Originally published as:

Angermann, L., Krause, S., Lewandowski, J. (2012): Application of heat pulse injections for investigating shallow hyporheic flow in a lowland river. - Water Resources Research, 48, W00P02

DOI: 10.1029/2012WR012564

Application of heat pulse injections for investigating shallow hyporheic flow in a lowland river

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Received 21 June 2012; revised 6 September 2012; accepted 7 September 2012; published 11 October 2012.

[1] Hyporheic zone processes can have significant impact on groundwater and surface water resources. Detailed knowledge of exchange flow patterns is crucial for understanding the ecohydrological and biogeochemical functioning of river corridors. In particular, small-scale hyporheic exchange flow is still poorly understood, partially because of the lack of adequate in situ monitoring technology. This paper investigates the spatial heterogeneity of hyporheic exchange flow in a lowland river at multiple scales. It demonstrates the conjunctive use of active heat pulse tracing at shallow depths (15 cm) and vertical hydraulic gradients (VHG) at 120–150 cm streambed depth for improving the understanding of hyporheic exchange flow processes. Generally positive VHG indicated a regional dominance of groundwater up-welling. High and temporally variable VHG were used to identify confined conditions caused by low conductivity layers in the subsurface (low connectivity), while locations with lower and temporally less variable VHG indicated free groundwater up-welling (high connectivity) in highly conductive sediments. A heat pulse sensor (HPS) was applied for identifying shallow hyporheic flow at three locations representative for high versus low streambed connectivity. Shallow hyporheic flow patterns were found to be spatially heterogeneous. Subsurface flow could only partially be explained by streambed topography. Surface water infiltration and horizontal flow coincided with inhibited groundwater up-welling, whereas locations with high streambed connectivity were characterized by increased up-welling. The combined information of spatiotemporal VHG variability and flow vector frequency distribution by HPS has the potential to improve the understanding of impacts of streambed topography and subsurface stratification on hyporheic flow patterns.

Citation: Angermann, L., S. Krause, and J. Lewandowski (2012), Application of heat pulse injections for investigating shallow hyporheic flow in a lowland river, *Water Resour. Res.*, 48, W00P02, doi:10.1029/2012WR012564.

1. Introduction

[2] The understanding of groundwater and surface water systems in hydrological sciences has experienced a paradigm shift in recent decades, progressing from defining rivers and aquifers as discrete, separate entities toward an understanding of groundwater and surface water as integral components of an aquifer-river continuum with strong mutual influences between river, aquifer, and the interconnecting hyporheic zone (HZ) [Brunke and Gonser, 1997; Bencala, 1993, 2000; Sophocleous, 2002; Krause *et al.*, 2011a].

[3] Exchange fluxes of water, solutes, and heat between surface and groundwater environments strongly impact biogeochemical and ecohydrological processes in the HZ [Dole-Olivier *et al.*, 1997; Malard *et al.*, 2003; Datry *et al.*, 2005; Boulton *et al.*, 1998, 2008]. The HZ, as the saturated interface between surface water and groundwater, has been recognized as an important buffer and refugial zone which provides fundamental ecosystem services and functioning [Brunke and Gonser, 1997; Boulton, 2007; Krause *et al.*, 2011a]. The often enhanced chemical reactivity found at aquifer-river interfaces [Boulton *et al.*, 1998; Fisher *et al.*, 1998; Mullholland *et al.*, 2000, 2008; Krause *et al.*, 2009; Pinay *et al.*, 2009; Lewandowski and Nützmann, 2010] is controlled by (1) steep redox-gradients (2) high abundances of organic matter and microorganisms [Jones *et al.*, 1995; Fisher *et al.*, 1998; Chafiq *et al.*, 1999; Duff and Triska, 1990, 2000; Hinkle *et al.*, 2001; Findlay, 2003; Hill and Cardaci, 2004; Storey *et al.*, 2004], and (3) hyporheic flow paths and hyporheic residence times [Bencala *et al.*, 1993; Jones *et al.*, 1995; Fisher *et al.*, 1998; Duff and Triska, 2000; Zarnetske *et al.*, 2011]. The comprehensive understanding of the ecohydrological and biogeochemical functioning of hyporheic streambed environments, therefore, requires detailed knowledge of magnitude, spatial patterns, and temporal dynamics of

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groundwater-surface water exchange within the streambed at multiple scales.

[4] Exchange flow at the aquifer-river interface is controlled by hydraulic head gradients and by the hydraulic conductivity of the streambed sediments. These parameters are determined by a variety of processes including streambed geomorphology, such as pool-riffle-step formations [Kasahara and Wondzell, 2003; Storey et al., 2003; Boano et al., 2007; Kaeser et al., 2009] and stream meanders [Boano et al., 2006; Cardenas, 2009], the hydrogeological setting [Fleckenstein et al., 2006; Frei et al., 2009], natural and artificial flow obstacles [Kasahara and Hill, 2008], and microtopography induced advective pumping [Thibodeaux and Boyle, 1987; Boano et al., 2007; Cardenas and Wilson, 2007a, 2007b; Tonina and Buffington, 2007]. Drivers and controls of exchange fluxes at the aquifer-river interface may act on variable spatiotemporal scales, resulting in exchange flow patterns that are often spatially very complex and vary from small (cm) to large (km) scales [White, 1993; Sophocleous, 2002; Storey et al., 2003; Krause et al., 2011a]. The experimental investigations of such complex flow patterns, therefore, require multidimensional exploration methods, covering a range of spatial and temporal scales [Palmer, 1993; White, 1993; 1993; Krause et al., 2011a]. Under field conditions the identification of potentially interacting processes controlling aquifer-river exchange fluxes and flow patterns in particular at small scales and in superficial sediments remains a challenge [Endreny and Lautz 2012; Krause et al., 2012a, 2012b; Munz et al., 2011].

1.1. Streambed Stratification Versus Topography Controls on Hyporheic Exchange Flow Patterns

[5] Hyporheic exchange fluxes at stream reach scales have been frequently described as being driven by the variability in pressure distributions in relation to streambed

topography and channel bed form [Boano et al., 2007; Cardenas and Wilson, 2007a, 2007b; Krause et al., 2011a, 2011b]. Experimental [Lautz et al., 2010; Endreny et al., 2011] and model-based investigations [Boano et al., 2007; Cardenas and Wilson, 2007a, 2007b] found that spatial patterns of hyporheic exchange were strongly influenced by dynamic pressure fields and advective pumping related to streambed topography and surface flow turbulences [Thibodeaux and Boyle, 1987; Elliot and Brooks, 1997a, 1997b].

