Hydraulic and thermal response of groundwater–surface water exchange to flooding in an experimental aquifer

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S U M M A R Y

The effects of flooding on interactions between streams and shallow aquifers are poorly studied due to the unpredictable nature of flooding events. This study examines groundwater–surface water interactions in a constructed experimental meander over several simulated flood events using hydrologic and thermal monitoring. Detailed steady-state hydraulic head measurements were taken of the stream surface and in 59 piezometer nests during baseflow, bankfull flow, and overbank flooding conditions, allowing for three-dimensional hydrologic characterization. Additionally, a lithium tracer test was conducted, and pressure transducers and thermistors were deployed to capture transient behavior. Results demonstrate extensive coupling between the stream and adjacent alluvial aquifer under all discharge conditions. Water table elevation responds rapidly to changes in stream stage and re-equilibrates to order of magnitude increases in discharge within about 1 h. Though flooding elevates the water table, steady-state hydraulic gradients within the meander are independent of stream stage. This results from the flow boundary condition imposed by the stream, which deepens in response to flooding but has a slope that is essentially independent of discharge. In addition to the stream boundary, flow within the meander is also controlled by loss of stream water to the subsurface, which directs hydraulic gradients towards the base of the meander. Finally, the temperature distribution within the meander during normal and bankfull flow mimics the water table distribution and results from advected warm stream water being progressively cooled by vertical conduction in the direction of groundwater flow. These findings suggest that areas of high reaction rates within meanders (hot spots) will be sensitive to whether a stream reach is gaining or losing water to the subsurface. Further, the location of these hot spots of is likely independent of stream stage, though other controls on reaction rates may be affected.

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1. Introduction

Groundwater–surface water exchange mediates thermal, biogeochemical, and ecological processes integral to aquatic systems comprising the stream–aquifer continuum (Findlay, 1995). Several processes have been shown to create hydraulic head gradients along the stream–aquifer interface and induce exchange between groundwater and surface water. Within a stream or river, bed topography and other submerged obstructions induce dynamic head gradients that generate predominantly vertical subsurface flow (Elliott and Brooks, 1997; Harvey and Bencala, 1993; Thibodeaux and Boyle, 1987). Such flow partially defines the hyporheic zone, which is demarcated by subsurface flowpaths that both begin and end at the stream–aquifer interface. While such vertical flow has received considerable attention, new work has begun to examine flow in the lateral direction. Several recent studies have modeled flow through meanders using stream surface elevation along a meander as a steady boundary condition (Boano et al., 2006; Cardenas, 2008; Revelli et al., 2008). These studies suggest that the drop in stream elevation around a meander creates a hydraulic gradient through the meander, thereby inducing hyporheic flow and creating an intra-meander hyporheic zone. A dominant control on this flow is channel sinuosity, which influences flow residence times (Cardenas, 2009a). Similarly, regional hydraulic gradients can significantly affect the extent of the intra-meander hyporheic zone, with hyporheic flow being limited to the very tip of a meander when streams are strongly gaining or losing (Cardenas, 2009b).

Although the first-order controlling effects of fluvial geomorphology on groundwater–surface water interactions are widely known (Kasahara and Wondzell, 2003; Larkin and Sharp, 1992), and the mechanics of groundwater–surface water exchange and its broader role in river corridors are becoming well understood,
there are still critical questions. In particular, little work has examined the effects of flooding on groundwater–surface water interactions. Previous work has documented the long-term adjustment of hyporheic flows to channel-changing floods and seasonal events (Wondzell and Swanson, 1996, 1999; Wroblicky et al., 1998). Additionally, work on the related topic of bank storage has examined hydraulics and chemistry during floods; however, these studies have generally looked at flow in and out of banks from a 1-dimensional or 2-dimensional framework and have mostly been concerned with the volume and timing of water storage (Chen and Chen, 2003; Cooper and Rorabaugh, 1963; Squillace, 1996; Todd, 1955). This work has made fundamental advances; however, we still know little about how flow and temperature near streams respond to flooding in three dimensions.

