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Revealing the spatial variability of water fluxes at the groundwater-surface water interface

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[1] There is widespread recognition that the groundwater-surface water interface can have significant influence on the pattern and form of the transfer of nutrient-rich groundwater to rivers. Characterizing and quantifying this influence is critical for successful management of water resources in many catchments, particularly those threatened by rising nitrate levels in groundwater. Building on previous experimental investigations in one such catchment: the River Leith, UK, we report on a multimeasurement, multiscale program aimed at developing a conceptualization of groundwater-surface water flow pathways along a 200 m reach. Key to this conceptualization is the quantification of vertical and horizontal water fluxes, which is achieved through a series of Darcian flow estimates coupled with in-stream piezometer tracer dilution tests. These data, enhanced by multilevel measurements of chloride concentration in riverbed pore water and water-borne geophysical surveying, reveal a contrast in the contribution of flow components along the reach. In the upper section of the reach, a localized connectivity to regional groundwater, that appears to suppress the hyporheic zone, is identified. Further downstream, horizontal (lateral and longitudinal) flows appear to contribute more to the total subsurface flow at the groundwater-surface water interface. Although variation in hydraulic conductivity of the riverbed is observed, localized variation that can account for the spatial variability in flow pathways is not evident. The study provides a hydrological conceptualization for the site, which is essential for future studies which address biogeochemical processes, in relation to nitrogen retention/release. Such a conceptualization would not have been possible without a multiexperimental program.

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1. Introduction

[2] The interface between groundwater and surface water, in particular the hyporheic zone, plays an important role in underpinning key ecosystem services such as maintenance of habitat and water quality [e.g., *Boulton et al.*, 2010; *Hester and Gooseff*, 2010]. Past research [e.g., *Smith*, 2005; *Wroblicky et al.*, 1998] has suggested that the hyporheic zone may have an important functional role with respect to groundwater-surface water interactions: this role may be particularly critical for reactive chemistries [e.g., *Seitzinger et al.*, 2006; *Storey et al.*, 2004]. Reactive chemistries in groundwater-fed rivers are highly sensitive to the physical hydrology and biogeochemistry of the hyporheic zone, although early research focused only on the uppermost few centimeters of the riverbed [e.g., *Jonsson et al.*, 2003]. More recently, understanding the potential for the hyporheic zone in river-catchment connections to attenuate human-modified reactive chemistries of groundwater-surface water fluxes, has been the focus of new research [e.g., *Boano et al.*, 2010; *Faulkner et al.*, 2012].

[3] A significant gap in our knowledge exists in understanding the relative importance of water flux from deeper (e.g., regional) groundwater in "conditioning" the hydrochemistry of the hyporheic zone and in governing its capacity to attenuate redox-sensitive nutrients such as nitrogen. In particular, recent work by *Stelzer and Bartsch* [2012], *Kennedy et al.* [2009], and *Flewelling et al.* [2012] identify the importance of the hyporheic zone in respect to nitrate enrichment. We recognize that water flux is not the only driver critical for transformations of reactive chemistries such as nitrogen. We identify other drivers that include prevailing redox conditions, supply of reactive substrates

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especially organic carbon, and the presence of bacteria that mediate nitrogen transformations [see, for example, *Storey et al.*, 2004; *Pretty et al.*, 2006]. We suggest that much can be gained by bringing together research on physical hydrology with that on nutrient biogeochemistry through focus on the quantification of the spatial variability in physical hydrology and its implications for rates and flux of reactive chemistries, especially nutrient transformations, at scales important for management decisions [*Heathwaite*, 2010].

[4] Our ultimate aim is to examine the physical and chemical drivers of exchange water fluxes and redox reactivity within and across the hyporheic zone to determine the spatial efficiency of nitrogen transformations, and to evaluate whether attenuation of contaminants in the hyporheic zone is a realistic expectation for the maintenance of the ecological quality of groundwater-fed rivers. A critical step in this process is to examine the spatial variability in water fluxes at the groundwater-surface water interface; this is the focus of our study. The hyporheic zone is normally described as a porous media within the riverbed where groundwater and surface water mixing take place. However, a number of authors [e.g., *Kennedy et al.*, 2009; *Malcolm et al.*, 2003; *Sophocleous*, 2002] found significant local heterogeneity at this interface: *Conant* [2004] described a key implication of such heterogeneity as "preferential discharge locations". Here we study the heterogeneity of water fluxes at this interface. We concentrate our efforts on a study reach of the River Leith in Cumbria, UK: a river of known groundwater connectivity (at the catchment scale) and within an area of known rising regional nitrate concentrations in groundwater [*Butcher et al.*, 2006, 2008].

[5] Previous experimental observations at the site [*Krause et al.*, 2009; *Käser et al.*, 2009] have highlighted heterogeneity of hydrological properties and chemical species, but have so far been focused on small subreaches (several meters in length). Here we report research from a major study that examined a 200 m reach of a groundwater-fed river, using a set of integrated measurements undertaken at multiple scales, in order to determine the principal flow pathways that may influence the magnitude and pattern of groundwater-derived nitrate within the reach. Unlike many previous studies, through appropriate experimental design, we provide measurements of vertical and horizontal flows within the river corridor. Although the use of multiple measurements in groundwater-surface water studies is not new [see, for example, *Malcolm et al.*, 2004], a novel element of the work is the fusion of a specific set of measurement types in our aim to develop a conceptualization of flow paths at the groundwater-surface water interface.

[6] We recognize that water fluxes and flow paths at the groundwater-surface water interface are regulated by a number of parameters including, the sediment structure, hydraulic conductivity and hydraulic head gradients and multiple-scale geomorphic features [*Wagner and Bretschko*, 2002; *Mutz and Rohde*, 2003; *Song et al.*, 2007; *Genereux et al.*, 2008; *Käser et al.*, 2009; *Chen*, 2011]. As a consequence of variable hydrogeologic and geomorphic controls on groundwater-surface water interaction, water flux is both vertical and horizontal [*Poole et al.*, 2008; *Käser et al.*, 2009]. Thus, calculation of water residence times at the interface and its impact on biogeochemical transformation based on either vertical or horizontal flows alone may result in under or overestimation of rates and spatial distribution of both hydrologic and biogeochemical processes that is of interest to hydrologists, limnologist, biogeochemists and ecologists. We argue here that it is critical to investigate both vertical and horizontal flux patterns at the reach scale to be able to accurately quantify water retention time and hence pollutant and nutrient attenuation rates and spatial variability. This conceptualization is needed to build site-specific models of nitrate transport and transformations at the groundwater-surface interface. We develop this conceptualization through the unique combination of a series of experimental tools, including riverbed point dilution tests and water-borne geophysical surveys. A novel element of the work is the measurement of vertical and horizontal flow components at the reach scale. We place particular emphasis on understanding the functioning of the hyporheic zone under base-flow conditions when the relative contribution of groundwater flux to river discharge is high and the consequences for water quality and freshwater ecology are particularly important [*Cotton et al.*, 2006; *Heathwaite et al.*, 2005].

2. Field Site and Methods

2.1. Study Site

[7] The study site is a 200 m long meander reach of the River Leith, a tributary of River Eden in Cumbria, UK, near the village of Cliburn (Figures 1 and 2). The River Eden controls direction of regional groundwater flow in the area (see Figure 1). The river is predominantly groundwater-fed [*Käser et al.*, 2009] through shallow Permo Triassic sandstones in

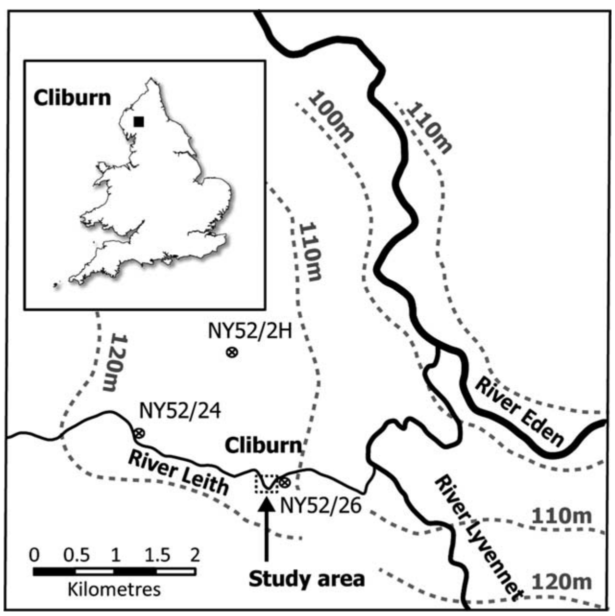

Figure 1. Location of Cliburn reach of the River Leith (inset map shows outline of England and Wales). Symbols indicate location of three Environment Agency boreholes for regional groundwater flow monitoring. Gray-dashed lines show September 2010 regional groundwater contours (Environment Agency, personal communication October 2012) in meters above mean sea level.

