2014

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Heeren, Derek M.; Fox, Garey A.; Fox, A. K.; Storm, Daniel E.; Miller, Ronald B.; and Mittelstet, Aaron R., "Divergence and flow direction as indicators of subsurface heterogeneity and stage-dependent storage in alluvial floodplains" (2014). *Biological Systems Engineering: Papers and Publications*. 363.

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Divergence and flow direction as indicators of subsurface heterogeneity and stage-dependent storage in alluvial floodplains

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Abstract

Assuming homogeneity in alluvial aquifers is convenient, but limits our ability to accurately predict stream-aquifer interactions. Research is needed on (i) identifying the presence of focused, as opposed to diffuse, groundwater discharge/recharge to streams and (ii) the magnitude and role of large-scale bank and transient storage in alluvial floodplains relative to changes in stream stage. The objective of this research was to document and quantify the effect of stage-dependent aquifer heterogeneity and bank storage relative to changes in stream stage using groundwater flow divergence and direction. Monitoring was performed in alluvial floodplains adjacent to the Barren Fork Creek and Honey Creek in northeastern Oklahoma. Based on results from subsurface electrical resistivity mapping, observation wells were installed in high and low electrical resistivity subsoils. Water levels in the wells were recorded real time using pressure transducers (August to October 2009). Divergence was used to quantify heterogeneity (i.e. variation in hydraulic conductivity, porosity, and/or aquifer thickness), and flow direction was used to assess the potential for large-scale (100 m) bank or transient storage. Areas of localized heterogeneity appeared to act as divergence zones allowing stream water to quickly enter the groundwater system, or as flow convergence zones draining a large groundwater area. Maximum divergence or convergence occurred with maximum rates of change in flow rates or stream stage. Flow directions in the groundwater changed considerably between base and high flows, suggesting that the floodplains acted as largescale bank storage zones, rapidly storing and releasing water during passage of a storm hydrograph. During storm events at both sites, the average groundwater direction changed by at least 90° from the average groundwater direction during baseflow. Aquifer heterogeneity in floodplains yields hyporheic flows that are more responsive and spatially and temporally complex than would be expected compared to more common assumptions of homogeneity.

Keywords: alluvial floodplains, direction, divergence, groundwater flow, heterogeneity, hydraulic gradient

Introduction

The importance of considering spatial heterogeneity in predicting flow and transport has been recognized, but quantifying heterogeneity, such as variability in the hydraulic conductivity, porosity, and aquifer thickness, has proven difficult. For example, Cardenas (2009) modeled stream-aquifer interactions within meander bands assuming homogenous hydraulic conductivity in the alluvial aquifer. However, assuming homogeneity limits our ability to accurately predict water and solute fluxes during stream-aquifer interactions in alluvial floodplains. Deposits of coarse alluvium in floodplains result in complex hydrologic pathways (Naiman *et al.*, 2005). Distal floodplain areas can be linked to modern channel flows by paleochannels, i.e. linear deposits of coarse-grained sediments (Stanford and Ward, 1992; Poole *et al.*, 1997, 2002; Amoros and Bornette, 2002; Naiman *et al.*, 2005). In fact, the Committee on Hydrologic Sciences of the National Research Council (NRC, 2004) specifically identi-

fied the need to determine the relative importance of diffuse *versus* focused recharge/discharge in hydrogeologic settings. Additional research is needed to document the occurrence of focused recharge and discharge relative to changes in stream stage.

Research has also documented that highly conductive alluvial systems can be zones of considerable bank storage (Chen and Chen, 2003) and correspondingly transient storage zones for water and nutrients, acting as a sink during high flow and a source during baseflow (Heeren *et al.*, 2011). However, limited data have been presented that documents and quantifies the transient nature of groundwater in alluvial floodplains, other than near-streambed hyporheic flow (Harvey *et al.*, 1996; Worman *et al.*, 2002; Stofleth *et al.*, 2008). Stream-aquifer interaction has been documented on spatial and temporal scales larger than typically associated with hyporheic exchanges (Larkin and Sharp, 1992; Covino *et al.*, 2011). Larkin and Sharp (1992) developed a conceptual model of predominant groundwater flow directions in alluvial valley aquifers.

