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Heterogeneity influences on stream water–groundwater interactions in a gravel-dominated floodplain

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Abstract Floodplains are composed of complex depositional patterns of ancient and recent stream sediments, and research is needed to address the manner in which coarse floodplain materials affect stream/groundwater exchange patterns. Efforts to understand the heterogeneity of aquifers have utilized numerous techniques typically focused on point-scale measurements; however, in highly heterogeneous settings, the ability to model heterogeneity is dependent on the data density and spatial distribution. The objective of this research was to investigate the correlation between broad-scale methodologies for detecting heterogeneity and the observed spatial variability in stream/groundwater interactions of gravel-dominated alluvial floodplains. More specifically, this study examined the correlation between electrical resistivity (ER) and alluvial groundwater patterns during a flood event at a site on Barren Fork Creek (BFC), in the Ozark ecoregion of Oklahoma, USA, where chert gravels

were common both as streambed and floodplain material. Water table elevations from groundwater monitoring wells for a flood event on May 1-5, 2009 were compared to electrical resistivity maps at various elevations. Areas with high electrical resistivity matched areas with lower water table slope at the same elevation. This research demonstrated that electrical resistivity approaches were capable of indicating heterogeneity in surface water/groundwater interactions, and that these heterogeneities were present even in an aquifer matrix characterized as highly conductive. Portions of gravel-dominated floodplain vadose zones characterized by high hydraulic conductivity features can result in heterogeneous flow patterns when the vadose zone of alluvial floodplains activates during storm events.

Key words alluvial floodplain; surface-ground water interaction; hydraulic conductivity; interpolations; streamflow; water table response

INTRODUCTION

Floodplains are generally planar landscape features that are composed of ancient and current stream sediments, which are typically higher in hydraulic conductivity than the adjacent hillslope sediments and can contain a variety of fluvial features (i.e., abandoned channels) (Woessner, 2000). The presence of a range of sediment textures within the floodplain can produce a complex, heterogeneous alluvial aquifer. The presence of coarse sediments with relatively high hydraulic conductivity can link surface flows to distal floodplain areas beyond the range of anticipated interaction (Sophocleous, 1991; Stanford and Ward, 1993; Poole et al., 1997; Amoros and Bornette, 2002; Naiman et al., 2005). If these structures are limited in extent, for example a linear feature such as a paleochannel, these surface water-groundwater interactions may be termed a "preferential flow path" (PFP). Additionally, a floodplain consisting mainly of high hydraulic conductivity sediments may exhibit high levels of broad-scale interaction between the stream and the groundwater throughout the floodplain; such a system may be termed a "high-flow domain" (Miller et al., 2014). Surface water-groundwater interactions are important as a habitat gradient in which organisms utilize and influence the movement of nutrients and energy (Findlay et al., 1993; Stanford and Ward, 1993; Battin, 1999; Hancock et al., 2005; Boulton et al., 2008). Surface water-groundwater interaction is also an important factor in the movement, fate and transport of environmental contaminants including excess nutrients (Kazezyilmaz-Alhan and Medina, 2006; Landmeyer et al., 2010; Heeren et al., 2011).

Efforts to understand the heterogeneity in alluvial floodplains have utilized numerous techniques (Li et al., 2008; Bohling et al., 2012). However, investigators have acknowledged that, in highly heterogeneous settings, the ability to model heterogeneity is dependent on the data density and spatial distribution (Miller et al., 2014). Investigations designed to investigate heterogeneity based on conventional methods thus need to balance data density on the one hand and the time and expense of data collection on the other (Iqbal et al., 2005). On the other hand, the performance of numeric models of heterogeneous systems can suffer from cell fabrics that are too fine, and research has been focused on methods to generalize heterogeneous systems to reduce model run time without compromising the realistic portrayal of the physical system (Fleckenstein and Fogg, 2008). There is still a need to assess the validity of broad-scale detection methods for heterogeneity and explicitly document their correlation to observed water level response during surface water-groundwater interactions.

Of particular interest in stream and alluvial floodplains systems in the Ozark ecoregion of Oklahoma is nutrient exchange between streams and shallow groundwater (Fox et al., 2011; Heeren et al., 2011; Mittelstet et al., 2011). The Ozark ecoregion is characterized by carbonate rock or sandstone plateaus dissected by steep-sided stream valleys. Erosion of the carbonate bedrock (primarily limestone) by slightly acidic water has left a large residuum of chert gravels in Ozark soils, and produced gravel bed streams with floodplains generally consisting of coarse chert gravels overlain by a mantle (1–300 cm) of gravelly loams or silt loams (Figure 1). Evidence of rapid and preferential flow within gravelly Ozark aquifers was found by Fuchs et al. (2009) at a floodplain site on Barren Fork Creek (BFC) in the Oklahoma Ozarks, who found that conservative tracers and dissolved phosphorus moved in preferential pathways significant distances within the gravel zone under an injection trench dug through the cohesive topsoil. Also, Heeren et al. (2010), working under natural flow conditions independently at the same site, found that phosphorus at concentrations similar to the stream could be found in alluvial groundwater samples taken simultaneously from monitoring wells placed more than 100 m from the stream.

Miller et al. (2014) used electrical resistivity imaging (ERI), along with limited core sampling, and hydraulic testing to survey the subsurface at the BFC site and found that the ER within the gravel subsoil was correlated to the fraction of fine material (<0.25 mm diameter). Higher ER was correlated to lower fractions of fine material. Isolated high flow regions were

identified by ER data within the vadose zone. It was inferred that large portions of the floodplain were characterized by broad-scale high hydraulic conductivity features with potentially significant consequences for the migration of solutes, but this research did not explicitly compare ER to observed water level responses.

It is important to understand the role of high hydraulic conductivity sediments, and how they affect stream/groundwater interactions. Connections between a stream and alluvial groundwater that occur over large distances could have important ramifications for stream water quality. Such connections might be most evident under rising stream stage conditions, when the groundwater gradient is forcing water into coarse sediments normally within the vadose zone. Given the established relationship between percent fines content in gravels and hydraulic conductivity, and the relationship described in Miller et al. (2014) between resistivity and fines content, it was hypothesized that heterogeneity within gravel would affect flood wave propagation through a gravel floodplain. Therefore, the objective of this research was to extrapolate an existing ER survey of the BFC site in order to create an estimate of the distribution of ER throughout the site, and compare ER maps to water table elevations during a storm event to understand how variations in the near-stream subsurface alluvium influence stream/groundwater interactions. The prevailing assumption was that limited heterogeneity can occur within coarse gravel alluvium; this research demonstrated otherwise.

MATERIALS AND METHODS

Study Area

The Ozark ecoregion consists of broad, uplifted plateaus of carbonate sedimentary rocks dissected by deep river canyons occupied by clear, high-gradient streams. Residual chert gravels derived from weathering of the bedrock is common in the streambeds, streambanks, and alluvial floodplains. The site, named after the adjacent Barren Fork creek, consists of gently-sloping pasture adjacent to the creek (35.90° N, -94.85° E), and will be referenced henceforth as the "BFC site". Bedrock within the watershed is primarily chert limestone (Keokuk/Reeds Spring formation), and soils at the BFC site are silts and silts/gravels (Razort and Elsah series) of fluvial origin. The alluvial floodplain itself consists of a mantle of silt and silt-gravel soils (1-3 m, Razort and Elsah soil series) overlying a deep, gravelly subsoil. The 1.2 ha site occupies the south bank of a northward bend in BFC (Figure 2), and is approximately 2.5 km downstream

from a US Geological Survey stream gage (7197000), hereafter termed "USGS gage". The median daily discharge at the USGS gage (61 yr) is 3.6 m^3s^{-1}. The site is located near the south wall of an alluvial valley, trending generally east to west and consisting of limestone valley walls (Figure 2). Caves, sinkholes, springs and other karst features associated with carbonate rocks that have the potential to affect the local hydrology are not known to exist within the study area; however, their presence cannot be ruled out.