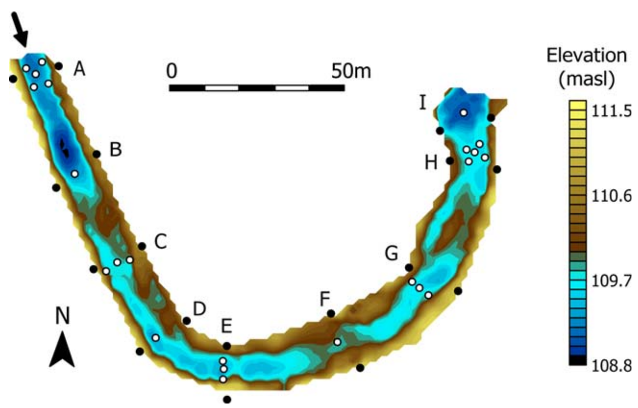

Figure 2. Layout of the field site and bed topography surveyed July 2010. River flow is from left to right (indicated by arrow). The circle symbols show location of piezometers at nine plots along the reach. White filled symbols indicate locations with three piezometers (drilled to depths: 100, 50, and 20 cm). Black filled symbols indicate locations of riparian piezometers. Elevations are shown in masl (meters above mean sea level).

the Eden valley; base flow during summer months is typically around 0.1 m^3/s [Käser *et al*., 2009], although summer discharges can rise to several m^3/s. The riverbed is mainly loose gravely alluvium overlying an unconsolidated *Penrith Sandstone* bedrock [*Allen et al*., 1997; *Krause et al*., 2009]. The riverbed topography along the reach reveals a sequence of riffles (sites C, F, G, H) and pools (sites A, B, I) with a noticeably deeper section in the upper 40 m of the reach (Figure 2).

[8] The reach was originally selected because of the known connectivity with the sandstone aquifer [*Seymour et al*., 2008] and close proximity to a flow gauging station (100 m downstream of the study reach). During 2006 and 2007 a series of piezometers were installed in the riverbed in an initial effort to study nitrate transport processes within the hyporheic zone at two contrasting sites along the study reach [*Krause et al*., 2009]. A further set of piezometers were installed in 2008 in the downstream 20 m of the reach and investigated by *Käser et al*.[2009] to support a study of small scale variation of water fluxes along a riffle-step-pool sequence. The initial piezometer design proved to be vulnerable to damage during high flow events, furthermore, installation to depths beyond 50 cm was challenging owing to the physical structure of the subsurface and a revision to both installation and piezometer design was necessary for continued investigation at the site and at depths below 50 cm. In 2009, we embarked on a new installation in order to capture more insight into hydrological and biogeochemical processes that influence the transport and transformation of nitrate within the system.

2.2. Piezometer Design and Installation

[9] New piezometers were constructed from 32 mm (outer) diameter uPVC (unplasticized polyvinyl chloride) pipe with a 2.5 mm wall thickness in order to withstand high-flow events. Each piezometer has a screen length of 6 cm perforated with 5 mm holes and wrapped with a 105 μm polyester mesh. Adopting a 35 mm sump in the new design allowed for minimum impact of fine sediment entering the piezometer. Furthermore, unlike the earlier design, a base plug was installed in the piezometer to prevent vertical flow within the piezometer, and thus permit more reliable estimates of horizontal flow from tracer dilution tests [cf. *Käser et al*., 2009]. Installation to a required depth was achieved by fabricating a HPDE drive point, which could be secured (glued) to the base of the piezometer (and act as the end plug). The drive point outer diameter is 51 mm, allowing a steel tube of similar outer diameter to be temporarily fitted over the piezometer tube and used to drive the piezometer to the required depth. This design is an extension of the approach used by *Rivett et al*. [2008].

[10] Since manual hammering of the steel tube is ineffective in a gravel bed like the River Leith, a petrol driven Cobra TT drill (Atlas Copco, Stockholm, Sweden), with appropriate auger attachment, was used to drill the riverbed to a desired depth and create a "pilot hole". Adopting a 60 mm diameter corer with the drill also permitted recovery of sediment samples for particle size analysis. Once the pilot hole had been drilled and cored, the piezometer tube, encased in a steel tube, was quickly driven to the prescribed depth using a 15 kg fence post hammer. When this depth was reached, the steel tube was then removed using a jacking device, allowing the sediments to collapse around the piezometer. The new installation procedure proved effective and allowed relatively rapid installation to depths of up to 100 cm below the water filled channel.

[11] A similar procedure was followed for installation of piezometers in river banks (outside the wetted channel). For each bankside piezometer a target depth of 50 cm below the bed of the center of the adjacent channel was selected in order to compare left and right bank hydraulic responses in a consistent manner. For these "bank" piezometers installation to a depth of up to 3 m was necessary, but achievable with the aforementioned approach. Since the wall of these cored holes did not always collapse around the piezometer tubes clean coarse sand was used to reseal the piezometer inside the bore to minimize any short circuit of water flows along the piezometers. All piezometers were left to stabilize for 20–30 days before monitoring of various hydrologic and biogeochemical parameters was commenced.

[12] Each piezometer tube was also fitted with multiple PTFE (polytetrafluoroethylene) tubes of 1.6 mm internal/3.2 mm outer diameter along the outside of the tube wall, prior to installation, to allow sampling of pore water. These narrow sampling tubes were installed at depths of 10, 20, 30, 50, 70, 100 cm below the riverbed in the channel piezometers and at the completion depth in the bank piezometers. The ends of the sampling tubes were wrapped in the 105 μm polyester mesh to avoid blocking with particles during water sampling. The top ends of the tubes were closed with tight fitting brass panel pins except during pore water sampling.

[13] A total of 87 piezometers were installed in June 2009 and June 2010 in the riverbed and adjacent left and right banks. Along the 200 m reach of the river, midchannel piezometers with screened interval at 100, 50 and 20 cm each were installed in nine sites (Figure 2). Head levels in the piezometers were surveyed at 2–4 week intervals during 2009 and 2010 using a narrow diameter manual

electronic dip meter. During these surveys, the local stream water level adjacent to the piezometer was also measured, thus allowing an assessment of total head (relative to datum) and a vertical hydraulic gradient.

[14] Measurement of in-stream piezometric head levels permit the quantification of the vertical component of flow within the riverbed. When used with head measurements from riparian piezometers, an assessment of lateral flow gradients can be achieved. Here we define lateral flow as the component of the flux vector orthogonal to the stream channel. As described later, tracer tests conducted in the piezometers also permits quantification of the horizontal component of flow, which is likely to be a combination of lateral and longitudinal (i.e., along the direction of the channel) flow.

2.3. Sediment Core Analysis

[15] Sediment samples from each core depth were air dried and sieved using the following sieve classes: 63, 45, 31.5, 16, 8, 4, 2, and 1 mm. Particle size distributions for sediment <1 mm were assessed with a laser diffraction technique (Malvern 2000 Mastersizer, Malvern Instruments Ltd., UK). Prior to analysis on the Mastersizer, samples (<1 mm) were digested with 30% hydrogen peroxide (H_2O_2), to remove organic content, and then shaken overnight in Calgon (a dispersing agent). The laser-derived distributions were then combined with sieve data.

2.4. Saturated Hydraulic Conductivity

[16] Saturated hydraulic conductivity was measured through both falling and rising slug tests, employing a water level logger (HOBO U20-001-01, Onset Corporation, USA) inside the piezometer. Saturated hydraulic conductivity was determined from duplicate rising and falling head tests using the Hvorslev method [*Hvorslev*, 1951]. Slug tests results were attempted on 83 piezometers out of the 87 and carried out during May to June 2009 and April to August 2010. For 12 of these piezometers, the response time was too long and precluded analysis for hydraulic conductivity.

2.5. Water-Borne Geophysical Surveys

[17] Near surface geophysical techniques are now widely used in hydrological investigations [e.g., *Robinson et al.*, 2008] and there has been a recent focus of attention to the role of geophysical techniques for mapping riverbed sediment architecture [e.g., *Crook et al.*, 2008]. Although geophysical techniques do not provide a direct measure of hydrological properties or states, they may be used to infer variation in them either through known or assumed petrophysical relationships. Previous high-resolution electrical resistivity imaging surveys were conducted along the Cliburn meander of the Leith by *Clifford and Binley* [2010]. Their work focused on several short (9 m) transects along the riverbed and highlighted contrasts in electrical conductivity, most notably elevated electrical conductivity at site C in Figure 2 (cf. "Site 7" in Figure 8 in *Clifford and Binley*, 2010). Assuming an "Archie-type" positive relationship [*Archie*, 1942] between porosity and electrical conductivity, and assuming insignificant variation in pore water electrical conductivity, we may attribute variation in electrical conductivity to changes in pore volume (porosity) and pore structure, which *may* in turn be related to permeability of the riverbed sediments. Alternatively, variation in bulk electrical conductivity may indicate spatial variability in fluid electrical conductivity, and hence help identify contrasts in groundwater sources.

