

FROM GRAIN TO FLOODPLAIN: EVALUATING HETEROGENEITY OF FLOODPLAIN
HYDROSTRATIGRAPHY USING SEDIMENTOLOGY, GEOPHYSICS,
AND REMOTE SENSING

by
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ABSTRACT

Floodplain stratigraphy, a major structural element of alluvial aquifers, is a fundamental component of floodplain heterogeneity, hydraulic conductivity, and connectivity. Watershed-scale hydrological models often simplify floodplains by modeling them as largely homogeneous, which inherently overlooks natural floodplain heterogeneity and anisotropy and their effects on hydrologic processes such as groundwater flow and transport and hyporheic exchange. This study, conducted in the East River Basin, Colorado, USA, combines point-, meander-, and floodplain-scale data to explore the importance of detailed field studies and physical representation of alluvial aquifers. We combine sediment core descriptions, hydraulic conductivity estimates from slug tests, ground-penetrating radar (GPR), historical maps of former channels, LiDAR-based elevation and Normalized Difference Vegetation Index data to infer 3-D fluvial stratigraphy. We compare and contrast stratigraphy of two meanders with disparate geometries to explore floodplain heterogeneity and connectivity controls on flow and transport. We identify buried point bars, former channels, and overbank deposits using GPR, corroborated by point sediment descriptions collected during piezometer installment and remotely sensed products. We map heterogeneous structural features that should control resultant flow and transport; orientation and connectivity of these features would control residence times important in hydrologic models.

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CHAPTER 1

INTRODUCTION

Rivers represent a dynamic equilibrium, constantly balancing water and sediment inputs and outputs, resulting in erosional and depositional processes and river migration. Meandering rivers transport and deposit sediment and in doing so, spatially organize materials, generally into discrete sediment packages, such as point-bar, overbank, and channel-fill deposits. As rivers migrate and change form, they create new sediment deposits and bedforms while abandoning others. Former bedforms provide a structural element to alluvial aquifers and packages of former floodplain sediment (strata) create natural heterogeneity and anisotropy at the floodplain scale (Figure 1.1; Miall 1996; Todd 1980; Van Den Berg & De Vries 2003).

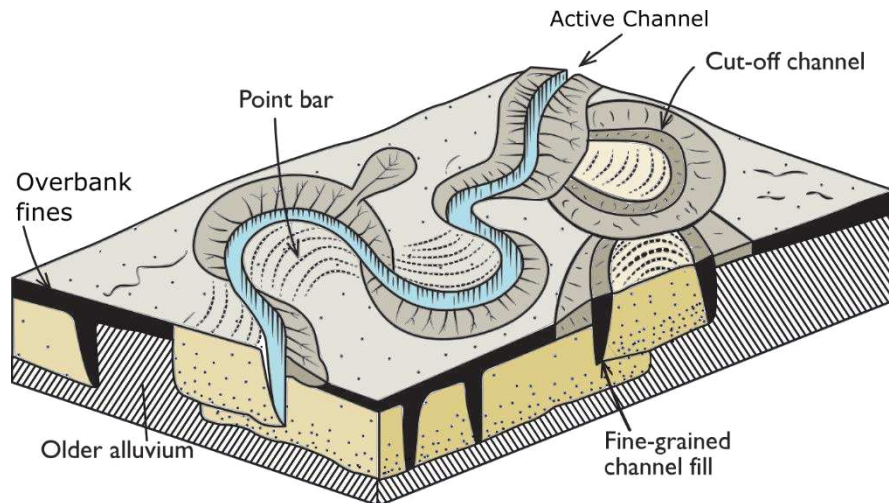


Figure 1.1: Cartoon depicting river migration, abandonment of former bedforms, and incorporation of fluvial strata into the alluvial aquifer. Note variability of homogeneity and heterogeneity within the sediment architecture.