Groundwater Monitoring

Observation wells were installed in the alluvial floodplain with a Geoprobe Systems direct-push drilling machine (6200 TMP, Kejr, Inc., Salina, KS). The wells were driven to refusal, a depth of 4 m to greater than 5 m, and included a 2 to 3 m screened section at the base with the remainder solid PVC. Bentonite clay was placed around the top of the well casing to prevent surface runoff from entering the borehole. The wells 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 April 2009 to April 2010, with one logger placed above the water table to account for changes in atmospheric pressure. 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 and produced water table elevation data (1 cm accuracy). Reference water table elevations were obtained with a water level indicator and laser level.

Well locations were surveyed using a TOPCON HiperLite Plus Real-Time Kinematic (RTK) 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 (OPUS). A set of 33 observation wells were originally installed and 23 were instrumented; however, active erosion of the streambank at the site carried away some wells and loggers. The position of the bank was surveyed with GPS on April 18, 2009, and 20 loggers were utilized in the current study for the storm pulse examined in this research. These observation wells were located throughout the floodplain including some wells located as close as 1 m and others as far as 180 m from the stream (Figure 3).

Flood Event

Several flood events of various magnitudes were recorded both at the USGS gage and within the monitoring well network during the overall monitoring period. A single event was chosen for this study that fit the following criteria: relatively isolated from previous or subsequent runoff events, well characterized by both the USGS gage and the well network, and peak stage was below the bank top and within the general elevation of the floodplain gravel layer (as determined by the exposed gravel at the bank face). The event began at 10:00 AM on 5/1/2009, peaked at 6:00 PM on May 3, 2009, and the recession continued until 10:00 PM on 5/5/2009, when a subsequent runoff event was recorded on the hydrograph (Table 1, Figure 4). The event had a peak flow of 204.5 $m^3 s^{-1}$, which corresponded to a return interval of approximately 1.25 yr recorded at the USGS gage (Table 1). The USGS gage hydrograph was adjusted for the travel time to the site by matching the flood peak recorded at the gage to the flood peak from the logger located closest to the bank face (0.5 m).

Electrical Resistivity Imaging

ER surveys were conducted at the floodplain sites for the purpose of characterizing the heterogeneity of the unconsolidated floodplain sediments, especially in the vadose zone between the ground surface and the alluvial water table. ERI has been utilized to determine preferential flow in complex vadose zone settings (Urish, 1981; Huntley, 1986; Webb et al., 2008). The ER data were collected using a SuperSting R8/IP Earth Resistivity Meter (Advanced GeoSciences Inc., Austin, TX) with a 56-electrode array. The system utilized stainless steel stakes and cables with seismic takeout electrodes to connect to the stakes.

Fourteen lines were collected at the BFC site between June 2008 and March 2009, each of which produced a two-dimensional vertical profile of the aquifer along the length of the electrode array. The various surveys employed electrode spacing of 0.5, 1.0, 1.5, 2.0 and 2.5 m with associated profile lengths of 28, 56, 84, 112, and 140 m, and depths of investigation of 7.5, 15.0, 17.0, 22.5 and 25.0 m, respectively. The ER survey electrode locations were spatially georeferenced with the same GPS equipment and methods utilized for the monitoring wells.

The ER sampling with the SuperSting R8/IP, and subsequent inversion utilized a proprietary routine (HF, Halihan and Fenstemaker, 2004; Halihan et al., 2005) that was developed to increase the definition of electrical insulators in the subsurface. The ER data were filtered for data with a repetition error of greater than 2%. Additional data were trimmed from

the data during the inversion process. The HF method has been utilized successfully within a number of applications published in the peer-reviewed literature (Halihan et al., 2005; Ong et al., 2010; Heeren et al., 2010; Heeren et al., 2011; Keppel et al., 2011; Halihan et al., 2011; Halihan et al., 2012; Christenson et al., 2012; Halihan et al., 2013; Heeren et al., 2014). Miller et al. (2014) also demonstrated that the HF method was more suitable for the gravel-dominated alluvial floodplains investigated in this study by comparing the technique to the standard dipole-dipole ER array and inversion. The HF method was found to have sensitivity which was greater than dipole-dipole by a factor that ranged from 6 to an order of magnitude in the depth ranges utilized in this their study. Therefore the HF method was selected for imaging the subsurface.

Water Table and ER Interpolations

Interpolation techniques were used to create site-wide estimates of both water table elevation and ER from the point values produced by the recording instruments. Interpolations of water table elevations were made using the "Minimum Curvature" method in Surfer 8 (Golden Software, Inc., Golden, CO), which approximates the surface by iteratively fitting a continuous curve (often visualized as a flexible sheet) under tension from the edges. This method produces a smooth surface that has been used in many earth science applications (Smith and Wessel, 1990). For this study, the minimum tension was utilized, which allowed the surface to best conform to the data. Interpolating requires iterating using different procedures to determine which approach best fits the observed water levels and also provides the most realistic water surface. The minimum tension provided the most realistic water surface profile for this study area. Interpolation schemes were available that more closely matched the water levels in the observation wells but did not produce a realistic water surface profile.

In addition, estimates of the general directional trend of the water table surface were derived by fitting a first-order polynomial to the water table elevations using the "Polynomial Regression" method in Surfer to create a best-fitting "plane", and then calculating the aspect (direction) of that planar surface (in degrees clockwise from north). These planar estimates were used to compare the changes in water table direction (i.e., general direction from which water was entering or leaving the studied floodplain) as the stream stage changed over time.

Interpolations of ER data were created using the conditional simulation interpolation method, which preserved the distribution of sampled characteristics across the site, using the geostatistical program GS+ (Gamma Design Software, LLC, Plainwell MI). The ER data (ρ, Ω-

m) were log-transformed to better approximate a normal distribution, then anisotropic variograms were prepared and used in the conditional simulation. The resulting grid of interpolated values was back-transformed to the original ER units. The final grid of ER values was imported into Surfer 8 for display.

Spatial Analysis

The factors controlling the movement of alluvial groundwater include (1) the local geometry and position of the stream channel within the floodplain, (2) the relation of stream stage to the water table, and (3) the distribution of hydraulic conductivities within the floodplain (Woessner, 2000). The first two factors were addressed with watershed maps, BFC hydrographs, and the water table directions derived from planar interpolations. The third factor was addressed by comparing water table interpolations with electrical resistivity maps. To address the question of whether the spatial variation in ER can indicate heterogeneous flow conditions within groundwater at the site, water table maps were overlaid on ER maps.

RESULTS AND DISCUSSION

Electrical Resistivity Interpolation

The ER data were positively skewed, indicating a large number of low values and a small number of extreme high values, and ER values ranged over several orders of magnitude (27-2,652 Ω-m) (Table 2). The contact resistance for most of the ER datasets was good, but the gravel surfaces had higher contact resistances of approximately 1500-2500 Ω. The inversion quality was good with data trimming of 4-9% and inversion root mean square (RMS) errors of 3-9% with most RMS errors between 5% and 8%.