[6] Previous investigations of the hydrogeological heterogeneity in lowland river streambeds found that the architecture of streambed strata can also significantly impact on groundwater-surface water exchange via the aquifer-river interface [Fleckenstein et al., 2006; Schornberg et al., 2010; Krause et al., 2012b]. At the River Tern (Figure 1), a characteristic lowland river in the UK, recent heat tracer studies at multiple scales, covering a pool-riffle-pool sequence [Krause et al., 2011b] and a larger 250 m long stream reach [Krause et al., 2012b], identified spatially heterogeneous patterns in exchange fluxes between aquifer and river to be controlled by the impact of streambed topography as well as streambed hydraulic conductivity. The combination of vertical hydraulic gradient (VHG) observations and fiber-optic distributed temperature sensing (FO-DTS) at the streambed [Krause et al., 2012b] provided evidence that groundwater up-welling patterns at stream reach scale can be strongly controlled by low conductivity peat and clay lenses in the streambed (Figure 2), resulting in up-welling inhibition at locations with flow confining layers but also preferential fluxes of enhanced groundwater up-welling where low conductivity strata were interrupted. Comparable heterogeneity in streambed stratification and conductivity has been found in other lowland river streambeds [Krause et al., 2008, 2011a] where spatially rather homogeneous

Figure 1. (a) Location of the River Tern field site in the UK. (b) Instrumentation at the River Tern with groundwater boreholes (GW), river stage recorders (SW), the streambed piezometer (P) network and indication of the location of selected representative sediment cores CI and CII. (c) Piezometer experimental design. PTS = Permotriassic sandstone.

Figure 2. Conceptual model of streambed hydrofacies impact on groundwater (GW) up-welling and hyporheic flow of surface water (SW) at the River Tern (top). Streambed core profiles (in cm) of selected representative streambed cores (for sampling locations see Figure 1), PTS = Permian sandstone (bottom).

groundwater up-welling from nonfractured aquifers is modulated by the spatially heterogeneous hydraulic conductivities of drift deposits in the streambed. However, although the combination of state of the art VHG observations, and FODTS monitored streambed temperature patterns has been successfully used for identifying groundwater up-welling in response to discrete changes in streambed conductivity (Figure 2), the applied methodologies did not account for possible implications on shallow hyporheic exchange flow patterns. As aforementioned, shallow hyporheic exchange flow in the uppermost streambed sediments, describing the downwelling and re-emerging of surface water from the hyporheic zone can be driven by multiple processes at a variety of spatial and temporal scales and is often characterized by a strong horizontal flow component. It may interfere or superimpose the rather vertical transit flow of for instance up-welling groundwater resulting in complex, three-dimensional flow pattern at the aquifer-river interface. As shallow hyporheic exchange is controlled by hydrostatic pressure head distributions and hydrodynamic forcing as advective pumping, it is likely that in particular large contrasts in groundwater up-welling patterns can have a significant effect on the spatial occurrence and extend of shallow hyporheic fluxes.

1.2. Aims and Objectives

[7] It is hypothesized that a reduction of groundwater up-welling in lowland rivers as caused by flow confining low conductivity streambed strata (peat or clay lenses) several tens of cm below the streambed surface may induce increased surface water downwelling and horizontal hyporheic flow into shallow streambed sediments above the low conductivity streambed strata. With the reduced forcing of up-welling groundwater, streambed topography induced pressure head distributions and advective pumping would become more important. Therefore, low conductivity streambed strata have the potential to significantly impact on overall transient storage in shallow streambed sediments with potential implications for riverine and riparian biogeochemical and ecohydrological functioning.

[8] The present study demonstrates the combined application of a novel heat pulse tracer methodology and observations of VHG for investigating the extent and spatial variability of shallow hyporheic flow at a lowland river section. The objectives of this study are to:

[9] 1. Investigate the spatial patterns of groundwater up-welling for a stream section based on observation of VHG patterns and temporal variability to discriminate between

streambed locations with intensive groundwater up-welling (associated with highly conductive streambed strata and lower VHG) versus locations with up-welling inhibition (corresponding with confining streambed strata and increased VHG of higher temporal variability) and provide a selection of representative sites for further investigation of shallow horizontal hyporheic flow.

[10] 2. Apply a novel heat pulse sensor (HPS) for investigating shallow hyporheic flow in the top 15 cm of the streambed at locations previously identified by VHG observations to represent characteristic conditions for groundwater up-welling and inhibited up-welling.

[11] 3. Compare flow vector frequency distributions observed by HPS in shallow streambed sediments of designated locations with VHG derived information on groundwater up-welling to test the hypothesis that local surface water infiltration and an increased hyporheic flow component may be expected at locations of inhibited groundwater up-welling.

[12] 4. Revise the conceptual understanding of streambed topography and streambed stratification as drivers of multi-dimensional hyporheic flow processes in lowland rivers.

2. Materials and Methods

2.1. Study Area and Field Site

[13] The field site ($2^{\circ}53'W$, $52^{\circ}86'N$) is located at the River Tern, an 852 km² tributary of the River Severn in the UK (Figure 1a). The local geology is dominated by Permian Sherwood Sandstone, which represents one of the UK's major groundwater aquifers. The investigated stream reach covers an approximately 250 m meandering section of the River Tern with its 5–8 m wide channel and the adjacent floodplain (Figure 1b). The meandering stream section with its streambed topography dominated by pool-riffle-pool sequences and partly vegetated side bars provides a representative example for characteristic lowland rivers in similar geological setting. The River Tern was selected by the UK Natural Environment Research Council (NERC) as a study area representative for lowland sandstone rivers

under the Lowland Catchment Research Programme (LOCAR; *Wheater and Peach*, 2004); and it is monitored by the Environment Agency as part of the Shropshire Groundwater Scheme [*Streetly and Shepley*, 2002]. Parts of the monitoring infrastructure of these previous projects provide baseline data for this study (groundwater observation boreholes, borehole logs, flow gauge).

[14] Riparian and streambed sediments are characterized by intensive spatial variability in material properties and stratification and a wide range of hydraulic conductivities [*Krause et al.*, 2012b]. While the majority of sediments in the research area vary from mid-sized gravels over different sizes of sands to fine silty material with hydraulic conductivities in the range of 10^{-3} – 10^{-5} m s⁻¹, hydraulic conductivity of clay and peat layers in the streambed was significantly lower with 10^{-8} – 10^{-9} m s⁻¹ [*Krause et al.*, 2012b]. As coring in the streambed would have permanently disturbed the investigated conditions and in particular altered the impact of peat and clay layers, Figure 2 displays the abundance, thickness, and depth of peat or clay structures in the research area for two example streambed sediment profiles. The high variability of low conductivity peat and clay structures in the riparian drift deposits as result of the postglacial depositional history is shown for seven riparian corelogs in Figure 3. Structure and geometry of the streambed peat or clay layers, which are common streambed features in lowland rivers, have been successfully identified by combined FO-DTS and VHG application at the field site [*Krause et al.*, 2012b]. Given their low hydraulic conductivities, these structures have been shown to have the potential to reduce the regional groundwater up-welling [*Krause et al.*, 2012b], causing flow confinement and potentially increased streambed residence times (Figure 2).

