to horizontal conductivity ratio, and porosity until the best-fit model for the time period was obtained. Best-fit models were similarly obtained for summer and winter model weeks. Finally, the changes in thermal profiles were compared between the three models using the extinction depth, changes

**Table 2** Representative value ranges for hydraulic and thermal parameters

Parameter	Value range	Initial values for models
Horizontal hydraulic conductivity ( $\text{m s}^{-1}$ )	–	$4.63 \times 10^{-3}$
Fine sand	$2 \times 10^{-7} - 2 \times 10^{-4a}$	–
Coarse sand	$9 \times 10^{-7} - 6 \times 10^{-3a}$	–
Gravel	$3 \times 10^{-4} - 3 \times 10^{-2a}$	–
Heat capacity ( $\text{J m}^{-3} \text{K}^{-1}$ )	–	–
Dry solid	$1.1 \times 10^6 - 1.3 \times 10^{6a}$	$1.2 \times 10^6$
Water	$4.2 \times 10^{6a}$	$4.2 \times 10^6$
Saturated solid	$2.5 \times 10^6 - 3.2 \times 10^{6a}$	–
Porosity	–	0.375
Sand	0.25–0.50 <sup>a</sup>	–
Sand and gravel	0.15–0.35 <sup>a</sup>	–
Dispersivity (m)	–	0.005
Vertical anisotropy ( $K_h/K_z$ )	–	10
Sand and gravel	3–5 <sup>b</sup>	–
Fine/medium sand	10–30 <sup>b</sup>	–
Fine sand and silt	30–100 <sup>b</sup>	–
Saturated thermal conductivity ( $\text{W m}^{-1} \text{K}^{-1}$ )	–	2.1
Saturated sediments (sand, loam, etc.)	1.4–2.2 <sup>a</sup>	–
Average saturated soil	2.9 <sup>a</sup>	–

<sup>a</sup> Values adapted from Weight (2008)

<sup>b</sup> Values adapted from Masterson et al. (2007)



in thermal and hydraulic parameters and changes in sensitivity of these parameters. Sensitivity analysis performed on the calibrated models examined fixed percent increment changes for all the parameters.

## Results

Over the 1-year duration of the study, stream-water temperature followed a sinusoidal curve with diurnal sinusoidal temperature imprints (Fig. 4). The yearly temperature range of about 30 °C was observed with highest temperatures observed between June and August, and lowest temperatures observed between December and February. This pattern closely follows the air temperature patterns presented in Table 1.

Stream stage was higher in spring and winter compared to summer and fall. The site was subjected to a series of storm events with periods of recession between the storm events. Recession periods from storm events to base flow were longer in winter compared to spring or fall. Despite higher average precipitation associated with summer storm events, rises in stage were smaller compared to spring, fall and winter.

Week-long periods in spring (April 1–7, 2009), summer (Jul 27–Aug 2, 2009) and winter (Jan 15–21, 2010) were selected where storm events were minor or absent and will be referred to as the spring, the summer, and the winter model periods hereafter. Stream temperature was on average about 8 °C during the spring model period and about 22 and 2 °C during the summer and winter model periods respectively (Table 3). The temperature of groundwater remained around 11 °C throughout the model periods with subtle changes over the seasons (Fig. 4). Lowest groundwater temperatures were

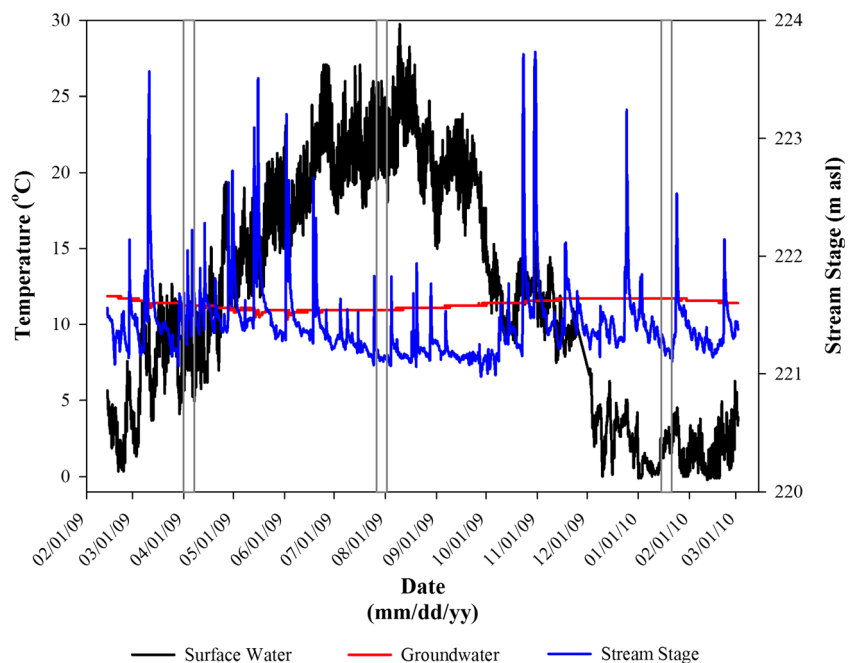
seen in early June and maximal temperatures occurred in December, representing a lag time of about 6 months with the air temperature.

During the spring model period, groundwater was on average 3 °C higher than surface water. Diurnal fluctuations give the week a stream temperature range of about 6 °C. The range of streambed temperature drastically decreased to about 0.7 °C for 30 cm depth and to less than 0.3 °C at 90 cm depth (Table 3). There were two small storm events during the week-long period that altered the diurnal fluctuations of stream temperatures. The effect of the second storm event is more noticeable with a subdued daily maximum on day 6 (Fig. 5a). A major storm event occurred about 3 weeks before the spring model period. The possible effect of these storm events on streambed will be discussed later.

For the summer model period, stream water had an average temperature 11 °C higher than the groundwater. Subsurface temperature fluctuations of over 1.2 °C were observed at 30 cm depth and about 0.3 °C at 150 cm depth. While groundwater temperature remained steady, stream temperatures experienced strong diurnal fluctuations resulting in a weekly temperature range of about 8 °C with an average temperature of about 22 °C. Average temperature inflection was observed at 90 cm. The stream remained at baseflow, experiencing no storms during the period (Fig. 5b). Groundwater temperatures remained steady during the period.