ER profiles show complex ER patterns within the soil profile. ER values of 100-300 Ω-m close to the surface of the profiles were interpreted as the surface Razort or Elsah soil layer. ER values below the surface layer did not show horizontal layering, but rather horizontal discontinuous zones of high and low ER (Figure 5). The pattern contained within the ER profiles was consistent with a transect passing through buried fluvial features that may include gravel bars and filled paleochannels. For instance, the high ER values on the BFC profiles (~2000 Ω-m) were similar to those observed at the surface of an ER profile from a gravel bar located near the BFC floodplain site (2000-5000 Ω-m), supporting the idea that high ER features within the floodplain may be clean gravels without fines (Miller et al., 2014). High ρ values

deeper (≥ 10 m and in the saturated zone) in the profiles were assumed to be produced by carbonate bedrock, although no cores were obtained from those depths to confirm that interpretation. No significant decrease in ER was observed below the water table.

Water Table Interpolations

The main down-valley trend in the vicinity of the site was difficult to determine, as the site was located in a bendway of the valley, but the range of down-valley directions for the straight section of the valley upstream of the site was approximately 215 to 225° (clockwise from north, Figure 2). Therefore it is likely that groundwater flowing with a down-valley trend will exhibit a general directional trend that is close to or greater than this range. The BFC approached the site at the northeast corner of the site at an angle of about 190°, and then passed along the northwest border of the site at an angle of about 228°, which is close to the down-valley trend direction. A water table general directional trend close to 190° would imply that the flows were influenced by the stream as it approached the site. Because the angle between the bank and the site is 228°, only directions greater than this are likely to produce "gaining stream" conditions in which water from the alluvial aquifer reenters the stream. A general directional angle of 228° would indicate water flowing along the bank, and less than 228° indicates water moving from the stream into the aquifer.

During the runoff event the stream stage increased 2.2 m (211.9 to 214.1 m, Figure 4), and the water table general direction ranged over a 104° span (138° to 242°, Figure 4). It is interesting to note in Figure 4 that a specific direction does not occur at a specific stream stage, but rather as a function of the change in stream stage. The initial period shown on the event hydrograph was a period of recession after the previous event peak 12 days earlier, in which the stream stage was decreasing slowly (Figure 4). The peak water table elevation, and the time of delay after the first recorded peak for each monitoring well were plotted against the well distance from the stream (Figure 6). The peak elevation in the wells farthest from the stream (180 m) was only about 0.25 m reduced from the peak recorded at the wells within 1 m of the stream; a condition that indicates that little energy was lost to the intervening aquifer material. The peak traveled rapidly as well, with the peak in the wells most distant from the stream occurring about 1.5 hr after the peak near the stream. Rapid water movement with little energy loss as water moved through the aquifer was a condition expected of a "high-flow" aquifer domain. The

general flow direction decreased (i.e., shifted counter-clockwise) as the stream stage increased, and increased (i.e., shifted clockwise) as the stream stage fell over the three peaks of the event. The minimum direction (138°) occurred at the event peak, and the maximum direction (242°) occurred at 12:30 AM May 4, 2009, during the initial hours of the hydrograph recession following the event peak.

The minimum direction (138°) corresponded to the high stage of the event peak and approximated the direction normal to the angle of the bank along the northwest site boundary, which indicated that water was being forced into the aquifer from the stream during the event peak (Figures 3 and 4). The maximum direction (242°) occurred during the initial recession limb of the stream hydrograph and when the water table was very high. Water table directions that were greater than the angle of the bank at the site (228°) indicated that water was flowing from the aquifer to the stream. This "gaining stream" condition persisted for 13 hours, between 10:30 PM 5/3/2009 and 12:00 PM 5/4/2009 (Figure 4), before shifting to a direction (~220°) that may be considered to constitute a down-valley trend. The alluvial aquifer thus appeared to respond primarily to local stream conditions at very high stream stages, when water flowed directly into the aquifer from the stream or from the aquifer into the stream. Down-valley flow prevailed below a threshold. The water table elevation was relatively high relative to the stream stage in this latter condition (Figure 4).

Spatial Analysis

An important element for understanding the hydraulic behavior of the alluvial aquifer was how the spatial variation of hydraulic conductivity affected the shape of the water table and thus the exchange of stream and groundwater. In this research, we utilized ER as a surrogate measure for hydraulic conductivity since previous research (Miller et al., 2014) reported that that the aquifer consisted of gravels with varying proportions of fine material, and that high ER within the floodplain corresponded to a lower proportion of fines.

Comparisons between the configuration of the water table and the ER were made by (1) selecting a point in time from the event hydrograph, (2) interpolating a planar "slice" of the aquifer by selecting ER data at various elevations within the shallow aquifer, (3) interpolating the water table from the well logger records at that time, and (4) overlaying the two interpolated maps. Several stream stages were chosen from the runoff hydrograph as marker elevations for

comparison, including baseflow, rising limb, falling limb, and a transition between falling and rising limbs (Table 1, Figure 4). Several of the elevations (212.5 and 213.6 m) included comparisons with two water table interpolations in order to assess whether the distribution of ER affected those water table directions differently.

The means and medians, as reported in Table 2, for the distributions of interpolated ER for each elevation "slice" increased with decreasing elevation, indicating the presence of gravels containing relatively few fines deeper within the floodplain's vadose zone. The distributions were also positively skewed and kurtotic, indicating that the aquifer was dominated by the lower ER values of approximately 30 to 250 Ω-m (Table 2). The vadose zone of the aquifer was hydraulically conductive and likely to respond quickly to stage changes in the stream. The lowest aggregate ER existed in the highest slice (213.6 m) close to the floodplain surface, a zone known from experience at the site to consist primarily of silt-dominated floodplain soils (Razort and Elsah) rather than gravels.

The spatial variation in ER was not distributed randomly at the different elevations, but rather was clustered which created distinct regions of high and low ER within the aquifer (Figure 7). The primary feature was a zone of high ER that appeared along the northwest edge of the site parallel to the stream in the lower three elevation interpolations (211.9, 212.2, and 212.5 m; Figure 7a, Figure 7b, and Figure 7c and 7d, respectively). This was hypothesized to be a large, gravel bar deposit with relatively, clean gravel deposited when the stream was positioned near that location. It should be noted again that these streams are geomorphically very active with streambank retreat rates of 30 m over some annual periods (Midgley et al., 2012). The sediments from this high ER feature were coarse, and the similarity of the particle size distribution of samples from this location to samples taken from the surface of a nearby gravel bar supported the hypothesis that the feature was a buried gravel bar (Miller et al., 2014). There was also a high ER feature that appeared close to the center of the site that was most obvious at 212.2 m (Figure 7b), but can be seen at the lowest elevation (211.9 m, Figure 7a) and also at the higher elevation (212.5 m, Figures 7c and 7d). Between the two high ER features was a zone of low ER, shown in light green and dark green tones, that was most obvious in Figures 7b, 7c and 7d. A possible interpretation of this feature was as paleochannel/depression between the two gravel bars which was filled after abandonment, primarily with finer sediments mixed with gravels.

In general, the water table maps associated with rising-limb conditions (Figures 7b, 7c, and 7e) had steeper gradients, as represented by more closely spaced contours, than the baseflow water table (Figure 7a), and the transitional and falling limb water tables (Figure 7d and 7f, respectively). As noted earlier, the primary driver for the configuration of the water table appeared to be the change in the stream stage, and the water table maps showed that type of change in direction. However some of the features within the contours appeared to be related to the differences in ER; principally, the high ER zones bordering the creek and in the center of the site corresponded to areas of low water table slope, recorded as wider spaces between the contours. Similarly, the low ER zone between the two high zones corresponded to a steeper slope characterized by more closely-spaced contours (Figure 7a-d). At stream stages near the event peak, the water table is at an elevation within the aquifer that lacks the zones of very high ER seen in the lower elevations (Figures 7e and 7f). The rising limb at this elevation showed consistent slope, implying that the variation in ER and potentially hydraulic conductivity was insufficient to affect the movement of water (Figure 7e); however, the falling limb showed that the higher ER along the northern and southeastern edges affected the water table in a characteristic way by reducing the slope (Figure 7f).

