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Impact of dam operations on hyporheic exchange in the riparian zone of a regulated river

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Abstract:

Dam operations commonly cause large, frequent fluctuations in river stage, which persist for long distances downstream. The stage fluctuations force river water into and out of the banks, defining lateral hyporheic exchange paths. To evaluate the penetration distance and rates of dam-induced hyporheic exchange, we monitored water-table elevation, temperature, and specific conductivity along a transect perpendicular to the Colorado River (Austin, Texas, USA), 15 km downstream of the Longhorn dam. Stage fluctuates daily by almost a metre. The daily hyporheic exchange volume per metre of bank is $1{\cdot}0$ m^3. Dam-induced hyporheic exchange penetrates several metres into the riparian aquifer, while water-table fluctuations propagate 30 m into the riparian aquifer. Water chemistry and temperature fluctuate near the channel in response to the flow oscillations. In the absence of dam operations, groundwater would flow steadily through the riparian aquifer towards the river, laterally limiting hyporheic exchange and stabilizing temperatures and water chemistry near the channel. Therefore, dam operations fundamentally change the hydrological, thermal, and geochemical dynamics of riparian aquifers and their hyporheic zones.

KEY WORDS hyporheic exchange; surface water-groundwater interaction; hydropeaking; dams; regulated rivers

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INTRODUCTION

Rivers and groundwater are intimately connected across multiple scales. River water infiltrates into the surrounding bed and banks, travels along short groundwater flow paths, and returns to the channel. These flow paths define the hyporheic zone, a critical ecological transition area between fluvial and groundwater ecosystems, which mediates the exchange of water, nutrients, contaminants, and heat (Stanford and Ward, 1988; Brunke and Gonser, 1997). The hyporheic zone functions as a reservoir for solutes and energy in rivers and facilitates important biogeochemical reactions. Hyporheic exchange therefore affects water quality at the watershed scale (Harvey and Fuller, 1998; Harvey and Wagner, 2000; Ensign and Doyle, 2006; Battin *et al.*, 2008).

Rivers can receive groundwater contributions or lose water to the surrounding aquifer, and these large-scale flow patterns impact the finer patterns of hyporheic exchange. For example, groundwater discharge to a gaining reach limits the size of the hyporheic zone (Cardenas and Wilson, 2007; Boano *et al.*, 2008). Since reaches alternate between gaining and losing conditions over seasons and flood events (Krause *et al.*, 2007; Lewandowski *et al.*, 2009), hyporheic exchange also varies (Wondzell and Swanson, 1996; Storey *et al.*, 2003). Some reaches alternate between gaining and losing conditions daily in response to stage fluctuations. Daily stage fluctuations can result from natural processes such as evapotranspiration and snowmelt, as well as human activities such as the release of water from hydroelectric dams. In these cases, the daily reversals in gaining and losing conditions represent a unique form of hyporheic exchange. The penetration distance of river water into the surrounding aquifer prior to flow-path reversal determines the size of the hyporheic zone. The frequency of flow-path oscillations determines hyporheic residence times.

Several authors have claimed that dam-induced river stage fluctuations impact hyporheic exchange (Winter *et al.*, 1998; Nilsson and Berggren, 2000; Hancock 2002), but few have presented measurements. Curry *et al.* (1994) monitored pressure and water chemistry within the riverbed at brook trout spawning sites in Ontario, Canada during a period of fluctuating dam releases. River stage fluctuations altered vertical hydraulic gradients and water chemistry. Arntzen *et al.* (2006) monitored riverbed pressure, temperature, and water chemistry in a reach of the Columbia River (Washington, USA), where hydroelectric dam operations induced daily stage fluctuations of up to two metres. They found that vertical head gradients, temperature, and water chemistry oscillated with river stage. They later estimated the rate of vertical hyporheic exchange in response to stage fluctuations (Fritz and Arntzen, 2007). Hanrahan (2008) monitored pressure and temperature in the bed of the Snake River downstream from Hells Canyon dam (Pacific Northwest, USA). They found that dam operations only induced vertical flux reversals in a few localized areas and hypothesized that

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pool-riffle morphology and steep channel gradients predominantly drove hyporheic exchange. Of these studies, only Fritz and Arntzen (2007) estimated a hyporheic exchange rate due to dam operations, and none of these studies estimated the size of the hyporheic exchange zone.

The goal of this study is to estimate the spatial extent and volumetric rate of dam-induced hyporheic exchange using high-resolution temporal data. We monitored pressure, temperature, and specific conductivity along a transect perpendicular to the Colorado River near Austin, Texas, USA. We show that dam-induced river stage fluctuations drive hyporheic exchange metres into the river bank and modulate temperatures and conductivity near the channel. Water-table fluctuations and groundwater flow reversals propagate tens of metres into the riparian aquifer. In the absence of dam operations, groundwater would predominantly flow towards the river (Larkin and Sharp, 1992) and limit the hyporheic zone laterally (Cardenas and Wilson 2007; Boano *et al*., 2008). Therefore, dam operations fundamentally change the hydrodynamics of the hyporheic zone and riparian aquifer.

STUDY SITE

The study area is located 10 km from downtown Austin, Texas within the 5-km^2 Hornsby Bend Center for Environmental Research, which borders the Colorado River (Figure 1). At Hornsby Bend, the Colorado River is a regulated fourth-order river. Upstream from Hornsby Bend, the Tom Miller dam releases water daily to generate hydroelectric power for the city of Austin. Between Hornsby Bend and the Tom Miller dam is the Longhorn dam, which forms a recreational lake as well as a cooling reservoir for an adjacent power plant. Two kilometres downstream from the Longhorn dam, river stage fluctuates by over 2·5 m due to dam operations. Fifteen kilometres downstream at the study site, river stage fluctuates by almost a metre.