[15] The research of this study focused on summer base flow conditions in July and August 2009. Meteorological data were recorded at the nearby Keele weather station ($2^{\circ}16'12.90''W$, $52^{\circ}59'55.86''N$). Mean annual precipitation close to the field site is 583 mm, although increased rainfall of up to 740 mm has been observed in the northern headwaters of the River Tern [*Hannah et al.*, 2009]. Precipitation

Figure 3. Borelogs for seven sediment cores of the riparian floodplain coinciding with the groundwater boreholes (GW1, GW2, GW3, GW7, GW8, GW9, GW10) of Figure 1. The stratification of the cores adjacent to the HPS focus sites (Figure 8) is shown for up to 300 cm below ground with ground levels in meters above sea level indicated for each location.

during the observation period was dominated by an extended wet period at the end of July and early August 2009 (Figure 4c) with monthly rainfall exceeding 100 mm. Several short but intensive storm events in August 2009 exceeded 10 mm h^{-1} of precipitation.

[16] Air temperature during the observation period varied by more than 20°C , with minimum temperatures of 5.2°C at night in the end of June 2009 and maximum temperatures of 26.4°C during the day in mid July 2009 (Figure 4a). Diurnal air temperature amplitudes varied substantially with maximum night–day temperature differences of 12.5°C in June and minimum night–day temperature differences of 2.3°C in the end of July. Diurnal surface water temperature amplitudes followed air temperature patterns and daily temperature oscillations were slightly dampened in shallow streambed sediments (5 cm) and not detectable within the groundwater (Figure 4b).

[17] Stream discharge data were provided by the Environment Agency from the nearby gauging station at Ternhill ($2^\circ55'12''\text{W}$, $52^\circ87'92''\text{N}$) which covers a catchment area of 92 km^2 . Mean discharge is $0.9 \text{ m}^3 \text{ s}^{-1}$ with a 95% exceedance (Q_{95}) of $0.4 \text{ m}^3 \text{ s}^{-1}$ and a 10% exceedance (Q_{10}) of $1.39 \text{ m}^3 \text{ s}^{-1}$ (data period 1972–2010, UK National River Flow Archive, http://www.ceh.ac.uk/data/nrfa/data/time_series.html?54044). Summer base flow conditions usually occur from May to October. During the observation period, mean daily river discharge for Ternhill (Q) over the

observation period averaged $0.91 \text{ m}^3 \text{ s}^{-1}$ (Figure 4c). Discharge dynamics varied significantly in time, ranging from minimum base flow discharges of $0.51 \text{ m}^3 \text{ s}^{-1}$ to maximum storm discharges of $1.77 \text{ m}^3 \text{ s}^{-1}$ (Figure 4c). The general summer base flow period was interrupted by a major discharge event in the end of July to early August with >20 days with discharges $>1.0 \text{ m}^3 \text{ s}^{-1}$ resulting from the prolonged rainfalls July/August 2009 (Figure 4c).

2.2. Data Collection

[18] Topographical surveys deploying differential GPS were carried out at the field site in June 2009. They cover the channel, streambed, and riparian zone. The resulting high-resolution digital elevation model of streambed and floodplain topography has a vertical precision of 1 cm and a horizontal precision of 25 cm. Differential GPS was also used for surveying the exact heights of installed groundwater boreholes and piezometers.

[19] During the observation period hydraulic heads and temperatures in groundwater and surface water were automatically recorded by combined pressure transducers and temperature probes (Solinst) which monitor surface water or groundwater head (i.e., water depth) at 5 to 15 min intervals (Figure 1b). Groundwater boreholes comprised a network of ten 3 m deep boreholes covering the riparian groundwater within the floodplain drift deposits (Figure 1b). Surface water levels were monitored at a downstream

Figure 4. Hydrometeorological conditions at the field site with (a) air temperatures; (b) surface water (SW3), groundwater (GW), and interstitial pore water at 5 cm depth (HZ 5 cm) temperatures; (c) precipitation and river discharge (EA Ternhill gauging station) for the period of 15 June 2009 to 30 August 2009.

located stream gauge (Figure 1b). Monitored groundwater and surface water pressure heads were corrected for barometric pressure fluctuations using an atmospheric pressure sensor located at groundwater borehole site GW7 (Figure 1b).

[20] Streambed mini piezometers were installed in the streambed sediments at depths between 120 and 150 cm below streambed surface (Figures 1b and 1c) to observe interstitial pore water pressure head distributions. They were installed in a longitudinal transect along the stream reach with several cross-sectional extensions toward the river banks (Figure 1b). Hydraulic heads in streambed piezometers were monitored manually, on an approximately fortnight basis throughout June and August 2009 using a graduated electric contact meter (dipmeter). In order to provide quality assurance for automatically logged pressure heads, manual dipmeter sampling was also carried out at the network of shallow riparian groundwater boreholes. As piezometers were installed to depths of 120–150 cm within the streambed, it is assumed that they are providing general information on aquifer–river exchange flow but do not account for obstacle driven or hydraulic pumping controlled shallow hyporheic exchange fluxes (Figure 1c).

[21] Vertical hydraulic gradients, indicating the potential strength and direction of exchange fluxes between groundwater and surface water, were determined from hydraulic head measurements in the streambed. VHGs were calculated by $\Delta h/\Delta l$, with Δh given by the elevation difference of the water tables observed inside and outside the piezometer and Δl given by the distance between the midscreen depth and the surface of the water–sediment interface. The accuracy of dipmeter based hydraulic head observations was approximately ± 3 mm head and accounts for uncertainties in the measurements introduced by turbulent flow conditions around the piezometers, which can affect the outside head estimates [Krause et al., 2009; Kaeser et al., 2009].

[22] Although VHGs measurements without further knowledge of the streambed hydraulic conductivity represent poor indicators of aquifer–river exchange fluxes, in combination with FO-DTS observed streambed temperature anomalies, variable patterns in streambed VHGs at the study site have been proven to successfully indicate spatial patterns in streambed sediment conductivities, which vary by five orders of magnitude [Krause et al., 2012b].

[23] By comparative analysis with FO-DTS observed streambed temperature anomalies, Krause et al. [2012b] found that increased VHGs (>0.4) correlate with areas of inhibited groundwater up-welling underneath flow confining streambed structures of low hydraulic conductivity (Figure 2) and low VHGs (<0.4) coincide with highly conductive streambed sediments. The combined application of VHGs observations and FO-DTS monitored temperature patterns lead to the development of a conceptual model of vertical exchange fluxes between aquifer and river [Krause et al., 2012b] that is believed to be representative for lowland rivers such as the River Tern (Figure 2). It assumes that within regionally rather homogeneous groundwater flow fields, strength and temporal variability of VHGs can present an indicator for the contrasting streambed conductivity patterns with inhibited groundwater up-welling due to flow-confining streambed strata and preferential groundwater up-welling at locations with high groundwater–surface water connectivity. However, VHGs interpretations consider groundwater–surface

water exchange processes as one dimensional. They give evidence about connectivity and interaction between river and aquifer only and do not provide any indication of potential shallow hyporheic exchange with possible increased horizontal flow in streambed sediments above flow-confining low conductivity layers.