[18] Extending the work of *Clifford and Binley* [2010], we conducted water-borne geophysical surveys along the 200 m reach using electromagnetic induction (EM) "terrain conductivity" meters. Such an approach was adopted to allow relatively easy coverage of the longitudinal and lateral variability in electrical conductivity. As demonstrated by, for example, *Mansoor et al.* [2006], EM methods can be used effectively to map variation in sediment properties beneath surface water bodies. The sensor provides an integrated measure of electrical conductivity over a depth that is a function of the separation between transmitter and receiver coils. The contribution of electrical conductivity over a vertical profile varies with depth, but may be assumed known [*McNeill*, 1980]. Knowledge of this signal sensitivity function together with known fluid electrical conductivity of the water column and depth of the water column, allows the estimation of the electrical conductivity of the riverbed sediments. From one measurement we may therefore estimate an effective electrical conductivity of the riverbed sediments.

[19] The cumulative sensitivity function for coils oriented in a horizontal plane is given by [*McNeill*, 1980]:

$$CS(z) = \frac{1}{(4z^2+1)^{0.5}}, \tag{1}$$

where z is the depth scaled by the coil separation.

[20] If we consider a two layer model: water column and riverbed, then the observed (apparent) electrical conductivity can be represented by:

$$\sigma_a = \sigma_w(1 - CS(z_w)) + \sigma_{rb}\ CS(z_w), \tag{2}$$

where $\sigma_w$and $\sigma_{rb}$are the electrical conductivities of the river water and riverbed, respectively, and z_w is the depth of the water column divided by the intercoil separation. Since σ_w and $z_w$are easily measured, we can use equation (2) to determine the riverbed conductivity, σ_{rb}.

[21] For the Leith surveys we employed two surveys. Using a Geonics EM38 (Geonics, Mississauga ON, Canada) with coil separation of 1 m, 70% of the signal can be attributed a depth of 1.59 m beneath the instrument (equation (1)). Using the 3.66 m coil separated Geonics EM31 (Geonics, Mississauga ON, Canada) the equivalent measure of depth of investigation increases to 5.82 m. For each survey, the instrument was installed on an inflatable raft, with the coils raised just a few cm above the top of the water column. Location within the reach was measured with a Trimble 58100 GPS receiver (Trimble, Sunnyvale, CA, USA); water depth variation during the survey was measured with an Onset HOBO U20-001-01 water level logger dragged along the riverbed; fluid conductivity measured with a Jenway 430 portable meter (Bibby Scientific, Staffs., UK).

2.6. Chloride Profiles

[22] During base-flow periods in summer 2009 and 2010 pore water was extracted using the multilevel sampling

tubes at a number of piezometers. Pore waters were analyzed for a number of chemical species, although here we concentrate on chloride as a conservative element that helps identify contrasting water pathways and/or origins when distinct differences in concentration are observed between stream water and deeper (100 cm) pore water. During each survey 40 ml of pore water was extracted at each depth, after purging the sample line, and filtered with a 0.45 μm surfactant-free cellulose acetate membrane. Samples were stored under cool conditions prior to analysis using ion chromatography (Dionex, Thermo Fisher UK Ltd.).

2.7. Vertical Flux Estimates

[23] Six measurement surveys of the piezometer heads and stage levels were undertaken between June 23 and 7 September 2010 under similar base-flow conditions, and are the focus of the assessment here. The head difference between piezometer head and river stage was used to calculate Darcian vertical water flux, q_v, following:

$$q_v = K_s \frac{\Delta h}{\Delta z}, \quad (3)$$

[24] where K_s is the saturated hydraulic conductivity (here derived from slug test analysis), Δh is the head difference and Δz is the vertical distance between riverbed and midscreen of the piezometer. Estimates of q_v for 20 cm deep piezometers were computed based on K_s values for the same piezometers. For estimates of q_v for 50 cm deep piezometers effective K_s values were computed using the harmonic mean of K_s in the 20 cm and 50 cm piezometers at the same location. Similarly, for 100 cm-based vertical flux estimates, the harmonic mean of K_s in the 20, 50, and 100 cm piezometers were used. We recognize that the slug test derived values of hydraulic conductivity will be biased towards horizontal permeability, and thus may be an overestimate in an anisotropic system.

2.8. Horizontal Flux Estimates

[25] Single well dilution (drift) tests allow estimation of local horizontal fluxes by monitoring, within a borehole, the dilution of a tracer injected along a screened interval [e.g., *Drost et al*., 1968]. Whilst such an approach has been used for studying flows within the riparian zone [e.g., *Lamontagne et al*., 2002], *Käser et al*. [2009] recently adopted the same approach in their investigation of horizontal fluxes within the riverbed. Extending the work of *Käser et al*.[2009] to the entire study reach, and adopting a design which prevents vertical movement within the piezometer, single well dilution tests were carried out in 53 piezometers during base-flow conditions in 2009 (August and September) and 2010 (July and September).

[26] A salt tracer was used and, assuming a linear relationship between solute concentration and electrical conductivity, the horizontal flux, q_h, is given by:

$$q_h = \frac{-\pi d}{4t\alpha} \ \ln\left(\frac{\mathrm{EC}\,(t) - \mathrm{EC}_b}{\mathrm{EC}_0 - \mathrm{EC}_b}\right), \quad (4)$$

[27] where t is time; d is the diameter of the piezometer screen; α accounts for the lateral influence of the monitored well within the flow field, which depend on the hydraulic characteristics of the sediments; EC(t) is the electrical conductivity at time, t; EC_0 is the electrical conductivity at time zero; EC_b is background electrical conductivity [*Käser et al*., 2009].

[28] Miniature four electrode sensors were fabricated for electrical conductivity sensing within the piezometer, which could be positioned within the screened section of the piezometer as detailed in *Käser et al*. [2009]. Connection of the electrode cables to a Campbell CR10X (Campbell Scientific Inc., Utah, USA) data logger allowed monitoring of a time series of electrical conductivity within the screened section of 10 piezometers simultaneously at the sampling interval.

[29] We used a KCl tracer (200 g/L) with an electrical conductivity of 200 mS/cm. For each tracer test a small volume (2.5–3 mL) of tracer was injected through a 1.6 mm diameter PTFE tube within the piezometer screened section resulting in a concentration of approximately 15 g/L; the tube was then flushed with 20 mL air in order to mix the tracer. The variation in electrical conductivity was recorded at 30 s intervals for at least 1 h prior to injection in order to ensure a stable background conductivity measurement; following injection, recording continued for a further 48 h. The decay curves of the conductivity measurements (e.g., Figure 3a) were then log transformed and a linear regression used to calculate flux q_h following equation (4). Figure 3b shows a typical response observed.

2.9. Statistical Analysis

[30] The saturated hydraulic conductivity, vertical flux and lateral flux data were log transformed prior to

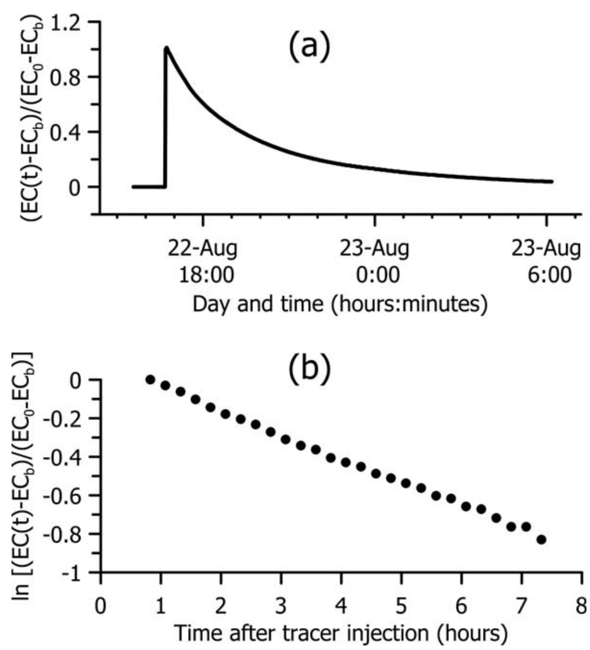

Figure 3. Example of solute dilution test in a riverbed piezometer. (a) Measured dilution of tracer (note that the actual points, sampled every 30 s, are shown not a fitted line). (b) log transformed relative dilution of the tracer. Note that in Figure 3b symbols show measurements every 0.5 h for illustration, although the entire data set was used for regression of the flux.

