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The hydraulic conductivity structure of gravel-dominated vadose zones within alluvial floodplains

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Abstract

The floodplains of many gravel-bed streams have a general stratigraphy that consists of a layer of topsoil covering gravel-dominated subsoil. Previous research has demonstrated that this stratigraphy can facilitate preferential groundwater flow through focused linear features, such as paleochannels, or gravelly regions within the vadose zone. These areas within the floodplain vadose zone may provide a route for interactions between the floodplain surface and alluvial groundwater, effectively extending the hyporheic zone across the floodplain during high stream stage. The objective of this research was to assess the structure and scale of texture heterogeneity within the vadose zone within the gravel subsoils of alluvial floodplains using resistivity data combined with hydraulic testing and sediment sampling of the vadose zone. Point-scale and broad-scale methodologies in combination can help us understand spatial heterogeneity in hydraulic conductivity without the need for a large number of invasive hydraulic tests. The evaluated sites in the Ozark region of the United States were selected due to previous investigations indicating that significant high conductivity flow zones existed in a matrix which include almost no clay content. Data indicated that resistivity corresponded with the fine content in the vadose zone and subsequently corresponds to the saturated hydraulic conductivity. Statistical analysis of resistivity data, and supported by data from the soil sampling and permeameter hydraulic testing, identified isolated high flow regions and zones that can be characterized as broad-scale high hydraulic conductivity features with potentially significant consequences for the migration of water and solutes and therefore are of biogeochemical and ecological significance.

Keywords: Electrical resistivity imaging, Floodplain, Gravel, Hydraulic conductivity, Permeameter, Vadose zone

Abbreviations: BFC, Barren Fork Creek; ER, electrical resistivity; ERI, electrical resistivity imaging; FC, Flint Creek; HC, Honey Creek; K_{sat} , saturated hydraulic conductivity; PFP, preferential flow pathway

1. Introduction

Heterogeneity in the subsurface has important ramifications for many aspects of hydrology, most commonly aquifer storage (Fetter, 2001) but also in understanding hyporheic flow in floodplains (Poole et al., 2006; Jones et al., 2008). Efforts to understand the heterogeneity of aquifers have utilized down-well flowmeters (Rehfeldt et al., 1992), tracer tests (Sudicky, 1986), pumping tests (Cardiff et al., 2009; Berg and Illman, 2011), and the joint analysis of those methods (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 density and spatial distribution of the data points (Rehfeldt et al., 1992; Slater, 2007; Alexander et al., 2011). Investigations designed to investigate heterogeneity based on conven-

tional 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). Additionally, 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).

Like the saturated zone, the vadose zone contains heterogeneity that is of research interest. For instance, Fuchs et al. (2009) and Heeren et al. (2010) investigated a horizontal preferential flow path (PFP) originating in a gravel subsoil layer within the vadose zone of an alluvial floodplain. The rate of flow and horizontal transport within the PFP limited the sorption of dissolved phosphorus (P), and consequently P was transported several meters through the subsurface gravel. How-

ever, investigations examining the heterogeneity of the vadose zone are limited relative to the saturated zone by the fact that pumping and tracer tests, which depend on groundwater movement, are impossible. Those studies must utilize other methods, including particle size analysis (Nimmo et al., 2004; Dann et al., 2009), air permeability (Dixon and Nichols, 2005; Chief et al., 2008) and permeameter testing (Xiang et al., 1997; Reynolds, 2010; Miller et al., 2011). When the vadose zone is dominated by coarse gravel such efforts are made more difficult by the problems associated with sampling in gravel: large particles resist penetration by and clog samplers, excavated pits are prone to collapse, and high hydraulic conductivity rates require large volumes of water for *in situ* permeameter testing.

Most applications of borehole permeameter methods involve fine grained soils that allow hand-augering of test holes and which only require a small water reservoir to maintain a constant head. However, in non-cohesive gravels, test holes are difficult to excavate by hand, unsupported holes are prone to collapse, and large volumes of water are necessary to maintain a constant head for the duration of a hydraulic test. To overcome the difficulties presented by coarse alluvial gravels, Miller et al. (2011) designed a steel permeameter that used a direct-push drilling machine to place a slotted-pipe at a specific sampling depth and a 3790 L (1000 gal) trailer-mounted water tank to maintain constant head conditions. A standard 48 by 3.25 in (1.22 by 0.082 m) section of Geoprobe Systems (Kejr Inc., Salina, KS) direct push pipe, with a 0.3 in (7.9 mm) wall thickness, was modified to create a screen by cutting 27 vertical slots 0.002 m (0.07 in.) wide by 0.203 m (8 in.) long arranged in three groups around the pipe perimeter and separated by solid (unslotted) areas. With this method, measurements can be made at successive depths at the same test location (Miller et al., 2011; Fox et al., 2012).

However, borehole permeameters quantify the soil hydraulics via hydraulic conductivity at a point scale with significant effort in gravel soils. A broad-scale non-invasive survey technique to investigate the subsurface would potentially allow rapid and broad spatial scale estimation at potentially fine spatial resolutions of vadose zone heterogeneity, compared to other point scale techniques which provide only local estimates of hydraulic properties (Bohling et al., 2007). The point-scale techniques can be used to ground-truth the broad-scale survey. Hydraulic tomography is commonly used for hydrogeologic characterization but requires the use of multiple pump tests to produce a series of pressure excitation/response data valid for saturated soils (Cardiff et al., 2009; Alexander et al., 2011; Berg and Illman, 2011).

Electrical resistivity imaging (ERI) is a rapid, non-invasive geophysical method that investigates the shallow subsurface by measuring how the materials affect a current of known amperage passing from one electrode to another through the ground (Telford et al., 1990; Herman, 2001; Milsom, 2003; Burger et al., 2006; Slater, 2007). A collinear array of multiple electrodes produces a dataset consisting of apparent resistivities at depths determined by the geometry of the electrode configuration (Herman, 2001). Mathematical inversion of the “apparent resistivity” creates a two dimensional (distance and depth) model of the resistivity of the earth material (Loke and Dahlin, 2002; Halihan et al., 2005; Slater, 2007). The resistivity of earth materials is non-unique, with many different materials having similar and overlapping resistivities (Zohdy et al., 1974; Burger et al., 2006), and interpreting the geology of subsurface resistivity patterns requires independent evidence, such as site stratigraphy, *in situ* or laboratory testing, or core samples.

ERI has been used to map floodplain fluvial sediments (Baines et al., 2002; Bersezio et al., 2007; Crook et al., 2008; Tye et al., 2011; and Ward et al., 2012), to detect gravel for commercial gravel prospecting (Auton, 1992; Beresnev et al., 2002), for geologic investigation of glacial deposits (Smith and Sjogren, 2006), for mapping buried

paleochannels (Gourry et al., 2003; Green et al., 2005), and for imaging hyporheic zone solute transport (Ward et al., 2010; Menichino et al., 2012). Anterrieu et al. (2010) found that two-dimensional ER profiles of a mining waste-rock pile correlated well with a model created from independently acquired data including cores, particle size distributions, and other geophysical surveys. ER techniques have also been applied to estimating the rock particle content of coarse, heterogeneous soils where the rock fragments are dispersed within a fine soil matrix (Huntley, 1986; Rey et al., 2006; Rey and Jongmans, 2007; Tetegan et al., 2012). These studies have shown that ERI can be used to detect gravel in contrast to other fine-grained sediment; thus the method has potential to reveal heterogeneity within the gravel of the floodplain vadose zone of the study sites.

Gravel soils tend to have high hydraulic conductivities and high resistivities, as well as the tendency to drain and achieve field capacity quickly (Lesmes and Friedman, 2005). Recent work has advanced understanding of gravel soils by modeling them as binary systems consisting of mixtures of coarse and fine elements where the particle size of the fine fraction is smaller than the pore size of the coarse fraction, and the coarse particles have no secondary porosity (Koltermann and Gorelick, 1995; Zhang et al., 2011). In those models, a “coarse porosity” maximum (ϕ_c) exists when the fine fraction is zero and the entire soil consists of self-supporting coarse sediment with large, open pores resulting in high electrical resistivity and high hydraulic conductivity, and a fine porosity maximum (ϕ_f) exists when the soil mixture contains only the fine fraction leading to a low electrical resistivity and low hydraulic conductivity (Kamann et al., 2007) (Figure 1). A porosity minimum (ϕ_{min}) exists within the binary model when the coarse fraction is self-supporting (particles resting on one another), but the pore spaces produced by the coarse particles are entirely occupied by the fine fraction. In this condition, the only open pores exist within the fine fraction, and the coarse fraction behaves as pore-free inclusions within a fine matrix effectively reducing the overall porosity. Constant-head flow tests on coarse/fine mixtures suggest that K_{sat} increases rapidly when the fine content decreases past the porosity minimum, and thus the fines content of gravel-dominated soils have a controlling effect on its hydraulic behavior (Koltermann and Gorelick, 1995; Kamann et al., 2007; Zhang et al., 2011). Alyamani and Sen (1993) have similarly acknowledged the role of fine particles in controlling hydraulic conductivity equation of sand aquifers by including the d_{10} and I_0 in their particle size distribution-based empirical hydraulic conductivity equation. Thus, for the special case of coarse vadose zone sediments, a saturated hydraulic conductivity/resistivity relationship may exist that does not rely on formation factor F and soil-water sampling.

The empirical relationship presented by Archie (1942), often referred to as “Archie’s Law”, represented a breakthrough in understanding the resistivity of fluid-saturated systems. A resistivity survey produces apparent resistivity values that “average” the resistivity of the entire path traveled by the current. This is problematic in the porous subsurface regions where the travel path may be complex and include one or more fluids and the current may travel preferentially through only one of the phases. Archie’s Law considers the solid particles to be insulators and the current to be carried exclusively by the saturating fluid, and that the resistivity of the solid particles is proportional to the ratio of the bulk resistivity and the resistivity of the fluid. Clay mineral particles are surrounded by an electrical double layer composed of cations that is capable of conducting electrical current, and thus Archie’s assumption of insulating particles is valid when clay minerals are absent.