[24] Therefore, measurements with a specially designed heat pulse sensor were conducted to extend process investigations into the shallow subsurface not covered by other methods. The HPS enables for the three-dimensional analysis of small-scale fluxes in the shallow subsurface of streambed sediments. It has been specifically developed for application in the hyporheic zone of sand-bed streams and covers a scale of few cm (diameter of investigated sediment cylinder 7 cm, Figure 5). The measuring principle is based on a short heat pulse emitted by a point source directly into the shallow sediment in 5 to 10 cm depth. Heating was applied with a power of 12 W for an interval of 60 s, resulting in an energy input of 720 J [Lewandowski et al., 2011]. The movement of the heat pulse through the sediment is traced by an array of 24 thermocouple-sensors surrounding the heat source three dimensionally. Each four sensors are attached to one separate sensor rod (Figure 5). This setup allows measuring hyporheic flow directly in the sediments

Figure 5. Instrumental setup of the heat pulse sensor. The heat source in the center is surrounded by an array of 24 temperature sensors, tracing the propagation of the heat pulse. Heater and sensors are attached to rods which are positioned and stabilized by a frame above the sediment water interface (not shown).

with minimal disturbance of the sediment specimen. The peak temperature increase $\Delta T_{\max}$ at the sensors was 0.22 to 1.20°C depending on flow velocity. Temperature changes were measured with an accuracy of approximately 0.005°C. The resulting breakthrough curves are analyzed with an analytical solution of the convective-dispersive heat transport equation in a cylindrical coordinate system (equation (1)) and provide evidence of pore water flow direction and velocity in the investigated streambed sediments [Angermann *et al.*, 2012]:

$$\Delta T_{(d,r,t)} = \frac{Q_i}{8\pi \frac{D_l}{R} t \rho_B c_B \sqrt{\pi \frac{D_l}{R} t}} \exp \left[-\frac{R(d - \frac{v_w}{R})^2}{4D_l t} - \frac{Rr^2}{4D_l t} \right]. \quad (1)$$

[25] Within equation (1) ΔT (°C) is the temperature increase relative to the ambient streambed temperature. d and r (m) are the longitudinal and radial coordinates of the cylindrical coordinate system that give the sensor's position relative to the fitted flow direction. Q_i is the thermal energy emitted by the heat source (J). For simplicity, the volumetric heat capacity of the bulk material $\rho_B c_B$ and the retardation factor R were assumed to be constant throughout the investigation site and defined to equal medium sand with a small fraction of organic matter. $\rho_B c_B$ was $3.3 \times 10^6 \text{ J m}^{-3} \text{ °C}^{-1}$, R was 1.7. The time after heat pulse injection is given with t (s). The longitudinal and transversal dispersion coefficients D_l and D_t ($\text{m}^2 \text{ s}^{-1}$) summarize the effects of dispersion and heat conduction. Both coefficients are velocity dependent and generally defined as the products of velocity and longitudinal and transversal dispersivity (m), respectively. Dispersivities are specific to the substrate and hard to quantify. They are fitted during the analysis procedure [Angermann *et al.*, 2012] and mean values were found to be 0.25 m (longitudinal, with a standard deviation of 0.43) and 0.11 m (transversal, with a standard deviation of 0.14). v_w (m s^{-1}) is the effective flow velocity (sometimes also called pore velocity; Darcy velocity divided by porosity).

[26] As discussed in detail by Lewandowski *et al.* [2011] and Angermann *et al.* [2012], some limitations of the methodology originate from the geometrical setup of the thermistors, and in particular the cylindrical alignment of the sensors around the heat source. As result of this setup, the propagation of the heat pulse is not covered ideally for vertical flow. When the vertical angle exceeds $\pm 50^\circ$, only the periphery of the heat pulse is recorded instead of its center. Flow velocity and direction have to be calculated from the break-through curves of the periphery of the heat pulse and the accuracy of the method is reduced from 5° in horizontal flow to 20° in vertical flow direction. However, this precision is assumed to be sufficient for in situ measurements so that all data were included in the evaluation of the measurement results.

[27] The analysis routine requires precise positions of the temperature sensors relative to the heat source, as errors in distances lead to a misinterpretation of the breakthrough curves. In comparison to the device presented by Lewandowski *et al.* [2011] and Angermann *et al.* [2012], who positioned a single sensor rod with a stencil, in the present study these rods were assembled to one device in order to cope with higher water depths. This setup ensures

very precise positioning, however, bending of single sensor rods due to subsurface obstacles or coarse substrate will not be prevented. Visual observation is necessary to prevent wrong measurements, and substrate properties limit the applicability of the device.

[28] A further problem exists at very low velocities, when the effect of heat conduction may exceed the convective heat transport. In such conditions it is not possible to discriminate both processes and quantify flow velocity with acceptable accuracy anymore. The threshold, i.e., the detection limit of the method, depends on the water content and the thermal properties of the sediment and can be recognized by the observation of the fitted parameters during data analysis [Angermann *et al.*, 2012]. According to this procedure, the detection limit was identified to be $1.0 \times 10^{-5} \text{ m s}^{-1}$. 21% of the analyzed heat pulse measurements at River Tern were below this threshold and are therefore of limited reliability but still considered in the analysis. Above this threshold, the method performs with an accuracy of approximately 10% in flow velocity.

[29] Heat pulse measurements were conducted at designated streambed sections of approximately 10 m length which were identified by VHG observations to be representative for characteristic streambed conditions, ranging from high aquifer-river conductivity to low conductivity and up-welling inhibition due to flow confining streambed structures. At each of the sites 18 to 21 HPS measurements were conducted. Measurements were carried out between 21.07.2009 and 19.08.2009 (Table 1) and arranged along several transects (named TR1 to TR7, enumerated upstream to downstream), most of them oriented across the streambed to cover the whole site. With this monitoring setup, the HPS observations aimed to provide a representative sample of the characteristic flow vector frequency distribution at the specific streambed locations but did not aim to monitor the entire hyporheic flow network in the research area which would have required a significantly higher number of HPS measurements.

3. Results and Discussion

3.1. Hydraulic Head Patterns in Groundwater and Surface Water

[30] Temporal variability in time series of surface water levels observed at stream gauges coincided with groundwater levels monitored at groundwater observation boreholes

Table 1. Timing of HPS Measurements in 2009^a

Date	HPS Measurements
21 July 2009	3.1–3.7
29 July 2009	2.16
31 July 2009	2.10–2.14
03 August 2009	2.3–2.9
06 August 2009	2.15, 2.17, 2.18
07 August 2009	2.1, 2.2
10 August 2009	3.8–3.13
12 August 2009	3.14–3.21
13 August 2009	1.10–1.17
19 August 2009	1.1–1.9

^aFor respective meteorological and hydrological conditions please see Figures 3 and 5.

Figure 6. Surface water (SW3) and selected groundwater levels (GW1, GW3, GW7) at the River Tern field site (Figure 1) for the investigation period from 15 June 2009 to 30 August 2009.