They suggested three types of regional groundwater flow systems: underflow-component dominated (flow parallel to the river and in the same direction as the streamflow), baseflow-component dominated (flow perpendicular to the river), and mixed flow. Also, limited research has been performed to investigate bank and transient storage relative to shifts in surface and subsurface flow conditions (Zarnetske *et al.*, 2007). Larkin and Sharp (1992) noted that near the river the type of groundwater flow system can be dynamic in time and space in response to changes in river stage.

Specific soil and hydrogeologic conditions can lead to circumstances where subsurface transport through alluvial floodplain subsoils may be important (Turner and Haygarth, 2000; Lacas *et al.*, 2005; Fuchs *et al.*, 2009; Fox *et al.*, 2011; Heeren *et al.*, 2011; Mittelstet *et al.*, 2011). Subsurface nutrient transport is promoted by spatial variability in hydraulic conductivity (Carlyle and Hill, 2001), preferential flow pathways (McCarty and Angier, 2001; Polyakov *et al.*, 2005; Fuchs *et al.*, 2009), and limited sorption capacity in riparian zone soils (Cooper *et al.*, 1995; Carlyle and Hill, 2001; Polyakov *et al.*, 2005). For example, phosphorus transport to streams has been assumed to primarily take place in surface runoff, resulting in a high emphasis on riparian buffer zones as a conservation practice (Reichenberger *et al.*, 2007; Sabbagh *et al.*, 2009; Fox *et al.*, 2010). Their effectiveness may be limited if a transport pathway through the subsurface circumvents the surface trapping objectives of the riparian buffer (Cooper *et al.*, 1995; Lacas *et al.*, 2005). Previous research in alluvial floodplains in northeastern Oklahoma demonstrated that subsurface transport of phosphorus was significant in localized flow paths (Fuchs *et al.*, 2009; Heeren *et al.*, 2010). More information is needed on floodplain-scale recharge/discharge and transient storage processes to understand the potential for subsurface nutrient transport.

The objective of this research was to document and quantify the impact of heterogeneity due to hydraulic conductivity, porosity, and aquifer thickness on groundwater flow patterns relative to changes in stream stage using flow divergence and direction. Divergence/convergence of pore flow velocity has been incorporated into groundwater contaminant transport modeling (Kavvas and Karakas, 1995), and divergence has been observed in subsurface flow models due to heterogeneity and anisotropy (Kumar *et al.*, 2009). However, the divergence (or convergence) of the water table gradient using field data has not been used to characterize heterogeneity in unconfined aquifers. This research strengthens arguments to consider spatial heterogeneity in conductivity, porosity, and aquifer thickness when studying flow interaction between streams and alluvial aquifers and that classification of alluvial aquifer systems must consider dynamic, hydrologic conditions during storm events. This research utilized data sets from two instrumented floodplain sites in the Ozark ecoregion of Oklahoma, but such results are expected to be applicable for gravel bed stream systems worldwide.

Materials and Methods

Barren Fork Creek and Honey Creek floodplain sites

The alluvial floodplain sites were located in the Ozark region of northeastern Oklahoma (Figure 1). Interest in these sites originated from a need to understand transport mechanisms of nutrients in these alluvial floodplains (Fox *et al.*, 2011; Heeren *et al.*, 2011; Mittelstet *et al.*, 2011). The Barren Fork Creek site (Figure 2a, latitude: 35.90°, longitude: -94.85°) was immediately downstream of the U.S. Geological Survey (USGS) gage station 07197000. With a watershed size of 845 km², the Barren Fork Creek was a fourth-order stream with a median daily flow of 3.6 m³ s⁻¹. The Honey Creek site (Figure 2b, latitude: 36.54°, longitude: -94.70°) was also located immediately downstream of a USGS gage station (07189542). As a third-order stream, Honey Creek site had a 0.54 m³ s⁻¹ median daily flow and a 150 km² watershed.