Because of this uncertainty, it remains unclear how nutrients are transformed during intra-meander flow, particularly under non-steady conditions. Modeling suggests that intra-meander hyporheic flow may significantly influence water chemistry (Boano et al., 2010), with reaction rates controlled by the amount of reactants delivered to the hyporheic zone and their residence times within it. This finding supports the concepts that both riparian and hyporheic zones contain biochemical “hot spots” where key reactants and organisms overlap and mix to create zones of pronounced reaction rates (Fuller and Harvey, 2000; Gu et al., 2008; Hill, 2000). However, it is unclear how reaction rates in these hot spots vary temporarily because riparian and hyporheic communities, their metabolism, and associated nutrient transformation rates are highly sensitive to hydrologic and biogeochemical perturbations (Findlay et al., 2003; Heffernan et al., 2010; Strauss and Lamberti, 2000). Changing conditions can result in “hot moments” where reaction rates are enhanced for a limited period of time (McClain et al., 2003). Flooding represents a likely driver of hot moments because floods may transport a disproportionate amount of nutrients and organic material compared to other times (Scott et al., 2006) or alter other conditions. Supporting the idea that flooding may trigger a hot moment is a study by Shibata et al. (2004) that demonstrates how a small flood can change the relative strengths of nitrogen sources and sinks in hillslopes and bank hyporheic zones.

Given the lack of previous work, a critical step toward understanding how flooding influences reaction rates in meanders is to first constrain the effects of flooding on groundwater–surface water processes in 3-dimensions. To this end, the objective of this study is to characterize the processes that control intra-meander flow and temperature and demonstrate the extent to which these processes are sensitive to flooding. Given our current lack of understanding concerning this behavior, this study is focused solely on flow and heat transport processes and does not address biogeochemistry explicitly. Our findings have biogeochemical implications; however, future studies should simultaneously consider hydrologic, ecologic and biogeochemical processes.

The exact timing of flood events in natural streams is difficult to predict accurately, which makes studying dynamics during floods challenging. To overcome this, we conducted experiments in an environment where stream discharge can be precisely controlled – the Outdoor Stream Laboratory (OSL).

2. The Outdoor Stream Laboratory and controlled floods

The OSL is a full-scale experimental stream–aquifer system at the Saint Anthony Falls Laboratory at the University of Minnesota. Measuring approximately 40 × 25 m, it is one of the largest facilities of its kind (Fig. 1a). The OSL, partly operated by the National Center for Earth-surface Dynamics, provides a unique environment to study the effects of flooding on a full-size stream–aquifer analog that mimics natural systems. The OSL is constructed on a sloping limestone dam spillway along the Mississippi River. Interlocking plastic sheet piles inside of stone walls contain the facility and help mitigate groundwater leakage. The aquifer substrate is predominantly coarse sand (D25 = 0.38 mm D50 = 0.64 mm D75 = 1.15 mm) excavated from a nearby construction site and has variable thickness between ~1.5 and 2 m. With an average width of ~2 m, the OSL’s stream completes one full meander within the facility and has ripples both upstream and downstream of the meander tip. Discharge is precisely controlled by an inlet valve and monitored with an upstream gauging station. By altering discharge, simulated high discharge events were conducted in both 2008 and 2009. In 2008, flow reached bankfull levels (Fig. 1b), while both bankfull and overbank events (Fig. 1c) were conducted in 2009. Stream discharge during bankfull flow was held constant at a level between 200 and 300 L s⁻¹, which was considerably higher than the baseflow discharge of about 35 L s⁻¹ (Fig. 2). Overbank floods reached about 1050 L s⁻¹.

Although the OSL largely replicates a natural stream meander, the lateral and vertical extent of the aquifer is unnaturally constrained. Because of this, groundwater–surface water interactions in the OSL are not influenced by region-scale hydraulic gradients directly. However, the facility is a plausible analogue for environments where intra-meander flow is strongly driven by the stream. Intuitively, one would expect this to be the case in many natural systems, especially those where the stream surface slope is relatively steep or in meanders that maintain close proximity to the stream (i.e. small or highly sinuous meanders).