(Figure 6). Water levels at selected riparian groundwater boreholes (GW1, GW3, GW7) and river gauges (SW3) were generally low during the base flow conditions with average levels ranging between 58.6 and 58.9 m asl (above sea level) (Figure 6). Base flow conditions were interrupted by a three-week episode of increased groundwater and surface water levels reaching values of up to 59.38 m asl (surface water) and 59.27 m asl (groundwater) in July 2009 (Figure 6), caused by precipitation events as seen in Figure 3b.

[31] Groundwater levels in the monitored boreholes GW1 and GW3 close to the stream as well as GW7 at further distance (Figure 1b) exceeded surface water levels for most of the observation period, indicating a general flow direction from riparian groundwater toward the stream. Inverse head gradients, indicating reversed flow conditions were only observed during the July/August storm event (Figure 4c), when surface water levels rose faster and higher than associated groundwater levels (Figure 6), causing surface water infiltration into the riparian groundwater.

However, such flow inversions did not reach the central parts of the floodplain as indicated at GW7.

3.2. VHГ for Characterizing Spatial Patterns in Groundwater Up-Welling

[32] Vertical hydraulic gradients observed at streambed piezometers were positive throughout the observation period, indicating a constant up-welling of groundwater into the investigated stream reach. With values ranging between 0.08 and 0.8, VHГ in the research area were found to be spatially highly variable (Figure 7). Although absolute values of VHГ at streambed piezometer locations varied at the three monitoring dates, spatial patterns remained rather constant throughout the observation period (Figure 7). The spatially and temporally very homogeneous VHГ in the east-west oriented up-stream (P1–P3) and down-stream (P28–P30) river sections (Figure 1b) were generally low with values never exceeding 0.2. In contrast, within the north-south oriented central section of the investigated

Figure 7. Spatial patterns of observed vertical hydraulic gradients in the River Tern streambed for three selected sampling dates (19 June 2009, 30 June 2009, and 31 July 2009) during the 15 June 2009 to 30 August 2009 observation period.

stream reach (P4–P27), VHG were spatially more variable with some dominant hotspots of increased VHG reaching values of up to 0.8 around the central section (P14–P22).

[33] The observed complex spatial patterns in monitored VHG are likely to result from the wide range and high spatial variability in hydraulic conductivities of the drift deposits forming the streambed sediments of the research area. As the sediment properties of the nonfractured Permian sandstone aquifer are spatially very homogeneous at the investigation scale, observed VHG patterns are more likely to relate to variable material properties of the drift sediments than to result from local or regional groundwater flow variability. Therefore, VHG observations of spatially isolated particularly high values of up to 0.8 (as in the central stream section) were interpreted as indicator of local inhibition of groundwater up-welling by flow confining streambed peat and clay lenses, whereas up-welling inhibition was considered to be less likely at locations with lower and spatially more homogeneous VHG (as for instance in the most up-stream and down-stream sections of the study area). VHG within the central river section expressed higher temporal variability with values varying by up to 0.3 between monitoring dates, whereas VHG at piezometer locations in the most up-stream and most down-stream river sections did not vary by more than 0.05 for different observation dates. Furthermore, given the depth of piezometer screening sections of between 120 and 150 cm below the streambed surface, a relevant impact of streambed topography induced advective pumping effects on the observed head gradients is not likely.

[34] Higher temporal variability in VHG within the central river section correlated with meteorological conditions, suggesting a higher impact of hydroclimatological forcing at the respective piezometer locations. As discussed in comparison with FO-DTS observed streambed temperatures in Krause *et al.* [2012b], such dynamics may be explained by the fact that flood conditions like at the end of July 2009 (Figure 4c) cause variable responses in groundwater and surface water (Figure 6) with peaks in groundwater heads usually tailing off slower than in surface water due to retention by riparian storage. In highly conductive sediments such induced changes in head differences between groundwater and surface water are expected to cause alteration in exchange fluxes over the aquifer–river interface which quickly equilibrate VHG. In contrast, underneath flow confining streambed structures alteration of VHG is likely to be more persistent as head differences are less compensated by exchange fluxes. As confirmed by comparison with streambed temperature observations [Krause *et al.*, 2012b], it can therefore be assumed, that VHG temporal variability qualifies as a further indicator for the existence of flow confining structures (e.g., in the central river section), whereas temporally stable VHG (as in the most up-stream and most down-stream sections) are likely to indicate highly conductive streambed sediments where groundwater–surface water head differences are faster equilibrated by exchange fluxes.

[35] As vertical hydraulic gradients were determined by head differences between river and 120–150 cm depth within the streambed, they do not account for potential local shallow hyporheic exchange as for instance streambed topography induced surface water downwelling and mixing with groundwater. In order to test if shallow horizontal

hyporheic fluxes are more pronounced at locations where the forcing of up-welling groundwater is inhibited, three key locations were selected based on VHG observations (Figure 7) in order to represent different streambed architecture and related exchange fluxes (Figure 8):

[36] Site HPS1—representing an isolated location of moderate and temporally stable VHG (suggesting groundwater up-welling through relatively conductive sediments) which is surrounded by up-stream and down-stream sections of high VHG suggesting flow confinement (Figure 7). HPS sampling points are located in direct vicinity of a major pool with >2 m water depth (Figure 8).

[37] Site HPS2—representing an area of complex VHG patterns with the highest and temporally most dynamic VHG's observed in close vicinity to locations with lower and temporally more stable VHG, suggesting spatially very heterogeneous streambed properties with low conductive and highly conductive sediments. Parts of the streambed are characterized by a pool-riffle-pool sequence (Figure 8) of which a longitudinal transect has been subject of investigation by Krause *et al.* [2011b].

[38] Site HPS3—representing a larger area (>20 m stream length) of low and temporally relatively stable VHG, suggesting intensive groundwater up-welling through highly conductive streambed sediments. The center of HPS3 covers an extended central riffle section and a partly vegetated mid-channel bar (Figure 8).

3.3. HPS Monitoring

[39] The applied HPS is a new technology that has only been tested under laboratory conditions and in the sandy streambed of the River Schlaube (Germany) prior to this study [Angermann *et al.*, 2012]. In addition to general limitations of the HPS heat sensor technology, size and site conditions at the investigated stream reach of the River Tern represented some practical challenges for the application and data analysis of the HPS.

[40] During the investigations at the River Tern, measurements in some locations were inhibited or disturbed by high water levels, coarse substrate, macrophyte growth, and high surface flow velocities. The HPS was deployed into the sediment manually and its correct position and the occurrence of sediment transport due to turbulence caused by the installation were observed via a hydroscope. Inclination of the device and the disturbance of the measurement by too high surface flow velocities could be identified and resulting errors accounted for. Areas with the above mentioned distinct properties could not be sampled and had to be ignored even though they might be a characteristic part of the investigated site, like, e.g., the deep and coarse thalweg at site HPS2. The negligence of entire structural units of a river section might affect the validity of the obtained flow vector frequency distribution.