While testing the electrical properties of pore fluid can be relatively straightforward in saturated systems, it is less so in the vadose zone,

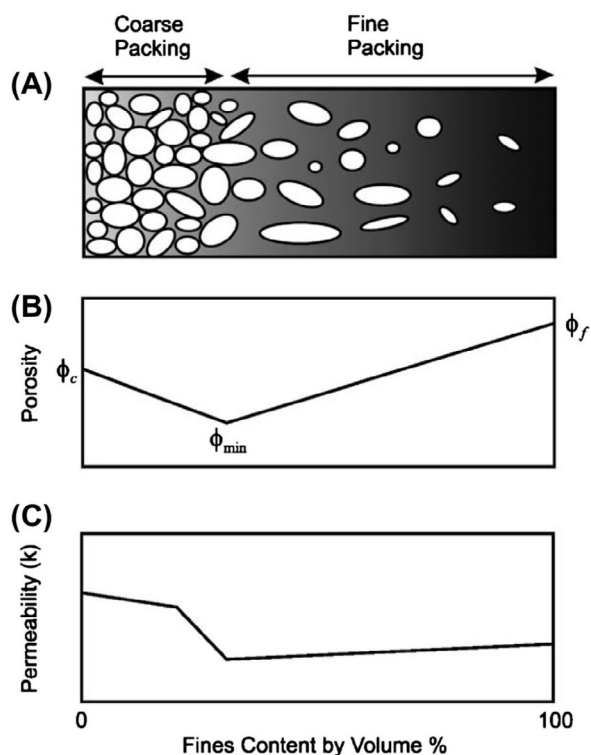


Figure 1. Ideal packing model for a volume of soil consisting of various coarse and fine phase fractions. The horizontal axis denotes the fraction of fine material within the volume as a percent of total volume. (a) Change in soil texture with the increase in fines; (b) the idealized change in porosity with the end members being the coarse (ϕ_c) and fine (ϕ_f) porosity maximums, and the porosity minimum (ϕ_{min}) occurring where the entire coarse pore space is filled with the fine phase; (c) variation in permeability with change in fine content. From Kamann et al. (2007) and used with permission from J. Wiley and Sons.

where pore fluid is held at less than atmospheric pressure, with some portion of the pore space occupied by air. In a gravel vadose zone the particle and pore sizes are much larger than in finer-textured systems. As was discussed above, gravel can be visualized as binary systems of coarse and fine phases, with the pore space containing air when the fine fraction is less than the pore volume. Because air is an electrical insulator, open pores within the gravel vadose zone will likely present a much more resistive target than pores filled with fines, which will likely retain a greater volume of water.

The Ozark ecoregion, which encompasses large areas of southwestern Missouri, northwestern Arkansas and eastern Oklahoma, and is characterized by carbonate bedrock plateaus, river valleys with steep bluffs, and clear gravel-bedded streams (Adamski et al., 1995). The predominance of gravel in the fluvial system has produced alluvial floodplains commonly composed of silt/sand topsoil and coarse gravel subsoil in distinct layers within the vadose zone (Figure 2). Research interest in this vadose zone heterogeneity heightened with the discovery of a PFP originating within the gravel subsoil that was capable of transporting dissolved P long distances without sorption (Fuchs et al., 2009; Heeren et al., 2010; Fox et al., 2011). Of interest in this research was estimating the properties of the unsaturated zone of alluvial floodplains to better understand ephemeral preferential flow paths when the unsaturated zone becomes inundated with water such as during storm events (Heeren et al., 2011, 2014). This was significant because the streams in the region have been shown to be sensitive to nutrient oversupply (e.g., Popova et al., 2006), and because land application of poultry litter from regional poultry production is a significant source of nonpoint P to waterways (Haggard et al., 2003; Storm et al.,

2010). Concern arose that those nonpoint pathways could include transport through the floodplain surface through the vadose zone and into the alluvial groundwater.

The underlying theme of this research is that the use of point-scale and broad-scale methodologies in combination can help to understand spatial heterogeneity in hydraulic conductivity without the need for a large number of invasive hydraulic tests. Therefore, the objective was to explore the appropriateness and develop the relationship between the broad-scale data density provided by ERI and the point-scale testing soil cores and hydraulic conductivity permeameter tests to determine the extent of vadose zone hydraulic heterogeneity in the studied gravel-dominated alluvial floodplains. The approach was to collect ER profiles of the floodplains, obtain coincident point-scale hydraulic measurements at locations on ER profiles and depths within the vadose zone, and then perform regression to determine the sediment controls on the resistivity (ρ) and hydraulic conductivity, with special attention paid to the proportion of fine and coarse material in the subsoil.

2. Methods and materials

2.1. Study sites

Three alluvial floodplain sites in the Ozark ecoregion of northeastern Oklahoma were investigated with ERI, soil coring and permeameter testing. The sites were named after the resident stream: a hay field adjacent to Barren Fork Creek (BFC), a riparian forest adjacent to Flint Creek (FC), and a riparian forest/hay field adjacent to Honey Creek (HC) (Figure 2a). Erosion of the carbonate bedrock by slightly acidic waters has left a large residuum of chert gravel in Ozark soils, with floodplains generally consisting of coarse chert gravel with an expected high electrical resistivity overlain by a mantle (1–300 cm) of gravelly loam or silt loam with an expected low electrical resistivity (Heeren et al., 2011; Midgley et al., 2012).

The sites had similar silt loam soils (dominantly Razort gravelly loam and Elsay very gravelly loam, NRCS, 2011) and dominant bedrock types (cherty limestone, mainly Keokuk/Reeds Spring formation) (Tables 1 and 2). Razort soil series was an alluvial soil occurring in the Ozark region (Oklahoma, Arkansas and Missouri) consisting of silt loam A and B horizons overlying a very gravelly silt loam C horizon. The Elsay soil series similarly occurred on floodplains and consisted of a silt loam A horizon containing chert gravel overlying a C horizon with 35–75% gravel content. The BFC site (35.90°N, –94.85°E) was an open pasture located on the outside of a bend on BFC. The HC site (36.54°N, –94.70°E) was located on the inside of a meander in HC and contained both pasture and riparian forest. The FC site (36.20°N, –94.71°E) occupied a narrow, forested floodplain adjacent to a 200-m relatively straight stretch of FC. The similarity of source materials for stream sediment, including similar bedrock and floodplain soils, allowed the assumption that the composition of the floodplain materials for the study sites would also be similar, despite the fact that they were from different watersheds and varied in stream order, land cover, land use and watershed size.

Soil moisture measurements from the vadose zone layers were not obtained at the time of the ER surveys, primarily because of the difficulty of obtaining samples from the gravels without direct-push coring equipment. However, it should be noted that the subsurface materials of primary interest were well-drained gravels, and that field measurements occurred several weeks after rainfall events when the soil profile was in a gravity-drained state such that gravel subsoils held minimal residual soil moisture content. Also, since investigations were performed in gravel soils close to the streams, the water table depths were assumed equivalent to water-level stage in the stream as determined from U.S. Geological Survey (USGS) gage station data, where gages were positioned within 2 km of each field site (Table 1).

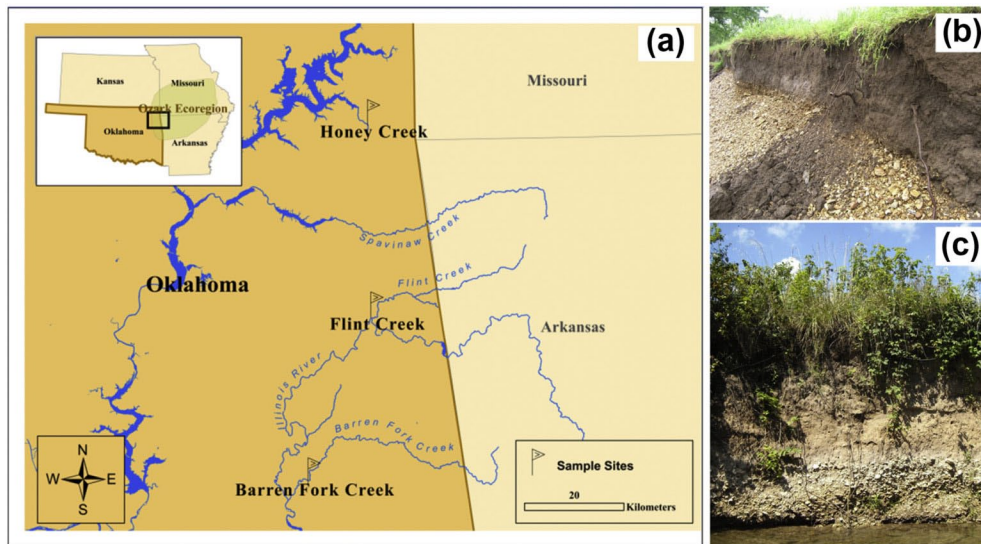


Figure 2. Map showing site locations (a) in Ozark region of eastern Oklahoma, and exposed streambanks with gravel subsoils at two of the studied floodplain sites [(b) Barren Fork Creek and (c) Honey Creek].

2.2. Electrical Resistivity Imaging (ERI)

We conducted ER surveys at each field site to characterize the heterogeneity of the unconsolidated floodplain sediments, focusing on the vadose zone. ER profile accuracy and precision was ensured through the use of a large number of electrodes, an appropriate electrode configuration for the depth of investigation, and the collection scheme. Apparent resistivity 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. The data acquisition and inversion utilized a proprietary Oklahoma State University method (HF, OSU Office of Intellectual Property, 2004; Halihan et al., 2005) to acquire apparent resistivity data and convert them

to a model resistivity profile. 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, 2011; Keppel et al., 2011; Halihan et al., 2011, 2012, 2013; Christenson et al., 2012; Heeren et al., 2014). The method has even been used and accepted by government agencies, such as the Department of Defense and U.S. EPA.