Both floodplain sites consisted of alluvial gravel deposits underlying a mantle of topsoil (Razort gravelly loam). Topsoil thickness ranged from 0.5 to 1.0 m at the Barren Fork Creek site and from 0.1 to 0.5 m at the Honey Creek site. Soil hydraulic studies on these soil types have shown that subtle changes in gravel content can lead to considerable differences in hydraulic conductivity (Sauer and Logsdon, 2002). Located on the outside of a meander bend, the stream is actively eroding away the study area along the Barren Fork Creek (Midgley *et al.*, 2012). Fuchs *et al.* (2009) described some of the soil and hydraulic characteristics of the Barren Fork Creek floodplain site, including estimates of hydraulic conductivity for the gravel subsoil between 140 and 230 m d⁻¹ based on falling head trench tests. The riparian area on Honey Creek was located on the inside of a meander bend, an area likely to be aggradational. There were no *a priori* hypotheses on the nature of divergence/convergence in these two different structure elements (outside *vs* inside of a meander) of a floodplain when selecting these sites.

Observation well installation and long-term monitoring

Based on a positive correlation between electrical resistivity and hydraulic conductivity (Miller, 2012), observation wells (Figure 2) were installed in both high and low electrical resistivity subsoils based on previous electrical resistivity results (Miller, 2012). Low resistivity areas were interpreted as fine sediments (i.e. sand, silt, and clay) possibly containing low percentages of gravel. Higher resistivity areas represented gravel-dominated soils and potential preferential groundwater flow pathways (Heeren *et al.*, 2010). No differences in background electrical conductivity were measured in the stream and groundwater at these sites. Observation wells were also placed around the boundary of the floodplain sites for this and other research projects (Heeren *et al.*, 2011; Mittelstet *et al.*, 2011). Limited wells were able to be placed throughout the middle section of the floodplain sites due to potential interference with landowner operations, especially at the Barren Fork Creek site.

Figure 1. Map showing locations of the two alluvial floodplain sites in eastern Oklahoma.

Figure 2. (a) Barren Fork Creek site, located near Tahlequah, OK, and (b) Honey Creek site, located near Grove, OK, showing observation wells placed in higher electrical resistivity and lower electrical resistivity subsoils. Low resistivity areas were interpreted as fine sediments (i.e. sand, silt, and clay) possibly containing low percentages of gravel. Higher resistivity areas represented gravel-dominated soils and potential preferential groundwater flow pathways. Also, observation wells were installed around the monitoring boundary of the sites. The white arrows indicate stream flow direction.

Using a Geoprobe Systems drilling machine (6200 TMP, Kejr, Inc., Salina, KS), observation wells were installed in the alluvial floodplains to a depth of approximately 3 to 5 m with a 2 to 3 m screened section at the base. Well locations were surveyed using a TOPCON Hip-Lite Plus Real-Time Kinematic global positioning system configured with a base station and rover unit (4 cm accuracy). These data were corrected for positional errors using the National Geodetic Survey Online Positioning User Service. Since the water table elevation data were more sensitive to measurement error than horizontal position, a laser level was used to determine the elevation at the top of each well (1 cm accuracy).

At each site, 24 observation wells (spaced at approximately 5 to 10 m) were instrumented with automated water level loggers (HoboWare, Onset Computer Corp., Cape Cod, MA, water level accuracy of 0.5 cm) to monitor water pressure and temperature at 5-min intervals from August 21, 2009 to October 15, 2009. One logger was placed above the water table at each site to account for changes in atmospheric pressure. Reference water table elevations, obtained with a water level indicator, were then calculated. The logger data were processed with HoboWare Pro software, which accounted for changes in atmospheric pressure as well as changes in water density due to temperature. The local USGS gage stations were used to analyze stream stage.