During the winter time period, the stream temperature was near 0 °C, and diurnal fluctuations were minimal, leading to a weekly temperature range of 2 °C. On average, groundwater temperatures were about 10 °C higher than the stream temperatures (Fig. 5c). Sub-surface temperature fluctuations were comparatively smaller than the summer model week, with

**Fig. 4** Stream temperature, stream stage, and groundwater temperature over the period of data collection. The narrow boxes outline the weeks simulated with the spring model, summer model, and winter model



**Table 3** Temperature ranges and average temperatures for different locations during the three model periods

Location	Spring model week		Summer model week		Winter model week	
	Range (°C)	Avg (°C)	Range (°C)	Avg (°C)	Range (°C)	Avg (°C)
Surface water	6.59	8.22±1.63	7.96	22.13±2.02	2.81	2.03±0.63
30 cm	0.69	8.63±0.20	1.22	18.26±0.28	0.53	6.12±0.15
60 cm	0.46	9.34±0.08	0.33	16.49±0.09	0.27	7.57±0.08
90 cm	0.25	9.40±0.04	0.24	15.26±0.09	0.13	8.47±0.03
150 cm	0.05	9.59±0.03	0.34	13.46±0.09	0.14	9.51±0.05
Groundwater	0.31	11.26±0.07	0	10.96	0	11.73

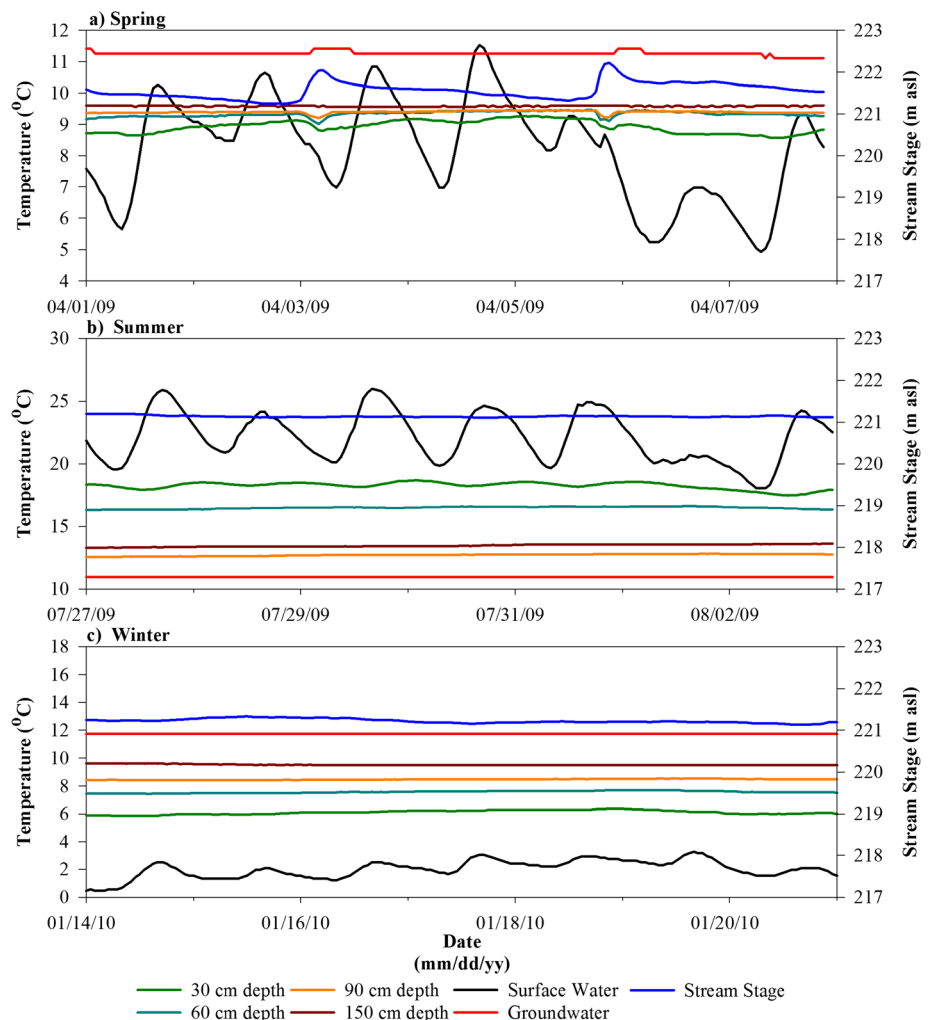
fluctuations of about 0.5 °C at 30 cm and of less than 0.15 °C at 90 cm depth. Average temperature inflection was observed at 30 cm depth. No storm events were seen during this time period.