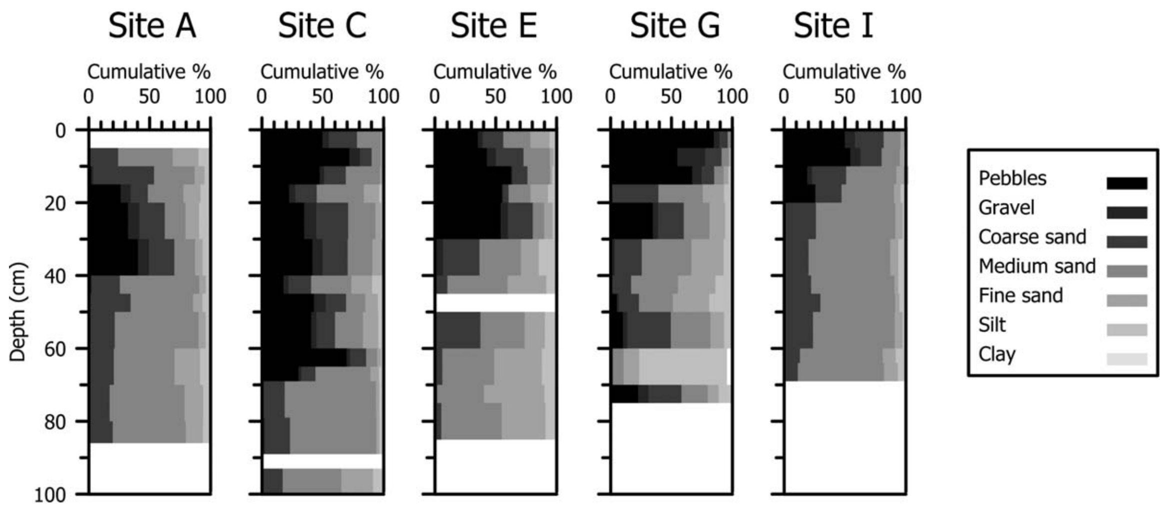

Figure 4. Example textural logs of riverbed cores along the reach. Blank sections indicated nonrecovery of sample. Elevations are shown in masl (meters above mean sea level).

parametric analysis of variance (ANOVA). The normality assumptions of these parametric analyses were tested using the Shapiro-Wilk test and one-way ANOVA was undertaken to test for differences in saturated hydraulic conductivity, vertical, and horizontal fluxes with depths at the 5% significance level.

3. Experimental Observations

3.1. Sediment Core Analysis

[31] Sediment cores extracted from the riverbed revealed a typical sediment profile of a loose pebble/gravel/sand fining with depth to a weathered sandstone base. The thickness of the superficial layer varied from 20 to 100 cm, consistent with *Krause et al.* [2009]. Figure 4 shows example variation in textural properties along the reach; all cores in this case were extracted from the center of the channel. The contrast between superficial sediments and the sandstone can be clearly seen in a number of textural profiles. For example, in Figure 4 at Site A a distinct boundary exists at a depth of 40 cm. Some variation in texture was noted across the channel (data not shown), some of which, at shallow depths, may be attributed to the impact of erosion of the steep right bank in the upper section of the reach (sites A to E).

3.2. Hydraulic Conductivity Variation

[32] Table 1 contains a summary of the hydraulic conductivity data. At each measurement depth we recorded roughly two orders of magnitude variation in K_s throughout the study reach. At all depths hydraulic conductivity showed a log normal distribution based on Shapiro-Wilk test of normality ($p = 0.15$), consistent with previous studies [e.g., *Bjerg et al.*, 1992; *Ryan and Boufadel*, 2007]. Hydraulic conductivity varied across the river reach, however, the mean differences in log transformed conductivity at 20, 50, and 100 cm depths were not statistically different from each other ($p > 0.05$) based on one-way ANOVA. When log transformed hydraulic conductivity at 20 and 50 cm depths in the riverbed were compared using pooled-variance *t*-test, a significant difference was found, showing that the hydraulic conductivity recorded at 20 cm depth was significantly higher than at 50 cm depths ($p = 0.04$). From observations of sediment texture, we anticipate that this extends to a total depth of 30–40 cm. *Song et al.* [2007]; *Chen* [2011]; and *Pretty et al.* [2006], amongst others, have also reported a decreasing trend in conductivity with depth in riverbed sediments.

[33] Table 2 reports the geometric mean hydraulic conductivities for in-channel and bank piezometers. For in-channel values, Table 2 shows a differentiation of values recorded in pool sections of the reach and other areas. The lower mean saturated hydraulic conductivities recorded at all depths in the pools relative to other features corroborates the findings of *Käser et al.* [2009] for a subreach of the same river. The geometric mean hydraulic conductivity at 20 cm depth was about two times higher in midchannel piezometers (data not shown) compared to channel margin piezometers. Such variation in hydraulic conductivity across channel may be a result of localized deposition of fine sediment. *Genereux et al.* [2008] speculated that more fine particles are deposited towards the channel margins

Table 1. Summary of Saturated Hydraulic Conductivity Values Determined From the Four Piezometer Types

		Hydraulic Conductivity (m/d)				
Piezometer Type	Sample Size	Minimum	First Quartile	Median	Third Quartile	Maximum
20 cm	20	0.151	0.726	1.466	2.483	11.190
50 cm	18	0.055	0.130	0.255	1.866	21.895
100 cm	18	0.064	0.372	1.381	2.130	4.864
Banks (50 cm)	15	0.144	0.260	0.488	1.463	9.602

Table 2. Geometric Mean of Saturated Hydraulic Conductivity, Horizontal and Vertical Fluxes in the Riverbed Sediments and River Banks[a]

	Depth Below Riverbed (cm)	Nonpool	Pool	Right Bank	Left Bank
Saturated Hydraulic Conductivity (m/d)	20	1.94 (11)	0.84 (9)		
	50	0.45 (10)	0.60 (8)	0.68 (6)	0.77 (9)
	100	1.07 (8)	0.78 (10)		
Vertical Flux (m/d)	20	0.15 (9)	0.18 (9)		
	50	0.03 (10)	0.07 (8)		
	100	0.02 (8)	0.05 (10)		
Horizontal Flux (m/d)	20	0.024 (10)	0.035 (8)		
	50	0.014 (6)	0.012 (7)	0.012 (4)	0.007 (6)
	100	0.013 (7)	0.011 (5)		

[a]The values in parenthesis show the number of samples. Right and left banks represent measurements at 50 cm depth relative to midchannel riverbed depth.

compared to midchannel that could be responsible for relatively higher conductivities in midchannel locations. We observed slumping of the river bank sediments on the eroding side of the river [see also, *Käser et al*., 2009] and fine sediment deposition on the opposite bank; this may account for lower hydraulic conductivities in channel margin piezometers compared to midchannel piezometers in the gravelly alluvium of the River Leith.

[34] Tests were also conducted to examine upstream-downstream variation. Comparisons of 20, 50, and 100 cm depth data from grouped sites A-D and E-I using *t*-tests revealed no significant difference between upstream and downstream grouped log hydraulic conductivity at the 5% level. Failure to resolve any difference is perhaps not surprising given the observed variability in the measured K_s values.

3.3. Hydraulic Head Variation

[35] As stated earlier, the general trend of regional groundwater flow is towards the River Eden, north east of the site (see Figure 1). Regional groundwater head data from three deep (50–60 m) Environment Agency monitoring boreholes (see Figure 1) for the period June–August 2010 were made available. Daily head levels remained stable ($\pm$0.1 m) throughout this period; triangulation of these measurements reveals an average bearing of 090.1° and an average gradient of 3.30%. In contrast, analysis of river bank (riparian) piezometers indicates an average local shallow groundwater flow direction of 045°, with an average gradient of 1.90%.

[36] Dipped in-channel piezometer levels throughout most base-flow surveys revealed upward vertical gradients throughout most of the reach. In the upstream part of the reach (sites A to E), vertical gradients were considerable (>10%, and as high as 35%). Only at (riffle) site H were downward gradients observed, usually in shallow piezometers. Figure 5a shows a composite map of hydraulic head measurements across the site on 9 September 2010 (other survey dates revealed similar patterns), based on an interpolation of four heads (0, 20, 50, 100 cm depth) at each of the nine sites. The profile reveals a clear dominance of vertical flow potential between sites B and F, downstream of which there is a transition from vertical to horizontal gradients. Figure 5b shows horizontal variation in heads 100 cm below the riverbed. Again, similar profiles were noted on other sampling dates under base-flow conditions. Clearly there is no measurement of lateral variation (within the channeled region) at all of the sites and thus this interpolated map must be interpreted with care; similarly the map only accounts for head variation within the perimeter of the channel. Nevertheless, the data do suggest that the potential for lateral flow exists in upstream sites and, given the domination of vertical gradients in the upstream section of the reach, hyporheic exchange may be limited in depth in the upper part of the reach, perhaps supporting the model-derived assertions of *Munz et al*. [2011]. Note that the apparent horizontal flow direction using only in-channel piezometers is consistent with the shallow groundwater flow direction inferred from the bank piezometers (see arrow in Figure 5b).