Divergence as an indicator of subsurface heterogeneity

Water table elevation data were analyzed with Matlab (The Mathworks, Natick, MA). Using 30-min intervals, a cubic interpolation was performed to determine the head for points in a two-dimensional well field grid. While a numerical flow model may be used to physically constrain the head data between wells, this would require

assumptions regarding aquifer heterogeneity before the divergence analysis could be performed. The contour maps were developed based on a 5-m grid to capture the scale of heterogeneity expected in the floodplain based on observations using electrical resistivity imaging (Miller, 2012) and to limit the influence of edge effects. For example, high hydraulic conductivity zones at these sites have been estimated to range from 3 to 20 m wide (Mittelstet *et al.*, 2011). Contour maps were plotted with equipotential lines using 2-cm spacing. The gradient (1 l^{-1}) of the scalar head field was calculated for each point in space using the following equation (Hunt, 1995):

$$\vec{i} = \nabla h = \frac{\partial h}{\partial x} \hat{x} + \frac{\partial h}{\partial y} \hat{y} \quad (1)$$

where $\vec{i}$ is the gradient vector, and h is the head (i.e. water table elevation). Streamlines were calculated as everywhere-tangent to the gradient or perpendicular to contour lines of h . The divergence of the gradient vector field was calculated using the following equation (Hunt, 1995):

$$\text{div}(\vec{i}) = \nabla \cdot \vec{i} = \frac{\partial i_x}{\partial x} + \frac{\partial i_y}{\partial y} \quad (2)$$

where $\text{div}(\vec{i})$ is the divergence of $\vec{i}$. The operator $\text{div}(\vec{i})$ is a scalar quantity with dimensions of l^{-1} and is essentially a measure of the change in gradient per unit length.

The divergence of a fluid velocity vector field, $\text{div}(\vec{v})$, represents mass sources and sinks. With dimensions of T^{-1} , $\text{div}(\vec{v})$ represents the net flux (outflow minus inflow) through the surface of a unit volume. This mathematical divergence, which is a composite of radial flow and changing gradient along a flow path, is similar to, but not exactly the same as, the common use of “flow divergence.” Due to continuity, the divergence of the velocity vector field for an incompressible fluid with no change in storage (e.g. a saturated porous media) must be zero (Hunt, 1995). This also applies to a two-dimensional representation of an unconfined aquifer if the change in water table elevation is slow relative to groundwater flow velocities. It is acknowledged that the alluvial aquifers at the study sites are characterized by three-dimensional heterogeneity (Heeren *et al.*, 2011); therefore, we expect the measured divergence to change when the water table rises and new layers of material become inundated. Velocity can be related to the gradient by:

$$\vec{v} = \frac{K}{\phi} \vec{i} \quad (3)$$

where K is saturated hydraulic conductivity, and ϕ is porosity. Assuming K and ϕ are homogenous, then

$$\text{div}(\vec{v}) = \frac{\partial}{\partial x} \left(\frac{K}{\phi} i_x \right) + \frac{\partial}{\partial y} \left(\frac{K}{\phi} i_y \right) = \frac{K}{\phi} \text{div}(\vec{i}) \quad (4)$$

If $\text{div}(\vec{v})$ is zero, then $\text{div}(\vec{i})$ must also be zero for an incompressible fluid with no change in storage. If $\text{div}(\vec{i})$ is non-zero, then K and/or ϕ (or the aquifer thickness)

must be heterogeneous, violating the assumption in Equation 4, or aquifer depth must be heterogeneous, violating the two-dimensional assumption. Therefore, when the change in water table elevation over time was small (e.g. baseflow conditions), the magnitude of $\text{div}(\vec{i})$ at each point in space was an indicator of the degree of heterogeneity at that point.

Statistics of the spatially distributed $\text{div}(\vec{i})$ were derived using the mean, median, standard deviation, and 5th and 95th percentiles at each time during the monitoring data. These statistics were correlated to the river stage derived from the USGS gages during the monitoring period.