Regional groundwater flow is towards the Colorado River at the study site (see Figure 4d in the work by Larkin and Sharp (1992)). The aquifer is composed of modern alluvial deposits, and the average aquifer conductivity is 10^{-4} m/s (Larkin and Sharp, 1992).

METHODS

Monitoring

Four bank piezometers and one river stage recorder were installed in September 2008 to monitor lateral hyporheic exchange near the channel (Figure 2). The distance between the river stage recorder and first bank piezometer is 5·91 m. The distances of bank piezometers from mean shoreline are 2·35, 5·08, 7·43, and 9·48 m. A terrace prevented installation of additional piezometers farther from the river. Bank piezometers were constructed using PVC of 5 cm diameter and screened through the water table. Screened intervals range from 1·2 to 2·4 m in length. The river stage recorder was constructed using 1·8 m of screened PVC and anchored in the riverbed with a 1·2 m section of solid PVC. We installed vented *in situ* Aqua Troll 200 probes inside bank piezometers and the river stage recorder to log pressure, temperature, and specific conductivity at 15-min intervals for 7 days. Probes were calibrated for temperature and specific conductivity. Piezometer elevations and bank topography were surveyed using a Sokkia Total Station.

Aquifer characterization

Our primary goal was to estimate hyporheic zone size and exchange rates in response to dam operations, which requires knowledge of aquifer hydraulic properties. We used daily head changes in the river and riparian aquifer to estimate hydraulic diffusivity (D). Analytical solutions for aquifer response to river stage fluctuations are numerous, and the practice of estimating aquifer hydraulic properties from water-table fluctuations is common (e.g. Rowe, 1960; Swamee and Singh, 2003; Srivastava, 2006).

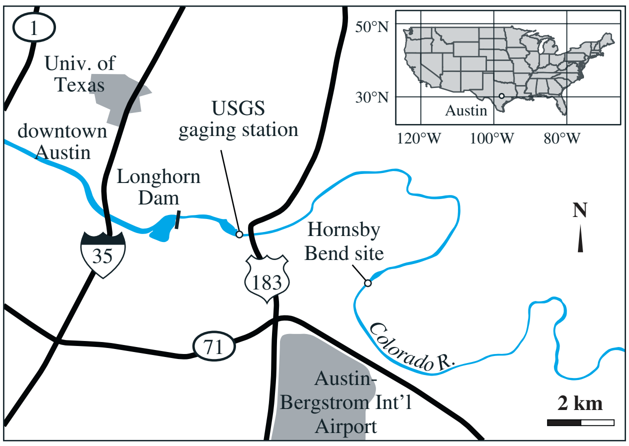

Figure 1. Location of study site on the Colorado River in relation to Austin, Texas, USA. USGS gaging station 08158000 is 2 km downstream from Longhorn dam, and the study site is another 13 km downstream

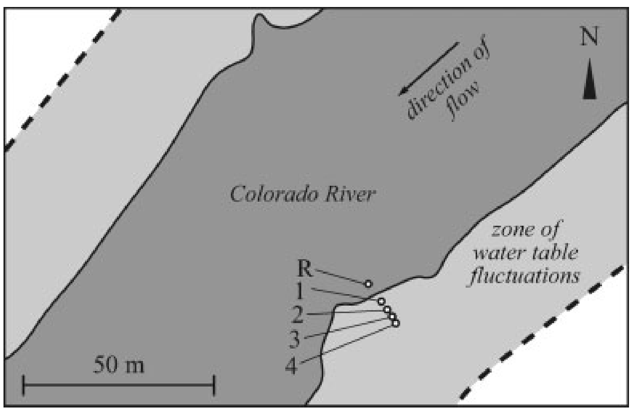

Figure 2. Map of Hornsby Bend piezometer transect. Bank piezometers are numbered in order of distance from the river, and the river stage recorder is denoted as (R). Dashed lines indicate the estimated extent of dam influence on the water table

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Consider a homogeneous, unconfined aquifer adjacent to a river with periodic stage fluctuations. The one-dimensional linearized Boussinesq equation describes transient flow in the aquifer:

$$\frac{\partial h}{\partial t} = D\frac{\partial^2 h}{\partial x^2}, \tag{1}$$

where h is hydraulic head, x is distance, and t is time. D is related to hydraulic conductivity (K) by:

$$D = \frac{Kb}{S_y}, \tag{2}$$

where S_y is specific yield, and b is saturated aquifer thickness. At the river bank ($x = 0$), the aquifer head equals the river stage:

$$h(x = 0, t) = A\sin(\omega t + \phi), \tag{3}$$

where ω is frequency, A is amplitude, and ϕ is phase. At large distance from the river, lateral flow in the aquifer approaches zero:

$$\frac{\partial h}{\partial x} = 0 \ at \ x \to \infty, \ \ 0 \le t \le \infty. \tag{4}$$

Singh (2004) presented an analytical solution for the head response in the semi-infinite aquifer:

$$h(x, t) = A\exp\left(-x\sqrt{\frac{\omega}{2D}}\right) \times \sin\left[-x\sqrt{\frac{\omega}{2D}} + \omega t + \phi\right]. \tag{5}$$

This solution assumes a straight river and a homogeneous aquifer that is significantly thicker than the amplitude of water-table fluctuations. Equation 5 also assumes negligible riverbed resistance. By superposition of individual frequency components, Equation 5 may represent the aquifer response to more complicated river stage signals.

To evaluate D at the Hornsby Bend site, we determined the amplitudes and phases of the daily (24-h) river stage and water-table oscillations at each piezometer using discrete fast Fourier transforms (FFTs). We then determined D values that minimized the least squares residual error (LSRE) between the measured and predicted heads. Kb was calculated from D (Equation 2) by assuming a typical value for S_y (Freeze and Cherry, 1979). K was calculated by assuming a reasonable estimate of b based on the approximate depth to a regional aquitard.