[37] Our piezometer array was designed to permit an assessment of lateral flow directions within the site. Figure 6a shows the variation in heads along the reach based on left

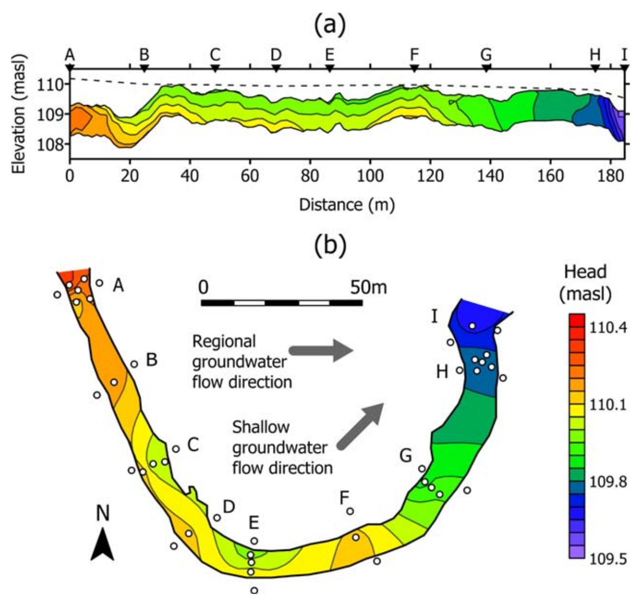

Figure 5. Interpolated map of heads in riverbed computed from piezometer dips on 9 September 2010 at all nine sites. (a) Vertical profile using 20, 50, and 100 cm deep piezometers dips and stage levels. The dashed line shows measured stage profile. (b) Horizontal profile using 100 cm deep piezometer dips. The symbols in Figure 5b show measurement locations. Note the exaggerated vertical scale in Figure 5a. Hydraulic heads are shown in masl (meters above mean sea level).

and right bank piezometers and the center channel piezometer, installed to a depth equivalent to that of the bank piezometers. Along the reach there is clear potential for lateral flow from both left and right banks. Using these measurements, and accounting for the location of each measurement, the variation in lateral gradients along the reach are shown in Figure 6b. A marked switch between right bank to left bank dominated gradient is clearly seen between sites D and E. For comparison, Figure 6b also shows the average head gradient based on the bank piezometers alone.

3.4. Water-Borne Geophysical Surveys

[38] Shallow (EM38) surveys were carried out in "zig-zag" traverses allowing the assessment of both lateral and longitudinal variation in electrical conductivity. Because of the physical size of the EM31 device the deep EM survey was only carried out in a longitudinal profile. Both EM surveys revealed significant variation in electrical conductivity across the reach. Throughout the survey the water depth varied spatially in the range 0–1.46 m, with an average water depth of 0.38 m. The measured stream water electrical conductivity during the survey was 596 μS/cm (59.6 mS/m). This measurement, combined with the water depth at a given location, allowed the computation of the first term on the right-hand side of equation (2), and thus the calculation of the riverbed conductivity.

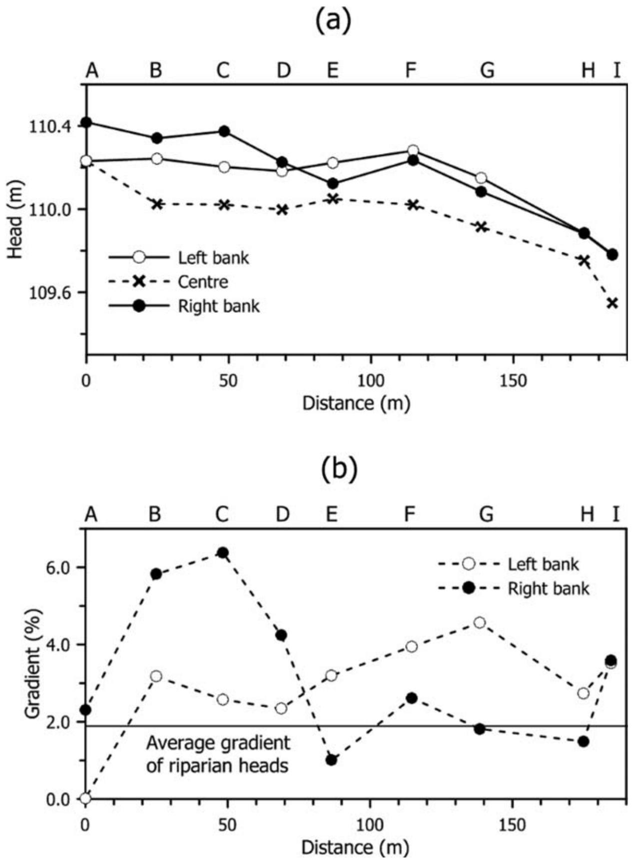

Figure 6. (a) Hydraulic heads in riparian piezometers and central channel piezometers (50 cm below riverbed) on 9 September 2010. (b) Lateral hydraulic gradients (i.e., normal to local channel direction) based on hydraulic head data displayed in Figure 6a. The solid horizontal line in Figure 6b is the average hydraulic gradient computed from the riparian piezometer hydraulic heads (direction is SW-NE—see Figure 5).

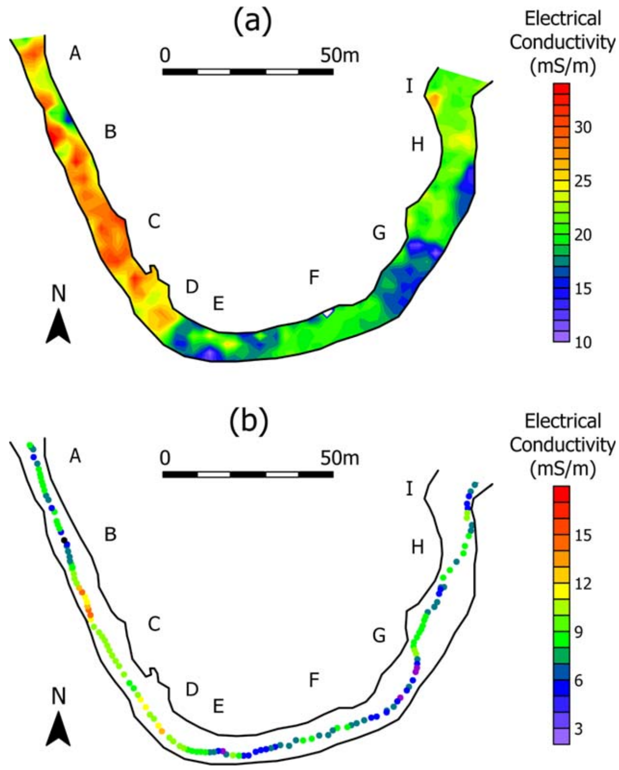

Figure 7. Maps of electrical conductivity of bed sediments inferred from EM surveys. (a) EM38 (shallow) survey. (b) EM31 (deep) survey.

[39] Figure 7a shows the EM38 derived map of riverbed electrical conductivity across the reach. A sharp contrast in electrical conductivity is apparent between sites D and E: at site D and upstream the electrical conductivities are significantly higher than at downstream locations. This is not an artifact of water column height: shallow depths, for example, show both high- and low-electrical conductivities (compare Figures 5a and 7). The high-electrical conductivity is due to different textural properties of the sediments (e.g., porosity) and/or contrasts in pore water conductivity: both of which may provide information about exchanges between groundwater and surface water.

[40] The elevated electrical conductivity at site C is consistent with low resistivities observed by *Clifford and Binley* [2010]. Also, as noted in Figure 4, coarser sediments were observed at depth in the core taken from site C. An initial interpretation, therefore, is that less compact (and possibly more permeable) sediments are prevalent in the upstream section of the reach. However, we note that attempt to differentiate upstream and downstream hydraulic conductivity was unsuccessful, as stated earlier.

[41] The EM31 survey (with an effective depth of investigation of over 5 m) in Figure 7b provides further evidence of contrasts in electrical conductivity: many of the features mirror the longitudinal changes observed in the shallow EM38 profile in Figure 7a.

3.5. Chloride Profiles

[42] Figure 8 shows an example sequence of chloride concentration profiles along the reach. These data are typical of profiles observed during base-flow conditions. The profiles reveal higher surface water chloride concentrations compared to 100 cm deep groundwater values with variable levels of mixing along the reach. In most of the profiles mixing is only evident in the 10 cm depth samples (if at all), with a progressive increase in mixing depth in the downstream direction, from site G. A comparison with the hydraulic head contours (Figure 5) reveal consistent behavior, i.e., a greater contribution of horizontal (longitudinal and/or lateral) flow in the downstream subreach (sites G to I).

[43] We also note in the chloride profiles that significantly higher chloride concentrations are seen at depth in sites B and C, These elevated concentrations are apparent in other datasets collected under base-flow conditions and suggests a potential localized groundwater source emerging at a depth of at least 100 cm, i.e., the delineation of emergent groundwater flow paths within this section of the reach. This is supported by the horizontal head gradients in Figure 5b. Note also that the electrical conductivity maps in Figure 7 show higher electrical conductivity in the upper reach, in particular near site C. It would appear, therefore, that between sites B and D the river is well connected to a groundwater body with a contrasting hydrochemical signature, reflected in the chloride profiles. Furthermore, given that the (~5 m deep) EM31 profile in Figure 7b reveals significantly elevated electrical conductivity in this region it appears reasonable to assume that the groundwater source is not shallow, and is likely to be associated with the regional sandstone aquifer.