In previous work to establish the quality of the HF method, Halihan et al. (2005) employed it as well as standard acquisition and inversion methods (i.e. Dipole-Dipole) on common electrode arrays (36 electrodes, 1.5 m spacing) to image a hydrocarbon plume at a previously remediated site. The remnant hydrocarbon was more resistive than the background soil, and subsequent coring along the line showed that the HF method was better able to image the spatial outlines of

Table 1. Watershed characteristics and geology of the study sites located in the Ozarks of northeast Oklahoma: Barren Fork Creek (BFC), Flint Creek (FC), and Honey Creek (HC).

Site	Watershed characteristics ^a			Geology ^c		
	USGS gage ^a	Watershed area ^b (km ²)	Median daily discharge ^d (m ³ s ⁻¹)	Formation	General rock type	Watershed area (%)
BFC	7,197,000	845	3.6	Keokuk/Reeds Spring Ada Bloyd and Hale	Cherty limestone Sandstone and shale limestone and shale	70 20 10
FC	7,196,000	300	1.6	Keokuk/Reeds Spring Pitkin, Fayetteville, Batesville	Cherty limestone Sandstone and shale	85 15
HC	7,189,542	150	0.54	Keokuk/Reeds Spring	Cherty limestone	100

a. US Geological Survey National Water Information System (USGS NWIS).

b. Based on 1-arcsecond National Elevation Dataset (US Geological Survey National Map, <http://seamless.usgs.gov>).

c. From Stoeser et al. (2005).

d. Calculated from USGS NWIS mean daily flow records for each gage.

Table 2. Land use and soil types for the Barren Fork Creek (BFC), Flint Creek (FC), and Honey Creek (HC).

Site	Site area (ha)	Primary land use	Soil series ^a	Percent of site area (%)
BFC	1.4	Hay Field	Razort Elsah	97 3
FC	0.6	Riparian Forest	Elsah Healing	62 38
HC	0.7	Riparian Forest and Hay Field	Razort	100

a. SSURGO Soils Database for Oklahoma (NRCS, 2011).

that resistivity contrast over a wider range of hydrocarbon concentrations than the other approaches. To test whether the HF method was suitable for this shallow, vadose zone study, a standard dipole–dipole (DD) ER array was run (56 electrodes, 1 and 2 m spacing) along with the HF method. After inverting the standard DD dataset in EarthImager 2D (version 2.4.0, Advanced GeoSciences Inc., Austin, TX) with standard surface settings, the relative sensitivity (unitless) for each inverted ER model was plotted. Greater sensitivity implies greater ability to accurately resolve spatial differences in resistivity, and the HF method was found to have sensitivity which was greater than DD by a factor that ranged from 6 to an order of magnitude in the depth ranges utilized in this study (Figure 3), and therefore the HF method was selected for imaging the subsurface.

We conducted the ER surveys at the sites between June 2008 and March 2009. Five lines were collected at the BFC site (Figure 4a), five at FC (Figure 4b), and six at the HC site (Figure 4c). The profiles at the BFC site employed electrode spacings of 1.0 m and 2.0 m, with associated depths of investigation of approximately 11.0 and 22.0 m, respectively. The larger depth of investigation at BFC was also used for objectives associated with a separate project at that site. The FC and HC sites utilized 1 m electrode spacing exclusively. Some of the lines were “roll-along” lines that consisted of a linear series of sequential ER profiles (1 m spacing) with at least one-half overlap of electrodes to produce longer, continuous ER profiles without the reduction in spatial resolution that would come with increased electrode spacing. The vadose zone extended no more than 3 m below the ground surface at each site and thus was well within the depth of investigation for our line spacings. The spatial resolution for 1 m and 2 m electrode spacings were approximately 0.5 m and 1 m, respectively. Since the hypothesized targets of interest were fluvial structures (i.e. in-filled channels, gravel bars) at the meter or tens of meters scales, these line configurations would yield data density and sensitivity appropriate for the study. While the ER profiles extended well below the groundwater surface, only the vadose zone data were used in the analysis and results, including vadose zone hydraulic conductivity measurements. Results were not extrapolated to the saturated zone.

A TopCon Hyper Lite Plus RTK GPS (Real-Time Kinematic, TopCon Positioning Systems, Inc. Livermore, CA) capable of 1.0 cm

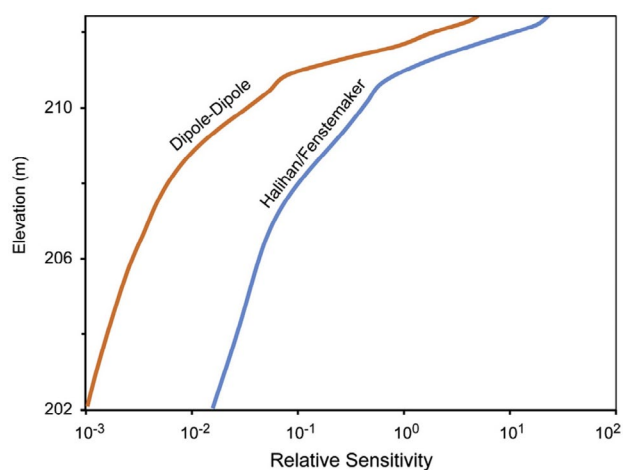


Figure 3. Plot of the change of relative sensitivity with depth for the conventional dipole-dipole (DD) and the Halihan/Fenstermaker (HF) methods for datasets collected from a common (1-m) array from the Barren Fork Creek study site. Relative sensitivity is a metric of the extent to which a change in point resistivity will change the measured apparent resistivity, and is thus a higher relative sensitivity indicates a better ability to detect differences in resistivity. Within the vadose zone (elevation >209 m), HF sensitivity is higher than DD by a factor of about 6 to greater than 10.

horizontal and 1.5 cm vertical precision was used to spatially locate the ER survey electrode locations. The raw GPS points were corrected using the National Geodetic Survey Online Positioning User Service (OPUS) Rapid Static correction, which can place points accurately within 3 cm horizontally and 5 cm vertically depending on the quality of the GPS data. The inverted resistivity data utilizing the Halihan et al. (2005) method were interpolated into grids and contoured utilizing the triangulation and linear interpolation function in Surfer 8 (Golden Software, Inc., Golden, CO), which is appropriate for regularly-spaced datasets. To distinguish them from the apparent resistivity measurements collected in the field, inverted (modeled) resistivity data are termed “ER data” when utilized as individual values, or “ER profiles” when interpolated into a two-dimensional image.

2.3. Soil cores and particle size analysis

We recovered seven direct push cores and one bucket sample from the floodplain subsurface at multiple locations on ER profiles. Core samples were collected at known depths with a Geoprobe Systems TM6200 (GeoProbe, Salina, KS) trailer-mounted probe direct-push drilling machine using a dual-tube core sampler 122 cm long with a 4.45 cm opening. The soil samples were obtained as the gravel permeameter was driven into place for hydraulic conductivity testing within the floodplain, thus the core samples were associated with both ρ and tested hydraulic conductivity at that location within the floodplain (Table 3). The RTK GPS spatially located the core sample locations on the ER profiles.

The sampler opening size limited the particle size that could be sampled, and large cobbles occasionally clogged the sampler, resulting in incomplete cores (about 30% recovery) for that depth interval. While those samples are unavoidably skewed toward smaller particle sizes than the full size range within the floodplain, they appear to consist of complete “undisturbed” samples of the interval until they clogged. Cores were not subsampled before processing, yielding a mass for each core of about 900 g, in order to obtain large enough samples for determining the size distribution of the coarse soils following Bunte and Abt (2001). Records of core depth were maintained, and the core samples could be associated with the nearest ρ value. In addition to samples recovered with the direct-push core sampler, one sample was collected from the Barren Fork Creek bank face sediments with a shovel and bucket after an extreme erosion event exposed the subsoil below the location of an ER profile (Fox et al., 2011). For simplicity, all soil samples are referred to as “cores”. The samples were weighed, dried in an oven at 70 °C then re-weighed. The dried samples were then disaggregated with a mortar and rubber pestle.

Cores were dry-sieved using a series of sieves from 16 to 0.25 mm, and the mass retained on each sieve was weighed. The size distribution of the particles retained on a 16 mm sieve was determined by measuring the “*b*”-axis of the particle with a digital caliper as an approximation of the retaining sieve dimension because that dimension largely controls whether a particle will pass a particular sieve (Bunte and Abt, 2001). Using this method, three or four size “bins” for each core were calculated to encompass the size range between the maximum-sized particle and 16 mm and the total coarse fraction was assigned to those bins by caliper measurement of the “*b*” axis, and the mass of each bin recorded.