Direction as an indicator of bank storage

The magnitude (R) and direction (θ) of the groundwater flow gradient at each point in the two-dimensional well field grid was derived using the following equations:

$$R = \sqrt{\left(\frac{\partial h}{\partial x} \right)^2 + \left(\frac{\partial h}{\partial y} \right)^2} \quad (5)$$

$$\theta = \tan^{-1} \left(\frac{\partial h / \partial y}{\partial h / \partial x} \right) \quad (6)$$

The arctangent function was in the range of $-\pi/2$ to $\pi/2$ radians. The direction (θ) in radians was then converted to a 0° (east) to (counterclockwise) 360° scale with the appropriate conversion factor for each quadrant. A weighted average direction (θ_w) was calculated for the entire floodplain based on the direction and magnitude (θ_i and R_i respectively) at each time during the monitoring period:

$$\theta_w = \frac{\sum \theta_i R_i}{\sum R_i} \quad (7)$$

Statistics of the R and θ were derived using the mean, weighted mean, median, standard deviation, and 5th and 95th percentiles, and these statistics were correlated to the stream stage during the monitoring period. Polar plots of the R and θ were created for specific flow events to investigate the change in the groundwater flow direction under baseflow and the rising limb, peak, and recession limb of streamflow hydrographs.

Results and Discussion

Water table elevations and contours

Contour patterns in the water table elevation plots at each site remained relatively similar during baseflow conditions, but changed during high flow events. Plots for baseflow conditions and during a large flow event at each site were selected to illustrate the range of contour patterns in the dataset (Figures 3 and 4). The highest gradients in the alluvial aquifer occurred during the rising limb of the hydrographs, when the stream stage was rising most quickly. The impact of aquifer heterogeneity could be seen quali-

Figure 3. Water table elevation contour plots for the Barren Fork Creek site during baseflow (a) on August 25, 2009 (12:00 PM), and also on the rising limb (b) on September 22, 2009 at 5:00 AM, peak (c) on September 22, 2009 at 10:00 AM, and recession limb (d) on September 23, 2009 at 1:30 AM of a streamflow hydrograph. The Barren Fork Creek is located at the top left corner of the plot as shown in Figure 2.

Figure 4. Water table elevation contour plots for the Honey Creek site during baseflow (a) on August 25, 2009 (12:00 PM), and also on the rising limb (b) on October 8, 2009 at 6:30 PM, peak (c) on October 9, 2009 at 7:00 AM, and recession limb (d) on October 9, 2009 at 7:00 PM of a streamflow hydrograph. Honey Creek is located around the bend at the bottom of the plot as shown in Figure 2.

tatively in the contour plots. For example, an area of localized heterogeneity (hydraulic conductivity, porosity, and/or aquifer thickness) can be seen along the Barren Fork

Creek (Figure 3) providing an inlet for stream water to enter the groundwater system. This area of focused recharge appears to be at point (80 m, 60 m), which is the location of

a hypothesized preferential flow pathway that was studied previously (Fuchs *et al.*, 2009; Heeren *et al.*, 2010). With homogeneous aquifer conditions, the highest water table elevation would be at the up-gradient end of the stream, located at the top-center of the contour plot; however, the highest water table elevation was in the zone of heterogeneity, where stream water could most readily enter the alluvial aquifer. At other times, the contour patterns indicated flow convergence zones (Figure 3), where a zone of heterogeneity, point (70 m, 10 m) to point (130 m, 10 m), appeared to be draining a large area of groundwater. At the Honey Creek site, there was a zone of heterogeneity that activated during the rising limb of flood events (Figure 4), creating a convergence zone that drained a large area of groundwater toward the northwest corner of the site. An interesting observation based on the water table elevation data was that the Barren Fork Creek was a losing stream at this field site, even during baseflow and falling limb conditions (Figure 3). This illustrated the complexity of stream-aquifer interactions in these coarse gravel alluvial aquifers. We hypothesize a flow pattern where water regularly left the stream at one point within the study area, traveled through the aquifer, and reentered further downstream outside of the study area. This would be equivalent to a large-scale hyporheic flow path, with its influence on the water table dependent on stream stage.