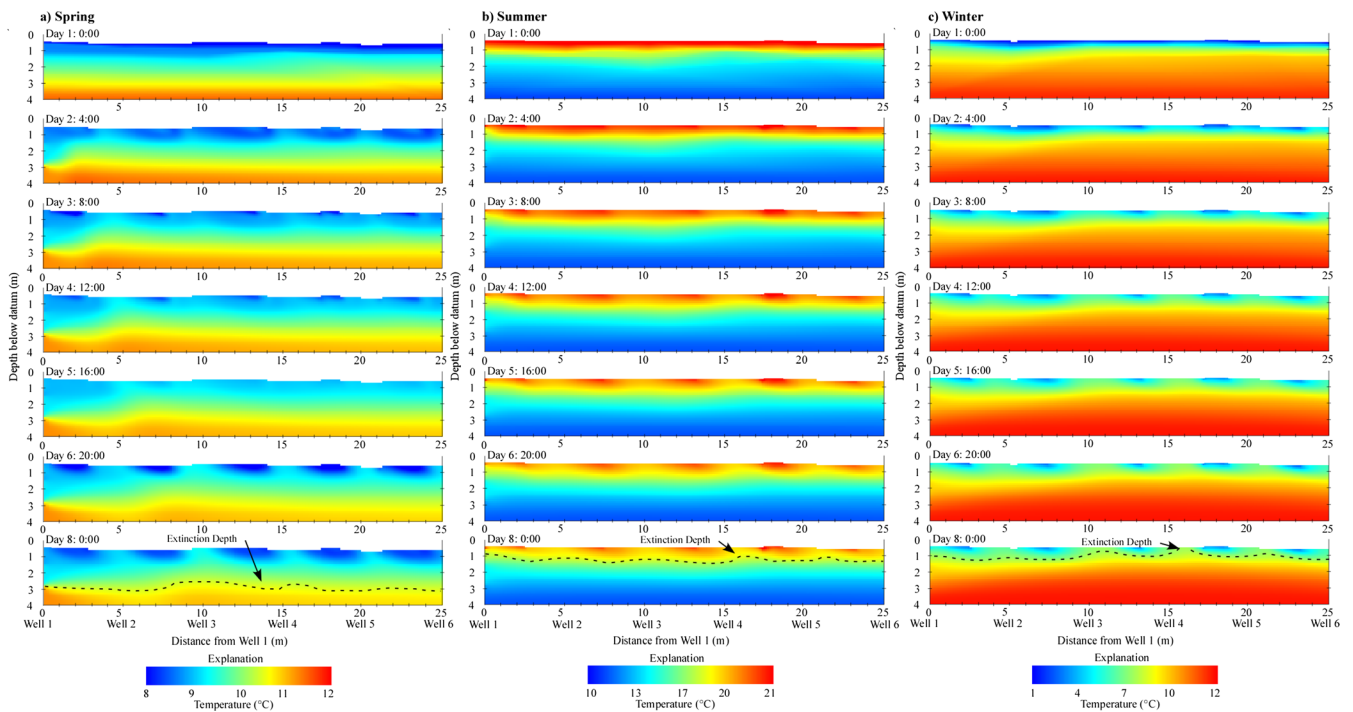
**Temperature models**

One of the most obvious contrasts represented by the simulated models is the difference in thermal gradient in the three

model periods (Fig. 6). A low thermal gradient with a temperature difference of 3 °C between surface and groundwater was observed for the spring time period, while comparatively large gradients with temperature ranges of about 11 and 10 °C were observed for the summer and winter time periods. A reversal of thermal gradient occurred from the winter model to the summer model with the groundwater at low thermal potential during the summer time period and the surface water at low thermal potential during the winter time period. The

**Fig. 5** Stage, surface and subsurface temperatures during the a spring, b summer, and c winter model weeks





**Fig. 6** Best fit VS2DH models simulating streambed temperature distribution for the **a** spring, **b** summer, and **c** winter model periods. *Dashed line* represents the extinction depth for each period and *arrows* show the extinction depth

temperature of groundwater remained constant at around 11 °C throughout the model periods. The models simulated stream induced hyporheic flow into the streambed across upper boundaries where boundary conditions in total head change. Stream induced temperature fluctuations were limited to the shallow substrate in all three models.

The transition between surface-controlled temperature and groundwater-controlled temperature may be thought of as the extinction depth as defined by Briggs et al. (2014). The extinction depth for the spring period was relatively deeper compared to those of summer and winter periods, while that of winter period was the shallowest (Fig. 6). Extinction front controls include (1) magnitude and direction of the fluid flux, (2) the thermal properties of the medium, and (3) the period of the surface temperature signal. The extinction depths allow for

a comparison of the effect of stream water temperature on shallow streambed temperatures.

A small model error was obtained for the spring period with a MAE of 0.2 °C compared to MAE of 0.4 and 0.3 °C for the summer and winter periods (Table 4). The MAEs for all three models are comparable to the range of logger accuracy, i.e. 0.2–0.47 °C. Lower observed temperature compared to the simulated temperature at 60 cm depth of well 4 and the model’s inability to represent daily fluctuations in temperature at 30 cm depth seems to have contributed to the MAE for the summer model. Similar difficulty in representing temperatures at 30 cm depth can also be observed to some extent in spring and winter models. Lower observed temperature compared to simulated temperature at 30 and 60 cm of well 2 contributes to the higher MAE for the winter model.

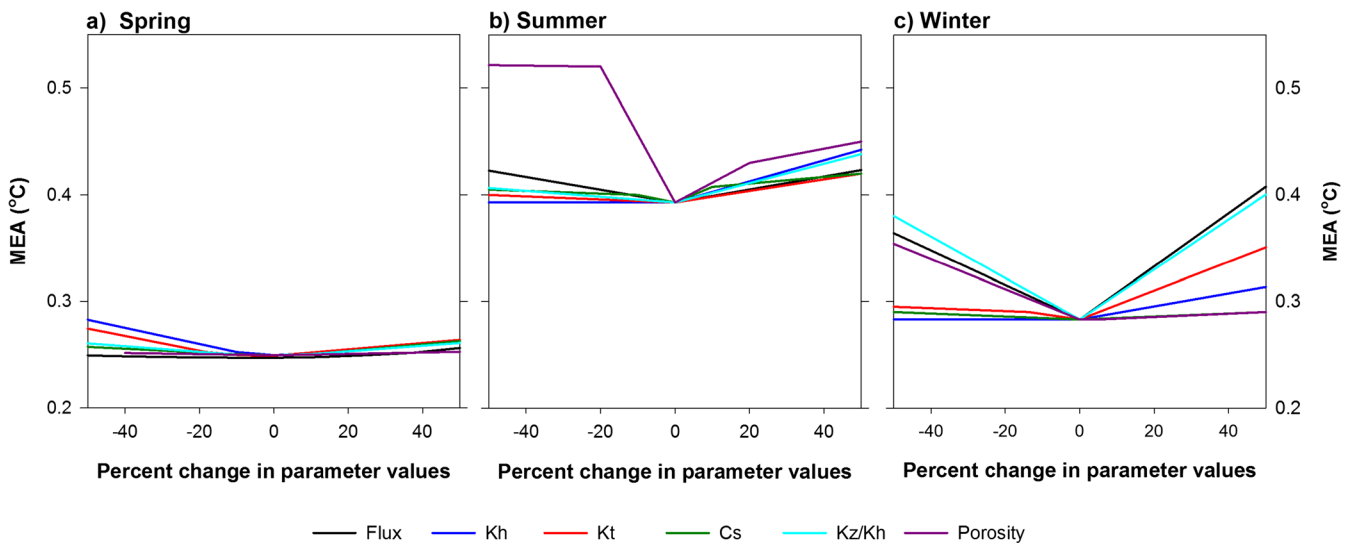
**Table 4** Parameter values for the best fit models for the spring, summer and winter time periods