The particle size distribution (PSD) of the mass passing the finest sieve (< 0.25 mm) was determined using a Cilas 1190 Laser Particle Size Analyzer (Cilas USA, Madison, WI). The Cilas 1190 measured the relative volume for particle size ranges of a representative sample. The particle size analyzer has a measurement reproducibility of better than 1% and accuracy error of less than 3% (Cilas USA, Madison, WI). The PSD of the mass passing 0.25 mm was calculated by multi-

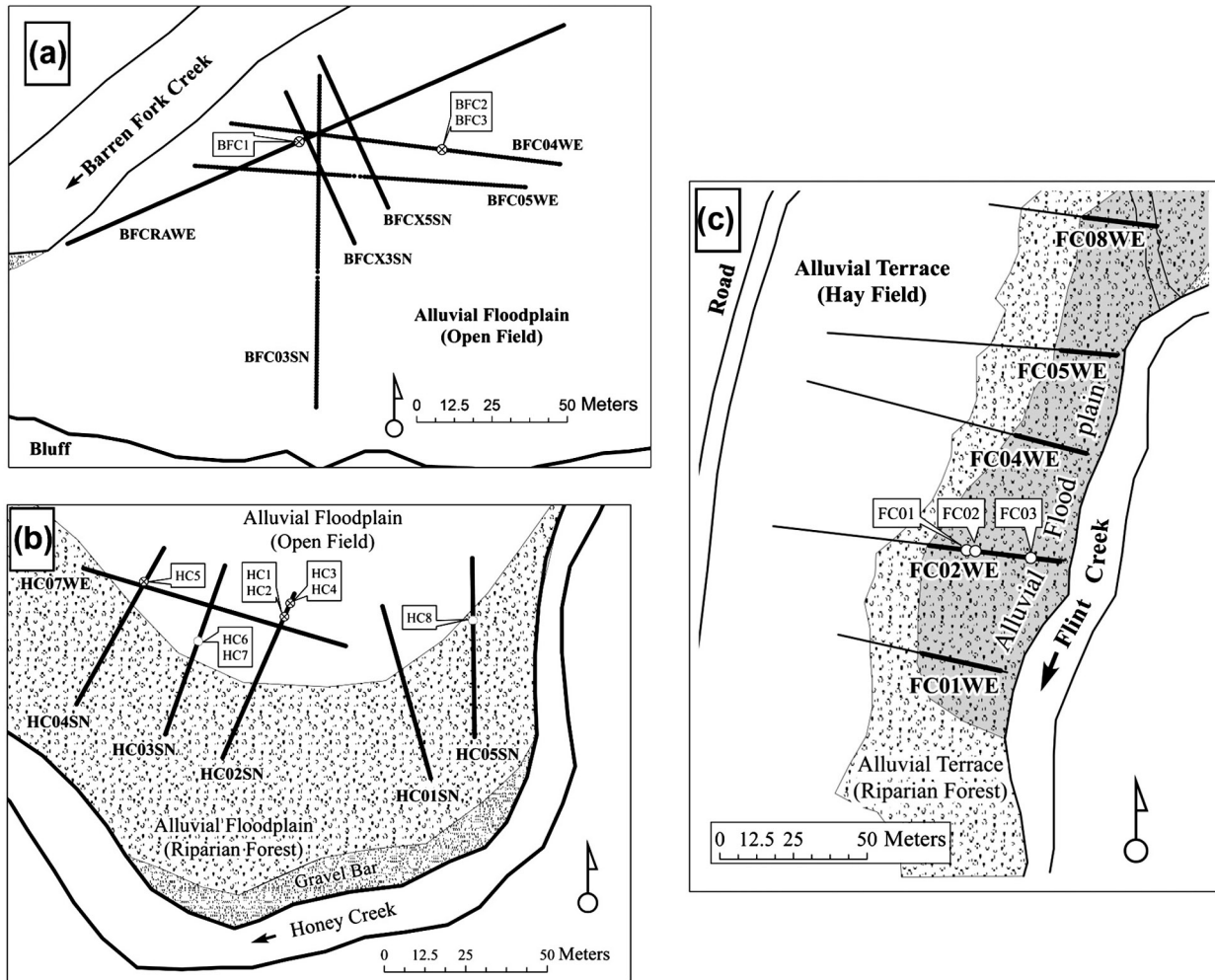


Figure 4. Maps of study ER lines at (a) Barren Fork Creek, (b) Honey Creek, and (c) Flint Creek. Crossed circles indicate locations of permeameter and core samples, open circles indicate locations of permeameter tests only. Thin lines on (c) indicate portions of ER lines not utilized in study.

Table 3. Collection depth and oven-dried mass of soil cores, resistivity, and saturated hydraulic conductivity estimated from borehole permeameter tests (Miller et al., 2011; Fox et al., 2012) from core locations at Barren Fork Creek (BFC), Flint Creek (FC), and Honey Creek (HC). Resistivity data are inverted (modeled) values. Core samples were not obtained from FC.

Site	Test	Ground surface elevation (m)	Depth of core (m)	Mass (g)	Proportion of fines	Resistivity (Ω m)	Saturated hydraulic conductivity ($m\ d^{-1}$)
BFC	BFC1	212.4	2.24	15,260	0.03	1263	400
	BFC2	212.2	2.05	890	0.28	180.5	130.7
	BFC3	212.2	1.67	1071	0.11	252.2	210.5
HC	HC1	236.2	1.75	684	0.08	415.6	844.2
	HC2	236.2	1.07	1128	0.74	71.1	0.2
	HC3	236.2	1.77	940	0.47	95.6	0.2
	HC4	236.5	1.67	791	0.34	325.1	12.9
	HC5	236.2	1.1	860	0.05	544.9	237.8
FC	FC1	266.1	1.17	*	*	293.7	121.3
	FC2	266.2	1.17	*	*	254.4	81.1
	FC3	266.2	1.17	*	*	224.9	55.9

plying the percent distribution from the sample by the total volume of the fine dry-sieved fraction.

The presence or absence of the proportion of the largest particles within a sample of coarse sediment could skew the resulting PSD. To reduce this effect, sample sizes that are ~ 1000 times the mass of the largest particle are recommended (Church et al., 1987). The core sampling method utilized in this study greatly constrained the mass of each sample obtained and thus represented a source of error in the resulting

PSDs, which are likely finer than the true subsoil. For this study the important feature of the PSD turned out to be the proportion of fine sediment relative to coarse particles within each sample, and since the bulk of the sampling was subject to the same error, and no alternate sampling was available, the resulting PSDs were used in the analysis.

X-ray diffraction (XRD) was used to determine the presence or absence of clay minerals in each sample (Moore and Reynolds, 1997; Poppe et al., 2001). The unique pattern produced by the diffraction

of a beam of X-rays through a mineral's crystal structure is described mathematically by Bragg's law

$$2d \sin \theta = n\lambda \quad (1)$$

which defines the relationship between the X-ray wavelength (λ), the X-ray angle of incidence (θ), and the separation distance between crystal layers (d). Diffractograms, which display X-ray intensity as a function of the angle 2θ , are commonly used to identify minerals. A representative sub-sample of the fine fraction (<0.25 mm) from each core was ground to a fine powder with a ball mill, suspended in tap water and centrifuged at 1500 rpm for 2 min to isolate clay-sized particles in the water column (Kittrick and Hope, 1963). A slide for XRD analysis was prepared by repeatedly applying water bearing the suspended clay-sized particles to the surface of a slide and allowing the water to evaporate. This procedure produces an oriented mount sufficient for establishing the presence or absence of clay minerals (Moore and Reynolds, 1997). Analyses were performed with a Phillips PW3020 computer-automated X-ray diffractometer (PANalytical B.V., Almelo, NED).

2.4. Permeameter testing of hydraulic conductivity

Field measurements of K_{sat} were collected at successive depths in the vadose zone of all three floodplain sites using a slotted section of the direct push pipe as a permeameter. The gravel permeameter device consisted of a section of direct push pipe with a screened interval near the base, a trailer-mounted water reservoir, and measurement instrumentation (Miller et al., 2011; Fox et al., 2012). The permeameter was placed at a chosen location along an ER profile and then driven to the desired depth within the floodplain gravel vadose zone using the Geo-Probe. Constant head within the permeameter was induced by flow at a known rate from the trailer mounted tank, and monitored with a vented pressure transducer within the permeameter.

The head and flow rate values for the permeameter tests were analyzed using the US Bureau of Reclamation Gravity Permeability Test Method 3 (USBR, 1985), which was suitable for cased permeameters in non-cohesive soils, and resulted in K_{sat} estimates ranging between 0.2 and 844.2 m d⁻¹ for targeted depths in the vadose zone (Miller et al., 2011; Fox et al., 2012). For this study it was primarily important to determine relative rather than absolute differences between the hydraulic conductivities at different locations within the study floodplains in order to establish the relationships used in the analysis. Therefore, although the method is suitable for determining hydraulic conductivity estimates to within an order of magnitude accuracy (USBR, 1985), Test Method 3 was the only method available that was suitable for cased boreholes in coarse gravel vadose zones, and therefore it was the method chosen.

2.5. Resistivity and unique flow domains

The ER profiles provide a two-dimensional vertical view into the floodplain subsurface showing heterogeneities within that subsurface. The "Grid Volume" function in Surfer 8 was used to calculate the area within the vadose zone of each profile greater than a given resistivity value. A series of those area calculations with increasing resistivities produced a cumulative distribution of resistivity by area. A normalized area was calculated by dividing each area by the total area for that line. This area was a function of the length of the line and the vertical distance between the ground surface and the water table at the line location. For example, a roll-along line at BFC had the greatest area, where the line consisted of multiple co-linear ER profiles (180 m in total) collected separately and then processed as a single unit.

2.6. Potential sources of error

The USBR Method 3 utilized for analyzing the permeameter tests in the coarse gravel subsoils was accurate to within an order of magnitude. These results were utilized in the study, but are likely best treated as indicators of relative difference, rather than precise estimates of hydraulic conductivity. While the ER surveys were conducted weeks after significant rainfall events, when the soil was likely in a gravity-drained state, actual soil moisture measurements were not obtained at that or any other time at those sites for the reasons stated previously. Therefore, any relationship that involves soil moisture determined in this study is like-wise best treated as a relative rather than a precise measure.

3. Results and discussion

3.1. Resistivity surveys

The resistivity surveys produced 12,270 individual vadose zone ER point estimates at BFC, 3425 at HC, and 880 at FC. ER values at the sites ranged over several orders of magnitude (11–12,000 Ω m) with the largest range found at BFC (Table 4). The contact resistance for most datasets was good, but the gravel surfaces had higher contact resistances of approximately 1500–2500 Ω . The inversion quality was good with RMS errors of 3–9%. The larger dynamic range in the Baren Fork site along with higher contact resistance caused the higher RMS errors. The ER data were positively skewed, indicating a large number of low values and a small number of extreme high values.

ER profiles from the sites show complex ER patterns within the vadose zone (Figure 5). It was noted during fieldwork that at times electrodes were placed directly into gravel. Examination of exposed streambanks showed that the surface Razort or Elsay soil layer was highly variable in thickness (from <0.1 to >1.0 m), and it was assumed that variability extended across the floodplains. The soil can be detected in some ER profiles where it is thicker than the spatial resolution (0.5 m for the 1 m spacing) and provides a resistivity contrast to the underlying gravel, for example at the left (west) of the BFCRAWE profile where was seen as ER data values of 100–300 Ω m close to the top of the profile (Figure 5). That lower resistivity soil layer is seen to thin and eventually disappear to the right (east) of the profile, leaving higher resistivity (gravel) at the profile surface. The soil layer was not resolved in all images, either because the layer thickness was less than the spatial resolution, or because the resistivity of the soil layer did not strongly contrast with the underlying gravel.