To independently estimate K, we conducted grain size analyses on sediment samples from the aquifer transect and performed three in-stream pneumatic slug tests near the river stage recorder (Cardenas and Zlotnik, 2003). Sediment samples from the aquifer transect were collected during hand-augering and were sieved to determine grain size distributions. Samples were oven-dried and vibrated through a stack of eight sieves. From the grain size distributions, we calculated K according to the methods of Hazen (1911) and Kozeny (1927). We could only conduct pneumatic slug tests in the riverbed because bank piezometers were screened through the water table. However, the gravel and cobble composition of riverbed sediment visually resembled gravel zones that we encountered in the aquifer while installing bank piezometers.

Estimation of volumetric flow rate

The volumetric flow rate across the bank per unit length of river (Q) is:

$$Q(t) = -Kb\frac{\partial h(x = 0, t)}{\partial x}. \tag{6}$$

Positive values indicate flow from the river into the aquifer. $Q(t)$ was estimated using our best-fit transmissivity (Kb) and head gradients between the river stage recorder and Bank Piezometer 1. In calculating the head gradient, we projected the head at the river stage recorder to the time-averaged shoreline location.

Estimation of hyporheic zone lateral extent

We estimated the extent of the hyporheic zone using two separate approaches based on the hydraulic and geochemical definitions of the hyporheic zone. In the hydraulic approach, the hyporheic zone comprises all flow paths that start and end in the channel. The advective penetration distance of river water into the surrounding aquifer prior to flow-path reversal thus determines the size of the hyporheic zone. We used particle tracking to estimate this penetration distance. Particle velocities, $v(x, t)$, were calculated as:

$$v(x, t) = -\frac{K}{n_e}\frac{\partial h(x, t)}{\partial x} \tag{7}$$

where n_e is effective porosity. Head gradients were estimated from water table elevation data.

In the geochemical approach, the hyporheic zone is the volume of sediment containing at least 10% river water (Triska *et al.*, 1989). Under this definition, hyporheic exchange includes both advective and dispersive mixing between river water and groundwater. We used mean specific conductivity ($\overline{S}$) to calculate the proportion of river water in each piezometer (W):

$$W = 100\frac{\overline{S}_{gw} - \overline{S}}{\overline{S}_{gw} - \overline{S}_{riv}}, \tag{8}$$

where $\overline{S}_{gw}$ is the mean specific conductivity of groundwater, and $\overline{S}_{riv}$ is the mean specific conductivity of river water. We assumed $\overline{S}_{gw}$ equals the mean specific conductivity at Bank Piezometer 4, and $\overline{S}_{riv}$ equals the mean specific conductivity at the river stage recorder.

RESULTS

Water-table fluctuations were damped and lagged at every bank piezometer relative to river stage fluctuations

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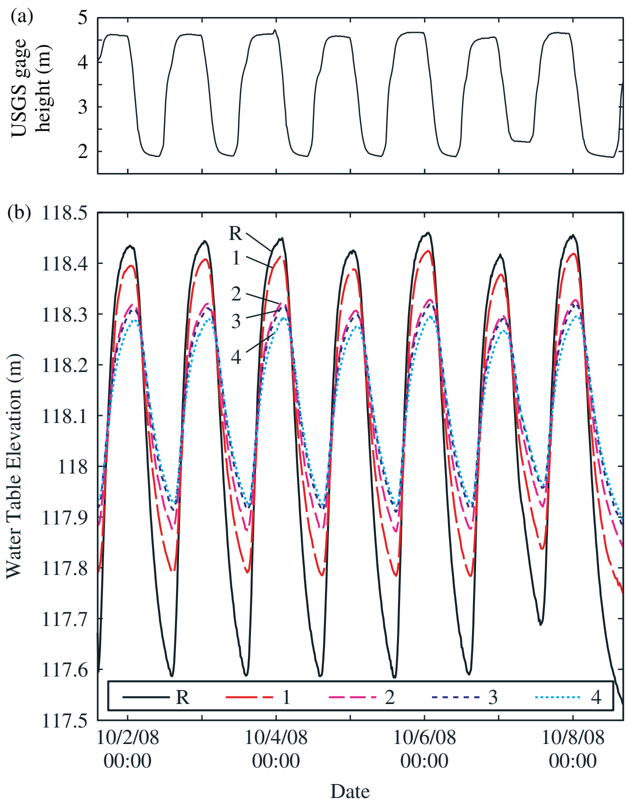

Figure 3. (a) Continuous USGS stream gage record from the Colorado River near Austin, Texas from October 1, 2008 at 14:15 to October 8, 2008 at 16:15. The gaging station is located 2 km downstream from Longhorn dam, and river stage fluctuates 2·5 m daily. (b) Water-table elevation time series at the study site. Distances of Bank Piezometers 1–4 from mean shoreline are 2·35, 5·08, 7·43, and 9·48 m, respectively. The study site is 15 km downstream from Longhorn dam, and river stage fluctuates 85 cm daily. Water-table fluctuations are attenuated and lagged relative to river stage fluctuations

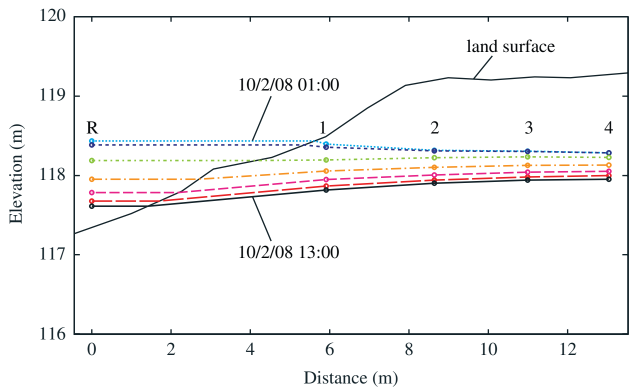

Figure 4. Water-table elevation profiles at 2-h intervals over a period of 12 h beginning on October 2, 2008 at 01:00. Land surface profile is also shown. Water-table fluctuations dissipate with distance from the river