Floodplain strata play a crucial role in the river corridor, because stratigraphic packages can create spatial variations in hydraulic conductivity, influencing subsurface solute and contaminant fate and transport (e.g. Gelhar & Axness 1983; Dagan 1988; Koltermann & Gorelick 1996; Fogg et al. 2000; Heeren et al. 2010), surface-

groundwater exchange processes (e.g. Poole et al. 2006; Jones et al. 2008; Buffington & Tonina 2009; Krause et al. 2013), river migration (e.g. Güneralp & Rhoads 2011; Motta et al. 2012), and connectivity between the river and surrounding aquifer (e.g. Pringle 2003; Freeman et al. 2007; Jencso et al. 2009; Argiroff et al. 2017). Importantly, stream-aquifer connectivity controls the movement and transport of solutes through the river corridor (e.g. Bencala 1993; Harvey & Fuller 1998; Battin et al. 2008; Boehlke et al. 2009; Miller et al. 2014; Savoy et al. 2017), affecting both stream and riparian health (e.g. Findlay 1995; Brunke & Gonser 1997; Boulton et al. 1998). This connectivity of hydrofacies—stratigraphic facies with internally consistent hydraulic properties—is an important attribute of geologic heterogeneity within the floodplain (Savoy et al. 2017). Connected high hydraulic conductivity sediment paths—preferential flow paths—can increase transport velocities and limit solute sorption processes in the alluvial aquifer (e.g. Fuchs et al. 2009; Heeren et al. 2010; Miller et al. 2014). Capturing the extent, orientation, and connectivity of floodplain strata in studies is key to evaluating physical and chemical hydrologic processes within the river corridor (e.g. Tockner et al. 1999; Larsen et al. 2012; Stone et al. 2017).

Despite the influence of floodplain stratigraphy and connectivity of hydrofacies on hydrologic processes in stream-aquifer systems, many hydrologic models simplify floodplain heterogeneity due to difficulties related to hydrogeomorphic field data collection, particularly at ecologically significant scales (Harvey & Gooseff 2015). For example, geostatistical approaches of mapping hydrofacies (e.g. Weissmann et al. 1999; Deutsch and Tran, 2002; dell’Arciprete et al. 2012; Modis & Sideri 2013; Perulero Serrano et al. 2014) do not necessarily incorporate 3-D sediment architecture of fluvial

deposits (Anderson et al. 1999) and do not always capture realistic geological heterogeneity (Savoy et al. 2017). Additionally, hydrologic models at the river-network scale often resort to permeability estimates of unconsolidated materials related to median grain sizes to represent alluvial aquifers (e.g. Boano et al. 2006; Revelli et al. 2008; Kiel & Cardenas 2014; Gomez-Velez et al. 2015). This simplification inherently reduces floodplains to homogeneous substrates and does not capture natural heterogeneity and connectivity of floodplain strata that might be important to predicting processes of interest. For example, the ecological impact of hyporheic exchange is attributed to the residence time of water in the hyporheic zone, as well as the number of turnover lengths per unit river (e.g. Gomez-Velez & Harvey 2014). At the channel-scale, heterogeneity of floodplain hydraulic conductivity can either facilitate preferential flow paths of hyporheic flow or can drive surface water exchanged in the floodplain back into the river channel (e.g. Tonina and Buffington, 2009). Both mechanisms ultimately shorten residence time of water in the hyporheic zone (e.g. as reviewed by Tonina and Buffington, 2009). If a singular grain size and homogeneous hydraulic conductivity are used to represent entire reaches of floodplains in hyporheic exchange models, these small-scale but frequent exchanges might be overlooked, ultimately underestimating ecologically significant hyporheic exchange. To bridge the divide between natural floodplain complexity and effective representation of floodplain hydrofacies in models, Savoy et al. (2017) recommended simplified facies models aimed at capturing important characteristics such as connectivity.

An open question is what methods exist to best estimate sediment facies and their connectivity at larger, ecologically significant scales (i.e. watershed and river

network- scale). Geomorphic field studies of fluvial stratigraphy commonly employ sediment core descriptions and geophysical methods to map sediment packages across floodplain and then use historical imagery (maps, aerial photographs, etc) to connect the strata to larger scale river migration. Ground-penetrating radar (GPR) is a common geophysical method used to map contacts between disparate sediment packages in alluvial systems (e.g. Jol & Smith 1991). GPR emits and receives electromagnetic waves, which reflect off differences in electromagnetic impedance of surveyed materials, or in non-magnetic materials, the dielectric permittivity. In fluvial settings, disparities in dielectric permittivity occur at the air-ground interface, changes in moisture, and contacts between different sediment compositions (Bridge 2009; Annan 2009). Radar facies—areas with similar GPR reflection characteristics from subsurface features—have been commonly used in fluvial environments to interpret fluvial stratigraphy and subsurface sediment architecture (e.g. Vandenberghe & Van Overmeeren 1999; Ekes & Hickin 2001; Skelly et al. 2003; Kostic & Aigner 2007; Słowik et al. 2016). Facies are classified by elements such as reflection amplitude, continuity, geometry, and degree of penetration (Van Overmeeren 1998).