[41] Due to the size and heterogeneity of the investigated river section, heat pulse measurements were conducted in different substrates reaching from fine sand at site HPS1 and the left bank of HPS2 over medium sand at HPS3 and the sand bank of HPS2 to coarse sand and fine gravel in the thalweg and downstream of HPS2. Substrate conditions at HPS1 and HPS3 were relatively homogeneous while HPS2 was characterized by a high diversity of substrate. Varying sediment types entail different parameter sets required for

Figure 8. (left) Location of the three identified HPS focus sites selected on the basis of characteristic VHGM measurements (open circles) to represent distinctive aquifer-river connectivity conditions. (right) Alignment of HPS monitoring points (mostly along transects TR1 through TR7) and bathymetry, indicated by dashed gray lines within the three focus sites.

the calculation of hyporheic flow velocities, i.e., the volumetric heat capacity of the sediment ρ_{BCB} and the retardation factor R . As it is practically not achievable to accurately quantify all of the required parameters for the entire area, as a simplification, average conditions for all parameter sets (equaling conditions of medium sands) were assumed for all measurements conducted in this study. The parameters used in this study were $3.3 \times 10^6 \text{ J m}^{-3} \text{ }^\circ\text{C}^{-1}$ for volumetric heat capacity and a retardation factor of 1.7. This parameter set equals conditions of medium sand with a small fraction of organic matter, while the range of possible values for the sediment types present in the investigated sites is approximately 2.6×10^6 to $3.6 \times 10^6 \text{ J m}^{-3} \text{ }^\circ\text{C}^{-1}$ for volumetric heat capacity (dependent on water and organic matter content) and 1.6 to 2.6 for the retardation factor, respectively. Such simplification has the potential to cause minor errors in flow velocity estimates and reduces the comparability between measurements of different sites, in particular if site specific fluxes differ only insignificantly. However, this limitation affects mainly site HPS2 with its spatially heterogeneous conditions while the comparability

within the more homogeneous sites HPS1 and HPS3 is not compromised.

3.4. HPS for Identifying Shallow Hyporheic Flow Paths

[42] As site HPS1 was located close to a pool of more than 2 m depth and as the HPS device had to be manually deployed at the sediment, experiments were restricted to locations where surface water depths were not exceeding 120 cm. Thus, heat pulse measurements were conducted along two cross-sectional transects up-stream of the pool (TR1 and TR2, Figure 8) and one longitudinal transect TR3 parallel to the eastern river bank (in flow direction).

[43] Site HPS1 was found to be dominated by upward directed flow (Figure 9). Highest flow velocities ($>4 \text{ m s}^{-1}$) occurred in upward direction. 24% of the measurements showed angles steeper than $+70^\circ$ relative to a horizontal plane and 47% of the measurements showed a vertical angle β steeper than $+30^\circ$, whereas only 18% showed clear surface water infiltration (i.e., $\beta < -30^\circ$). Nevertheless, 35% of the measured flow directions were between -30° and $+30^\circ$

Figure 9. Relative frequencies (percentage of the whole amount of measurements at one site (%), radial axis) of absolute hyporheic flow velocities (gray scale) and vertical angles (in 20° clusters) measured with the heat pulse sensor at the three HPS sites.

(Figure 9) and thus, were categorized as horizontal flow. Hence, although the up-welling was dominant at the study site, horizontal flow was found to be still relevant. Figure 10 shows the horizontal orientation of hyporheic flow which in case of HPS1 is dominated by a clockwise deviation from the reference flow direction (i.e., the approximated main surface flow direction).

[44] The HPS measurements indicated that the downstream tailing end and the thalweg side of a sand bar located directly up-stream of the pool (Figure 11) were characterized by up-welling with angles of 38° (measurement 1.2 in transect TR1, Figure 11) and close to 90° (80° and 88°, measurements 1.3 and 1.4 of the same transect). Measurements 1.5 and 1.6 of transect TR1 were located in the thalweg and showed horizontal flow parallel to surface flow, whereas the measurements 1.7, 1.8, and 1.9 of transect TR2 indicated surface water infiltration into the sediment of the tailing end of the sand bank perpendicular to

surface flow direction (−26°, −75° and −73°). Heat pulse measurements 1.10, 1.11, and 1.12 of transect TR2, located in the thalweg and at the Eastern bank, as well as 1.13 to 1.17 along the longitudinal transect TR3 close to the Eastern bank indicated lateral groundwater exfiltration from the bank into the river (Figure 11).

[45] The observed patterns of surface water downwelling into flow obstructions beside the thalweg, predominantly along the Western sand bar as well as groundwater up-welling from the down-stream Eastern sandy river bank can be largely explained by topography induced pressure field variation.

[46] Due to coarser substrate in the area of the main thalweg close to the right bank at site HPS2, heat pulse measurements were restricted to manageable sediment conditions where the HPS could be inserted into the streambed. Hence, HPS measurements were concentrated on the remaining area of site HPS2, which was characterized by a pool-riffle-pool

Figure 10. Relative frequencies (percentage of the whole amount of measurements at one site (%), radial axis) of horizontal hyporheic flow components, i.e., the horizontal projections of flow vectors. The gray scale indicates horizontal velocity and the angles show the horizontal deviation (in 20° clusters) from the main surface flow direction, acting as reference for every single HPS site.

Figure 11. Spatial pattern of shallow hyporheic flow at the three HPS sites (HPS1 to HPS3) for 7 transsects (TR1 to TR7). Symbol size indicates the (left) vertical component of hyporheic flow, arrows show horizontal projections of the flow vector, i.e., (right) horizontal flow components (compare Figure 10). Maps in the background show schematics of the local streambed topography (HPS1 and HPS2) and macrophyte stands (HPS3). Heat pulse measurements are labeled, with the first digit referring to the HPS location and the digit right of the period referring to the respective sample location, numbered sequentially for every site up-stream to down-stream.

section with a sandy central riffle bar which separated the main (Western) thalweg from a minor, secondary thalweg on the eastern side of the stream. Downstream of the site, the streambed widened with a general flattening of the streambed topography (Figures 8 and 11).

[47] Hyporheic flow velocities at site HPS2 were found to be the lowest of all three investigated sites. The average flow velocity at site HPS2 was $1.6 \times 10^{-5} \text{ m s}^{-1}$ in comparison to 2.4×10^{-5} and $2.8 \times 10^{-5} \text{ m s}^{-1}$ at site HPS1 and HPS3, respectively. Flow in the streambed was dominated by horizontal fluxes (72%), with only 11% of downward and 17% of upward flow (Figure 9). Slow horizontal flow with vertical angles β around 0° was found in front of the upstream crest of a sandy riffle around the central part of transect TR4 (Figure 11). At these locations, the main flow direction in the streambed was surprisingly found to be directed against the surface water flow direction with a slight tendency toward the main thalweg (measurements 2.1 and 2.2; 2.5, 2.6, and 2.7 of transect TR4 and 2.10 of transect TR5). The further downstream part of the streambed including transect TR5 was characterized by an area of moderate up-welling, with angles between 14° and 36° (measurements 2.11, 2.12, 2.13 of transect TR5, and 2.16). This area included the central section of the sandy riffle between main and secondary thalweg. Negative vertical angles indicating downwelling fluxes of surface water were mainly found within the area of the secondary thalweg (measurements 2.9, 2.14, 2.17, and 2.18).