ER data values below the surface layer at the sites do not show horizontal layering, but rather horizontally discontinuous zones of high and low ρ . The pattern contained within the ER profiles is consistent with a transect passing through discontinuous features that may include buried fluvial structures such as gravel bars or filled paleochannels. The high ρ 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, Figure 6), supporting the conjecture that the high resistivity features within the floodplain may consist of clean gravel without fines similar to the bar surface. 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.

3.2. Soil samples

The core samples obtained from the sites did not contain significant layering; therefore they were processed in their entirety, and it was assumed

Table 4. Statistics for ER data from surveys at Barren Fork Creek, Honey Creek, and Flint Creek. All values except sample number (n) and skewness are reported as Ω m, or as \log_{10} Ω m.

	Barren Fork Creek		Honey Creek		Flint Creek	
	ER	\log_{10} ER	ER	\log_{10} ER	ER	\log_{10} ER
n	12,799	12,799	3823	3823	1297	1297
Mean	494	2.48	281	2.39	271	2.37
Standard deviation	611	0.42	152	0.24	156	0.24
Minimum	11	1.04	30	1.48	33	1.52
Median	266	2.43	247	2.39	235	2.37
Maximum	11,830	4.07	1190	3.08	1220	3.09
Skewness	5.06	0.27	1.23	-0.19	1.65	-0.15

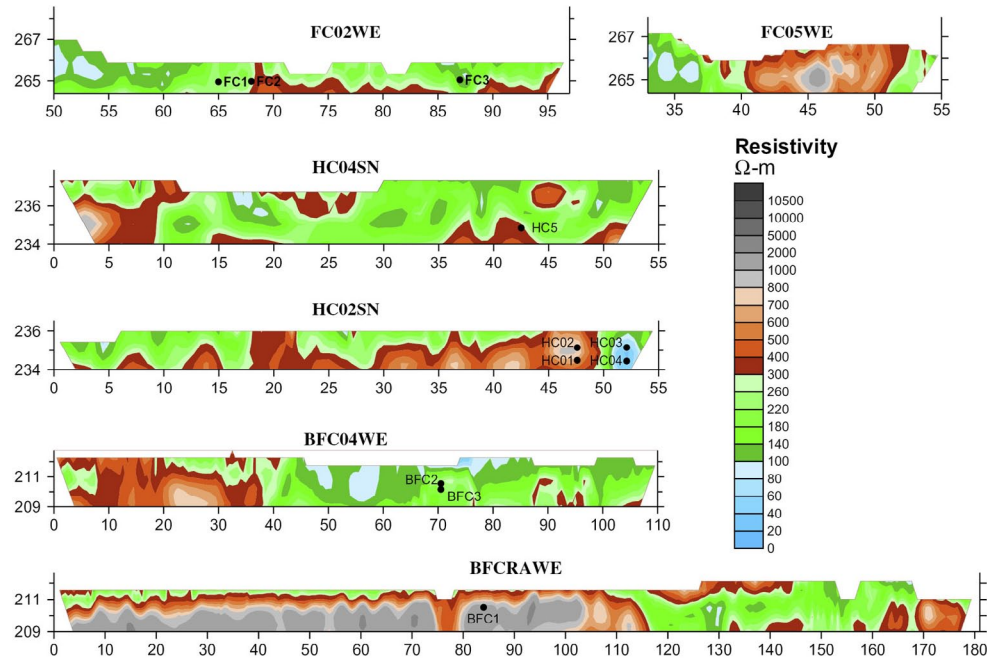


Figure 5. Example electrical resistivity (ER) profiles from the vadose zone floodplain sites at Flint Creek (FC02WE and FC05WE, Honey Creek (HC04SN and HC02SN), Barren Fork Creek (BFC04WE and BFCRAWE). The Flint Creek profiles extend onto the stream terrace, but only the floodplain segment is shown. Permeameter and PSD test locations are indicated with solid circles.

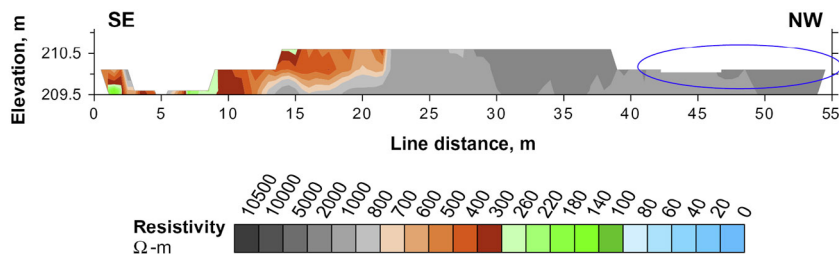


Figure 6. Electrical resistivity (ER) profile of gravel bar located approximately 100 m upstream of the Barren Fork Creek (BFC) study site. The profile is oriented across the gravel bar and perpendicular to Barren Fork Creek in a southwest to northeast direction, terminating at the stream edge at right of profile. Clean gravel is exposed at the surface of the gravel bar along the stream (circled area at the right of the profile).

the PSDs produced were a valid representation of the subsurface soil texture at those points. The PSDs of the core samples showed a range of textures (Figure 7), with most having large fractions of sand or coarser particles. Cores with the coarsest textures were predominantly gravel-sized (>2 mm), with one core at BFC and one core at HC having 93% and 89% larger than 2 mm by mass, respectively. The finest textures were predominantly silt (0.002–0.05 mm), with two cores at HC having 65% and 46% in this size range. The coarse (sand-sized and larger) particles in the samples were identified as chert fragments. Random slide mounts of the material passing the 0.25 mm sieve were analyzed with

X-ray diffraction. XRD diffractograms plot the intensity of X-rays diffracted through the crystal lattice of minerals against 2θ (twice the angle of incidence). According to Bragg's law, the 2θ is related to the spacing within the crystal lattice and thus is a characteristic of particular minerals. The XRD diffractograms for the clay-sized fraction collected at the study sites showed minimal intensities at 8.8 and 12.2, 2θ , indicating no presence of the clay minerals illite and kaolinite, and slight intensity at 20.7 and 26.7 2θ , indicating the presence of quartz (Poppe et al., 2001). Kaolinite and illite are clays common in the weathering of limestone, and their absence is likely due to their long settling time kept them in

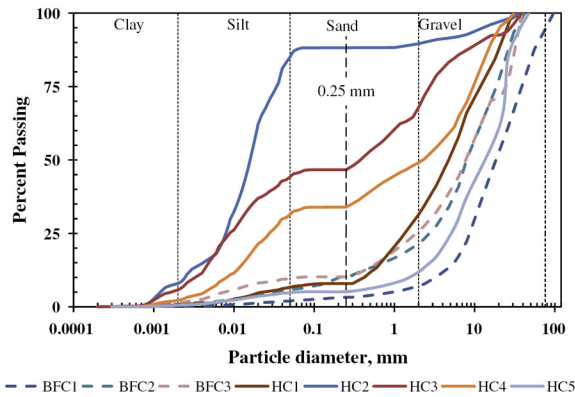


Figure 7. Particle size distributions for floodplain vadose zone samples obtained at various depths below ground surface. Samples from Barren Fork Creek are shown with dashed lines, and samples from Honey Creek are shown with solid lines.

suspension in the energetic fluvial environment that produced the gravel deposits. Likewise, quartz is the principal mineral of the chert gravel present in the subsoil, so its presence as finely-divided sand-, silt- and clay-sized particles is expected. Considering the chert-rich source material in the watershed, and the chert coarse material in the samples, the fine material in the core samples was assumed to be primarily finely-divided chert fragments, not clay mineral particles.

The USDA soil classification for the cores ranged from “cobbly gravel” for the coarsest sample (BFC1, Table 3) to “silt” for the finest sample (HC2, Table 3). Most of the other samples would be classified as “extremely gravelly sandy loam” or “extremely gravelly loamy sand”. Many of the PSDs exhibited a flat trend near the particle size of 0.25 mm (within the range of a sand particle), implying that many of the cores consisted of particle sizes that were either much larger or smaller than 0.25, and further indicating that the soils resembled the artificial binary models of coarse and fine phases that have been designed to better understand gravel soils (Koltermann and Gorelick, 1995; Kamann et al., 2007; Zhang et al., 2011). For this study, the soils were assumed to be mixtures of coarse and fine elements with the division placed at 0.25 mm due to the above-mentioned observations in the PSDs, with the proportion of the PSD < 0.25 mm (fine fraction, f) used to describe each soil.

3.3. Resistivity and fine fraction

A plot of the ER measured at the core location and the fine fraction of the core showed a negative relationship in which ρ decreased as the f of the core PSD increased, with a power function best fitting the relationship (Figure 8). Linear regression of the log-transformed variables was significant at $\alpha = 0.05$ ($P = 0.001$) with $R^2 = 0.85$. According to Archie’s Law, in unconsolidated materials, when clay minerals are not present, the sediment particles themselves are generally insulators, and current is conducted by pore fluid; therefore, areas with high soil moisture will have lower resistance to electrical current (Archie, 1942; McNeill, 1980). Since clay minerals were determined to be rare in these subsoils, it was assumed that the resistance of the subsurface was primarily controlled by the moisture retention of the gravel subsoil. All ER data were collected under gravity-drained conditions, so it was assumed that the largest pore spaces in the coarse gravel were filled with air or by fine material, and that soil moisture was generally retained in the small pore spaces produced by the fine particles.