3. Methods

3.1. Piezometer network, water level monitoring, and analysis

A nested piezometer network was constructed to monitor the OSL aquifer. A total of 59 nests were installed, each with three piezometers, one screened through the water table for all flow conditions, one screened from about 30 cm through 60 cm below the water table during baseflow, and one screened from about 100 through 130 cm below the water table during baseflow. The piezometers used are composed of polyvinyl chloride (PVC), with an inner diameter of 3.2 cm, an outer diameter of 4.2 cm, and a 0.15 mm screen slot size. Forty-one nests were installed within the point bar...
along hypothesized hyporheic flow paths to give dense coverage, while the remaining piezometers were installed on the opposite bank to provide broader general coverage. Piezometers were installed with a fencepost driver after first hand augering 10 cm diameter holes down to the water table. In some instances, piezometers became stuck during installation, in which case a concrete vibrator was used to induce liquefaction of the sediment below the water table as weight was applied to the piezometer. Following installation, the holes were back-filled and the piezometers were developed by discharging approximately 25 L of water through each. The top of each piezometer casing was surveyed to sub-mm accuracy using a reflector-less digital Sokkia total station. Stream surface elevation, both during baseflow and bankfull flow, was also surveyed.

In 2008 and 2009, steady-state water levels within the piezometers were measured during baseflow conditions using an electronic water level tape. In 2008, measurements were also collected before, during, and after simulated bankfull flow events. Measurements were not taken during overbank flooding because the water table was above the ground surface. To ensure that water level had reached steady state during bankfull conditions, repeat measurements in four piezometers within the meander were taken every 2 min until no changes were observed. We then waited an additional hour before measuring all of the piezometers by hand.

To generate water table maps, these measurements were combined with the surveyed stream elevations along each bank and interpolated using kriging with a linear variogram and no weighting factor. Several piezometers near the tip of the meander were not measured during bankfull flow, which is reflected in the map.

To assess transient changes within the aquifer during floods, a 72-h drip releasing a LiCl stock evenly across the upstream section of the stream was started on 29 September 08. The tracer release resulted in an in-stream Li concentration of 117 µg L⁻¹ when fully mixed with the constant 35.1 L s⁻¹ discharge. Groundwater samples were collected before, during, and after the drip from select medium depth piezometers using a peristaltic pump. The pump was purged with sample well water for 3 or more pump volumes prior to sample collection. Samples were stored in propylene bottles acidified with nitric acid to prevent sorption and flocculation.

Li concentrations of the undiluted samples were determined using a 7500ce Quadrupole inductively coupled plasma mass spectrometer (Agilent). ²⁶⁶Be and ⁴⁵Sc were both tested as internal standards, although the ⁴⁵Sc proved superior. Several Li spikes and duplicate samples were introduced during the analysis for quality assurance. The detection limit for the method used is 0.09 µg L⁻¹ Li, and the average relative standard deviation for the samples (n = 109) was 1.11%.

3.2. Tracer test

A tracer test using lithium was completed in the fall of 2008 during constant baseflow discharge to characterize flow within the aquifer. Several studies have used Li as a tracer successfully to characterize hydrologic characteristics of aquifers and groundwater–surface water interactions (Bencala et al., 1990; Leblanc et al., 1991). Li is a reactive tracer, except at low pH where it does not strongly sorb. At our study site, pH is approximately neutral and Li sorbs to the sediment. We used Li as a tracer rather than a conservative tracer such as Cl because Li can be measured accurately at low concentrations and consequently we could use a more diluted drip that would not harm biota associated with other studies within the OSL. Although the use of a reactive tracer is not ideal, Li is suitable for this study because the analysis of the tracer test for this study is more qualitative than quantitative.