To give GPR transects and interpreted fluvial stratigraphy broader geomorphic context, the data can be compared to former channel locations based on historical images, maps, or satellite imagery (e.g. Poole et al. 2002; Słowik 2016). However, these remotely sensed data forms can be both spatially and temporally limited, particularly in remote regions, and generally do not capture channels abandoned prior to when the datasets were collected. An alternative approach to mapping former channels is to leverage vegetative and elevation disparities between abandoned

channels and the surrounding floodplain. Reoccurring disturbance of abandoned channels caused by reoccupation during flooding, and the gradual nature of vegetative colonization create contrasting vegetation densities and distributions between former channels and the adjacent floodplain (Poole et al. 2002; Greco et al. 2007; Bätz et al. 2016). Recently, remote multispectral data have been employed to identify vegetative succession in fluvial systems (e.g. Hamilton et al. 2007). Vegetation differences between established and colonizing vegetation, as well as bare and wetted soils, can be captured in Normalized Difference Vegetation Index of multispectral data (NVDI; Bertoldi et al. 2011). NDVI is the ratio of the difference between near-infrared (NIR) and red wavelength bands (RED) and the sum of the wavelengths, and is sensitive to photosynthetic activity of plants, as well as bare soil, and standing water (e.g. reviewed in Xie et al. 2008). Additionally, persistent spatial mosaic patterns develop naturally during vegetative succession, as different plant species adapt to fluvial morphodynamics, such as erosion, inundation, and groundwater depths (Egger et al. 2015). The progressive nature (Poole et al. 2002; Greco et al. 2007; Bätz et al. 2016) and patch dynamics (Latterell et al. 2006) of vegetative succession allows for the use of vegetation distributions to infer hydrogeomorphic features, such as former channels or gravel bars, that may not be captured by historical imagery alone (Gurnell et al. 2003; Greco et al. 2007).

In addition to vegetative differences between the abandoned channels and the surrounding floodplain, there are long-term elevation disparities. Flooding fills abandoned channels with fine sediment and slowly raises the abandoned channel elevation to that of the surrounding floodplain (Toonen et al. 2012). In the interim,

former channels are generally lower than the adjacent floodplain and can exhibit surface depression and ponded water features, which LiDAR and NDVI can illuminate, respectively.

The study presented here maps 3-D sediment architecture and the spatial extent of and relations between fluvial deposits to distill floodplain strata into packages related to current and former channel features in the floodplain. The goal of this multiscale study is to present a framework to bridge the divide between detailed field studies and large-scale physical representation of alluvial aquifers, and explore the importance of these detailed data on hydrology in a highly meandering river system in a montane floodplain of the East River in the Rocky Mountains, Colorado, USA. The East River is an ideal natural laboratory for this work, because the river is actively meandering, allowing for observable connections between recent floodplain evolution and sedimentology, at a scale between meander-scale field investigations and basin-scale mapping efforts. We relate sediment core descriptions, estimates of hydraulic conductivity from slug tests, GPR, and maps of abandoned channels based on historical imagery within the current floodplain to infer 3-D river meander stratigraphy in the East River Basin. In addition to tracking former river migration using historical photography, we expand the catalogue of former channel locations by combining light detection and ranging (LiDAR), National Agriculture Imagery Program (NAIP) images, and WorldView-2 (WV-2) 8-band multispectral data, the latter of which we use to estimate NDVI. One-dimensional lateral transect studies of fluvial succession have used elevation and land cover disparities to identify former channels (Greco et al. 2007), but to our knowledge, our study is the first to corroborate elevation and vegetation signatures to map

abandoned channels across a floodplain. We compare and contrast stratigraphy of two meanders with disparate geometries to estimate floodplain heterogeneity and strata connectivity using surficial features, such as former channel locations and meander geometry, and consider hydrological impacts of assigning representative soil characteristics, such as hydraulic conductivity, to inferred hydrofacies.