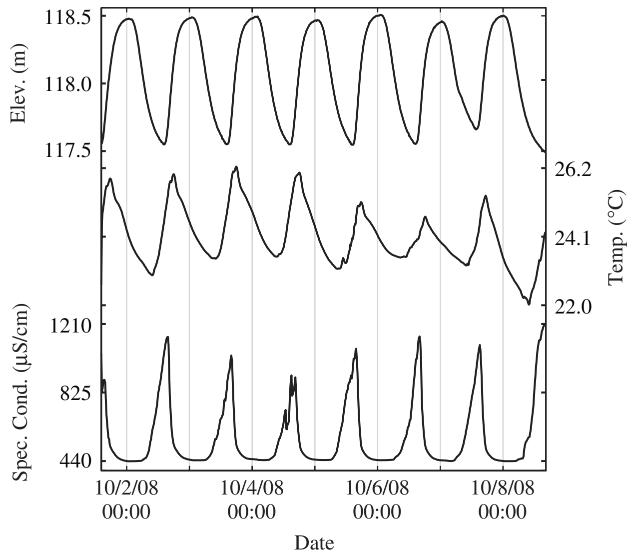

Figure 5. River stage, temperature, and specific conductivity time series. Temperature peaks in the early evening. Specific conductivity maxima coincide with river stage minima

(Figures 3 and 4). Amplitudes of water-table fluctuations, in order of increasing piezometer distance from the river, were 71%, 50%, 44%, and 40% of the river stage amplitude. Phases were $-5°$, $-11°$, $-14°$, and $-18°$, respectively, which correspond to lags of 22, 42, 57, and 70 min.

The daily temperature variation in the river was approximately 3 °C (Figure 5). River temperature was warmest during the evening and coolest during the morning, consistent with typical diurnal river temperature patterns. In the riparian aquifer, daily temperature fluctuations were subtle in all piezometers except one (Figure 6). The piezometer closest to the river (Bank Piezometer 1) recorded daily temperature fluctuations of up to 0·5 °C (Figure 6).

Specific conductivity in the river fluctuated daily by approximately 600 μS/cm (Figures 5 and 7). Conductivity peaks coincided roughly with river stage minima (Figure 5). Within the aquifer, Bank Piezometer 1 recorded the greatest specific conductivity fluctuations (∼500 μS/cm) (Figure 7). The fluctuations were lagged and attenuated relative to the river conductivity signal. Specific conductivity in Bank Piezometer 3 fluctuated daily by approximately 35 μS/cm, while specific conductivity in Bank Piezometers 2 and 4 varied negligibly over the monitoring period (Figure 7). Mean specific conductivities at the bank piezometers increased with distance from the river.

Best-fit D values for individual bank piezometers based on LSRE analysis were $4{\cdot}0 \times 10^{-3}$, $4{\cdot}2 \times 10^{-3}$, $6{\cdot}2 \times 10^{-3}$, and $7{\cdot}7 \times 10^{-3}$ m^2/s, respectively. To estimate K, we assumed a typical S_y value for sandy gravel of 0·25 (Freeze and Cherry, 1979). The aquifer thickness (b) is unknown but is likely between 1 and 10 m, based on cuttings from a nearby well that penetrates a regional aquitard. Assuming D equals $5{\cdot}5 \times 10^{-3}$ m^2/s, K lies between $1{\cdot}4 \times 10^{-4}$ m/s and $1{\cdot}4 \times 10^{-3}$ m/s (Equation 2). For comparison, pneumatic slug tests in the riverbed yielded K values of $6{\cdot}5 \times 10^{-4}$ m/s, $7{\cdot}9 \times 10^{-4}$ m/s, and $9{\cdot}7 \times 10^{-4}$ m/s.

Grain size analyses yielded smaller K estimates. Sediment samples were predominantly composed of poorly sorted pea-gravel and sand (Figure 8). Although cobble-dominated samples were collected from the piezometer transect, they were too coarse for sieving. Of the

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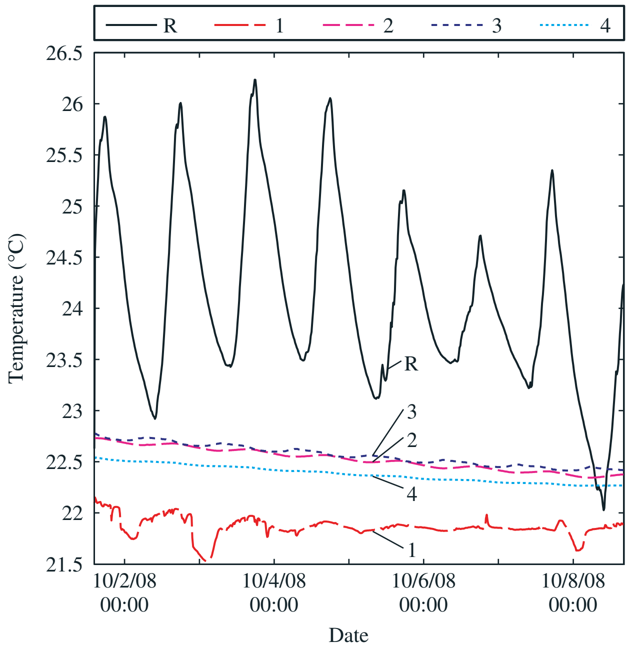

Figure 6. Temperature time series in the river stage recorder and four bank piezometers. Distances of Bank Piezometers 1–4 from mean shoreline are 2·35, 5·08, 7·43, and 9·48 m, respectively. Temperature fluctuations in Bank Piezometer 1 are attenuated relative to temperature fluctuations in the river. Temperatures at the other bank piezometers are more stable