Fig. 2. (a) Hydrograph of bankfull flow event in 2008 and (b) corresponding stream water surface elevation. Typical bankfull events reached maximum discharges between 200 and 300 L s⁻¹, while overbank floods (not shown) reached 1050 L s⁻¹. Events were typically between 8 and 10 h in duration.

Fig. 3. Locations of thermistors (solid circles) and pressure transducers (open squares) deployed in 2009 to monitor water temperature and pressure.
3.3. Temperature monitoring

In 2009, temperature within the aquifer was monitored in 5-min intervals using HOBO U12 data loggers (Onset) with TMCG-HD thermistors (Sensortechnik, accuracy ±0.25 °C, precision 0.05 °C). The thermistors were installed to provide both vertical and lateral temperature distributions within the aquifer. Eleven separate fully screened piezometers were monitored, each with four thermistors installed in a vertical profile at 20, 60, 80 and 90 cm above the bottom of each well (Fig. 3). Data quality was assured by having thermistors log in a room with constant air temperature, which indicated offsets were minimal. Temperature distributions were generated by linear interpolation. For simplicity, results from thermistors nearest 0.3 m and 0.9 m below the baseflow water table were used to provide horizontal temperature distributions at shallow and deep intervals. This was necessitated by the fact that thermistor elevations were referenced to the bottom of each well, but the wells in turn were not at uniform elevation. Upstream temperature was monitored with a CSI-107 thermistor (Campbell), while air temperature was measured with the same Baro Troll data logger used to monitor atmospheric pressure.

4. Results

4.1. Groundwater hydraulics of the experimental facility

The water table and groundwater flow paths in the OSL during baseflow in 2008 and 2009 conditions are starkly similar to those modeled for intra-meander hyporheic zones in losing sinuous rivers (Cardenas, 2009b). In both years, the water table during baseflow suggests intra-meander hyporheic flow is minimal and concentrated to the meander tip (Fig. 4a and c), observations that are consistent with previous field observations and modeling studies (Boano et al., 2006; Peterson and Sickbert, 2006; Revelli et al., 2008). In contrast, significant flow is directed into the aquifer on both the upstream and downstream sides of the meander, suggesting the OSL stream is losing rather than gaining or neutral. This was not part of the original facility design and is likely driven by leaks in the sheet piles confining the OSL. This leakage appears to evolve from 2008 to 2009, which causes the baseflow water tables to change.

The vertical head gradient distributions for both 2008 and 2009 show spatially distinct zones of slight upwelling and downwelling groundwater (Fig. 5). In general, areas with negative head gradients (i.e., downward groundwater flux) occur near the boundaries of the OSL, most notably in the southeast corner in 2009. Gradients within the meander are near zero in most cases, indicating neither dominant downward nor upward flux of groundwater. These observations are consistent with leakage along OSL boundaries, and the fact that strong downward gradient in the southeast corner is present in 2009 but not 2008 confirm that leakage evolved in a way consistent with the observed water table change.

Rough estimates for leakage within the main meander can be calculated using Darcy’s Law:

\[ Q_w = -K \frac{dh}{dA} A \]  

where \( Q_w \) is total discharge (m³ s⁻¹), \( K \) is hydraulic conductivity (m⁻¹), \( \frac{dh}{dA} \) is the hydraulic head gradient, and \( A \) is the area through which flow occurs (m²). If we assume leakage only occurs at the wall, we can estimate leakage as the discharge through the 238.9 m contour within main the meander because none of the flow through it returns to the stream. Along the contour, \( K \approx 1.4 \times 10^{-4} \) m s⁻¹ (Nowinski et al., 2011), \( \frac{dh}{dA} \approx 0.03 \), and \( A \approx 26.5 \) m² (17.7 m × 1.5 m). This results in a loss of about \( 1.1 \times 10^{-4} \) m³ s⁻¹ within the main meander, or about 0.3% of stream discharge under baseflow conditions. For 2009, \( A \) decreases to about 5.3 m² (3.5 m × 1.5 m), but \( \frac{dh}{dA} \) increases to about 0.05, and \( K \) increases to about \( 3.0 \times 10^{-4} \) m s⁻¹ (Nowinski et al., 2011). With these values, leakage decreases slightly to about \( 8.0 \times 10^{-5} \) m³ s⁻¹, or 0.2% of baseflow stream discharge.