[48] Low hyporheic flow velocities at a central sand bank in contrast to downwelling in the secondary thalweg, as shown in Figure 11, were also reported by Angermann *et al.* [2012] and Rosenberry and Pitlick [2009]. Both reported the same relation between hyporheic flow in sand bank and thalweg in different rivers. These patterns are indicative for the impact of surface flow velocities on hyporheic flow, dominating over the pressure of upwelling groundwater. The moderate up-welling at the central crest and the Western area of the sandy riffle coincided with expectations of dominant flow directions caused by topography induced hyporheic fluxes. However, observations of dominant groundwater up-welling at the head of the riffle section and downwelling down-stream of the riffle tail at the field site did not coincide with the previously described typical exchange flow patterns caused by topography induced pressure field distributions [Kasahara and Wondzell, 2003; Boano *et al.*, 2007; Tonina and Buffington, 2007; Kasahara and Hill, 2008]. VHG observations at site HPS2 indicated significantly increased VHG in piezometers P19–23 (Figures 1b and 7) which has been interpreted as result of confined groundwater conditions under flow confining streambed peat and clay strata. These locations spatially coincided with the HPS measurements at transect TR5 and downstream (Figure 8). The coincidence of VHG evidence for confined up-welling and HPS observed downward fluxes provided strong evidence that the existence of low conductivity streambed strata in this area induced increased surface water infiltration and horizontal flow and thus increased shallow hyporheic transient storage. The observed hyporheic fluxes of opposite direction than the surface water flow in front of the riffle head cannot be explained by streambed topography but were related to preferential groundwater up-welling in the neighborhood of flow

confining streambed strata. This implies that the preferential groundwater up-welling acted against and exceeded the topography induced flow at the riffle head.

[49] Overall, spatial patterns of streambed flow at site HPS2 can only partially be explained by streambed topography induced pressure field distribution but, in comparison with VHG observations, provide evidence for the strong impact of streambed stratification on hyporheic exchange fluxes, including preferential surface water infiltration and horizontal fluxes above flow confining low conductivity streambed strata.

[50] Site HPS3, located in the most down-stream part of the research area, was confined by two sharp river bends. The site was characterized by a clearly defined thalweg of about 1 m water depth on the Southern side, a midchannel sandy riffle section and two sections of dense macrophyte growth (dominated by *Ranunculus*) in the central section of the sandy riffle and the sandy Southern river bank (Figures 8 and 11). Dense macrophyte roots partially restricted HPS measurements at some locations by preventing a correct deployment and spacing of the sensors.

[51] Heat pulse measurements in the HPS3 area were conducted during two periods of different flow conditions. Measurements 3.1 to 3.7 at the northern end of transect TR6 (Figure 11) and the up-stream sampling locations were carried out during the major discharge event in the end of July (Figure 4c). Due to high surface flow velocities and water levels, HPS measurements during this period were restricted to an area close to the Northern river bank. The four most upstream measurements 3.1 to 3.4 were located at a gently sloped section of the bank which was not completely submerged during base flow conditions. During flood conditions these sediments were submerged and formed a bay at the concave bank that acted as dead water zone. Streambed pore water fluxes in this area were found to be weakly up-welling, but also showed a strong horizontal flow tendency in opposite direction than surface flow in the main channel. Streambed flow in the northern end of transect TR6 (3.5, 3.6, and 3.7) were weakly up-welling and downwelling at one location with a horizontal flow component following the surface water flow direction. The majority of HPS measurements at site HPS3 were conducted during base flow conditions at the same time as measurements at sites HPS1 and HPS2. Measurements of this period covered the two major transects TR6 and TR7 across the entire river (measurements 3.8 to 3.21). 29% of these measurements indicated upward flow, 71% showed horizontal flow whereas no measurement showed angles steeper than -10° downward flow (Figure 9).

[52] The steepest positive vertical angles of 84° and 64° at transect TR6 and 86° and 62° at transect TR7 were found at the slopes from the central riffle toward the thalweg (3.11 and 3.13 at transect TR6, 3.17 and 3.21 at transect TR7), while horizontal flow with moderate positive angles characterized the bottom of the thalweg. Measurement 3.12 located at the deepest point along transect TR6 showed a vertical angle β of 29° . The deepest measurement of transect TR7 and the two adjacent measurements, 3.18, 3.19, and 3.20, showed values of 5° , 0° , and 16° , respectively.

[53] The general up-welling flow at site HPS3 was influenced by macrophyte growth at the streambed surface, representing a flow barrier in the surface water body, forcing a

flow separation into two subchannels. High velocities in the thalweg force water into the sediment and cause a diversion of the vertical flow of up-welling groundwater. Strong horizontal fluxes in the dead water zone during the high discharge period were interpreted as the results of backflow and eddy formation in this area and represent further indication of the impact of streambed morphological structures on hyporheic flow and the propagation of surface flow patterns into the subsurface.

3.5. Intersite Heterogeneity of Shallow Hyporheic Flow Paths

[54] Heat pulse measurements at all three sites detected high heterogeneity of fluxes in the streambed at the investigated spatial scale and shallow hyporheic flow opposed the main surface flow direction in several cases. HPS measurements indicated that, in contrast to flow patterns observed by 120–150 cm deep streambed piezometers, fluxes in the shallow streambed were not only characterized by groundwater up-welling but complex patterns of upward and downward fluxes as well as a substantial horizontal flow component, indicating a strong impact of streambed topography and heterogeneous patterns of streambed hydraulic conductivity on hyporheic exchange flow. At all three sites, horizontal flow and downwelling were detected mainly at the deepest points of the thalweg and in case of HPS1 and HPS3, steep vertical angles were found at the slopes of the thalweg.

[55] Although in many cases monitored hyporheic flow vectors can be at least partially explained by streambed topography, we observed significant differences in flow vector frequency distributions between the three investigated sites that were related to small-scale streambed structural heterogeneity and related spatially diverse patterns of streambed hydraulic conductivities.

[56] As sites HPS1 and HPS3 showed both relatively low and temporally stable VHGs, it was assumed that groundwater and surface water at these sites were well connected with no evidence of flow confining streambed strata. At both sites, this assumption was partly confirmed by HPS measurements which identified predominately positive vertical flow, indicating groundwater exfiltration. According to nonparametric Wilcoxon-Mann-Whitney tests for non-normally distributed sample populations, HPS2 and HPS3 sites were found to differ significantly in regard to their absolute flow velocities. While HPS1 and HPS2 differed significantly in their vertical flow velocities and absolute vertical angles, difference between HPS1 and HPS3 were not significant (all tests for 95% significance levels). With a mean absolute vertical angle $|\beta|$ of 29° , site HPS3 showed a stronger horizontal flow component in comparison to HPS1 with a $|\beta|$ of 46° . This fact has been interpreted as an indication of a more immediate interaction between surface water and groundwater at site HPS1 with predominantly vertical exchange and some lateral groundwater exfiltration through the bank. As the highly conductive sediments at site HPS1 are known to be in direct vicinity to confining streambed strata restricting exchange between groundwater and surface water up-stream and down-stream of the site, the increased hydraulic conductivity of streambed sediments at site HPS1 presents a window of preferential exchange flow between aquifer and river. The HPS results coincide with observed streambed temperature anomalies

at this location observed in Krause *et al.* [2012b]. In contrast, HPS3 represents an extended stream section with no known presence of flow confining peat layers in the streambed and evidently lower VHGs than HPS1, implying high hydraulic conductivity but also no particular hydraulic forcing and preferential groundwater upwelling due to nearby flow restriction as seen at HPS1. Thus, with a hydrologically less constrained situation, weaker driving forces of hyporheic exchange as for instance macrophytes structures were likely to gain more impact.