SUMMARY AND CONCLUSIONS

This research demonstrated that electrical resistivity approaches were capable of indicating heterogeneity in surface water/groundwater interactions, and that these heterogeneities were present even in an aquifer matrix characterized as highly conductive. The study compared area-wide heterogeneity in electrical resistivity, a more broad-level and rapid measurement indicator of potential hydraulic conductivity, to the changing water table as alluvial groundwater responded to a flood event. A storm runoff pulse passed the site during May 1-5, 2009 featuring 2.2 m of stream stage increase, which caused the water table to rise into the gravel-dominated vadose zone at the site. Water table maps, corresponding to the times when stream elevation matched the selected electrical resistivity profiles at specific elevations, were prepared from groundwater elevation data obtained from pressure transducers placed in observation wells at the site. When the flood peaks from each monitoring well were compared, it appeared that there was little attenuation of the energy of the storm pulse even at the furthest point in the site; at 180 m the flood peak had only dropped 0.25 m and was delayed by 1.5 hours. This lends credence to

the idea that the floodplain was a “high-flow” domain. Further evidence for a “high-flow” domain was provided by the direction of the water table as estimated by a fitted plane. The direction of the water table shifted over time as the storm pulse travelled through the floodplain, beginning in a direction similar to the stream as it approached the site, changing until it was normal to the stream during the rising limb, and shifting to a direction that was nearly parallel to the valley axis during the recession limb. This indicates that the water table responds on a time scale of hours, similar to that of the storm pulse itself, and that the groundwater responds primarily to regional factors. Despite the floodplain as a whole constituting a “high-flow” domain, areas of heterogeneity existed within the floodplain, characterized by high electrical resistivity. These results have important implications both for the movement of environmentally sensitive solutes such as excess nutrients (i.e., phosphorus in the Ozark ecoregion) between streams and groundwater in alluvial floodplains, and for the use of rapid, non-invasive methods such as ERI for detecting heterogeneity in floodplains.

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REFERENCES

Amoros, C. & Bornette, G. (2002) Connectivity and biocomplexity in waterbodies of riverine floodplains. *Freshwater Biol.* **47**, 761-776.

Battin, T. J. (1999) Hydrologic flow paths control dissolved organic carbon fluxes and metabolism in an alpine stream hyporheic zone. *Water Resources Res.* **35**(10), 3159-3169.

Bohling, G. C., Liu, G., Knobbe, S. K., Reboulet, E. C., Hynde, D. W., Dietrich, P., Butler, Jr, J. J. (2012). Geostaatistical analysis of centimeter-scale conductivity variations at the MADE site. *Water Resour. Res.* **48**, W02525, doi: 10.1029/2011WR010791.

Boulton, A. J., Fenwick, G. D., Hancock, P. J. & Harvey, M. S. (2008) Biodiversity, functional roles, and ecosystem services of groundwater invertebrates. *Invertebrate Systematics* **22**, 103-116.

Christenson, M. D., Kambhu, A. & Comfort, S. D. (2012) Using slow-release permanganate candles to remove TCE from a low permeable aquifer at a former landfill. *Chemosphere* **89**(6), 680-687, doi: 10.1016/j.chemosphere.2012.06.009.

Findlay, S., Strayer, S., Goumbala, C. & Gould, K. (1993) Metabolism of steamwater dissolved organic carbon in the shallow hyporheic zone. *Limnol. Oceanogr.* **38**(7), 1793-1499.

Fleckenstein, J. H., Fogg, G. E. (2008). Efficient upcscaling of hydraulic conductivity in heterogeneous alluvial aquifers. *Hydrogeol. J.* **16**, 1239-1250.

Fox, G. A., Heeren, D. M., Miller, R. B., Mittelstet, A. R. & Storm, D. E. (2011) Flow and transport experiments for a streambank seep originating from a preferential flow pathway. *J. Hydrol.* **403**(3-4), 360-366.

Fox, G. A., Miller, R. B., Heeren, D. M., Storm, D. E. & Halihan, T. (2012). Reply to comment on "Design and application of a direct-push vadose zone gravel permeameter". Ground Water **50**(4), 512-513.

Fuchs, J. W., Fox, G. A., Storm, D. E., Penn, C. J. & Brown, G. O. (2009) Subsurface transport of phosphorus in riparian floodplains: Influence of preferential flow paths. *J. Environ. Qual.* **38**(2), 473–484.

Halihan, T. & Fenstemaker, T. (2004) Proprietary electrical resistivity imaging method. 2.0 Ed. Oklahoma State University Office of Intellectual Property, Stillwater, OK.

Halihan, T., Love, A., Keppel, M. & Berens, V. (2013) Analysis of subsurface mound spring connectivity in shale of the western margin of the Great Artesian Basin, South Australia. *Hydrogeology J.* **21**(7), doi: 10.1007/s10040-013-1034-8.

Halihan, T., Albano, J., Comfort, S. D. & Zlotnik, V. A. (2011) Electrical resistivity imaging of a permanganate injection during in situ treatment of RDX-contaminated ground water. *Ground Water Monit. R.* **32**(1), 43-52, doi: 10.1111/j1745-6592.2011.01361.x.

Halihan, T., McDonald, S., Patey, P. & Stonecipher, M. (2012) ERI evaluation of injectates used at a dry cleaning site. *Remediation* **22**(3), 79-91, doi 10.1002/rem.21322.

Halihan, T., Paxton, S., Graham, I., Fenstemaker, T. & Riley M. (2005) Post-remediation evaluation of a LNAPL site using electrical resistivity imaging. *J. Environ. Modeling* **7**, 283-287.

Hancock, P. J., Boulton A. J. & Humphreys, W. F. (2005) Aquifers and hyporheic zones: Towards an ecological understanding of groundwater. *Hydrogeology J.* **13**, 98-111.

Heeren, D. M., Fox, G. A., Fox, A. K., Storm, D. E., Miller, R. B. & Mittelstet, A. R. (2014) Divergence and flow direction as indicators of subsurface heterogeneity and stage-dependent storage in alluvial floodplains. *Hydrol. Proc.* **28**(3), 1307-1317, doi: 10.1002/hyp.9674.

Heeren, D. M., Fox, G. A., Miller, R. B., Storm, D. E., Fox, A. K., Penn, C. J., Halihan, T. & Mittelstet, A. R. (2011) Stage-dependent transient storage of phosphorus in alluvial floodplains. *Hydrol. Proc.* **25**(20), 3230-3243, doi: 10.1002/hyp.8054.

Heeren, D. M., Miller, R. B., Fox, G. A., Storm, D. E., Penn, C. J. & Halihan, T. (2010) Preferential flow path effects on subsurface contaminant transport in alluvial floodplains. *Trans. ASABE* **53**(1), 127–136.

Huntley, D. (1986) Relations between permeability and electrical resistivity in granular aquifers. *Ground Water* **24**(4), 466-474.

Iqbal J., Thompson, J. A., Jenkins, J. N., Owens, P. R., Whisler, F. D. (2005). Spatial variability analysis of soil physical properties of alluvial soils. *Soil Sci. Soc. Am. J.* **69**, 1338-1350.