4.2. Hydraulic response to flooding

The stream surface slope varies spatially under both baseflow and high discharge (Fig. 2) due to the riffle–pool–riffle sequence in the stream. However, during flooding, the stream surface elevation has a very similar configuration to that during baseflow, but elevated by 4–8 cm (Fig. 2). Breaks in stream slope are located in approximately the same areas in both cases, but migrate upstream slightly during flooding.

Similar to the stream surface, the water table configuration at steady state during flooding is very similar to that during baseflow, but elevated by an average of roughly 7 cm (Fig. 4b). Head gradients within the aquifer are also very similar during both baseflow and high discharge. In fact, quantiles of the kriged gradient magnitude during bankfull discharge are 0.017 for 25%, 0.034 for 50%,
from the upstream side to the downstream side (Fig. 6a). During the start of the drip, while well 31 reached a peak of 8.4 µg L⁻¹ after 12 days.

On the downstream side of the meander, both the piezometers near its tip and base showed pronounced Li peaks that occur before the upstream peaks (wells 35 and 12); however, concentrations near the middle (well 28) increased only slightly. In well 35, Li concentration reached 37.5 µg L⁻¹ 5 days after the start of the drip. Similarly, concentration in well 12 also peaked 5 days after the start, reaching 32.3 µg L⁻¹. In well 28, however, concentrations rose to only 5.9 µg L⁻¹ 23 days after the start.

In contrast to the piezometers near the banks of the stream, Li concentration within the middle of the meander did not show a significant response to the tracer test. Only the piezometer nearest the tip of the meander (well 33) increased above background level, with a peak concentration of 5.9 µg L⁻¹ occurring 16 days following the start of the drip.

4.4. Thermal response to flooding

During the 2 days where overbank and bankfull events were conducted (22–23 July 2009) upstream temperature was fairly constant, with a mean value of 23.1 °C (Fig. 8). This was similar to the mean air temperature (23.3 °C) during the period. Under baseflow conditions, temperature within the shallow portion of the aquifer decreased linearly across the meander from a high of about 22.6 °C near the upstream bank to a low of about 20.0 °C near the downstream bank (Fig. 9a). During bankfull flow, this distribution remained largely unchanged; however, the maximum temperature decreased to about 21.5 °C (Fig. 9b). In contrast, the temperature distribution within the shallow aquifer during overbank flooding changed dramatically, with most areas experiencing a temperature decrease (Fig. 9c). The temperature gradient that was present during baseflow and bankfull flow was also disrupted, and temperature gradients generally became smaller.

Unlike the shallow aquifer, temperature within the deeper sections of the aquifer was nearly constant regardless of stream discharge (Fig. 10). Temperature decreased linearly across the meander from a high of near 20.8 °C on the upstream side to 18.4 °C on the downstream side during baseflow discharge, bankfull flow, and overbank flooding. In addition, the temperature in the aquifer was cooler in all cases in the deeper portions of the aquifer.

5. Discussion

5.1. Groundwater hydraulics and transport

The above results help delineate the extent to which stream stage influences hydrologic processes within stream meanders. Clearly, changing discharge can alter steady state groundwater flow for limited periods of time, and in extreme flooding events, the thermal regime. However, these results are remarkable in the degree to which most hydrologic characteristics within the meander remain unchanged regardless of discharge. Specifically, the water table configuration during baseflow and flooding is essentially the same, only with different absolute elevations. The apparent reason for this consistency is that the slope of the stream surface itself is independent of stream discharge, hence the config-
duration of the dominant boundary condition driving flow in the meander does not change.