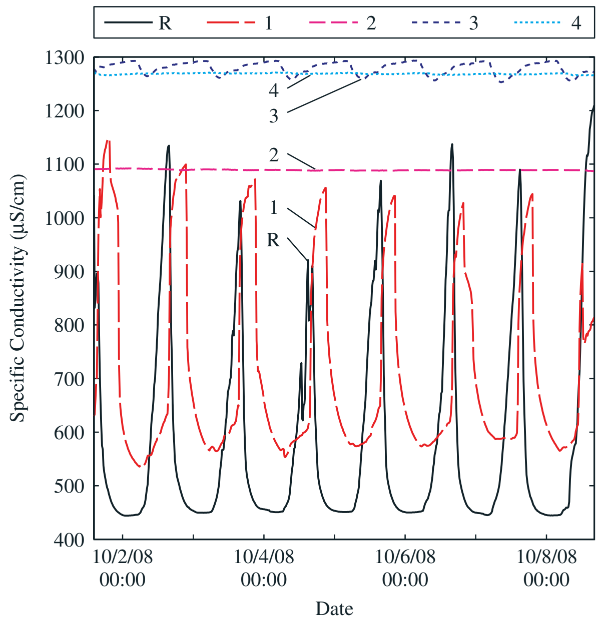

Figure 7. Specific conductivity time series in the river stage recorder and four bank piezometers. Distances of Bank Piezometers 1–4 from mean shoreline are 2·35, 5·08, 7·43, and 9·48 m, respectively. Fluctuations are greatest at Bank Piezometer 1. Specific conductivity in Bank Piezometer 2 remained stable, likely due to mud on the sensor that we observed during retrieval

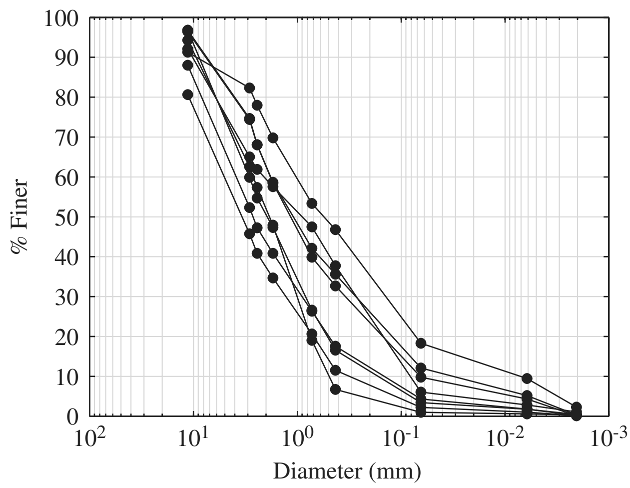

Figure 8. Grain size distributions for aquifer sediment samples. The samples are predominantly sandy, although we also collected cobble-dominated samples that were too coarse for dry-sieving

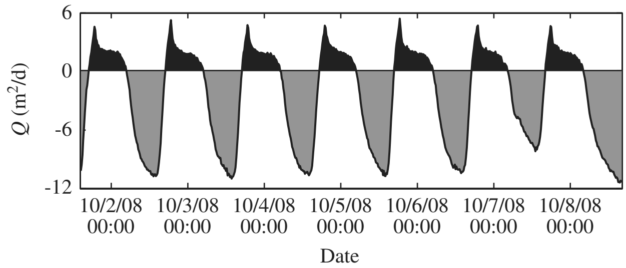

Figure 9. Volumetric flow rate between the river and riparian aquifer per metre of bank. Positive values indicate flow from the channel into the bank. The black shaded area represents the volume of water pumped through the hyporheic zone per metre of bank over the duration of the study. The gray shaded area represents the total volume of water discharged into the river (both hyporheic water and baseflow)

analysed samples, mean grain size diameters ranged from 0·5 to 2·5 mm (coarse to very coarse sand). The average K estimate from the grain size distributions is $2{\cdot}6 \times 10^{-5}$ m/s.

Assuming a value of $1{\cdot}4 \times 10^{-3}$ m^2/s for Kb, the instantaneous flow rate across the bank per unit length of river ranged from -11 m^2/day to 5·4 m^2/day during the monitoring period (Equation 6, Figure 9). By integration of $|Q(t)|$ through time, the daily exchange volume per unit length of bank averaged 4·5 m^3 (3·5 m^3 towards the river and 1·0 m^3 away from the river). The river was therefore gaining on average, in agreement with the study by Larkin and Sharp (1992). Of the 3·5 m^3 of water that discharged daily from each metre of bank, 2·5 m^3 was baseflow, and 1·0 m^3 was hyporheic water associated with dam operations. The flow direction was from the river into the riparian aquifer from evening until early morning and from the aquifer to the river from early morning until evening.

Assuming our range of K values from the Boussinesq model and an effective porosity of 0·25, we estimate that the hydraulically defined hyporheic zone (based on particle tracking) extends 0·41 to 5·1 m into the riparian aquifer. The geochemically defined hyporheic zone extends at least 5 m into the riparian aquifer: Bank Piezometer 2 is located 5·08 m from mean shoreline and contains 26% river water (Equation 8). Therefore, the size of the hyporheic zone depends on the definition of hyporheic exchange.

While the hyporheic zone extends a few metres into the riparian aquifer, water-table fluctuations propagate

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much farther. The penetration distance of water-table fluctuations into the riparian aquifer can be estimated from Equation 5, given D. The amplitude of water-table fluctuations in the aquifer decays to 10% of the river stage amplitude at a distance x_{10} satisfied by:

$$0{\cdot}1 = \exp\left(-x_{10}\sqrt{\frac{\omega}{2D}}\right). \quad (9)$$

Assuming D equals $5{\cdot}5 \times 10^{-3}$ m^2/s, x_{10} is 28 m at Hornsby Bend (Figure 2).