Divergence as an indicator of subsurface heterogeneity

Divergence was used as an indicator of aquifer heterogeneity (i.e. variation in hydraulic conductivity, porosity, and/or aquifer thickness). Depth to refusal during well installation ranged from 4.0 m to greater than 5.0 m at the Barren Fork Creek site and from 2.5 to 3.5 m at the Honey Creek site. Porosity ranged from 0.2 to 0.5 (Fuchs *et al.*, 2009). In contrast, variability in K is likely to cover orders of magnitude (Miller, 2012). Therefore, it is hypothesized that variation in K was the primary source of measured divergence, although all three likely contributed.

The mean divergence data were plotted over time relative to the stream stage at both the Barren Fork Creek and the Honey Creek sites (Figure 5). During baseflow conditions, divergence provided a direct measure of aquifer heterogeneity since the change in water table elevation over time was small. For example, a divergence of 0.00002 m^{-1} would correlate to approximately 2% change in gradient across the well field with parallel flow lines. At both sites, divergence was positive during baseflow conditions (Figure 5).

Divergence only became negative under rapidly changing water table conditions (Figure 5). During high flow events, the change in water table elevation over time may have become significant, violating the two-dimensional aquifer flow assumption, in which case these divergence data would reflect both aquifer heterogeneity and change in storage (i.e. water table elevation). For example, at the Honey Creek site on the rising limb of the October 9th event, the mean divergence decreased to $-3 \times 10^{-4} \text{ m}^{-1}$ (a large convergence), followed by an increase in divergence

Figure 5. Divergence data for the Barren Fork Creek site (a) and the Honey Creek site (b). The dashed line indicates zero divergence.

to approximately $1 \times 10^{-4} \text{ m}^{-1}$, before returning to baseflow levels of divergence (Figure 5). On the rising limb, measured divergence was negative due to the rapidly changing water table. Since the mean divergence was negative during the rising limb, we were unable to determine whether it was primarily due to aquifer heterogeneity or the rapidly rising water table. As the water table declined, layers of aquifer material became unsaturated and no longer affected flow patterns, leading to increasing divergence.

At the Barren Fork Creek site, the mean divergence was positive during the rising limb of one of the high flow events (September 6, 2009), indicating that the positive divergence due to aquifer heterogeneity was greater than the negative divergence due to water table rise. Mean divergence levels in the groundwater approached $5 \times 10^{-5} \text{ m}^{-1}$ (Figure 5) during the rising limb of this high flow event. The positive divergence data is likely from the area of focused recharge, point (80 m, 60 m), providing an inlet for stream water to enter the groundwater system. This zone of high hydraulic conductivity is consistent with previous electrical resistivity data (Heeren *et al.*, 2010; Miller, 2012).

It appeared that both the convergence zone and the divergence zone at the Barren Fork Creek site could be active at the same time. At baseflow, both were active, but which one had the dominant effect on the mean divergence depended on the water table elevation. At higher stream stage, the convergence zone had a greater impact on groundwater flow, resulting in a negative mean divergence. At baseflow, the divergence zone had a greater impact, resulting in a positive mean divergence (Figure 5).

Direction as an indicator of bank storage

The average groundwater flow direction at each flood-plain site changed considerably between baseflow and

Figure 6. Average direction of the hydraulic gradient at the (a) Barren Fork and (b) Honey Creek field sites from August 21, 2009 to October 15, 2009.

storm events (Figure 6), which concurs with a number of previous studies (Harvey and Bencala, 1993; Wondzell and Swanson, 1996; Malard *et al.*, 1999; Jones *et al.*, 2008; Sawyer *et al.*, 2009; and Francis *et al.*, 2010). At the Barren Fork Creek site, the average weighted gradient direction in the floodplain was approximately south-southwest (240° to 260°) during baseflow conditions, but changed to a south-southeastern direction (280° to 320°) during large flow events and then quickly to the west (140° to 190°) during the recession or falling limb of the storm events (Figure 7). At the Honey Creek site, the average weighted groundwater direction was west-southwest (160° to 190°) (i.e. across the meander bend and directed back towards the stream) during baseflow conditions. During storm events, the average direction changed to northwest (i.e. 130° to 150° or away from the meander bend) at the peak and then south-southwest on the falling or recession limb (Figure 8). Variability in groundwater flow directions increased with stage across both of the two-dimensional well grids during flow events (Figures 7 and 8), suggesting that the flow patterns in these floodplains became more complex during high stream stage.