The negative power relationship between ρ and f was similar to the volumetric water content/ ρ relationship shown in Samouelian et

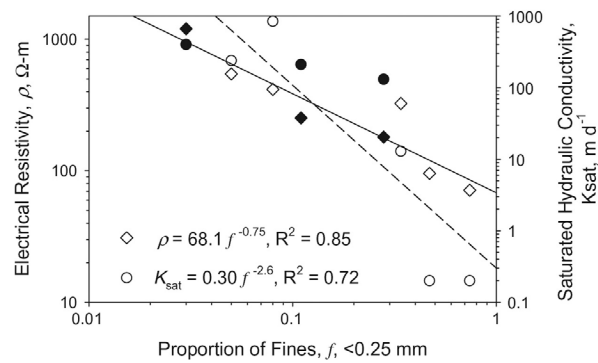


Figure 8. Relationship between fine fraction (f , <0.25 mm) and resistivity (ρ , Ω -m, diamond symbols, solid line) and field-measured saturated hydraulic conductivity (K_{sat} , m d^{-1} , circle symbols, dashed line) from subsoil samples taken from locations on Barren Fork Creek (filled symbols) and Honey Creek (open symbols) floodplains. No Flint Creek samples were acquired for fine content analyses at the time of sampling.

al. (2005), suggesting that the f/ρ relationship was similarly responding to the moisture retained within the f since fine-textured soils have greater moisture-holding capability compared to gravel (Fetter, 2001). It was hypothesized that the variation in ρ of the gravel soils was due to the variation of the fine fraction. The actual moisture content/ ρ relationship was unknown, since soil moisture measurements were not taken at the time of the survey. However, subsurface materials were mostly well-drained gravels with ER measurements that typically occurred several weeks after rainfall events, leaving the profiles in a near gravity drained state especially in the upper portions of the soil profiles. It therefore seems likely that areas with higher proportions of f within the coarse subsoil had low ρ because the fines were able to retain relatively more water within its pore spaces than coarser soils, and that areas with low f had high ρ because the pore space within the coarse matrix was occupied by air.

3.4. Fine fraction and hydraulic conductivity

Field permeameter testing was conducted at locations on ER profiles and some were conducted at locations from which the cores were obtained (Table 3). The ER data grid has a horizontal and vertical spacing of 0.5 m, for 1-m electrode spacing, and ρ -values were taken from the nearest grid point. Thus each hydraulic test was associated with a ρ value, and tests at BFC and HC were associated with both ρ and particle size distribution. A plot of the f and hydraulic conductivity showed a negative power relationship similar to that between the f and measured ρ (Figure 8) in which low fractions of fine material corresponded to high hydraulic conductivity. Linear regression of the log-transformed variables shows that the regression was significant at $\alpha = 0.05$ ($P = 0.008$) with $R^2 = 0.72$. The accuracy limitations of the hydraulic testing method means that only the general nature of the hydraulic conductivity/ f relationship is known, but it is a significant feature of this relationship that hydraulic conductivity increases as f decreases. The highest hydraulic conductivities are found in gravel subsoil classified as “cobbly gravel” and lower conductivities in subsoil classified as “gravelly silt loam”. Existing work on artificial binary soil mixtures of coarse and fine elements found that the hydraulic conductivity of the mixtures was highest when the fine fraction approached zero and the material consisted of coarse material with interconnected, open pores (Koltermann and Gorelick, 1995; Kamann et al., 2007; Zhang et al., 2011) (Figure 1). The hydraulic behavior of the naturally-occurring gravel soils present in the floodplain appeared to be similar to that of artificially-constructed coarse soils, suggesting that the fraction

of f within the gravel subsoil was an important factor in describing the hydraulic behavior of those coarse soils.

Hydraulic testing was conducted at Flint Creek, but no core samples were collected (Table 3). To test whether the f/K_{sat} and f/ρ relationships held for that site as well, a linear regression of the \log_{10} transformed ρ and K_{sat} for the BFC and HC cores was prepared, with the understanding that the hydraulic testing method could estimate hydraulic conductivity to an order of magnitude, and thus the ρ/K_{sat} relationship was best understood as an estimate of the general nature of the relationship. The regression was used to predict K_{sat} for the Flint Creek tests based on the ρ at the hydraulic test locations, and that predicted value was compared to the *in situ* observed values. The slope for the observed/ predicted relationship was 0.91 ($R^2 = 0.99$), which is close to 1, indicating a very close predictive correspondence, indicating that the $f-\rho-K_{sat}$ system for Flint Creek vadose zone gravels was similar to those of Honey Creek and Barren Fork Creek.

3.5. Resistivity and unique flow domains

The range of ρ spanned nearly four orders of magnitude at BFC, and about three orders of magnitude at HC and FC, with BFC having relatively large fractions greater than 1000 Ω m (Figs. 9a–c). The fractional area of the BFC ER profiles fell into two distinct configurations: a more uniform distribution of relatively lower ER data resulting in a steeper curve and a more variable distribution of ER data resulting in a shallower curve (Figure 9a). The HC fractional area plot showed ER profiles with similar patterns, all of which had resistivity that fell into a range similar to that of the low-resistivity BFC profiles (100–1000

Ω m) (Figure 9b). The FC fractional area plot showed that most of the ER profiles fell within the range of 100–1000 Ω m, but the patterns of the ER profiles were not consistently similar (Figure 9c).

The “high- ρ ” BFC fractional area curves appeared unique among all ER profiles in having a large area of high resistivity (Figure 9d). The 16th percentile denotes one standard deviation (in a normal distribution) and is a standard representative of large values in a distribution (Bunte and Abt, 2001). The 16th percentile calculated from the ER profile was chosen to be representative of the high ρ for each profile (Figure 9d). To test whether the high- ρ BFC ER profiles were significantly different than the other ER profiles, a two-sample *t*-test was performed on the 16th percentile values. The mean of the 16th percentile of the high- ρ BFC ER profiles was significantly different when compared to the mean of all other 16th percentiles ($P = 0.000$, $\alpha = 0.05$). Analysis of Variance (ANOVA) and Tukey’s pairwise comparison of the differences between the individual sites (high- ρ BFC, low- ρ BFC, HC and FC) showed that the high- ρ BFC ER profiles were significantly different than all other sites (95% confidence level) but that the remaining sites were not statistically different. The high- ρ BFC ER profiles had approximately 15% of the area with ρ greater than 1000 Ω m, while the remaining profiles had an equivalent area with ρ above only 300 Ω m.

The three study sites represent a range of gravel floodplain vadose zones within the Ozark ecoregion. All of the ER profiles from those floodplain vadose zones had median ρ which suggested hydraulic conductivities in the tens of m/d at minimum (Figure 9). If those areas were spatially connected across the floodplain, they could form broad-scale ephemeral flow domains, and become active during high stream stages when the floodplain is saturated (Fuchs et al., 2009; Fox

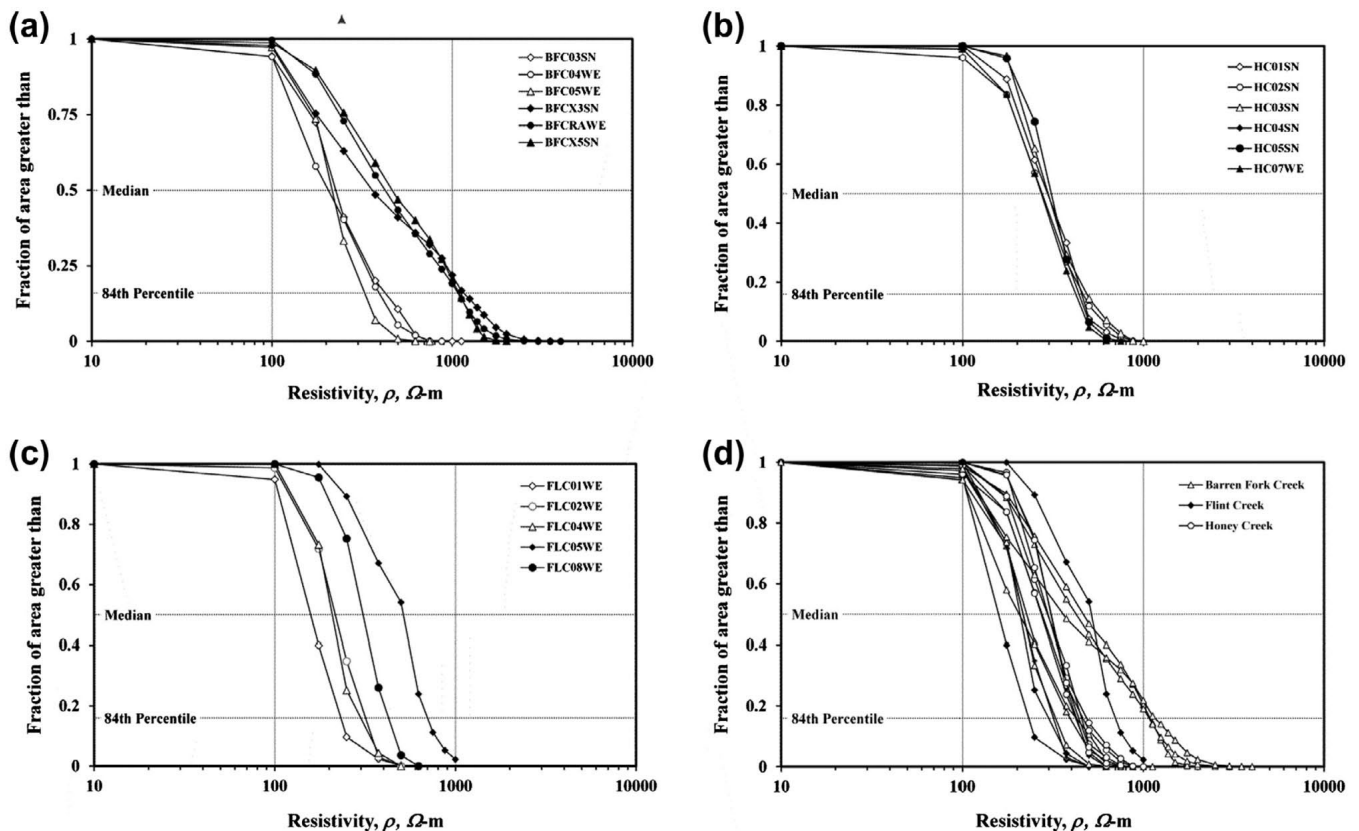


Figure 9. Relative percentage of area for electrical resistivity (ER) profiles from Barren Fork Creek (a), Honey Creek (b), Flint Creek (c), and all sites combined (d). The vertical axis of each plot is percentage of area of the individual ER profiles, and the horizontal axis is inverted ER (Ω m).

et al., 2011; Heeren et al., 2011; Mittelstet et al., 2011), and small domains within the highest- ρ ER profiles could form ephemeral preferential flow paths within the gravel floodplain as observed by Heeren et al. (2011). Similar observations of heterogeneity and preferential flow pathways have been reported in alluvial floodplains worldwide. Stanford and Ward (1992) and Amoros and Bornette (2002) point out that surface water/ground water exchanges link river channels to the entire floodplain even during periods of base flow. Hydrologic pathways become complex with deposits of coarse alluvium (Naiman et al., 2005). These preferential flow pathways should be considered when conceptualizing subsurface transport in alluvial floodplains, as this heterogeneity can influence flow and mass transport rates and therefore be of biogeochemical and ecological significance (Poole et al., 2006; Kalbus et al., 2009).