Kazezyilmaz-Alhan, C. M. & Medina, M. A. (2006) Stream solute transport incorporating hyporheic zone processes. *J. Hydrol.* **329**, 26-38.

Keppel, M. N., Clarke, J. D. A., Halihan, T., Love, A. J. & Werner, A. D. (2011) Mound springs in the arid Lake Eyre South Region of South Australia: A new depositional tufa model and its controls. *Sediment. Geol.*, doi: 10.1016/j.sedgeo.2011.08.001.

Landmeyer, J. E., Bradley, P. M., Trego, D. A., Hale K. G. & Haas, J. E. (2010) MTBE, TBA, and TAME attenuation in diverse hyporheic zones. *Ground Water* **48**(1), 30-41.

Li, W., Englert, A., Cirpka, O.A., Vereecken, H. (2008). Three-dimensional geostatistical inversion of flowmeter and pumping test data. *Ground Water* **46**(2), 193-201.

Midgley, T., Fox, G. A., Heeren, D. M. (2012). Evaluation of the Bank Stability and Toe Erosion Model (BSTEM) for predicting lateral streambank retreat on composite streambanks. *Geomorphology* **145-146**, 107-114.

Miller, R. B., Heeren, D. M., Fox, G. A., Halihan, T., Storm, D. E. & Mittelstet, A. R. (2014) The hydraulic conductivity structure of gravel-dominated vadose zones within alluvial floodplains. *Journal of Hydrology* (In Press), doi: 10.1016/j.jhydrol.2014.03.046.

Miller, R. B., Heeren, D. M., Fox, G. A., Storm, D. E. & Halihan, T. (2011) Design and application of a direct-push in-situ gravel permeameter. *Ground Water* **49**(6), 920-925, doi: 10.1111/j.1745-6584.2010.00796.x.

Mittelstet, A. R., Heeren, D. M., Fox, G. A., Storm, D. E., White, M. J. & Miller, R. B. (2011) Comparison of subsurface and surface runoff phosphorus transport rates in alluvial floodplains. *Ag. Ecosyst. Environ.* **141**, 417-425.

Naiman, R., Descamps, H. & McClain, M. (2005) Catchments and the physical template. In: *Riparia: Ecology, Conservation, and Management of Streamside Communities* (ed. by R. J. Naiman). Elsevier Academic Press, Boston, MA.

Ong, J. B., Lane, Jr., J. W., Zlotnik, V. A., Halihan, T. & White, E. A. (2010) Combined use of frequency-domain electromagnetic and electrical resistivity surveys to delineate near-lake groundwater flow in the semi-arid Nebraska Sand Hills, USA. *Hydrogeology J.* **18**, 1539-1545.

Poole, G., Naiman, R., Pastor, J. & Stanford, J. (1997) Uses and limitations of ground penetrating RADAR in two riparian systems. In: *Groundwater/Surface Water Ecotones: Biological and Hydrological Interaction and Management Options* (ed. by J. Gibert, J. Mathieu & F. Fournier). Cambridge University Press, Cambridge.

Smith, W. H. F. & Wessel, P. (1990) Gridding with continuous curvature splines in tension. *Geophysics* **55**(3), 293-305.

Sophocleous, M. A. (1991) Stream-floodwave propagation through the Great Bend alluvial aquifer, Kansas: Field measurements and numerical simulations. *J. Hydrol.* **124**, 207-228.

Stanford, J. A. & Ward, J. V. (1993) An ecosystem perspective of alluvial rivers: Connectivity and the hyporheic corridor. *J. North Am. Benthol. Soc.* **12**(1), 48-60.

Urish, D. W. (1981) Electrical resistivity-hydraulic conductivity relationships in glacial outwash aquifers. *Water Resour. Res.* **17**(5), 1401-1408, doi: 10.1029/WR017i005p01401.

U.S. Bureau of Reclamation (USBR) (1985) *Ground Water Manual: A Water Resources Technical Manual* Revised Reprint. Denver, CO. U.S. Department of the Interior, Bureau of Reclamation.

Webb, G., Tyler, S. W., Collord, J., van Zyl, D., Halihan, T., Turrentine, J. & Fenstemaker, T. (2008) Field scale analysis of flow mechanisms in highly heterogeneous mining media. *Vadose Zone J.* **7**(2), 899-908. doi:10.2136/vzj2007.0130.

Woessner, W. W. (2000) Stream and fluvial plain ground water interactions: Rescaling hydrogeologic thought. *Ground Water* **38**(3), 423-429.

Figure Captions

Figure 1. Location of Barren Fork Creek floodplain site and watershed within the Ozark ecoregion of Oklahoma (top) and the streambank at the BFC site showing unconsolidated gravels underlying silt loam topsoils (bottom).

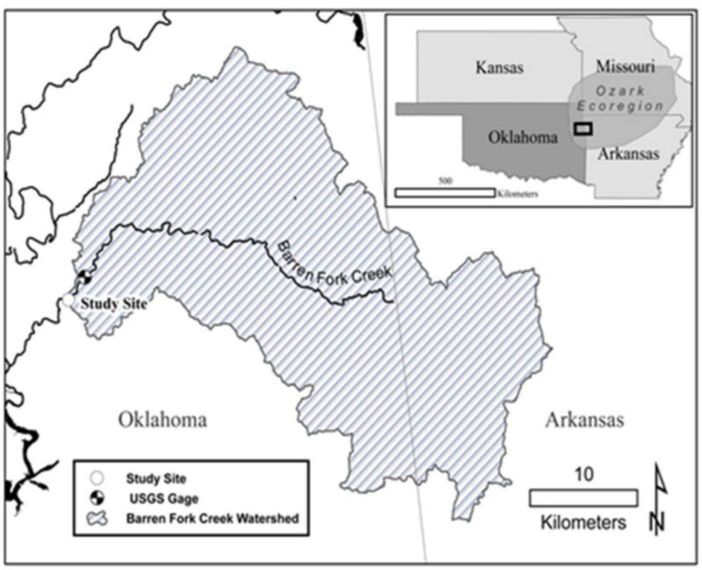


Figure 2. Map of the BFC floodplain in the vicinity of the site. The terrain includes a flat floodplain surface bounded by steep limestone bluffs. BFC flows northeast to southwest. The angle of the bank at the site relative to the stream is 228^{o}. Colored arrows indicate the general direction (degrees, clockwise from north) of the valley axis (blue arrow, 215-225 ^{o}) of BFC where it enters the site (yellow arrow, 190 ^{o}) and the direction normal to the bank where it is adjacent to the site (green arrow, 138 ^{o}).