[57] HPS2, in contrast, represented an area of complex VHGs pattern, suggesting spatially very heterogeneous streambed properties with a mosaic of highly conductive sediments as well as flow confining peat or clay layers. The observed dominance of horizontal flow with only few cases of vertical flow and relatively slow hyporheic flow velocities supports the hypothesis of potential increase in surface water infiltration and horizontal flow above flow confining streambed strata which inhibit or at least reduce the pressure of up-welling groundwater in the uppermost streambed sediments overlying low conductivity streambed structures.

[58] A number of previous studies emphasized the impact of streambed topography induced dynamic pressure fields and advective pumping on hyporheic exchange fluxes [Boano *et al.*, 2007; Cardenas and Wilson 2007a, 2007b; Tonina and Buffington, 2007; Endreny *et al.*, 2011; Endreny and Lautz., 2012]. However, the combined VHGs and HPS measurements of this study provide evidence that at the research site, spatial patterns of high versus low conductivity streambed strata, and in particular the size and extent of confining peat and clay lenses, can have significant impact on near surface aquifer-river exchange fluxes in the streambed and may in some cases even superimpose the impact of streambed topography on hyporheic exchange fluxes. The fact that streambed conductivity significantly controls groundwater-surface water exchange has been previously investigated in experimental and model-based studies [Palmer, 1993; Wondzell *et al.*, 2009; Krause *et al.*, 2011a]. However, the results of this study indicate that in particular small-scale patterns of streambed conductivity and the combined impact of complex streambed stratification and topography on hyporheic exchange flow patterns are not sufficiently understood. The results of this study improved the conceptual understanding of drivers and controls of hyporheic exchange fluxes (Figure 2) and in particular highlight the impact of flow confining low conductivity streambed strata on shallow hyporheic flow. With its capacity to improve understanding of flow processes on local scales of cm to dm, the combination of HPS and VHGs provides a useful approach that complements other methods as the previously suggested combination of FO-DTS and VHGs [Krause *et al.*, 2012b] which targets the identification of flow pattern at a spatial scale of several hundred meters. The knowledge gain might be useful for a further process-based upscaling and to build up profound models.

[59] These enhancements in process understanding will need to be considered for the improvement of quantitative assessments of multiscale implications of different drivers and controls of groundwater-surface water exchange fluxes. In order to ensure the efficient design of streambed restoration measures aiming at alteration of streambed topography or streambed structure and stratification [Boulton *et al.*, 1998;

Boulton, 2007; Kasahara and Hill, 2008], future research is necessary to investigate the interaction of streambed structure and surface topography in controlling hyporheic exchange fluxes. As demonstrated in this study, the application of HPS technology has a great potential to complement the portfolio of experimental methods for improving mechanistic process understanding in particular of small-scale flow patterns in streambed sediments.

4. Conclusions

[60] The results of this study provide evidence that, in addition to pressure head distributions and topography induced advective pumping, streambed stratification and in particular spatial patterns of high versus low conductivity streambed strata can have substantial impact on exchange flow patterns between groundwater and surface water. Combination of VHG observations and novel HPS measurements of small-scale streambed fluxes indicate that low conductivity lenses of peat and clay at streambed depths of several tens of cm can locally reduce the impact of upwelling groundwater pressure on streambed pore waters above the confining structures and thus, support preferential surface water downwelling and horizontal streambed flow. Hence, the research results confirm the hypothesis that low conductivity streambed sediment structures not only inhibit groundwater upwelling but can also support the enhancement of surface water downwelling and horizontal pore water flow at locations where the upwelling force of the groundwater is impeded. Such increase in surface water downwelling and horizontal streambed fluxes does enhance hyporheic transient storage with potential implications for the biogeochemical cycling and ecohydrological functioning of hyporheic zones. The results furthermore indicate the possibility of enhanced groundwater upwelling where highly conductive streambed sediments in close vicinity to confining streambed structures provide a preferential flow path for fast upwelling of the semiconfined groundwater. The results of this study provide new mechanistic understanding of aquifer-river exchange flow processes in dependence of streambed stratification and topography. They provide evidence that in lowland rivers with complex patterns of variable hydraulic conductivities in streambed as at the investigated field site, the impact of flow confining streambed strata can superimpose the effects of streambed topography induced hyporheic exchange. This improved process understanding has substantial implications for river restoration and management. As residence time distribution and mixing ratios of groundwater and surface water in the streambed provide a unique hyporheic environment where structural alteration can have significant implications for its ecohydrological and biogeochemical functioning, consideration of the improved process knowledge will be crucial in particular for river restoration programs altering the streambed sediment structure.

[61] The field application of the HPS demonstrated the potential as well as limitation of this novel technology for tracing streambed flow patterns at small scales. Additionally, heat pulse measurements have been found to provide valuable insight into drivers and controls of shallow streambed pore water fluxes at the aquifer-river interface when compared with VHG observations at greater depth. While HPS observations

did not in every case support a fully conclusive explanation of measured single flow vectors, the identified flow vector frequency distributions indicate that shallow hyporheic flow is strongly controlled by streambed conductivity and patterns in groundwater upwelling, supporting initial hypothesis formulated on the basis of VHG observations.

[62] The HPS identified patterns of shallow hyporheic flow contribute highly valuable information to the study of groundwater-surface water interaction and demonstrate a high potential to complement larger scale pressure head or passive heat tracing methods. The comparison of HPS and VHG results provided further evidence that in particular the combination of experimental methods at different scales has the potential to significantly improve mechanistic process understanding, whereas the application of each singular method would have not allowed conclusive differentiation between fluxes, their drivers, and controls. Future research will be required to implement the improved process knowledge into existing modeling approaches for quantification and up-scaling to management relevant scales.

[63] **Acknowledgments.** The authors wish to acknowledge substantial field work support by E. Naden (formerly University of Keele) for some of the HPS measurements as well as from M. Munz (University of Potsdam) and C. Tecklenburg (Helmholtz Centre for Geoscience Research Potsdam) for the installation of the streambed piezometer network and sediment coring. We thank the UK Natural Environment Research Council (NERC-NE/I016120/1) as well as the Royal Geographical Society for funding parts of the research presented in this paper. The authors would like to thank the editor Graham Sanders as well as the anonymous reviewers for their helpful comments and constructive criticism.

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