4. Conclusions

ERI was used to investigate the extent of vadose zone heterogeneity within three gravel-dominated floodplain sites and detected differences within the gravel subsoil confirmed based on core samples that showed differences in fines content within the gravel. Additionally, *in situ* tests with a gravel permeameter showed that the fines content was correlated to the hydraulic conductivity of the subsoil. In general, the methods used above in tandem were able to identify heterogeneity within the gravel vadose zones of the alluvial floodplains, not identifiable from using the methods individually. Fractional area plots of ER profiles followed two distributions: “low- ρ ” ER profiles shown at all three sites and “high- ρ ” shown for some of the BFC ER profiles. The “low- ρ ” ER profiles indicated that large portions of the studied floodplain vadose zones may be broad-scale ephemeral flow domains with potentially significant consequences for the movement of environmentally sensitive materials such as excess nutrients (i.e., phosphorus in the Ozark ecoregion). Those “low- ρ ” ER profiles appeared at all three study sites based on the ER, PSD, and permeameter hydraulic testing, and may be typical for Ozark gravel floodplain vadose zones. The “high- ρ ” ER profiles contained small areas with extremely high ρ (limited fine content in the PSD and higher hydraulic conductivities based on permeameter testing), a configuration which, if they are spatially connected across the floodplain, could constitute PFPs and transport environmentally sensitive materials without attenuation under the appropriate conditions. The “high- ρ ” ER profiles were unique to BFC for this study, but it seems unlikely that the BFC site was unique within the Ozarks, and further work is required to determine the fluvial, geologic and geomorphic factors that contribute to creating similar high- ρ features. Future work should also be devoted to assessing spatial connections across the floodplain, and modeling the implications of the large-scale ephemeral domains on hyporheic exchange and contaminant transport in stream systems during both base flow and high flow conditions.

While this study continues to expand the application of the geophysical method ER to hydrology, the relationships between fines content and ρ and hydraulic conductivity outlined in this study are fundamentally empirical and are not directly applicable to other vadose zones. Further, in this study a direct relationship between ρ and hydraulic conductivity was valid only in a relative sense (i.e. higher ρ corresponds to higher hydraulic conductivity), given the uncertainty associated with the hydraulic testing method. However, given that the stream energy required to mobilize and rework gravel is incompatible with depositing clays, it is likely that gravel-dominated floodplains in other regions would exhibit a similar relationship between ρ and hydraulic conductivity within the vadose zone.

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References

- Adamski, J.C., Petersen, J.C., Freiwald, D.A., Davis, J.V., 1995. Environmental and Hydrologic Setting of the Ozark Plateaus Study Unit, Arkansas, Kansas, Missouri, and Oklahoma. USGS Water-Resources Investigations Report 94-4022.
- Alexander, M., Berg, S.J., Illman, W.A., 2011. Field study of hydrogeologic characterization methods in a heterogeneous aquifer. *Ground Water* 49 (3), 365–382.
- Alyamani, M.S., Sen, Z., 1993. Determination of hydraulic conductivity from complete grain-size distribution curves. *Ground Water* 31 (4), 551–555.
- Amoros, C., Bornette, G., 2002. Connectivity and biocomplexity in waterbodies of riverine floodplains. *Freshw. Biol.* 47, 761–776.
- Anterrieu, O., Choteau, M., Aubertin, M., 2010. Geophysical characterization of the large-scale internal structure of a waste rock pile from a hard rock mine. *Bull. Eng. Geol. Environ.* 69, 533–548.
- Archie, G., 1942. The Electrical Resistivity Log as an Aid in Determining Some Reservoir Characteristics. *Petroleum Technology Technical Paper* 1422.
- Auton, C., 1992. The utility of conductivity surveying and resistivity sounding in evaluating sand and gravel deposits and mapping drift sequences in northeast Scotland. *Eng. Geol.* 32, 11–28.
- Baines, D., Smith, D.G., Froese, D.G., Bauman, P., Nimeck, G., 2002. Electrical resistivity ground imaging (ERGI): a new tool for mapping the lithology and geometry of channel-belts and valley-fills. *Sedimentology* 49, 441–449.
- Beresnev, I.A., Hruby, C.E., Davis, C.A., 2002. The use of multi-electrode resistivity imaging in gravel prospecting. *J. Appl. Geophys.* 49, 245–254.
- Berg, S.J., Illman, W.A., 2011. Three-dimensional transient hydraulic tomography in a highly heterogeneous glaciofluvial aquifer-aquitard system. *Water Resour. Res.* 47, W10507. doi: 10.1029/2011WR010616
- Bersezio, R., Giudici, M., Mele, M., 2007. Combining sedimentological and geophysical data for high-resolution 3-D mapping of fluvial architectural elements in the Quaternary Po Plain (Italy). *Sediment. Geol.* 202 (1–2), 230–248.
- Bohling, G.C., Butler Jr., J.J., Zhan, X., Knoll, M.D., 2007. A field assessment of the value of steady state hydraulic tomography for characterization of aquifer heterogeneities. *Water Resour. Res.* 43, W05430. doi: 10.1029/2011WR010791
- Bohling, G.C., Liu, G., Knobbe, S.K., Reboulet, E.C., Hynde, D.W., Dietrich, P., Butler Jr., J.J., 2012. Geostatistical analysis of centimeter-scale conductivity variations at the MADE site. *Water Resour. Res.* 48, W02525. doi: 10.1029/2011WR010791
- Bunte, K., Abt, S., 2001. Sampling Surface and Subsurface Particle-Size Distributions in Wadable Gravel- and Cobble-Bed Streams for Analyses in Sediment Transport, Hydraulics, and Streambed Monitoring. USFS General Technical Report 74, USDA, Rocky Mtn. Res. Sta. Fort Collins CO.
- Burger, H., Sheehan, A., Jones, C., 2006. Introduction to Applied Geophysics: Exploring the Shallow Subsurface. W. W. Norton and Co., New York, NY.
- Cardiff, M., Barrash, W., Kitanidis, P.K., Malama, B., Revil, A., Straface, S., Rizzo, E., 2009. A potential-based inversion of unconfined steady state hydraulic tomography. *Ground Water* 47 (2), 259–270.
- Chief, K., Ferre, T.P.A., Hinnell, A.C., 2008. The effects of anisotropy on *in situ* air permeability measurements. *Vadose Zone J.* 7, 941–947.