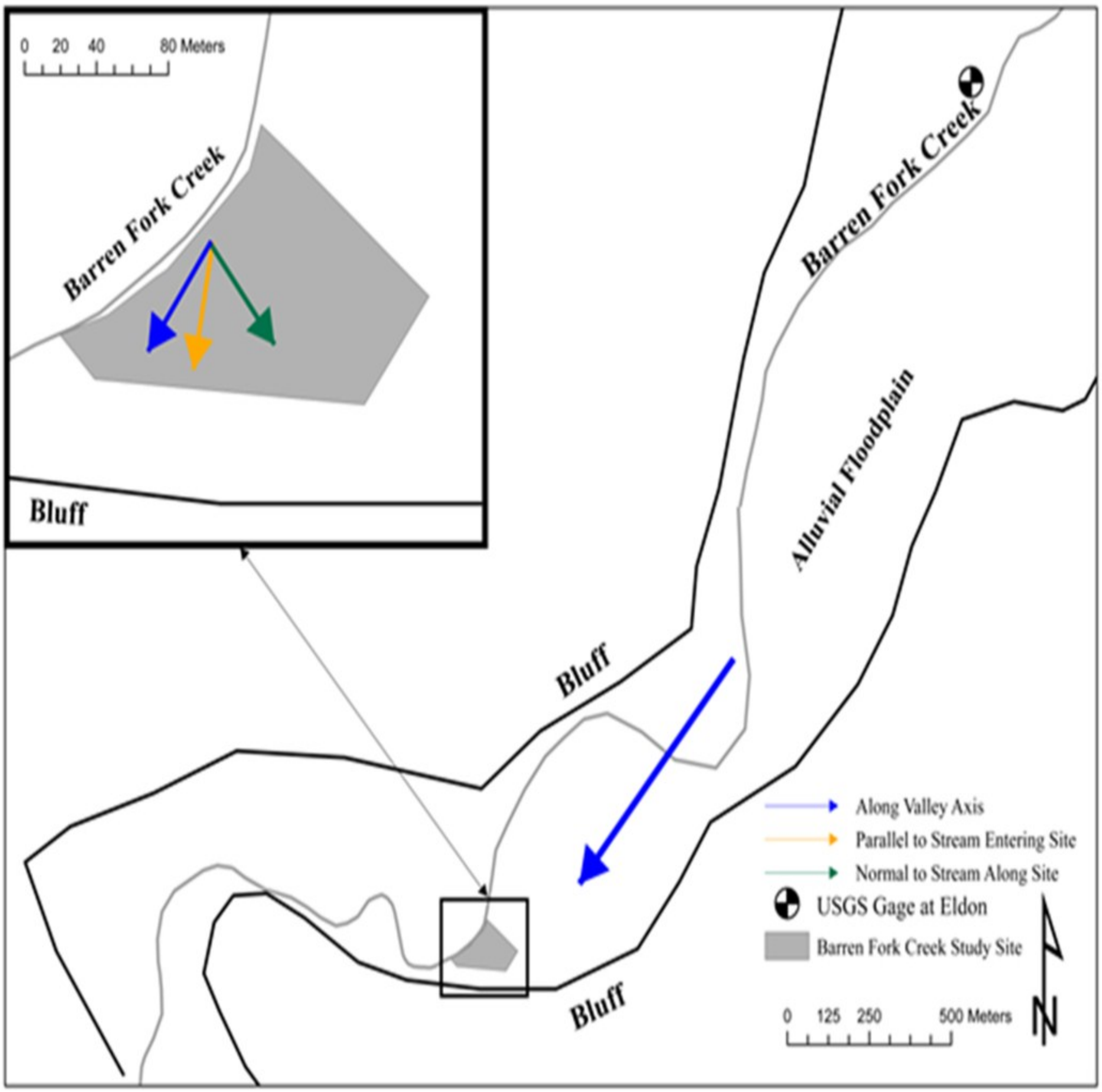
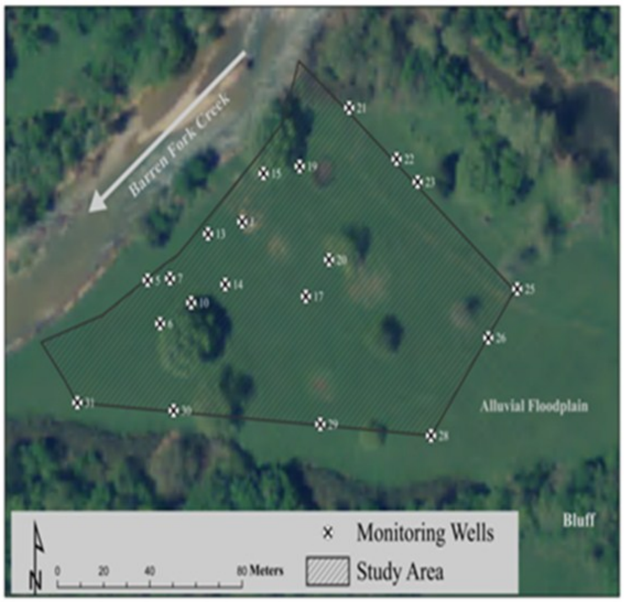


Figure 3. Locations of monitoring wells (top) and electrical resistivity lines (bottom) within Barren Fork Creek floodplain site. Image was acquired as part of the US Department of Agriculture National Agricultural Imagery Program (NAIP) in 2008. Streambank erosion that occurred subsequent to the studied flood event is evident in the difference between the stream bank in the image and that mapped for the study area (May, 2009). BFC1, BFC2, and BFC3 refer to hydraulic conductivity tests conducted by Miller et al. (2014) and confirmations of electrical resistivity to coarse, gravel substrates.



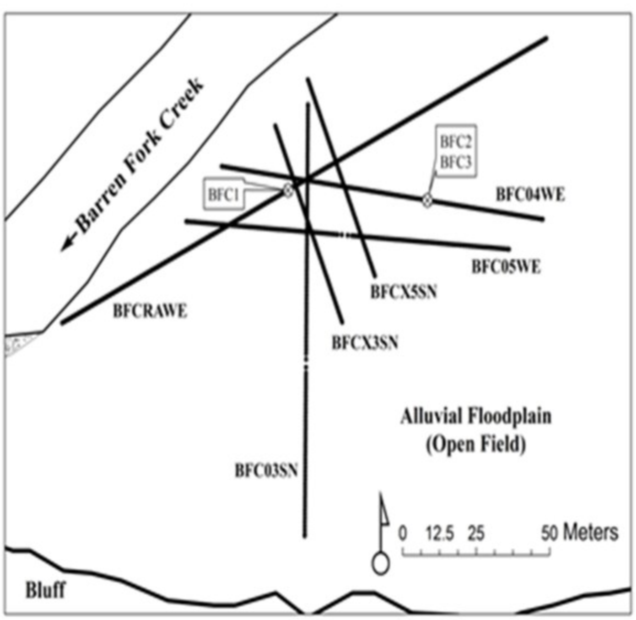


Figure 4. The stream hydrograph, water table elevation in Well 20 (46 m from stream bank), and change in water table direction with change in storm hydrograph for storm runoff event May 1-5, 2009 on BFC, OK (30 minute stage record, USGS gage 7197000). Open circles indicate times for water table/ER comparisons, and actual values are shown in Table 1. "Stream Elevation" is the adjusted hydrograph derived from USGS gage, "Well 20 Elevation" is the water table elevation recorded by the water level logger, "Water Table Direction" is the angle of aspect derived from planar interpolations of water table elevations (i.e., general direction from which water was entering or leaving the studied floodplain), and the "Gaining Stream Threshold Angle" was calculated from the angle (clockwise from north) of the stream bank at the site.