For natural meanders where intra-meander groundwater flow is invariant of stream stage at steady state, the hydraulic diffusivity (the ratio of permeability to specific storage) and size of meanders could considerably affect the time needed to reach equilibrium following a change in discharge. In this study, the meander has fairly high permeability and is smaller than meanders in larger rivers. Because of this, steady state was reached within about an hour following dramatic changes in stream discharge. This is a short amount of time relative to the duration of a natural flood, which could persist for days. Hypothetically, however, a larger meander with lower permeability would take significantly longer to equilibrate, and in some cases might not do so at all during the course of a flood. Similarly, the specific storage of aquifers will also influence their response, and aquifers with higher specific storage than the OSL will also take longer to equilibrate. Consequently, the size, permeability, and specific storage of a meander are crucial controls on how significantly discharge influences intra-meander hyporheic flow.

The rapid transport of the tracer indicates not only the spatial extent of the hyporheic zone at the OSL but also its temporal range. The tracer was transported several meters away from the bank into the meander in a few days. This is typically within the scope of most in-stream field tracer tests. Thus, our results are consistent with bank or lateral hyporheic zones causing tailing of solutes released in the stream to generate non-exponential residence time distributions in subsurface transient storage zones (Cardenas, 2009a, 2009b; Revelli et al., 2008).

The tracer results also indicate the strong sensitivity of near-stream groundwater and hyporheic flow to gaining or losing stream conditions. For example, the OSL was designed to be a neutral (neither gaining nor losing) channel. However, it appears that local leaks through the sheet piles that bound the OSL alluvium...
large enough to affect the groundwater flow regime in the meander, an effect we previously verified with a groundwater flow model (Nowinski et al., 2011). The flow distribution we observed is remarkably similar to the model results in Cardenas (2009b), where losing stream conditions were driven by regional groundwater gradients directed away from the stream. Thus, while our artificial system is not influenced by regional gradients, it appears the leaks mimic their effects.

Both our observed water table and several simulated by Cardenas suggest that in losing conditions stream water can infiltrate into meanders along most of their banks, thereby limiting the extent of the intra-meander hyporheic zone. The tracer test helps verify this is the case because Li peaks were observed in both the upstream and downstream sides of the meander, but not in the middle. This is contrary to “normal” intra-meander flow, where peaks would be expected first near the upstream bank, then the middle of the meander, then on the downstream side. In fact, the tracer actually reached wells 12 and 35 on the downstream side of the meander before the upstream side, likely due to head gradients near those wells.

These observations suggest that detailed knowledge of near-stream aquifer conditions is critical to understanding hyporheic flows. While this seems obvious, we note that without a dense piezometer network it is easy to misinterpret the vast area of the point bar as a hyporheic zone. For example, if one were to interpret flow based solely on the steady state 2D water table profiles in Fig. 6, the head gradient would suggest hyporheic flow through the meander is dominant. However, the water table maps in Fig. 4 show that this is not the case, and the majority of the flow is directed inward into the meander and does not return to the stream.

5.2. Thermal dynamics

Under baseflow conditions, the observed stream and aquifer temperatures suggest that intra-meander temperature at our site is predominately controlled by two heat transport processes: advection of stream water into the meander and vertical conduction within it. During baseflow, shallow aquifer temperatures are highest in the upstream region of the meander where flow initiates (Fig. 4c) and are similar to the mean upstream temperature. This suggests that advection of stream water dominates the thermal regime where it first enters the meander. However, temperatures decrease across the meander in approximately the same direction of groundwater flow, which suggests the water is being cooled as it flows within the meander. Because mean air temperature is similar to the upstream temperature, the only plausible source of this cooling is downward conduction of heat resulting from the relatively steep temperature gradient between the shallow and deep groundwater temperatures (\( \sim 0.3 \ C \ m^{-1} \)). In contrast, the observed lateral temperature gradient within the meander appears to persist because it is much lower (\( \sim 0.2 \ C \ m^{-1} \)), and lateral conduction is not significant enough to the laterally redistribute the heat introduced into the meander by advection.