Parameter	Spring	Summer	Winter
Vertical flux (m s <sup>-1</sup> )	1.5×10 <sup>-7</sup>	7.58×10 <sup>-7</sup>	2.3×10 <sup>-6</sup>
Horizontal hydraulic conductivity (m s <sup>-1</sup> )	2.8×10 <sup>-3</sup>	1.1×10 <sup>-3</sup>	2.0×10 <sup>-3</sup>
Sat. thermal conductivity (W m <sup>-1</sup> K <sup>-1</sup> )	2.51 <sup>a</sup>	1.38 <sup>b</sup>	1.38 <sup>b</sup>
Dry heat capacity (J m <sup>-3</sup> K <sup>-1</sup> )	1.1×10 <sup>6</sup>	1.3×10 <sup>6</sup>	1.3×10 <sup>6</sup>
$K_h/K_z$	9.1	100 <sup>a</sup>	28
Porosity	0.22	0.5 <sup>a</sup>	0.5 <sup>a</sup>
MAE (°C)	0.2	0.4	0.3

<sup>a</sup> Indicates that the parameter value was the highest value used

<sup>b</sup> Indicates that the parameter value was the lowest value used





**Fig. 7** Comparison of observed and simulated temperatures for the depths of 30 cm (blue diamond), 60 cm (green triangle), 90 cm (red square), and 150 cm (yellow diamond) in well 3 for a spring, b summer, and c winter model periods. Dotted line represents 1:1 line

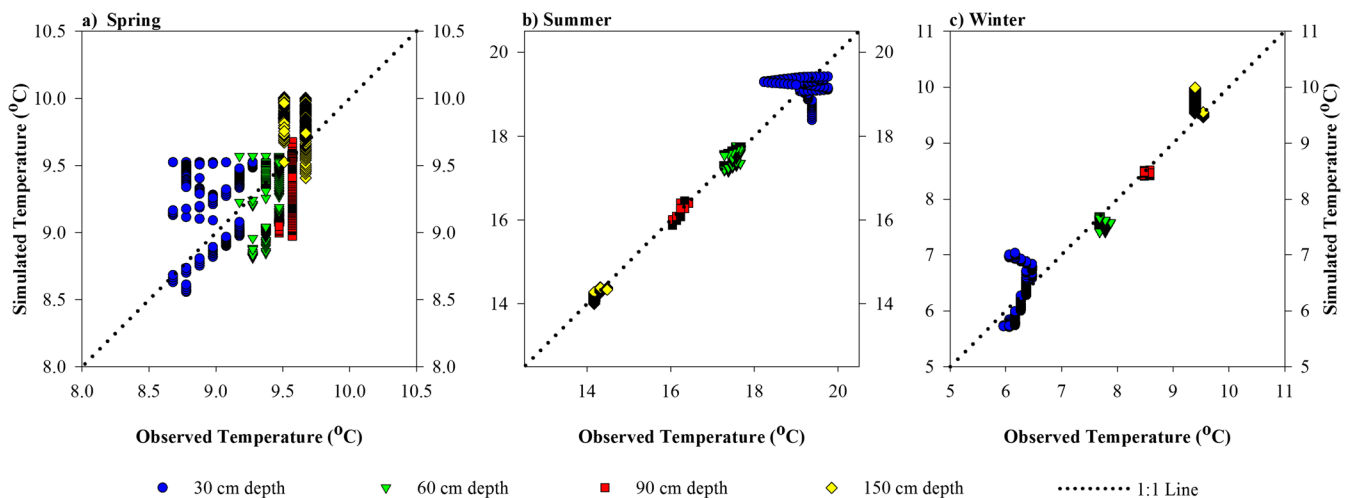
The thermal models, despite some pockets of higher errors, do an adequate job of simulating the observed temperatures (Fig. 7). Simulated temperatures better represent the temperatures measured for deeper positions. Larger deviations from observed temperatures are seen at 30 cm depth. It should be noted that for the spring model, the range of temperatures is narrow and the overlapping of simulated temperatures are likely a result of similar temperatures across the depth of the streambed.

**Parameter values and sensitivity analysis**

Comparison of parameters from the best-fit spring, summer and winter periods suggest some changes in these parameters between the models. It is, however, important to account for the sensitivity of these parameters before making interpretations. Vertical flux of over  $2 \times 10^{-6} \text{ m s}^{-1}$  was seen for the

winter model period, while fluxes of  $1.5 \times 10^{-7}$  and  $7.58 \times 10^{-7} \text{ m s}^{-1}$  for spring and summer model periods resulted in the best-fit model (Table 4). Hydraulic conductivity values were highest for spring model at  $2.8 \times 10^{-3} \text{ m s}^{-1}$  and lowest for the summer model with a value of  $1.1 \times 10^{-3} \text{ m s}^{-1}$ . Saturated thermal conductivity and dry heat capacity values have a narrow range in earth materials and are not expected to change significantly over seasons. As a result, adjustments to these values were halted if the model values exceeded the range of accepted values. The models optimized for a vertical anisotropy value of 100 for summer and values less than 30 for spring and winter. A high value for porosity was seen for the best-fit summer model.