DISCUSSION

Hyporheic response to dam operations

Dam-induced stage fluctuations at Hornsby Bend force water in and out of the riparian aquifer daily. Flow paths that start and end in the channel constitute a zone of lateral hyporheic exchange. The residence times (hours) and path lengths (centimetres to metres) of dam-induced hyporheic exchange are similar to hyporheic exchange induced by natural processes (Harvey and Wagner, 2000). However, a fundamental difference exists between them. Dam-induced hyporheic exchange is intrinsically transient because unsteady pressures at the sediment–water interface drive the exchange. In contrast, 'natural' hyporheic exchange (due, for example, to flow over an undulating riverbed) is comparatively steady because time-averaged pressure gradients at the sediment–water interface drive the exchange.

We can estimate the fraction of river water exchanged through the hyporheic zone between Longhorn dam and Hornsby Bend using the daily hyporheic exchange volume at Hornsby Bend. One cubic metre of river water enters and exits the hyporheic zone daily per unit length of bank. The daily exchange volume likely increases with proximity to the Longhorn dam, since river stage fluctuations increase from 85 cm at Hornsby Bend to 2·5 m near the dam (Figure 3). Nevertheless, if we assume a uniform exchange volume, dam operations pump 30 000 m^3 of river water through the hyporheic zone each day along the 15-km reach. The mean discharge recorded at USGS gaging station 08 158 000 is 30 m^3/s (Figure 1). Therefore, at least 1% of all river water passes through the hyporheic zone in the 15-km reach. Water-table fluctuations also persist downstream of Hornsby Bend, providing new opportunities for hyporheic exchange.

Three potential sources of error may impact our estimate of the daily hyporheic exchange volume, which depends on our *Kb* estimate from the analytical solution to the linearized Boussinesq equation. First, the analytical solution assumes a no-flow boundary at a large distance from the river. A more appropriate boundary condition would be constant flow towards the river. Using a no-flow boundary in place of a constant flow boundary tends to underestimate *Kb*. Our calculated dam-driven hyporheic flux is therefore a minimum estimate. Second, the analytical solution is one-dimensional and assumes a fixed shoreline. In reality, seepage directions in the river banks may include vertical or along-channel components. For example, the anomalously cool mean temperature at Bank Piezometer 1 may be due to multi-dimensional flow paths (Figure 6). Third, the analytical solution assumes the phreatic aquifer is significantly thicker than the amplitude of river stage fluctuations. Hornberger *et al.* (1970) showed for an analogous problem that the linearized Boussinesq approximation is adequate for amplitudes up to 50% of the mean aquifer thickness. At Hornsby Bend, the amplitude is 40% of our minimum aquifer thickness estimate.

Chemical and thermal implications

Dam operations fundamentally alter hydrologic flow paths within the riparian aquifer up to 30 m from the river at Hornsby Bend (Figure 10). In the absence of dam operations, groundwater would flow steadily towards the river through the riparian aquifer like water flowing through gills (Larkin and Sharp, 1992). The steady discharge of groundwater would limit lateral hyporheic exchange (Cardenas and Wilson, 2007; Boano *et al.*, 2008). In contrast, dam operations induce stage fluctuations that drive river water in and out of the riparian aquifer. The unsteady, reversing flow of water through the riparian aquifer is like the flow of air through lungs (Figure 10). We discuss in the following text how this fundamental change in riparian aquifer behaviour may influence water chemistry and temperature downstream of dams, based on our observations along the Hornsby Bend transect (Table I).

At Hornsby Bend, water chemistry fluctuates in both the river and hyporheic zone (Figure 7). The specific conductivity of river water is greatest in between dam releases during baseflow periods and lowest during dam releases when reservoir water dominates the flow (Figure 5). Hyporheic exchange transports the specific conductivity signal into the riparian aquifer. As a result, the specific conductivity of pore water near the channel is damped and lagged relative to the specific conductivity of the river (Figure 7). In the absence of dam operations, river and riparian aquifer chemistry would fluctuate less (Figure 10, Table I). The specific conductivity of river water would consistently resemble values measured during baseflow periods. Since steady groundwater discharge to the river would limit lateral hyporheic exchange, the specific conductivity of pore water near the channel would consistently resemble the specific conductivity of groundwater.

Along steady groundwater flow paths, a ladder of redox conditions would also regulate nitrogen, sulfur, and organic carbon inputs to the river (Hill, 2000). We hypothesize that dam operations cause fluctuations in biogeochemical reaction rates within the riparian aquifer. The temporary convergence of chemically distinct waters, such as surface water and groundwater, can increase biogeochemical activity (McClain *et al.*, 2003). These periods of increased activity, termed hot moments, specifically occur when a temporal change in hydrologic flow

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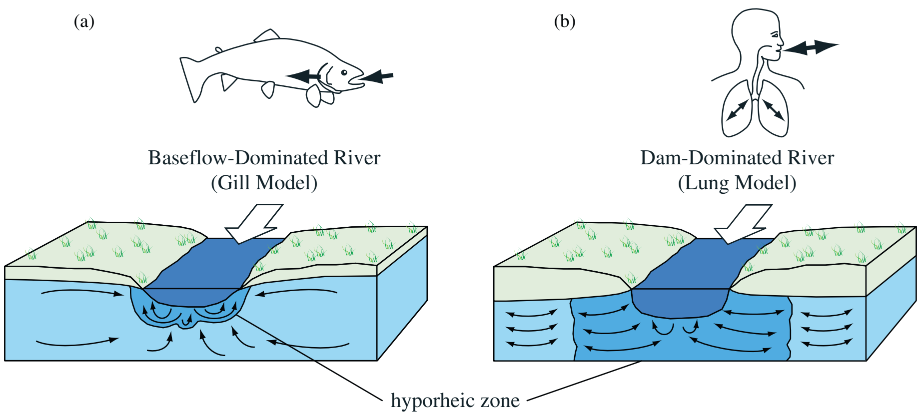

Figure 10. (a) Conceptual model of a natural river-groundwater system in a reach dominated by baseflow. During most of the year, groundwater flows steadily through the riparian aquifer in one direction like water through a gill. Groundwater discharge to the river limits the size of the hyporheic zone. (b) Conceptual model of a river-groundwater system downstream of a dam. Due to frequent stage fluctuations, river water flows in and out of the riparian aquifer like air flowing in and out of lungs. The hyporheic zone includes all flow paths that start and end in the channel

paths introduces a limiting ingredient that allows the biogeochemical reaction to proceed. The reaction continues as long as a continuous source of reactants persists (typically, as long as the flow paths continue to converge). Examples of processes that trigger hot moments include arroyo floods, storms, and periods of intense snowmelt (McClain *et al.*, 2003). A goal for future research is to determine whether dam operations frequently induce hot moments.