The change in shallow aquifer temperature during overbank flooding demonstrates that floods are capable disrupting the apparent balance between advection and vertical conduction described above. The upstream portion of the shallow aquifer generally became cooler, though temperatures near its base generally increased. This, and the fact that stream temperature was \( 1-2 \ C \) higher than the shallow aquifer temperatures suggest that the altered temperature distribution was influenced not only by the influx of stream water, but also by the redistribution of water within...
it. Specifically, it appears the flooding event may have induced flow of cooler deep groundwater into areas where temperatures decreased. This effect was not observed during bankfull flow, which indicates a relatively large perturbation is required for this redistribution to occur.

5.3. Implications for nutrient cycling

The hydraulic and thermal results of this study have several implications for guiding future field and modeling studies of nutrient cycling within meanders and interpreting their results. Most notably, our results suggest that the spatial distribution of nutrient transformation within a meander will be highly sensitive to whether a stream is gaining, losing, and neutral because reactants will be delivered to different areas of the meander. For instance, we would expect a zone of high reaction rates (hot spot) to be present along the entire perimeter of the OSL meander aquifer owing to the stream losing condition. This is in contrast to model results in Boano et al. (2010), where the stream was neutral and resulted in pronounced denitrification rates along the upstream meander bank.

Unlike gaining or losing conditions, stream stage is unlikely to influence the general location of hot spots because intra-meander meander flow appears to be largely insensitive to changes in stream discharge. Similarly, the observed range of temperatures within our meander was small (~4 °C) and not strongly affected by flooding, which makes flood-induced temperature fluctuations unlikely to influence reaction rates. However, our observations suggest flooding may influence nutrient transformation in meanders in other ways. Specifically, flooding may enhance bulk nutrient transformation by increasing the volume the biochemical hot spots and delivery of reactants to them. This is suggested by our observation that a thicker portion of the meander aquifer receives hyporheic water during a flood since the water table rises and equilibrates with the river. As a result, reaction rates per plan-view area in the aquifer will be larger during flood conditions if microbial activity and other conditions in the expanded hot spot are similar to the original. Similarly, although steady state flow within the meander is largely invariant of stream stage, the steepened hydraulic gradients at the onset of flooding enhances flow to the periphery of the meander where reaction rates are expected to be highest. This could temporarily enhance bulk nutrient transformation.

Finally, it is important to note that the destination of transformed nutrients will also be highly sensitive to gaining and losing stream conditions. Where intra-meander flow is hyporheic and/or a stream is gaining, intra-meander nutrient transformation has the potential to affect stream-water quality. In contrast, intra-meander nutrient transformation in losing streams has the potential to influence groundwater quality, but its affect on stream-water quality will be limited.

6. Summary and conclusions

Hydraulic head and pressure measurements at an experimental stream–aquifer analogue demonstrate how the configuration of intra-meander flow is sensitive to whether a stream is gaining, losing, or neutral. The stream in our experimental stream–aquifer is losing, which reduces the extent of the hyporheic zone and causes stream water to flow into the meander from both the upstream and downstream banks. This finding suggests the location of bio-
chemically reactive hot spots is similarly dependent on whether a stream is gaining or losing. Measurements taken during simulated bankfull flow and overbank floods suggest that steady-state hydraulic gradients within the floodplain are largely independent of stream stage because the driver of flow, stream surface slope, is also independent of stage. However, there are transitional periods at the onset and end of flooding where the configuration changes. At the onset, there is an increased flux of stream water into the aquifer that is focused close to the stream banks. The thickened water table and this increase in flux of water may enhance reaction rates by expanding the biochemical hot spot and reactants that are delivered to it, though future work is required to determine if this is the case. Finally, we found water advected into the meander was cooled in the direction of flow by vertical conduction with cooler deep groundwater. This results in a temperature distribution within the meander that generally mimics the water table during baseflow and bankfull flow. This distribution was altered during overbank flow, most likely by the redistribution of water within the meander. Such temperature changes could also influence reaction rates in some meanders, however, the range of temperatures observed in our meander (−4 °C) suggests the effect would be minimal.

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References