Sensitivity of the models to the various thermal and hydraulic parameters changed from one model period to another (Fig. 8). A proper understanding of the sensitivity of the model to these parameters is critical in making proper



**Fig. 8** Model sensitivity analysis for a spring, b summer, and c winter models

interpretation of the parameter values. The spring model sensitivity was very small compared to the accuracy of the loggers used with MAE changes of less than 0.03 °C with a 50 % change in parameter values of the best-fit model. No change in MAE was observed with changes in lower values of horizontal hydraulic conductivity ( $K_h$ ) for the summer and winter models; however, a 50 % increase in  $K_h$  resulted in a MAE increase of 0.05 °C for the summer model. The summer model showed some of the highest sensitivities to lower values of porosity and higher values of thermal conductivity with MAE changes of over 0.1 °C with a 50 % change in the parameter values. Comparable sensitivities were observed for vertical flux and anisotropy in the winter model.

## Discussion

### Nature of seasonal changes

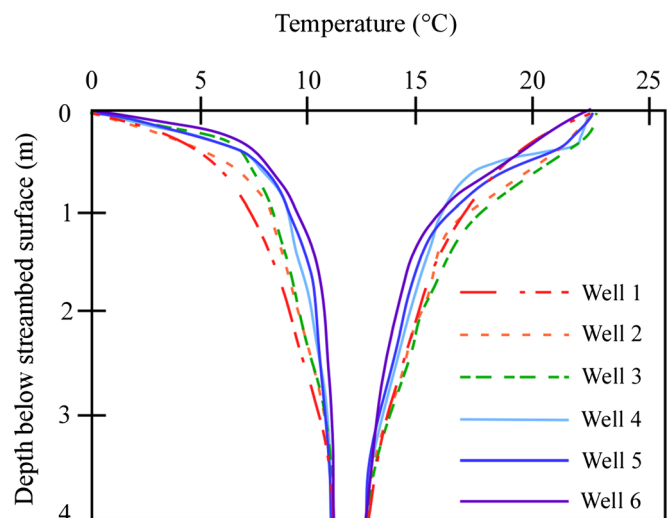
The distribution of substrate temperature is based on stream and ground temperatures, fluctuations of stream temperatures and the nature of water flow from stream water to groundwater and vice versa. The importance of stream-water fluctuations on streambed was evident by the fact that temperature fluctuations in the substrate were higher for the summer and spring than during the winter and were dependent on the range of temperature of stream water itself (Table 3). While some of the temperature fluctuation for the spring model period can be due to the small storm events that occurred during the time period, the higher fluctuations in streambed temperatures during the summer is a marked difference from the winter. Hence, the greater temperature range of over 0.3 °C observed at 150 cm depth during the summer model period suggests a greater influence of stream-water temperatures on the substrate (strong influence from downwelling) compared to less than

0.3 °C observed at 60 cm depth during the winter model period (stronger influence of upwelling).

The upward-contracting temperature envelope for the wells in the site suggests that the stretch experienced upwelling (Constantz 2008; Stonestrom and Constantz 2003) with annual temperature fluctuations of less than 8 and 3 °C at depths of 150 cm and 4 m respectively. In a system without advection, annual temperature fluctuations of about 12 °C at a depth of 150 cm and of about 4 °C at a depth of 4 m are seen (Lapham 1989). Comparison of temperature envelopes for various wells over the period of 1 year suggests a comparatively higher upwelling in wells 4–6 as compared to wells 1–3 (Fig. 9). The difference in temperature envelopes spatially is likely topographically driven because of riffle and pool sequence, although spatial heterogeneity could also contribute to the trend. Relatively higher upwelling in wells 5 and 6 falls in the downstream pool where hyporheic flows were expected to rejoin the stream (Cardenas 2007; Cardenas et al. 2004; Tonina and Buffington 2007). Less upwelling is associated with areas where stream water enters the hyporheic zone and more upwelling is associated with areas where hyporheic water reenters the stream. It can thus be implied that upwelling is influenced by topography-induced flow.

In a complimentary study during the same period, Oware (2010) found that there was significant difference in the after-storm substrate temperature change (ASSTC) from 30 to 150 cm depth during the winter, while no significant difference between the two depths were found during the summer. The similarity in ASSTC over the various depths during the summer was attributed to the warmer summer water penetrating the streambed deeper than the cold winter water. Although density differences and associated viscosity differences could be a major factor in the deeper penetration to stream water in the substrate as put forward by Oware (2010), similar roles of density and viscosity change may not be inferred to nonstorm time periods, where temperature difference in the streambed

**Fig. 9** Temperature envelopes for wells 1–6



from summer to winter at 60 cm depth is less than 10 °C and at 90 cm is only about 4 °C. A greater role in the comparatively deeper penetration of temperature fluctuations during the summer in both storm and nonstorm time periods could be attributed to the smaller upwelling during the summer that allows for deeper hyporheic flows compared to winter.

Comparison of vertical flux across the bottom boundary suggests a higher upward flux value for the winter model period. Higher stream stage, which correlates to faster stream-water velocity, has been correlated to greater upwelling in streambeds (Sickbert and Peterson 2014). While stage during the winter model period is slightly higher than that during the summer model period, no relation can be established between stage and upward flux over all three models because stage was on average higher during the spring model period than that during the winter model period (Fig. 4). Possible explanation could be found in the difference in evapotranspiration rates between summer and winter and in the nature in which recharge occurs from winter snow (Constantz 1998). The rate of evapotranspiration is very high during summer compared to winter, and the loss of groundwater from shallow aquifers to the atmosphere due to evapotranspiration can lower the water table, and consequently lower the hydraulic gradient between the aquifer and the gaining stream. This could also result in relatively smaller baseflow into the stream during summer as compared to winter.

Walton (1965) suggested that evapotranspiration in Illinois is so high during the summer that little precipitation percolates into the water table except during periods of excessive rainfall. Snow pack during the winter provides for a sustained source of recharge into shallow aquifers and can be expected to result in higher baseflow to streams even when there is no precipitation. Although no soil temperatures are available in the vicinity of the study site, soil temperatures recorded in nearby Peoria, IL for the time period between Jan 15–21 (winter model period) suggested that the ground went through diurnal freeze and thaw cycles at a depth of 0.1 m (Water and Atmospheric Resources Monitoring Program 2010). Ground freezing was absent during the model period at a depth of 0.2 m, illustrating the moderating effect of warmer groundwater on the ground surface temperatures. A detailed study on the snow melt and ground freezing patterns in the study site during the winter will be needed to better establish the role of snow pack on baseflow. Because precipitation in the region is higher during the summer than winter, the importance of decreased evapotranspiration and snowpack melting during winter on increasing baseflow could be higher than the vertical flux values reflect.