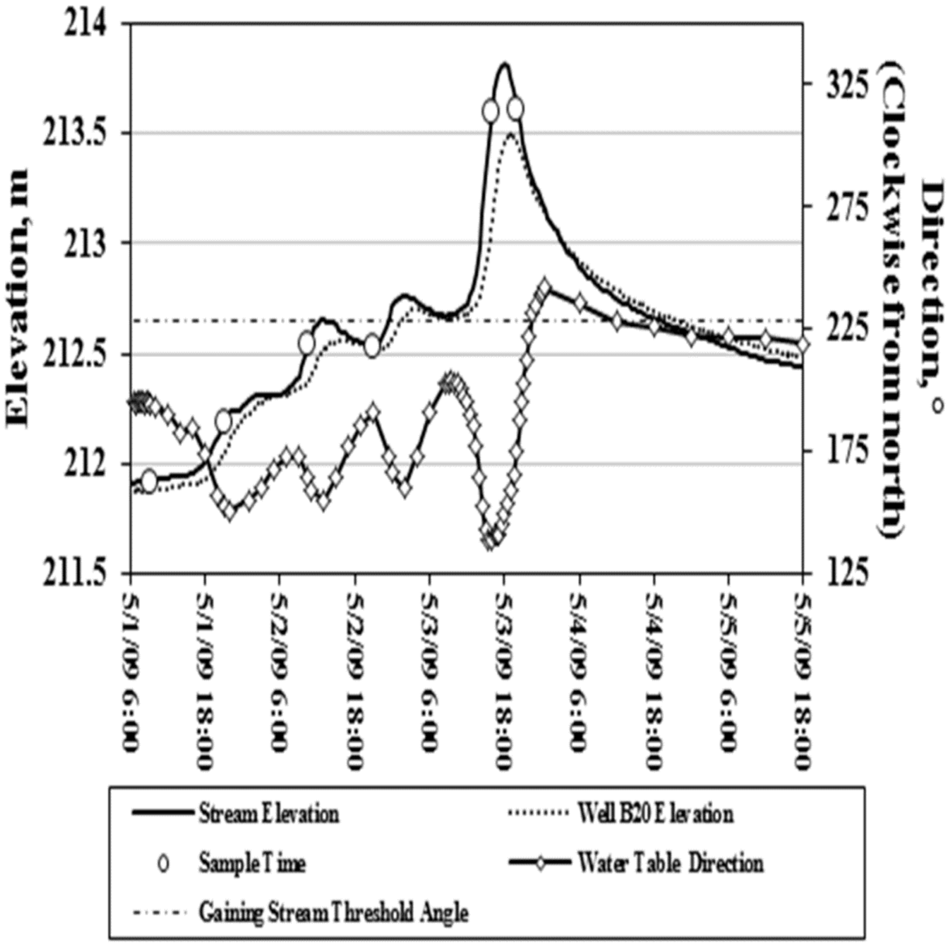


Figure 5. Example ER profiles from Barren Fork Creek (BFC) site. (a) ER profile (1 m electrode spacing) from the BFC site, and (b) ER profile from a gravel bar near the study site. Blue line indicates the position of the water table at the beginning of the May 2009 runoff event. Note the similarity of the resistivity of the feature in (a) marked by the X to the gravel bar in (a) inside the ellipse.

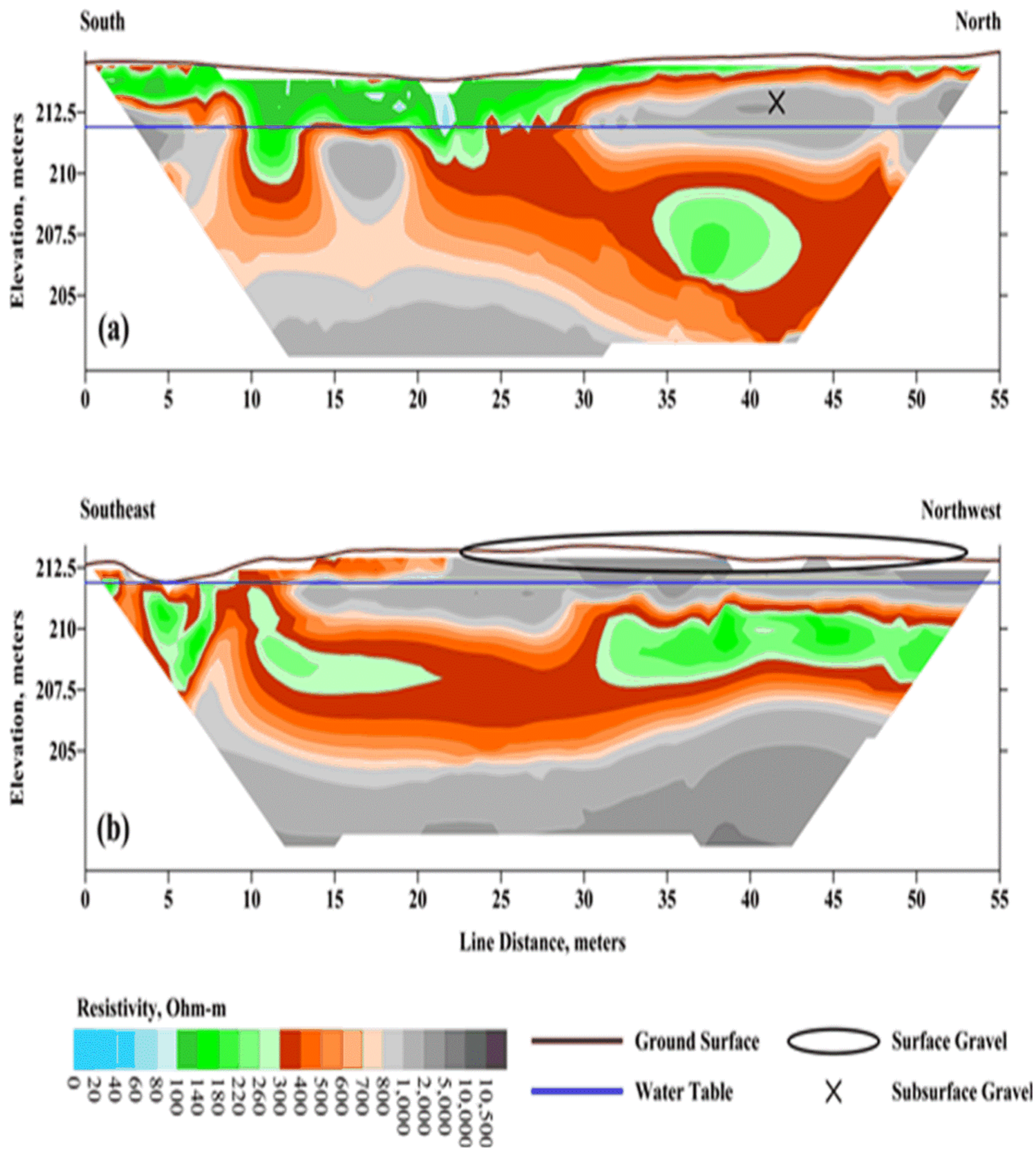


Figure 6. Elevation of maximum flood peak (m) measured in monitoring wells, and delay in time (min) between event peak in the stream and the monitoring well versus distance (m) from the stream. The regressions are significant ($\alpha = 0.05$), and R^2 for the event maximum is 0.68 and for the peak delay is 0.91.