Dam operations may also impact riparian aquifer temperatures near the channel. At Hornsby Bend, temperatures fluctuated daily up to 0·5 °C at Bank Piezometer 1. Thermal conduction between the atmosphere and land cannot explain the observed temperature oscillations. Given a typical thermal diffusivity of 7×10^{-7} m^2/s for saturated sediment, a 24-h atmospheric temperature signal would only penetrate 0·25 m into the ground. In unsaturated sediment, the penetration depth is even less. All temperature sensors were located more than a metre below land surface. Although coupled fluid flow and heat transport simulations are beyond the scope of this paper, we suggest that thermal advection associated with hyporheic exchange influences the temperature signal near the channel. In the absence of dam operations, the flow of groundwater to the river would stabilize temperatures near the channel.

Finally, dam operations may impact river temperature by periodically interrupting the flow of groundwater to the river. Groundwater discharge locally buffers temperature in the channel, creating thermal refuges for fish (Arrigoni *et al.*, 2008). At Hornsby Bend, groundwater only discharges to the river in between dam releases.

Table I. Conceptual model of impact of dam operations on a river-aquifer system dominated by baseflow

Variable	The Gill model: baseflow-dominated rivers [modified after Hill (2000)]	The Lung model: rivers influenced by dam operations (this study)
Water-table variation	Moderate to small	Rapid, large fluctuations
Hydrologic link to river	Steady groundwater flow towards river	Flow directions regularly reverse over short timescales
	Flood events can temporarily perturb flow paths and rates	
Hyporheic zone characteristics	Laterally limited by groundwater discharge to the river	Laterally extensive
	Flows driven by bedform topography, river morphology, and in-stream structures	Flows driven by river stage fluctuations downstream of dams
	Flows are generally steady but may vary in response to floods, seasons, and morphodynamics	Flows are unsteady and reverse at timescales of hours to days
Example of effects on river temperature	Hyporheic and groundwater inputs steadily influence river temperatures	Hyporheic and groundwater inputs only influence river temperatures when flow direction is towards river
Example of effects on river chemistry	Riparian aquifer is a sink for NO_3 and SO_4 and a source for DOC	Uncertain but likely a function of competition between hyporheic residence times and reaction kinetics
Example of effects on riparian aquifer chemistry	Variation in redox conditions with distance along groundwater flow paths	Uncertain but likely variable in both time and space

DOC, dissolved organic carbon

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In unregulated, baseflow-dominated rivers, continuous groundwater discharge would potentially provide more persistent thermal refuges in the channel.

CONCLUSIONS AND RECOMMENDATIONS

At Hornsby Bend, dam operations pump river water in and out of the riparian aquifer daily. Flow paths that begin and end in the river form a hyporheic zone that extends approximately 1–5 m into the riparian aquifer. The daily volume of water that passes through the hyporheic zone is 1 m^3 per metre of bank. To our knowledge, this is the first study to estimate these metrics for lateral hyporheic exchange due to dam operations. The same method could be used to characterize hyporheic exchange due to any frequent river stage fluctuation.

Worldwide, over half of all large rivers are regulated by dams (Nilsson *et al*., 2005). A clear direction for future research is to characterize how the hyporheic response to dam operations influences thermal budgets, biogeochemistry, and ecology at both the reach scale and watershed scale. For example, how do flow reversals in the riparian aquifer impact redox conditions and biogeochemical zonation? What is the resulting impact on solute transport at the watershed scale? Are steady or oscillating groundwater flows more beneficial to the biogeochemistry and ecology of rivers, and under what conditions? Investigating the answers to these fundamental questions will help assess the impact of dams on watershed health.

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REFERENCES

Arntzen EV, Geist DR, Dresel PE. 2006. Effects of fluctuating river flow on groundwater/surface water mixing in the hyporheic zone of a regulated, large cobble bed river. *River Research and Applications* **22**: 937–946. DOI: 10·1002/rra.947.

Arrigoni AS, Poole GC, Mertes LAK, O'Daniel SJ, Woessner WW, Thomas SA. 2008. Buffered, lagged, or cooled? Disentangling hyporheic influences on temperature cycles in stream channels. *Water Resources Research* **44**: 1–13. DOI: 10·1029/2007WR006480.

Battin TJ, Kaplan LA, Findlay S, Hopkinson CS, Marti E, Packman AI, Newbold JD, Sabater F. 2008. Biophysical controls on organic carbon fluxes in fluvial networks. *Nature Geoscience* **1**: 95–100. DOI: 10·1038/ngeo101.

Boano F, Revelli R, Ridolfi L. 2008. Reduction of the hyporheic zone volume due to the stream-aquifer interaction. *Geophysical Research Letters* **35**: L09401. DOI: 10·1029/2008GL033554.

Brunke M, Gonser T. 1997. The ecological significance of exchange processes between rivers and groundwater. *Freshwater Biology* **37**: 1–33.

Cardenas MB, Wilson JL. 2007. Exchange across a sediment-water interface with ambient groundwater discharge. *Journal of Hydrology* **346**: 69–80. DOI: 10·1016/j.jhydrol.2007·08·019.