Specifically, the temperature envelope on the winter side showed inflection at 30 cm depth as compared to the 90 cm depth for the summer side, suggesting a greater influence of groundwater temperatures on the streambed substrate than during the summer. This difference was also represented by

the models (Fig. 6b,c) where streambed temperature fluctuations were observed at a relatively deeper depth in the summer model compared to the winter model. The streambed is thus more stable during the winter because of the greater upwelling than during the summer. This also results in the lower model errors for the winter model period at shallow depths compared to the summer model period.

### Seasonal changes of hydraulic parameters and thermal transport mechanisms

Changes in  $K_h$  between the three models were small and the model sensitivity to  $K_h$  in its range is small compared to the accuracy of the loggers, suggesting there was no discernible difference in overall  $K_h$  values between the models. However, the variation observed in the fitted  $K_h/K_z$  indicate that the  $K_z$  values do change, with the smallest values observed in the summer. Temporal variation in streambed  $K$  values that have been attributed to deposition and erosion induced changes in the streambed have been observed in other locations (Constantz 2008; Genereux et al. 2008; Hatch et al. 2010; Simpson and Meixner 2012; Su et al. 2004). Genereux et al. (2008) further proposed that lower  $K_h$  values were associated with streambed areas with higher amounts of finer-sized particles, and that decrease in hydraulic conductivity with depth may be expected if the streambed is homogeneous. Simpson and Meixner (2012) found that deposition associated with the recession limb of storm flows resulted in decreases of greater than 84 % in  $K_z$ . Prolonged deposition of fine-grained materials in the upper streambed will lead to continual lowering of the  $K_z$  (Su et al. 2004). In a study done on the 600-m stretch of LKC immediately upstream on the study site, Peterson et al. (2008) reported significant mobilization of the LKC streambed during larger storm events. Similar to reported findings by Su et al. (2004), Genereux et al. (2008), and Hatch et al. (2010), the storm events generated higher  $K$  values in the top portion of the streambed, up to 30 cm, that decreased over time as the particles realigned and compacted and as pore spaces between coarse sediments were filled in by finer particles.

Higher positive errors during summer and some negative errors during winter at 30 cm depth tells us that stream temperature influence is more prominent at 30 cm depth than the deeper substrate and suggests the top of the streambed is more dynamic and witnesses changes in conductivity more than the lower depths of the streambed. Importantly, the layer of streambed mobilized is very shallow (30 cm) compared to the streambed thickness (4 m) used in the model, implying that the actual changes in hydraulic conductivity of the dynamic streambed top could be significant, although not represented by the models built in this study. While the model allows for heterogeneity, the model assumed  $K$  to be homogeneous and does not represent the variation of hydraulic

conductivity within the shallow substrate, the heat transfer to shallow substrate due to hyporheic flows cannot be represented accurately by the model.

Simulated temperature values were comparable to the observed values (Fig. 7) and are in line with error ranges in previous studies using VS2DH (Conlon et al. 2003; Constantz et al. 2003) and other unspecified numerical models (Hoffmann et al. 2003; Prudic et al. 2003). It is important to note that all of these aforementioned models were used to simulate either smaller range of depth or lateral distance and this difference in set up is one of the most distinct differences for the model used in this study. The model does not simulate the temperature at 30 cm as well as it simulates the deeper streambed temperatures because model calibration is based on MAE of all the observation points. Thus, the dynamic shallow substrate with only four observation points is not as well represented compared to the deeper substrate with 12 observation points. Consequently the more dynamic and, likely, more hydraulically and thermally different upper layer is not as well represented by the model.

A study of thermal Peclet numbers in the streambed provides valuable information of the temporal and spatial dynamics of the hyporheic zone. When an interaction of a one dimensional (1D) conductive thermal field occurs with a 1D advective flow field in the same plane, the distribution of conduction-induced temperatures is altered and the nature of this alteration can be represented by a ratio of advective to conductive heat transfer known as thermal Peclet number (Pe):

$$Pe = \frac{\rho_w c_w q l}{k_t} \quad (2)$$

where,  $\rho_w c_w$  is the volumetric heat capacity of water,  $l$  is the length scale,  $q$  is the advective velocity associated with the length scale and  $k_t$  is thermal conductivity of the saturated medium. Pe equal to 1 implies an equal role of advective and conductive heat transfer,  $Pe > 1$  suggests dominance of advective transport over conductive transport and  $Pe < 1$  suggests that conduction is the dominant mode of heat transport.

Peclet numbers (Table 5) calculated from the simulated data suggest that advection plays a more dominant role in heat transport than conduction in the streambed. Temperature curve displacement in the upward direction suggests that the

direction of advection is in the upward direction in all three-model time periods hinting that the stretch is perennially a gaining stretch. Higher Pe for the winter model period as compared to the spring and summer time period may validate the higher upward flux values obtained for the winter model, but caution should be used since the calculations employ the simulated data rather than field data. Also, the variations in the Peclet numbers either represent differences in thermal conductivity or advective rates along the stretch of the study area. Since thermal conductivity changes little between earth materials, some spatial heterogeneity may be causing variations in the vertical hydraulic fluxes.

## Conclusion

LKC goes through temperature reversals between summer and winter with the stream at higher temperature than groundwater during the summer and at lower temperature during the winter. During spring and likely during fall, the streambed has a low temperature gradient with very little variation in temperature from stream water to groundwater. Streambed temperature curves contract closer to the surface during the winter than during the summer and the observation is pronounced in areas with higher upwelling like the upstream reaches of pools in a riffle and pool sequence. Numerical models using VS2DH are not sensitive to parameter changes for the spring suggesting the limitation of using the model to simulate low temperature gradient periods. Vertical flux in the upward direction is lower during the wetter summer season than during the drier winter likely due to greater evapotranspiration during the summer and sustained percolation into the shallow aquifer from the snow during the winter. The yearlong upwelling in the study area has a stabilizing effect on most of the streambed. The effect of thermally variant stream temperature is restricted to the shallow substrate and during upwelling conditions like the one seen in the site, no overall change in other hydraulic and thermal parameters is observed.