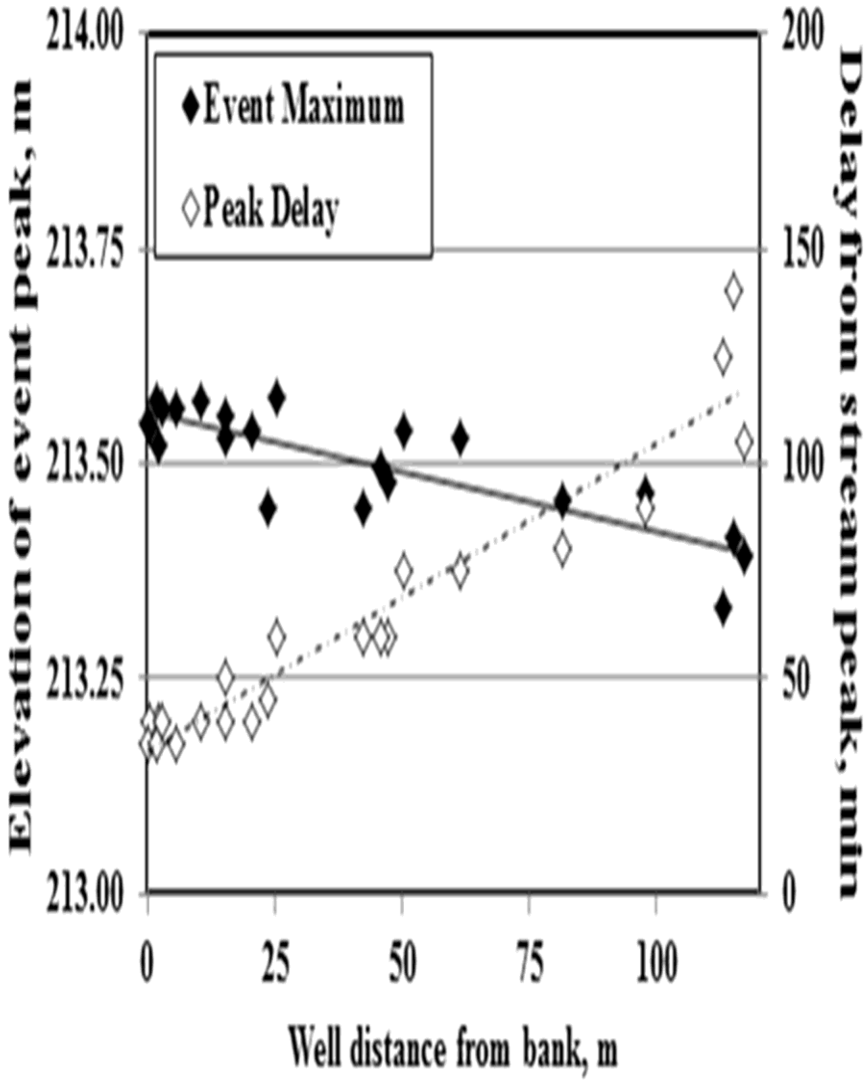


Figure 7. Comparison of interpolated electrical resistivity and water table maps for selected times and stream stages during runoff event May 1-5, 2009 at BFC site. Map comparisons shown are (a) baseflow (5/1/2009 09:00, 211.9 m), (b) rising limb (5/1/2009 21:00, 212.2 m), (c) rising limb (5/2/2009 10:30, 212.5 m), (d) transition (5/2/2009 21:00, 212.5 m), (e) rising limb (5/3/2009 16:00, 213.6 m), and (f) falling limb (5/3/2009 20:00, 213.6 m).

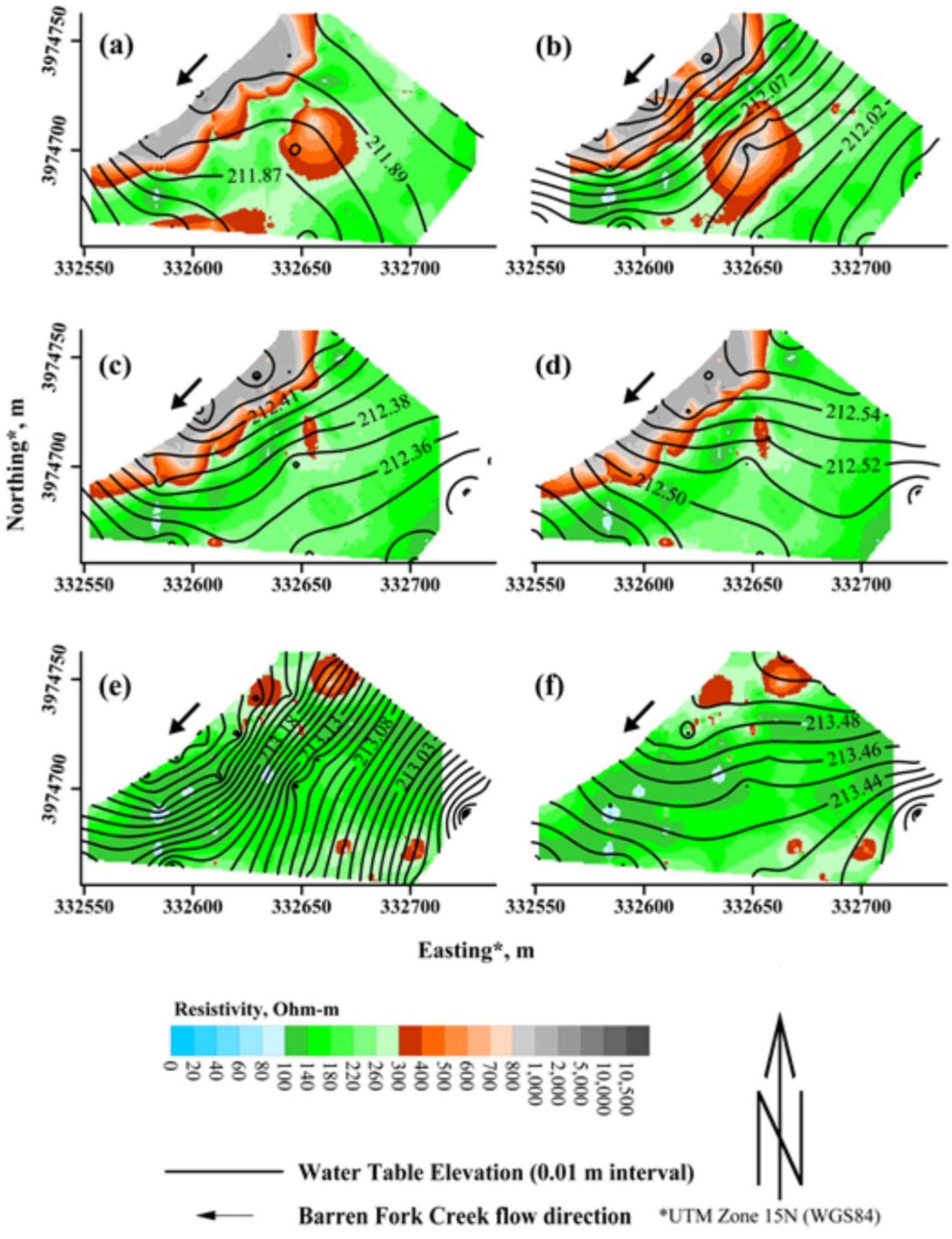


Table 1. Time and stream stage for water table interpolations. Stream stage was taken from USGS gage Barren Fork at Eldon (7197000) 30-minute stage record. Event hydrograph with sample times is shown in Figure 4.

Sample Time	Stream Stage (m)
5/1/2009 9:00	211.9
5/1/2009 21:00	212.2
5/2/2009 10:30	212.5
5/2/2009 21:00	212.5
5/3/2009 16:00	213.6
5/3/2009 20:00	213.6

Table 2. ER statistics for interpolated elevation "slices" of the BFC site alluvial aquifer. "Elevation" is of the interpolated plane within the alluvial floodplain in meters above mean sea level and corresponds to the stream stage elevation in Table 1.

Elevation (m)	Mean (Ω-m)	Standard Deviation (Ω-m)	Minimum (Ω-m)	Median (Ω-m)	Maximum (Ω-m)	Skewness	Kurtosis
211.9	391.1	340.2	48.0	257.0	2652.1	2.4	5.8
212.2	356.0	257.4	29.5	252.8	2467.1	1.9	3.6
212.5	346.4	340.5	33.3	227.4	2587.2	2.6	6.7
213.6	187.7	75.3	27.1	165.1	906.1	1.8	6.3