The changes in average groundwater direction indicated the occurrence of considerable bank storage within the floodplain: groundwater flow direction changed as water moved rapidly into the floodplain during the rising limb of the streamflow hydrograph and then returned to its original average direction as water drained through zones of heterogeneity during the recession of the hydrograph (Figures 6, 7, and 8). Similar to the flow divergence, the change in the direction at both field sites appeared to

be a function of the rate of change in stream flow rate or stage, with higher rates of stream stage change correlating to greater variations in the average groundwater direction compared to the direction under baseflow conditions. In fact, the maximum deviation in average groundwater gradient from the average gradient under baseflow conditions occurred slightly before the peak of the streamflow hydrograph at both sites. The average groundwater gradient after the peak shifted quickly back to the average gradient during baseflow conditions (Figures 7 and 8).

Research implications

As discussed by Packman and Bencala (2000), the surface and subsurface hydrological interactions in alluvial floodplains can be viewed from either the perspective of the surface stream or of the subsurface aquifer. First, utilizing the surface stream viewpoint, the interaction is commonly idealized using a transient storage model that simulates hyporheic storage in an aggregate fashion as a well-mixed but immobile system (Bencala and Walters, 1983). Harvey *et al.* (1996) and others have suggested this idealization captures rapid hyporheic transport (i.e. near streambed exchange) but cannot capture exchange with the more extensive alluvium. The groundwater divergence and direction results in this research further verify these conclusions and more intensely emphasize the importance of considering stream exchanges beyond the near-streambed zone (i.e. larger scale bank or transient storage) relative to changes in stream stage. Changes in discharge and stream stage are known to create bank storage and also suggested to influence near-streambed transient storage (D'Angelo *et al.*, 1993; Harvey and Bencala, 1993; Morrice *et al.*, 1997; Worman *et al.*, 2002; Zarnetske *et al.*, 2007; Stofleth *et al.*, 2008). However, as noted by Zarnetske *et al.* (2007), "... the overall understanding of how they [perturbations in discharge, elevation of channel stage, and water table] correlate to in-channel and hyporheic storage dynamics is still unclear," especially for larger scale interactions with the alluvium.

Aquifer heterogeneity yields hyporheic flows that are more responsive (well connected to the stream) and spatially and temporally complex than would be expected from common assumptions of homogeneity. The presence of large-scale storage at high stream stage may have a direct impact on the transport of in-stream contaminant loads as the stream water interacts with the alluvial groundwater in floodplains throughout the watershed. Future work should be devoted to creating models capable of handling both the near-streambed and larger scale storage to quantify implications of this larger scale exchange on solute and contaminant transport in stream systems during both baseflow and high flow conditions.

Viewing the surface–subsurface interaction from the perspective of the subsurface aquifer, we consider hyporheic exchange as the mixing of stream-derived and aquifer-derived water. This perspective generally relies heavily on the use of numerical groundwater flow models to

Figure 7. Polar plots of the hydraulic gradients during a hydrograph at the Barren Fork Creek (BF) field site at baseflow on August 25, 2009 and during a storm event on September 22–23, 2009. These polar plots are constructed with the magnitude and direction of the groundwater gradient for all points in the floodplain grid. The line indicates the direction of the weighted average groundwater gradient: (a) baseflow at 258°, (b) rising limb at 253°, (c) peak flow at 280°, and (d) recession limb at 179°.

describe reach-scale groundwater flow pathways during bank storage events (Packman and Bencala, 2000; Chen and Chen, 2003; Poole *et al.*, 2006). Particle tracking models may be used to determine the extent of penetration of stream-derived water into the aquifer (Wroblicky *et al.*, 1998). The stream is commonly idealized as a boundary that controls subsurface flow. Of course any modeling effort is dependent on the ability to adequately parameterize input data and specify the appropriate boundary conditions; for example, groundwater heads are typically measured throughout study areas to calibrate the model (Chen and Chen, 2003).