3.6. Vertical and Horizontal Fluxes

[44] Figure 9a shows a summary of Darcian vertical flux estimates derived from hydraulic gradients and measured hydraulic conductivity. The values are log normally distributed according to the Shapiro-Wilk test of normality ($p = 0.09$). Vertical flux at 20 cm depth is significantly ($p < 0.05$) larger than at 50 and 100 cm depths based on one-way ANOVA . We attribute such a gain in flux to larger lateral contributions. This is supported by the horizontal flux data (summarized in Figure 9b). These data also follow a log normal distribution (Shapiro-Wilk, $p = 0.83$), with the flux at 20 cm depth appearing significantly ($p < 0.05$) higher than at 50 and 100 cm depths below the riverbed and at an equivalent 50 cm depth in the banks.

[45] We recognize that the measured K_s values may not be truly representative of the effective hydraulic conductivity for vertical flux estimates. We sampled K_s only at depths of 20, 50, and 100 cm: armoring or hydraulic sealing of the riverbed at shallower depths may lead to much lower hydraulic conductivities. Such a mechanism could be spatially and temporally variable (e.g., by mobility of bed armor during high-flow events), thus contributing to the observed variation in vertical gradients across the site.

[46] As expected, hydraulic conductivity correlates significantly ($r^2 = 0.44$, $p < 0.05$) with vertical fluxes; such a link is in agreement with the results of *Käser et al.* [2009], see also *Essaid et al.* [2008] and *Hyun et al.* [2011]. A weak ($r^2 = 0.27$), but statistically significant ($p < 0.05$), linear relationship was also observed between saturated hydraulic conductivity and horizontal flux in the riverbed sediments. *Wroblicky et al.* [1998] and *Wagner and Bretschko* [2002] also reported similar finding of significant control of conductivity over the magnitude and spatial distribution of horizontal fluxes in the hyporheic zone. Unlike vertical fluxes derived from hydraulic conductivity data, horizontal fluxes were derived here using dilution of a conservative tracer in the piezometers.

[47] Computation of lateral fluxes at each of the nine plots using head gradients is challenging because of the

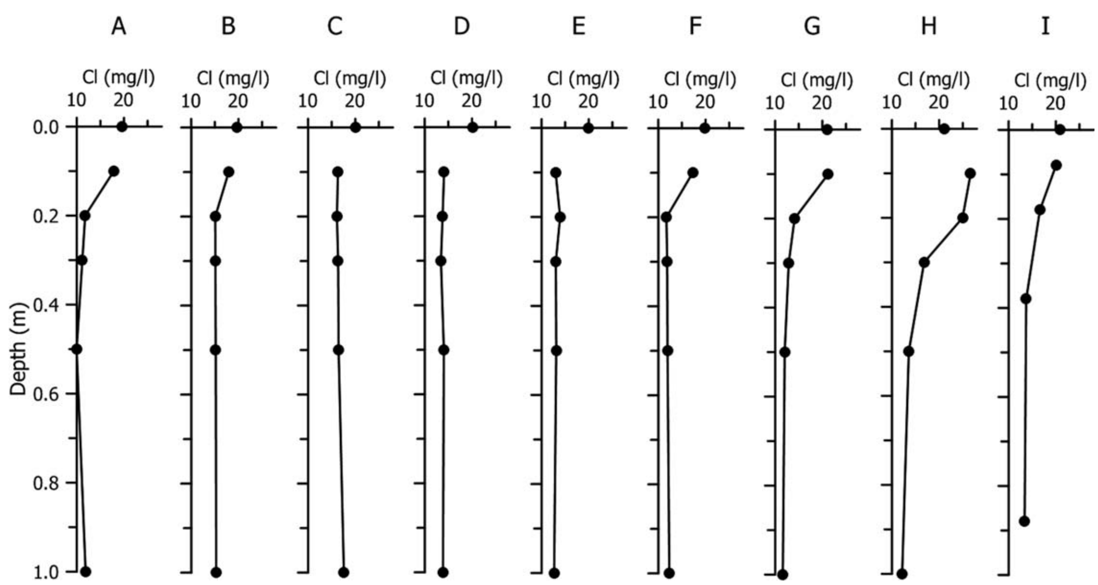

Figure 8. Vertical profiles of chloride concentration measured in the center of the channel in July 2009 under base-flow conditions. Surface water chloride concentration measurements (at each location) are shown as zero depth.

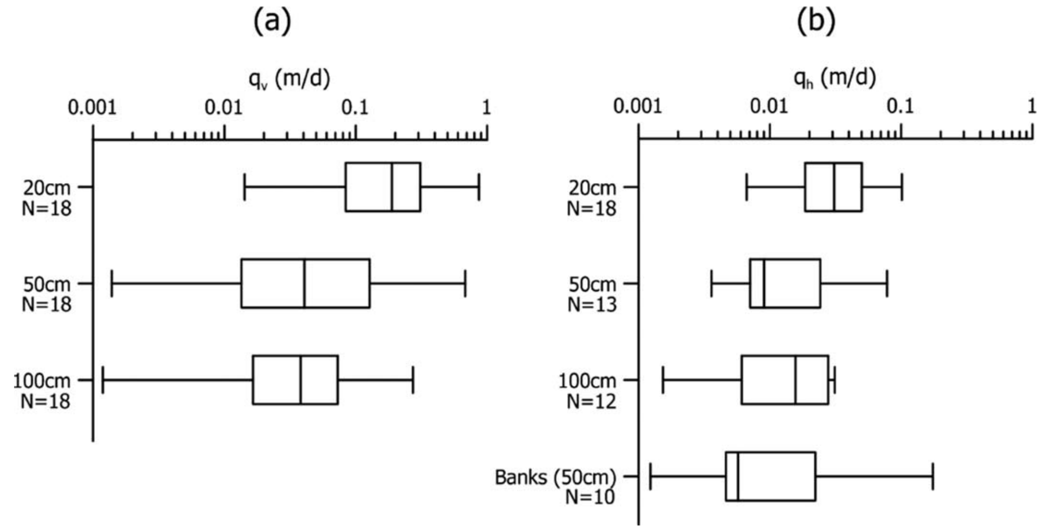

Figure 9. Flux estimates. (a) Vertical flux determined from Darcy estimates. (b) Horizontal flux computed from dilution tests. Box and whiskers show extremes and interquartile range. The vertical line indicates the median value.

lack of measurements of hydraulic conductivity from some of the bank piezometers. If we adopt the mean hydraulic conductivity for left and right bank piezometers (see Table 2), the lateral gradients reported in Figure 6 equate to an average lateral flux of 0.020 m/d from the left bank and 0.025 m/d from the right bank. The magnitude of these figures is consistent with the horizontal flux estimates through tracer dilution reported in Figure 9b, suggesting that, on average, the horizontal flux measured is dominated by a lateral component. For comparison, the equivalent calculation based on the inferred local groundwater flow direction gives a horizontal flux of 0.014 m/d in the SW-NE direction.

[48] Although data coverage is somewhat sparse along the reach (despite the large number of piezometers), we have attempted to examine the spatial patterns of vertical and horizontal fluxes. Figure 10 shows interpolated maps of vertical fluxes and horizontal fluxes at 100 cm depths. Clearly such maps are subject to significant uncertainty, however, general trends can be seen. Most notable is that elevated fluxes (both horizontal and vertical) appear to exist in the region of site C. In this region there is little (or no) mixing of chloride data apparent (Figure 8), suggesting that vertical upwelling suppresses hyporheic exchange in this part of the reach. This is consistent with the observed fluxes: despite horizontal fluxes being higher in this area, they are still typically one order of magnitude smaller than the vertical fluxes.