Overall, second to temporal changes in thermal gradients, vertical flux was the most dominant parameter in imparting differing temperature profiles to the streambed. Seasonal changes in other parameters like hydraulic conductivity, porosity, vertical anisotropy and heat capacity

**Table 5** Table of Peclet numbers for various wells during the three model periods.

Spring <sup>a</sup>					Summer <sup>b</sup>					Winter <sup>c</sup>			
W1	W2	W3	W4	W5	W1	W2	W3	W5	W6	W1	W3	W5	W6
4	2	3	4	0.5	3	2	2	2	3.5	3.5	3.5	3.75	4

<sup>a</sup> Well 6 malfunctioned

<sup>b</sup> Well 4 and 6 had malfunctioning loggers at 90 cm and 150 cm respectively

<sup>c</sup> Well 2 had malfunctioning logger at 150 cm



are very small or nonexistent and determining any changes in these parameters will require controlled laboratory experiments.

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## References

- Bartolino JR, Niswonger RG (1999) Numerical simulation of vertical ground-water flux of the Rio Grande from ground-water temperature profiles, central New Mexico. US Geol Surv Water Resour Invest Rep 99-4212
- Bastola H (2010) Identifying season changes in streambed thermal profile in a third order agricultural stream using 2D thermal modeling. MSc Thesis, Illinois State University, USA
- Beach V, Peterson EW (2013) Variation of hyporheic temperature profiles in a low gradient third-order agricultural stream: a statistical approach. *Open J Modern Hydrol* 3:55–66. doi:10.4236/ojmh.2013.32008
- Briggs MA, Lautz LK, McKenzie JM, Gordon RP, Hare DK (2012) Using high-resolution distributed temperature sensing to quantify spatial and temporal variability in vertical hyporheic flux. *Water Resour Res* 48, W02527. doi:10.1029/2011WR011227
- Briggs MA, Lautz LK, Buckley SF, Lane JW (2014) Practical limitations on the use of diurnal temperature signals to quantify groundwater upwelling. *J Hydrol* 519(Part B):1739–1751. doi:10.1016/j.jhydrol.2014.09.030
- Caissie D, Kurylyk BL, St-Hilaire A, El-Jabi N, MacQuarrie KTB (2014) Streambed temperature dynamics and corresponding heat fluxes in small streams experiencing seasonal ice cover. *J Hydrol* 519(Part B): 1441–1452. doi:10.1016/j.jhydrol.2014.09.034
- Cardenas MB (2007) Potential contribution of topography-driven regional groundwater flow to fractal stream chemistry: residence time distribution analysis of Toth flow. *Geophys Res Lett* 34: doi:10.1029/2006GL029126
- Cardenas MB, Wilson JL, Zlotnik VA (2004) Impact of heterogeneity, bed forms, and stream curvature on subchannel hyporheic exchange. *Water Resour Res* 40: doi: 10.1029/2004WR003008
- Conlon T, Lee K, Risley J (2003) Heat tracing in streams in the central Willamette Basin, Oregon. In: Stonestrom DA, Constantz J (eds) Heat as a tool for studying the movement of ground water in streams. US Geol Surv Circ 1260, pp 29–34
- Constantz J (1998) Interaction between stream temperature, streamflow, and groundwater exchanges in alpine streams. *Water Resour Res* 34: 1609–1615
- Constantz J (2008) Heat as a tracer to determine streambed water exchanges. *Water Resour Res* 44(4), W00D10. doi: 10.1029/2008WR006996
- Constantz J, Stonestrom DA (2003) Heat as a tracer of water movement near streams, chap 1. In: Stonestrom DA, Constantz J (eds) Heat as a tool for studying the movement of ground water near streams. US Geol Surv Circ 1260, pp 1–6
- Constantz J, Thomas CL (1996) The use of streambed temperature profiles to estimate the depth, duration, and rate of percolation beneath arroyos. *Water Resour Res* 32:3597–3602
- Constantz J, Stonestrom D, Stewart AE, Niswonger R, Smith TR (2001) Analysis of streambed temperatures in ephemeral channels to determine streamflow frequency and duration. *Water Resour Res* 37: 317–328
- Constantz J, Stewart AE, Niswonger R, Sarma L (2002) Analysis of temperature profiles for investigating stream losses beneath ephemeral channels. *Water Resour Res* 38(12):1316. doi: 10.1029/2001WR001221
- Constantz J, Cox MH, Su GW (2003) Comparison of heat and bromide as ground water tracers near streams. *Ground Water* 41:647–656
- Essaid HI, Zamora CM, McCarthy KA, Vogel JR, Wilson JT (2008) Using heat to characterize streambed water flux variability in four stream reaches. *J Environ Qual* 37:1010–1023. doi:10.2134/jeq2006.0448
- Fanelli RM, Lautz LK (2008) Patterns of water, heat, and solute flux through streambeds around small dams. *Ground Water* 46:671–687. doi:10.1111/j.1745-6584.2008.00461.x
- Genereux DP, Leahy S, Mitasova H, Kennedy CD, Corbett DR (2008) Spatial and temporal variability of streambed hydraulic conductivity in West Bear Creek, North Carolina, USA. *J Hydrol* 358:332–353. doi:10.1016/j.jhydrol.2008.06.017
- Gordon RP, Lautz LK, Briggs MA, McKenzie JM (2012) Automated calculation of vertical pore-water flux from field temperature time series using the VFLUX method and computer program. *J Hydrol* 420–421:142–158. doi:10.1016/j.jhydrol.2011.11.053
- Hatch CE, Fisher AT, Revenaugh JS, Constantz J, Ruehl C (2006) Quantifying surface water–groundwater interactions using time series analysis of streambed thermal records: method development. *Water Resour Res* 42, W10410. doi 10.1029/2005WR004787
- Hatch CE, Fisher AT, Ruehl CR, Stemler G (2010) Spatial and temporal variations in streambed hydraulic conductivity quantified with time-series thermal methods. *J Hydrol* 389:276–288. doi:10.1016/j.jhydrol.2010.05.046
- Healy RW, Ronan AD (1996) Documentation of computer program VS2DH for simulation of energy transport in variably saturated porous media: modification of the U.S. Geological Survey's computer program VS2DT. United States Geological Survey, Denver, CO, p 36
- Hoffmann JP, Blasch KW, Ferre TP (2003) Combined use of heat and soil-water content to determine stream/ground-water exchanges, Rillito Creek, Tucson, Arizona. In: Stonestrom DA, Constantz J (eds) Heat as a tool for studying the movement of ground water in streams. US Geol Surv Circ 1260, pp 47–56
- Kenoyer GJ, Anderson MP (1989) Groundwater's dynamic role in regulating acidity and chemistry in a precipitation-dominated lake. *J Hydrol* 109:287–306. doi:10.1016/0022-1694(89)90020-6
- Lapham WW (1989) Use of temperature profiles beneath streams to determine rates of vertical ground-water flow and vertical hydraulic conductivity. US Geological Survey, Reston, VA, 35 pp
- Masterson JP, Sorenson JR, Stone JR, Moran S, Hougham A (2007) Hydrogeology and simulated ground-water flow in the salt pond region of southern Rhode Island. Scientific Investigations Report 2006–527, US Geological Survey, Reston, VA, p 56
- NOAA (2010) Climatological Data, Illinois. 115(1):3
- Oware E (2010) The impact of storm on thermal transport in the hyporheic zone of a low-gradient third-order sand and gravel bedded stream. MSc Thesis, Illinois State University, USA
- Peterson EW, Benning C (2013) Factors influencing nitrate within a low-gradient agricultural stream. *Environ Earth Sci* 68:1233–1245. doi: 10.1007/s12665-012-1821-x
- Peterson EW, Sickbert TB (2003) Assessment of stream water bypass through a meander neck in a flood plain. *Geol Soc Am Abstr Programs* 35:376
- Peterson EW, Sickbert TB (2006) Stream water bypass through a meander neck, laterally extending the hyporheic zone. *Hydrogeol J* 14: 1443–1451. doi:10.1007/s10040-006-0050-3
- Peterson EW, Sickbert TB, Moore SL (2008) High frequency stream bed mobility of a low-gradient agricultural stream with implications on the hyporheic zone. *Hydrol Process* 22:4239–4248. doi:10.1002/hyp.7031