- 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
- Church, M., McLean, D.G., Walcott, J.F., 1987. River bed gravels: Sampling and analysis. In: Thorne, C.R., Bathurst, J.C., Hey, R.D. (Eds.), *Sediment Transport in Gravel-Bed Rivers*. John Wiley and Sons, Chichester, UK.
- Crook, N., Binley, A., Knight, R., Robinson, D., Zarnetsky, J., Haggerty, R., 2008. Electrical resistivity imaging of the architecture of substream sediments. *Water Resour. Res.* 44. doi: 10.1029/2008WR006968
- Dann, R., Close, M., Flintoft, M., Hector, R., Barlow, H., Thomas, S., Francis, G., 2009. Characterization and estimation of hydraulic properties in an alluvial gravel aquifer. *Vadose Zone J.* 8, 651–663.
- Dixon, K.L., Nichols, R.L., 2005. Permeability estimation from transient vadose zone pumping tests in shallow coastal-plain sediments. *Environ. Geosci.* 12 (4), 279–289.
- Fetter, C.W., 2001. *Applied Hydrogeology*, forth ed. Prentice Hall, Upper Saddle River, NJ.
- Fleckenstein, J.H., Fogg, G.E., 2008. Efficient upscaling 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” by Miller, Heeren, Fox, Storm and Halihan, November-December 2011, 49(6): 920–925. *Ground Water* 50 (4), 512–513.
- Fuchs, J.W., Fox, G.A., Storm, D.E., Penn, C., Brown, G.O., 2009. Subsurface transport of phosphorus in riparian floodplains: Influence of preferential flow paths. *J. Environ. Qual.* 38 (2), 473–484.
- Gourry, J.-C., Vermeersch, F., Garcin, M., Girod, D., 2003. Contribution of geophysics to the study of alluvial deposits: A case study in the Val d’Avaray area of the River Loire, France. *J. Appl. Geosci.* 54, 35–49.
- Green, R., Klar, R., Prikryl, J., 2005. Use of Integrated Geophysics to Characterize Paleo-fluvial Environments, *Geotechnical Special Publication 138: Site Characterization and Modeling*. ASCE, New York, NY.
- Haggard, B.E., Moore, P.A., Chaubey, I., Stanley, E.H., 2003. Nitrogen and phosphorus concentrations and export from an Ozark Plateau catchment in the United States. *Biosyst. Eng.* 86 (1), 75–85.
- Halihan, T., Paxton, S., Graham, I., Fenstemaker, T., Riley, M., 2005. Post-remediation evaluation of a LNAPL site using electrical resistivity imaging. *J. Environ. Model.* 7, 283–287.
- 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. Rev.* 32 (1), 43–52. doi: 10.1111/j.1745-6592.2011.01361.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. <http://dx.doi.org/10.1002/rem.21322>
- 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. *Hydrogeol. J.* 21 (7). doi: 10.1007/s10040-013-1034-8
- Heeren, D.M., Miller, R., Fox, G.A., Storm, D.E., Penn, C.J., Halihan, T., 2010. Preferential flow path effects on subsurface contaminant transport in alluvial floodplains. *T. ASABE* 53 (1), 127–136.
- 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., 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. Process.* 28 (3), 1307–1317. doi: 10.1002/hyp.9674
- Herman, R., 2001. An introduction to electrical resistivity in geophysics. *Am. J. Phys.* 69 (9), 943–952.
- 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.
- Jones, K.L., Poole, G.C., Woessner, W.W., Vitale, M.V., Boer, B.R., O’Daniel, S.J., Thomas, S.A., Geffen, B.A., 2008. Geomorphology, hydrology and aquatic vegetation drive seasonal hyporheic flow patterns across a gravel-dominated floodplain. *Hydrol. Proc.* 22, 2105–2113.
- Kalbus, E., Schmidt, C., Molson, J.W., Reinstorf, F., Schirmer, M., 2009. Influence of aquifer and streambed heterogeneity on the distribution of groundwater discharge. *Hydrol. Earth Syst. Sci.* 13 (1), 69–77.
- Kamann, P., Ritzi, R., Dominic, D., Conrad, C., 2007. Porosity and permeability in sediment mixtures. *Ground Water* 45, 429–438.
- 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
- Kittrick, J., Hope, E., 1963. A procedure for the particle-size separation of soils for X-ray diffraction analysis. *Soil Sci.* 96, 319–325.
- Koltermann, C., Gorelick, S., 1995. Fractional packing model for hydraulic conductivity derived from sediment mixtures. *Water Resour. Res.* 13, 3283–3297.
- Lesmes, D.P., Friedman, S.P., 2005. Relationships between the electrical and hydrogeological properties of rocks and soils. In: Rubin, Y., Hubbard, S.S. (eds.), *Hydrogeophysics*. Springer, Dordrecht.
- 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.
- Loke, M.H., Dahlin, T., 2002. A comparison of the Gauss–Newton and quasi-Newton methods in resistivity imaging inversion. *J. Appl. Geophys.* 49, 149–162.
- McNeill, J.D., 1980. *Electrical Conductivity of Soils and Rocks*. Technical Note TN-5. Geonics Limited, Ontario, Canada.
- Menichino, G.T., Ward, A.S., Hester, E.T., 2012. Macropores as preferential flow paths in meander bends. *Hydrol. Proc.* doi: 10.1002/hyp.9573
- 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., 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
- Milsom, J., 2003. *Field Geophysics*, third ed. John Wiley and Sons Ltd., West Sussex, GB.
- 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. *Agric. Ecosyst. Environ.* 141, 417–425.
- Moore, D., Reynolds, R., 1997. *X-Ray Diffraction and the Identification and Analysis of Clay Minerals*, second ed. Oxford University Press, New York, NY.
- Naiman, R.J., Decamps, H., McClain, M.E., 2005. Catchments and the physical template. In: *Riparia: Ecology, Conservation, and Management of Streamside Communities*. Elsevier Academic Press, Boston, MA, pp. 19–48.
- Nimmo, J.R., Rousseau, J.P., Perkins, K.S., Stollenwerk, K.G., Glynn, P.D., Bartholomay, R.C., Knobel, L.L., 2004. Hydraulic and geochemical framework of the Idaho Engineering and Environmental Laboratory vadose zone. *Vadose Zone J.* 3, 6–34.
- NRCS, 2011. United States Department of Agriculture. Soil Survey Geographic (SSURGO) Database for Cherokee and Delaware Counties, OK. (Accessed Apr. 2, 2011) <http://soildatamart.nrcs.usda.gov>

- 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. *Hydrogeol. J.* 18, 1539–1545.
- OSU Office of Intellectual Property, 2004. Halihan/Fenstermaker Proprietary Electrical Resistivity Imaging Method. 2.0 Ed. Oklahoma State University Office of Intellectual Property, Stillwater, OK.
- Poole, G.C., Stanford, J.A., Running, S.W., Frissell, C.A., 2006. Multiscale geomorphic drivers of groundwater flow paths: Subsurface hydrologic dynamics and hyporheic habitat diversity. *J. North Am. Benthol. Soc.* 25 (2), 288–303.
- Popova, Y.A., Keyworth, V.G., Haggard, B.E., Storm, D.E., Lynch, R.A., Payton, M.E., 2006. Stream nutrient limitation and sediment interactions in the Eucha–Spavinaw basin. *J. Soil Water Conserv.* 61 (2), 105–115.
- Poppe, L.J., Paskevich, V.F., Hathaway, J.C., Blackwood, D.S., 2001. A Laboratory Manual for X-Ray Diffraction. USGS OFR 01-041. <http://pubs.usgs.gov/openfile/of01-041/index.htm>
- Rehfeldt, K.R., Boggs, J.M., Gelhar, L.W., 1992. Field study of dispersion in a heterogeneous aquifer 3. Geostatistical analysis of hydraulic conductivity. *Water Resour. Res.* 28 (12), 3309–3324.
- Rey, E.D., Jongmans, D., 2007. A 2D numerical study of the effect of particle shape and orientation on resistivity in shallow formations. *Geophys. Res. Lett.* 34, F9–F17.
- Rey, E., Jongmans, D., Gotteland, P., Garambois, S., 2006. Characterization of soils with stony inclusions using geoelectrical measurements. *J. Appl. Geophys.* 58, 188–201.
- Reynolds, W.D., 2010. Measuring soil hydraulic conductivity properties using a cased borehole permeameter: Steady flow analyses. *Vadose Zone J.* 9, 637–652.
- Samouelian, A., Cousin, I., Tabbagh, A., Bruand, A., Richard, G., 2005. Electrical resistivity in soil science: A review. *Soil Till. Res.* 83, 173–193.
- Slater, L., 2007. Near surface electrical characterization of hydraulic conductivity: from petrophysical properties to aquifer geometries—A review. *Surv. Geophys.* 28 (3), 169–197.
- Smith, R., Sjogren, D., 2006. An evaluation of electrical resistivity imaging (ERI) in Quaternary sediments, southern Alberta, Canada. *Geosphere* 2 (6), 287–298.
- Stanford, J.A., Ward, J.V., 1992. Management of aquatic resources in large catchments: Recognizing interaction between ecosystem connectivity and environmental disturbance. In: Naiman, R.J. (ed.), *Watershed Management*. Springer-Verlag, New York, pp. 91–124.
- Stoeser, D., Green, G., Morath, L., Heran, W., Wilson, A., Moore, D., Van Gosen, B., 2005. Preliminary Integrated Geologic Map Databases for the United States: Central States: Montana, Wyoming, Colorado, New Mexico, North Dakota, South Dakota, Nebraska, Kansas, Oklahoma, Texas, Iowa, Missouri, Arkansas, and Louisiana. USGS Open File Report 2005-1351. <http://pubs.usgs.gov/of/2005/1351/>
- Storm, D.E., Busteed, P.R., Mittelstet, A.R., White, M.J., 2010. Hydrologic Modeling of the Oklahoma/Arkansas Illinois River basin Using SWAT 2005 Final Report. Oklahoma Department of Environmental Quality.
- Sudicky, E.A., 1986. A natural gradient experiment on solute transport in a sand aquifer: Spatial variability of hydraulic conductivity and its role in the dispersion process. *Water Resour. Res.* 22 (13), 2069–2082.
- Telford, W.M., Geldart, L.P., Sheriff, R.E., 1990. *Applied Geophysics*, second ed. Cambridge University Press, New York, NY.
- Tetegan, M., Pasquier, C., Besson, A., Nicoulaud, B., Bouthier, A., Bourenane, H., Desbourdes, C., King, D., Cousin, I., 2012. Field-scale estimation of the volume percentage of rock fragments in stony soils by electrical resistivity. *Catena* 92, 67–74.
- Tye, A., Hessler, H., Ambrose, K., Williams, J., Tragheim, D., Scheib, A., Raines, M., Kuras, O., 2011. Using integrated near-surface geophysical surveys to aid mapping and interpretation of geology in an alluvial landscape within a 3-D soil-geology framework. *Near Surf. Geophys.* 9, 15–31.
- U.S. Bureau of Reclamation (USBR), 1985. *Ground Water Manual: A Water Resources Technical Manual, Revised Reprint*. U.S. Department of the Interior, Bureau of Reclamation, Denver, CO.
- Ward, A.S., Gooseff, M.N., Singha, K., 2010. Imaging hyporheic zone solute transport using electrical resistivity. *Hydrol. Proc.* 24 (7), 948–953.
- Ward, A.S., Gooseff, M.N., Singha, K., 2012. How does subsurface characterization affect simulations of hyporheic exchange? *Ground Water*. doi: 10.1111/j.1745-6584.2012.00911.x
- Xiang, J., Scanlon, B.R., Mullican III, W.F., Chen, L., Goldsmith, R.S., 1997. A multistep constant-head borehole test to determine field saturated hydraulic conductivity of layered soils. *Adv. Water Res.* 20 (1), 45–57.
- Zhang, Z., Ward, A., Keller, J., 2011. Determining the porosity and saturated hydraulic conductivity of binary mixtures. *Vadose Zone J.* 10, 313–321.
- Zohdy, A.A.R., Eaton, G.P., Mabey, D.R., 1974. Application of surface geophysics to ground-water investigations. In: *Application of Surface Geophysics to Ground-water Investigations of the United State Geological Survey*. Book 2 (chapter D-1).