[49] In order to further investigate the contrasting pattern of fluxes in upstream and downstream sections of the reach, Figure 11 shows a comparison of vertical and horizontal flux data for upstream and downstream sites. In Figure 11a, data from upstream sites C and E reveal the dominance of the vertical flow component throughout the three sampled depths. Measurements from these sites show vertical flux to be typically one order of magnitude greater than the horizontal component. In contrast, at sites G, H, and I, despite significant scatter in the data, the dominance of vertical flow is not evident. At these sites the vertical flow appears smaller, with some evidence of greater horizontal flow at particular sites. These contrasting trends further highlight the change in flow pattern throughout the reach. Note also

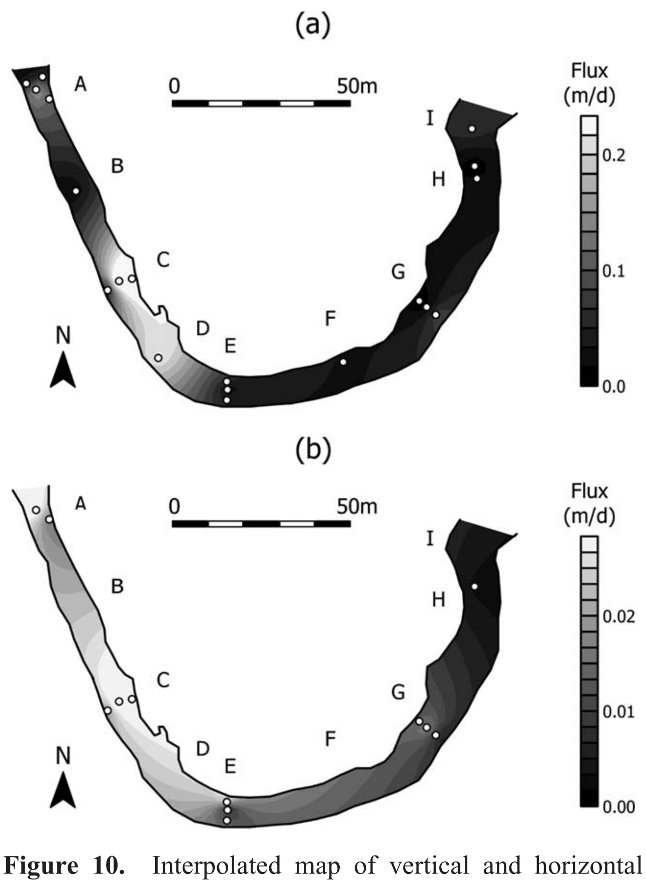

Figure 10. Interpolated map of vertical and horizontal fluxes at 100 cm depth. (a) Vertical flux based on mean gradient between June and September 2010 under base-flow conditions. (b) Horizontal fluxes during summer 2010 field campaign. The symbols show location of piezometer used for interpolation in each plot.

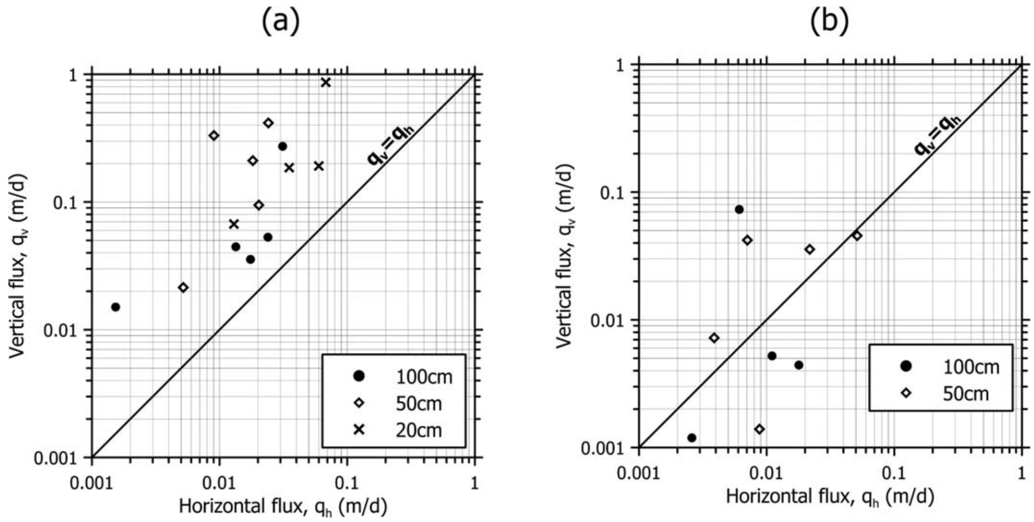

Figure 11. Comparison of measured vertical and horizontal fluxes. (a) Data from sampled piezometers in upstream sites C and E. (b) Data from sampled piezometers in downstream sites G, H, and I.

that if we consider the computed hydraulic conductivities to be an under-estimate of effective values because of hydraulic sealing of shallow sediments not examined via slug tests, then horizontal fluxes would appear even more significant in terms of the contribution to the overall water balance within the reach.

4. Discussion

[50] Our ultimate aim, from working at the River Leith site, is to develop a conceptual model of nitrate transport between surface and groundwater. The core objective is to identify critical mechanisms, for example flow path lengths and residence times, that influence the fate of reactive nitrogen species at this interface, thus improving our ability to identify, assess and manage potentially vulnerable water resources. One of the foundations of such a model is an equivalent conceptual model of water fluxes within the river corridor: the focus of this study. Our field-based observations suggest that vertical, longitudinal, and lateral fluxes occur in different proportions along the 200 m study reach. Synthesizing measurements from a range of experimental techniques, Figure 12 shows a sketch conceptualization of subsurface flow paths within the reach. From our observations we infer an increase in the horizontal flow component in the downstream section of the reach, and consequently a greater likelihood of hyporheic exchange.

[51] The measured component of vertical fluxes across the site shows an increasing trend in the vertical, implying that a greater contribution of horizontal flux must exist in the shallow sediments, and this is, indeed, supported by other observations. *Jensen and Engesgaard* [2011] also observed increasing contributions of horizontal flow at shallow depth in their study of a reach of the Skjern River, Denmark. *Vogt et al*. [2010] have also inferred contrasting contributions of horizontal flow from observations of vertical flow in riverbeds.

[52] Close examination of chloride data from multilevel sampling ports along the reach reveal variability in an effective depth of mixing of surface and groundwater. In three (adjacent) locations (sites C, D, E) little (or no) mixing depth is measurable suggesting that groundwater is either isolated from, or extremely well connected to, the river. Measured hydraulic head profiles along the reach provide independent supporting evidence of high potential groundwater flux with little apparent horizontal flux component. It would appear that in sections of the reach, predominantly vertical groundwater flux plays a significant role in suppressing shallow hyporheic exchange flow.

[53] In their recent study, *Munz et al*. [2011] calibrated a subreach groundwater flow model of the site, using data from an older piezometer array, and used this model to examine the potential for shallow surface water infiltration. They concluded that such infiltration is unlikely to occur. Our findings here add support to the modeling study of *Munz et al*. [2011]: their area of study is close to site C, from where we show clear evidence, from multiple sources and using multiple approaches, of enhanced vertical flux component, suppressing hyporheic exchange. However, our larger scale reach study with enhanced piezometric

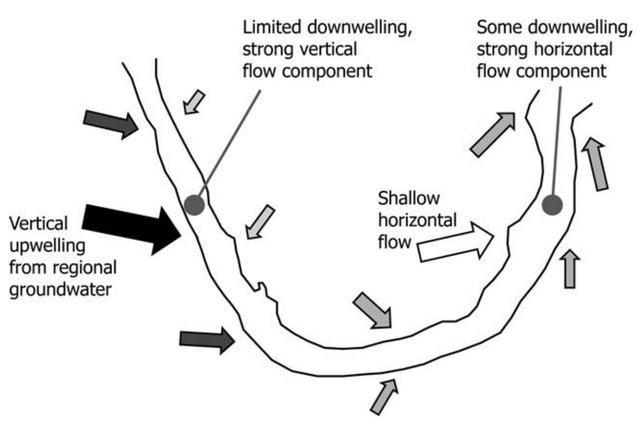

Figure 12. Schematic of conceptualization of subsurface water fluxes within the reach. Arrow sizes indicate relative magnitude of flux. Arrows are shaded in gray scale—dark shading indicates dominance of vertical flow component; white indicates dominance of horizontal flow component.

observations is able to reveal that such limited vertical exchange is probably localized to approximately 20% of the 200 m study reach, and is likely to be the result of enhanced connectivity to deeper groundwater. In this section of the reach we anticipate that the opportunities for biogeochemical processing, may be limited given the potentially short residence times, at least under the conditions monitored here (i.e., base flow). A vertical flux of 0.3 m/d, assuming a porosity of 30%, equates to a travel time of 1 day through 1 m sediment. However, we recognize that residence times alone will not control biogeochemical activity.

[54] As illustrated in Figure 12, the potential for vertical flux diminishes with downstream distance from A to I. Hydraulic head contours show a clear progression from predominantly vertical to horizontally dominated flux. Chloride profiles offer clear supporting evidence of such a transition: the depth of vertical mixing appears to increase with distance in the downstream direction. In the upper section of the reach, where longitudinal hydraulic gradients are weak, horizontal fluxes were recorded. In this upper section of the meander, it would be intuitive to assume that the potential for exchange into the adjacent left bank exists (i.e., discharge from the stream), however, recharge to the stream from deeper groundwater appears to dominate. In fact, hydraulic head data from riparian piezometers reveal gradients towards the stream (at a depth equivalent to 50 cm below the center of the channel) in both left and right banks along the reach, although there appears to be a gradient of local groundwater across the stream at this location.

[55] From site E and downstream, based on lateral gradients, connectivity to the local, shallow groundwater appears greater on the left bank, with a more longitudinal direction of flow on the right bank of the stream. Toward the end of the study reach, downwelling of surface water, and consequently a significant hyporheic zone thickness, is apparent, as illustrated clearly by the profiles of chloride concentration.