The divergence and direction results from the two floodplain studies reported in this research indicate the necessity of considering horizontal zones of aquifer heterogeneity and anisotropy within alluvial floodplains to adequately simulate larger scale heterogeneity over a range of water table elevations, as noted by Poole *et al.* (2006). Larkin and Sharp (1992) noted no relationship between the alluvial aquifer systems and the aquifer hydraulic characteristics in their studied and modeled re-

gional groundwater flow systems, but it is apparent from this research the impact of heterogeneity on local (near-stream) exchanges during dynamic hydrologic conditions. According to their classification, the BF site completely switches from an underflow- to a baseflow-component dominated system during a single high flow event. If interested in the localized connection between a river and stream system, the impact of stage-dependent heterogeneity must be considered. One potential source of long, continuous horizontal zones of heterogeneity is buried gravel bars (Heeren *et al.*, 2010). Also, groundwater heads measured for flow calibration should be measured during both baseflow and high stream stage events when the stage-dependent heterogeneity activates. Neglecting heterogeneity and their stage-dependent activation may limit the amount of useful information derived from groundwater flow models due to an inappropriate representation of the flow system. An analysis of groundwater flow divergence at a new study site may be useful for justifying a three-dimensional characterization of the aquifer before performing a modeling study.

Figure 8. Polar plots of the hydraulic gradients during a hydrograph at the Honey Creek (HC) field site at baseflow on August 25, 2009 and during a storm event on October 8–9, 2009. These polar plots are constructed with the magnitude and direction of the groundwater gradient for all points in the floodplain grid. The line indicates the direction of the weighted average groundwater gradient: (a) baseflow at 186°, (b) rising limb at 191°, (c) peak flow at 145°, and (d) recession limb at 236°.

Summary and Conclusions

The assumptions of uniform, homogeneous stream/aquifer interaction and only localized, near-streambed hyporheic interactions were not realistic in the studied alluvial floodplains. The activity of stage-dependent zones of heterogeneity depended on the elevation of the water table and the interaction between the stream and the groundwater. It appeared that heterogeneous regions acted as divergence zones, allowing stream water to quickly enter the groundwater system or as flow convergence zones draining a large groundwater area. A method was developed to quantify aquifer heterogeneity using divergence of the water table gradient. During baseflow conditions, non-zero results indicated areas of spatial heterogeneity in hydraulic conductivity, porosity, and/or aquifer thickness consistent with previous research. During high flow events, divergence was a measure of both aquifer heterogeneity and the rapidly changing water table elevation (aquifer storage). At the Barren Fork Creek site, stream water consistently flowed into the alluvial aquifer, even

during baseflow conditions. Bank storage occurred in these alluvial floodplains systems as highlighted by the fact that the average groundwater flow direction at each floodplain site changed considerably between baseflow and storm events. Such storage may have a direct impact on the transport of in-stream contaminant loads as the stream water interacts with the alluvial groundwater along floodplains throughout the watershed. More research needs to be performed in additional alluvial floodplain sites with coarse, gravel material to document the occurrence of such heterogeneity and their impact on contaminant fate and transport.

Acknowledgments — This research is based upon work supported by the Oklahoma Conservation Commission with a U.S. Environmental Protection Agency Region VI 319 grant. The project described in this publication was also supported by Grant/Cooperative Agreement Number G10AP00137 from the United States Geological Survey. Its contents are solely the responsibility of the authors and do not necessarily represent the official views of the USGS. This material was also developed under STAR Fellowship Assistance Agreement no. FP-917333 awarded

by the U.S. Environmental Protection Agency (EPA). It has not been formally reviewed by EPA. The views expressed in this paper are solely those of the authors, and EPA does not endorse any products or commercial services mentioned in this paper. The authors acknowledge Mr. Dan Butler of the Oklahoma Conservation Commission, retired, and Mr. Bill Berry for providing access to the alluvial floodplain properties. The authors also acknowledge Sharla Lovern and Jason Walker, Biosystems and Agricultural Engineering, Oklahoma State University, for assisting with field work.

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