- Prudic D, Niswonger R, Wood J, Henkelman K (2003) Trout Creek: estimating flow duration and seepage losses along an intermittent stream tributary to the Humboldt River, Lander and Humboldt counties, Nevada. In: Stonestrom DA, Constantz J (eds) Heat as a tool for studying the movement of ground water near streams. US Geol Surv Circ 1260, pp 58–71
- Schmidt C, Conant B Jr, Bayer-Raich M, Schirmer M (2007) Evaluation and field-scale application of an analytical method to quantify groundwater discharge using mapped streambed temperatures. *J Hydrol* 347:292–307. doi:10.1016/j.jhydrol.2007.08.022
- Sickbert TB, Peterson EW (2014) The effect of surface water velocity on hyporheic interchange. *J Water Resour Prot* 6:327–336. doi:10.4236/jwarp.2014.64035
- Silliman SE, Booth DF (1993) Analysis of time-series measurements of sediment temperature for identification of gaining vs. losing portions of Juday Creek, Indiana. *J Hydrol* 146:131–148
- Silliman SE, Ramirez J, McCabe RL (1995) Quantifying downflow through creek sediments using temperature time series: one-dimensional solution incorporating measured surface temperature. *J Hydrol* 167:99–119. doi:10.1016/0022-1694(94)02613-g
- Simpson SC, Meixner T (2012) Modeling effects of floods on streambed hydraulic conductivity and groundwater–surface water interactions. *Water Resour Res* 48, W02515
- Stallman RW (1963) Computation of ground-water velocity from temperature data. US Geol Surv Water Suppl Pap 1544:36–46
- Stonestrom DA, Constantz J (2003) Heat as a tool for studying the movement of ground water near streams. US Geol Surv Circ 1260, 96 pp
- Su GW, Jasperse J, Seymour D, Constantz J (2004) Estimation of hydraulic conductivity in an alluvial system using temperatures. *Ground Water* 42:890–901
- Suzuki S (1960) Percolation measurements based on heat flow through soil with special reference to paddy fields. *J Geophys Res* 65:2883–2885
- Swanson TE, Cardenas MB (2010) Diel heat transport within the hyporheic zone of a pool–riffle–pool sequence of a losing stream and evaluation of models for fluid flux estimation using heat. *Limnol Oceanogr* 55:1741–1754. doi:10.4319/lo.2010.55.4.1741
- Swanson TE, Cardenas MB (2011) Ex-Stream: a MATLAB program for calculating fluid flux through sediment–water interfaces based on steady and transient temperature profiles. *Comput Geosci* 37:1664–1669
- Tonina D, Buffington JM (2007) Hyporheic exchange in gravel bed rivers with pool–riffle morphology: laboratory experiments and three-dimensional modeling. *Water Resour Res* 43, W01421
- Van der Hoven SJ, Fromm NJ, Peterson EW (2008) Quantifying nitrogen cycling beneath a meander of a low gradient, N-impacted, agricultural stream using tracers and numerical modelling. *Hydrol Process* 22:1206–1215. doi:10.1002/hyp.6691
- Walton WC (1965) Ground water recharge and runoff in Illinois. In: Survey ISW (ed) Department of registration and education. State of Illinois, Urbana, IL, 55 pp
- Water and Atmospheric Resources Monitoring Program (2010) Illinois Climate Network. Illinois State Water Survey, Champaign, IL
- Weight D (2008) Hydrogeology field manual, 2nd edn. McGraw Hill Publisher Professional, New York
- Winter TC (1986) Effect of ground-water recharge on configuration of the water table beneath sand dunes and on seepage in lakes in the Sandhills of Nebraska, USA. *J Hydrol* 86:221–237. doi:10.1016/0022-1694(86)90166-6