[56] Geophysical surveys also provide powerful evidence of contrasting upstream and downstream sections. One may interpret the variation in electrical conductivity as a result of contrasts in the fabric of the riverbed sediments, however, spatial differences in hydraulic conductivity could not be resolved within the variability of all slug test measurements, although differences in textural properties were noted in sediment core logs (Figure 4). The measured chloride concentration profiles reveal the cause of the contrast in bulk electrical conductivity: elevated chloride concentrations in the upper reach sediment pore waters show clearly a contrast in pore fluid chloride chemistry. For example, the 1 m depth chloride concentration at site C is over 50% greater than that at site G. We cannot dismiss possible contrasts in the textural properties of the riverbed sediments as having an influence on the bulk electrical conductivity, and although chloride is not the only ionic species within the pore water, we believe that this marked contrast in fluid chemistry is the dominant cause of bulk electrical changes along the reach. Assuming chloride is conservative, this contrast indicates a potentially localized source of groundwater feeding the river. One may postulate a different source of such pore water chemistry: downwelling driven by the bed form between sites B and C, perhaps enabling the ingress of surface water with high-chloride concentration into deeper sediments at site C. However, upward vertical fluxes are clearly strong in this area and we may assume that any such downwelling, if it occurs, would be correlated with major variations in river discharge linked to rainfall events and short lived. Furthermore, the deep ($\sim$5 m deep) EM31 surveys of electrical conductivity (Figure 7b) show, conclusively, evidence of elevated high electrical conductivity integrated over a depth much greater than any possible downwelling of surface water. It would appear, therefore, that elevated chloride concentrations are a result of localized direct connectivity to the sandstone aquifer. This is consistent with the direction of local groundwater flow, observed from riparian piezometers.

[57] Such "preferential discharge locations", as labeled by *Conant* [2004], may focus groundwater, suppress hyporheic exchange, and lead to short residence time pathways. The short residence time may be an important physical constraint on the capacity for nitrate removal by heterotrophic denitrification that, consequently, could be an important biogeochemical function of the hyporheic zone [*Flewelling et al*., 2012; *Zarnetske et al*., 2011]. However, rates and magnitude of nitrate removal will also depend on the oxygen content of groundwater, on the supply of reactants (labile carbon and nitrate) and the presence/absence of denitrifying bacteria [*Stelzer et al*., 2011; *Lansdown et al*., 2012]. Although potentially small in area, these localized discharge zones may be critical in influencing river stream water quality and ecological status, particularly under baseflow conditions. Identification of these features may, therefore, be important at the catchment scale as we seek to assess the potential impact of rising levels of nitrates in some groundwater bodies [e.g., *Butcher et al*., 2006, 2008].

[58] Our observations highlight the potential complexity of water flow pathways at the groundwater-surface water interface. In our study reach, it appears that transitions from predominantly vertical flow to horizontal flow can occur over relatively short spatial scales. This has implications for the use and validity of relatively simple 1-D models for describing solute transport in such systems, particularly when one considers reactive transport processes along a flow line. Furthermore, the assessment of solute exchange properties through less distributed measurements, e.g., instream tracers, is challenging given the apparent complexity in subsurface flow pathways. We believe that only through a series of distributed measurement (as used here) can the water flux distribution be realized. Such an assessment appears essential for reliable interpretation of reactive solute transport behavior. In a companion study of biogeochemical controls of nitrate transport on the same study reach, C. M. Heppell et al. (Interpreting spatial patterns in redox and coupled water-nitrogen fluxes in the streambed of a gaining river reach, submitted to *Biogeochemistry*, 2013) have identified that the distribution of vertical and lateral water fluxes within the reach influences the pattern of redox-sensitive chemical species in the riverbed: oxic conditions appear associated with the localized upwelling zone and reducing conditions with enhanced horizontal flows, observed within the downstream zone.

[59] A key contribution of this work is the assessment of the horizontal flux component through in-stream dilution tests. Assessing horizontal hydraulic gradients through

piezometer networks is challenging because of the resolution over short flow path lengths; such approaches are also extremely invasive because of the extensive network of piezometers required. Many experimental approaches offer only a lumped assessment of groundwater recharge or infiltration (see for example the comparative study by *Briggs et al*. [2012]) and fail to provide directional information about fluxes. In contrast, more sophisticated multisensor techniques [e.g., *Angermann et al*., 2012] are likely to be limited to relatively fine grained streambed environments, due to the nature of the necessary installation procedures. In this work, we utilized single conventional piezometers for the assessment of vertical and horizontal flow.

[60] The observations reported here were made under base-flow conditions. Under these conditions hydraulic gradients appear relatively stable, however, under high-river flow events one may expect differences in response [*Malcolm et al*., 2006; *Puckett et al*., 2008]. *Käser et al*. [2009], working on earlier data from the downstream sites H and I, noted stable vertical hydraulic gradients under changing river head: groundwater and surface water heads appeared to be, in the main, synchronized. However, in the upstream sites (A to E) of the longer reach considered here, steep banks may cause short term suppression or reversal of vertical gradients under a rapidly rising stage supplied, in part, by less damped rainfall-runoff in the headwater. Such an effect could lead to an enhanced hyporheic zone mixing depth in this section of the reach, perhaps through short duration high-stage events. It is, therefore, necessary to explore the temporal variability of fluxes in order to establish any dynamics in the hyporheic zone, particularly in relation to its biogeochemical function.

5. Conclusions

[61] Building on previous experimental investigations at the River Leith site [*Krause et al*., 2009; *Käser et al*., 2009] we developed a new experimental observation network consisting of piezometer arrays and multilevel pore fluid samplers, designed to offer more insight into the hydrological exchanges at the groundwater-surface water interface, and ultimately help in the development of a conceptual model of nitrate transport. Earlier modeling studies at the site [*Munz et al*., 2011] have highlighted the potential for limited surface water infiltration at the site; we have examined this using experimental data over a 200 m reach.

[62] From our network of piezometers we have shown, under base-flow conditions, that distinct contrasts in flow directions exist at the site. A zone of predominantly upwelling flow is apparent within the upstream section of the reach. By comparing measured vertical and horizontal components of flow in the riverbed sediments we see a transition in the relative contribution of horizontal flow along the reach, which results in greater potential for hyporheic exchange flow in the downstream section of the reach. Such observations have implications on nitrate transport mechanisms within this groundwater fed system.

[63] We observed depth variation in hydraulic conductivity, however, longitudinal variations are not apparent. We deduce from this that the localized upwelling is a result of greater connectivity to a local or regional groundwater body. Concentration profiles of pore water chloride observed to a depth of 100 cm below the riverbed reveal areas of limited (or no) surface water mixing, and thus provide further evidence of enhanced vertical flow. Electrical geophysical surveys conducted at the site help to delineate a zone of contrasting groundwater chemistry in this enhanced upwelling region, and highlight the likelihood of a deep (beyond several meters) source. Indirect observations of vertical and horizontal flux add further evidence of localized upwelling. It would appear that the direct connectivity to groundwater exists over only 20% of the study reach, the remainder having more lateral connectivity with shallow groundwater/soil water. Such an area of enhanced upwelling, and resultant limited hyporheic exchange, despite existing bed-form features, may be significant in the overall nitrogen balance in the river as we recognize the potential for rising groundwater nitrate levels. This study provides a hydrological conceptualization for the site, at least under base flow, and, therefore, a necessary framework for our further studies (for example, Heppell et al., submitted manuscript, 2013) which address biogeochemical processes, in relation to nitrogen retention/release, at the site.

[64] Our conceptualization would not have been possible without a multiexperimental programme. Hydraulic head data, although critical in any subsurface hydrological study, can often be inconclusive when used alone. A combination of small-scale sampling of a conservative pore water and river water solute, point dilution tests and geophysical surveys, combined, have allowed us to rationalize our interpretations and remove ambiguities in conceptual models. As we seek to assess dominant processes at the catchment scale, rather than the study reach scale, we are constrained in the use of all these tools. It is unclear if localized zones of river-groundwater connectivity are prevalent at the catchment scale and clearly such high density of sampling is impractical. However, in this study at least, the geophysical survey offers relatively rapid screening of a site. We may expect to see such methods used more widely as reconnaissance tools, allowing the focusing of efforts through more carefully selected piezometer and network sampling.

[65] We note that our conceptualization is based purely on observations under relatively low flows. Localized upwelling of groundwater may be a relatively insignificant contributor of the overall chemical balance under high flows and we recognize that hyporheic exchange processes may be transformed under such conditions due to changes in stage and groundwater head transition times and also the bed infiltration induced by turbulent flow. These mechanisms will be the focus of further study at the site.

[66] Finally, we highlight the use of single well tracer dilution tests for characterizing the horizontal flow component in riverbed sediments. We believe that such an approach offers potentially valuable information, for example, in testing the validity of commonly used 1-D models at the groundwater-surface water interface. And, given its deployment requires only the provision of conventional piezometers, we anticipate future studies adopting this method.

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