Cardenas MB, Zlotnik VA. 2003. Three-dimensional model of modern channel bend deposits. *Water Resources Research* **39**: 1141–1153. DOI: 10·1029/WR001383.

Curry RA, Gehrels J, Noakes DLG, Swainson R. 1994. Effects of river flow fluctuations on groundwater discharge through brook trout, Salvelinus fontinalis, spawning and incubation habitats. *Hydrobiologia* **277**: 121–134.

Ensign SH, Doyle MW. 2006. Nutrient spiraling in streams and river networks. *Journal of Geophysical Research-Biogeosciences* **111**: 1–13. DOI: 10·1029/2005jg000114.

Freeze RA, Cherry JA. 1979. *Groundwater*. Prentice-Hall: Englewood Cliffs, NJ, 604.

Fritz BG, Arntzen EV. 2007. Effect of rapidly changing river stage on uranium flux through the hyporheic zone. *Groundwater* **45**: 753–760. DOI: 10·1111/j.1745–6584·2007·00365.x.

Hancock PJ. 2002. Human impacts on the stream-groundwater exchange zone. *Environmental Management* **29**: 763–781. DOI: 10·1007/s00267-001-0064-5.

Hanrahan TP. 2008. Effects of river discharge on hyporheic exchange flows in salmon spawning areas of a large gravel-bed river. *Hydrological Processes* **22**: 127–141. DOI: 10·1002/hyp.6605.

Harvey JW, Fuller CC. 1998. Effect of enhanced manganese oxidation in the hyporheic zone on basin-scale geochemical mass balance. *Water Resources Research* **34**: 623–636.

Harvey JW, Wagner BJ. 2000. Quantifying hydrologic interactions between streams and their subsurface hyporheic zones. In *Streams and Ground Waters*, Jones JB, Mulholland PJ(eds). Academic Press: San Diego, CA: 1–43.

Hazen A. 1911. Discussion: dams on sand foundations: transactions. *American Society of Civil Engineers* **73**: 199.

Hill AR. 2000. Stream chemistry and riparian zones. In: *Streams and Ground Waters*, Jones JB, Mulholland PJ (eds). Academic Press: San Diego: 83–110.

Hornberger GM, Ebert J, Remson I. 1970. Numerical solution of the Boussinesq equation for aquifer-stream interaction. *Water Resources Research* **6**: 601–608.

Kozeny J. 1927. Uber kapillare leitung des wassers im boden (On capillary flow of water in soil). *Wien, Akad. Wiss.* **136**: 271–306.

Krause S, Bronstert A, Zehe E. 2007. Groundwater-surface water interactions in a North German lowland floodplain–Implications for the river discharge dynamics and riparian water balance. *Journal of Hydrology* **347**: 404–417. DOI: 10·1016/j.jhydrol.2007·09·028.

Larkin RG, Sharp JM. 1992. On the relationship between river-basin geomorphology, aquifer hydraulics, and ground-water flow direction in alluvial aquifers. *Geological Society of America Bulletin* **104**: 1608–1620.

Lewandowski J, Lischeid G, Nützmann G. 2009. Drivers of water level fluctuations and hydrological exchange between groundwater and surface water at the lowland River Spree (Germany): Field study and statistical analyses. *Hydrological Processes*. DOI: 10·1002/hyp.7277.

McClain ME, Boyer EW, Dent CL, Gergel SE, Grimm NB, Groffman PM, Hart SC, Harvey JW, Johnston CA, Mayorga E, McDowell WH, Pinay G. 2003. Biogeochemical hot spots and hot moments at the interface of terrestrial and aquatic ecosystems. *Ecosystems* **6**: 301–312. DOI: 10·1007/s10021-003-0161-9.

Nilsson C, Berggren K. 2000. Alterations of riparian ecosystems caused by river regulation. *Bioscience* **50**: 783–792.

Nilsson C, Reidy CA, Dynesius M, Revenga C. 2005. Fragmentation and flow regulation of the world's largest river systems. *Science* **308**: 405–408. DOI: 10·1126/science.1107887.

Rowe PP. 1960. An equation for estimating transmissibility and coefficient of storage from river-level fluctuations. *Journal of Geophysical Research* **65**: 3419–3424.

Singh SK. 2004. Aquifer response to sinusoidal or arbitrary stage of semipervious stream. *Journal of Hydraulic Engineering* **130**: 1108–1118. DOI: 10·1061/(ASCE)0733–9429(2004)130 : 11(1108).

Srivastava R. 2006. Aquifer diffusivity estimation from response to stream stage variation. *Journal of Hydrologic Engineering* **11**: 273–277. DOI: 10·1061/(ASCE)1084-0699(2006)11 : 3(273).

Stanford JA, Ward JV. 1988. The hyporheic habitat of river ecosystems. *Nature* **335**: 64–66.

Storey RG, Howard KWF, Williams DD. 2003. Factors controlling riffle-scale hyporheic exchange flows and their seasonal changes in a gaining stream: a three-dimensional groundwater flow model. *Water Resources Research* **39**: 1034–1050. DOI: 10·1029/2002WR001367.

Swamee PK, Singh SK. 2003. Estimation of aquifer diffusivity from stream stage variation. *Journal of Hydrologic Engineering* **8**: 20–24. DOI: 10·1061/(ASCE)1084-0699(2003)8 : 1(20).

Triska FJ, Kennedy VC, Avanzio RJ, Zellweger GW, Bencala KE. 1989. Retention and transport of nutrients in a third-order stream in northwestern California: Hyporheic processes. *Ecology* **70**: 1893–1905.

Winter TC, Harvey JW, Franke OL, Alley WA. 1998. Groundwater and surface water: A single resource. *U.S. Geological Survey Circular* **1139**: 1–79.

Wondzell SM, Swanson FJ. 1996. Seasonal and storm dynamics of the hyporheic zone of a 4th-order mountain stream. I: Hydrologic processes. *Journal of the North American Benthological Society* **15**: 3–19.