CHAPTER 2

STUDY AREA DESCRIPTION

2.1 Upper East River Basin

The East River flows southeast through a subalpine valley near Mount Crested Butte in the West Elk Mountain Range, Colorado, USA. The drainage basin ranges in elevation from 4090 m at the headwaters to 2440 m at the East River's confluence with the Taylor River, forming the Gunnison River. Our study focuses on the upper portion of

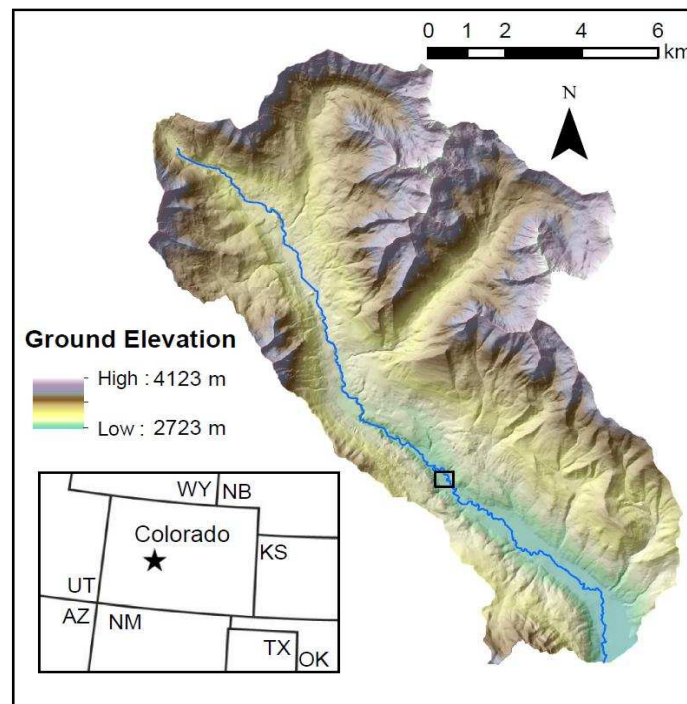


Figure 2.1: Upper portion of the East River basin and shaded topography. Black box denotes study area within the basin.

the East River, from the headwaters to the confluence with Brush Creek (Figure 2.1).

This portion of the valley's surficial geology has been heavily impacted by alpine glaciation related to the Last Glacial Maximum. The surficial deposits that comprise the East River's sediment source vary downstream, from rock glaciers, talus, and landslide

deposits in the headwaters, to lateral moraine deposits flanking the lower gradient floodplains downstream. Localized areas of Cretaceous Mancos Shale are exposed in-channel and exhibit varying degrees of metamorphism (Gaskill et al. 1991). The channel morphology varies greatly along the river's path, alternating between sinuous, unconfined reaches to straighter, more confined, channelized flow. The sinuosity of the East River channel in the upper catchment (shown in blue in Figure 2.1) is 1.8. Sinuosity is calculated by dividing the longitudinal length of the river along the midline by the shortest distance between apexes at the meander neck (Rosgen 1996).

The upper East River valley, typical of other montane river valleys in Colorado (e.g., Andrews 1984; Mueller & Pitlick 2005; Livers & Wohl 2015), has been greatly impacted by Pinedale glaciation. Previously glaciated montane systems are characterized by shallow down-valley gradients (Wohl 2004) inherited by alpine glacial processes (Brardinoni & Hassan 2007; Livers & Wohl 2015); heterogeneous floodplain sediments sourced from landslides and moraine deposits (Brardinoni & Hassan 2007; Livers & Wohl 2015); and channel morphologies that reflect previous climates and sediment supplies (Andrews 1984; Wohl 2004; Livers & Wohl 2015). Similar to other Colorado rivers that source former glacial deposits in snowmelt-dominated systems (Andrews 1984; Wohl 2004; Livers & Wohl 2015), the East River floodplain is comprised of heterogeneous sediments, from silts and sands to large gravels and boulders.

The East River's hydrology is a snowmelt-dominated system, consisting of high flows in late spring and consistently low flows beginning in late summer and continuing through winter and early spring. There is currently no USGS gauge along the upper

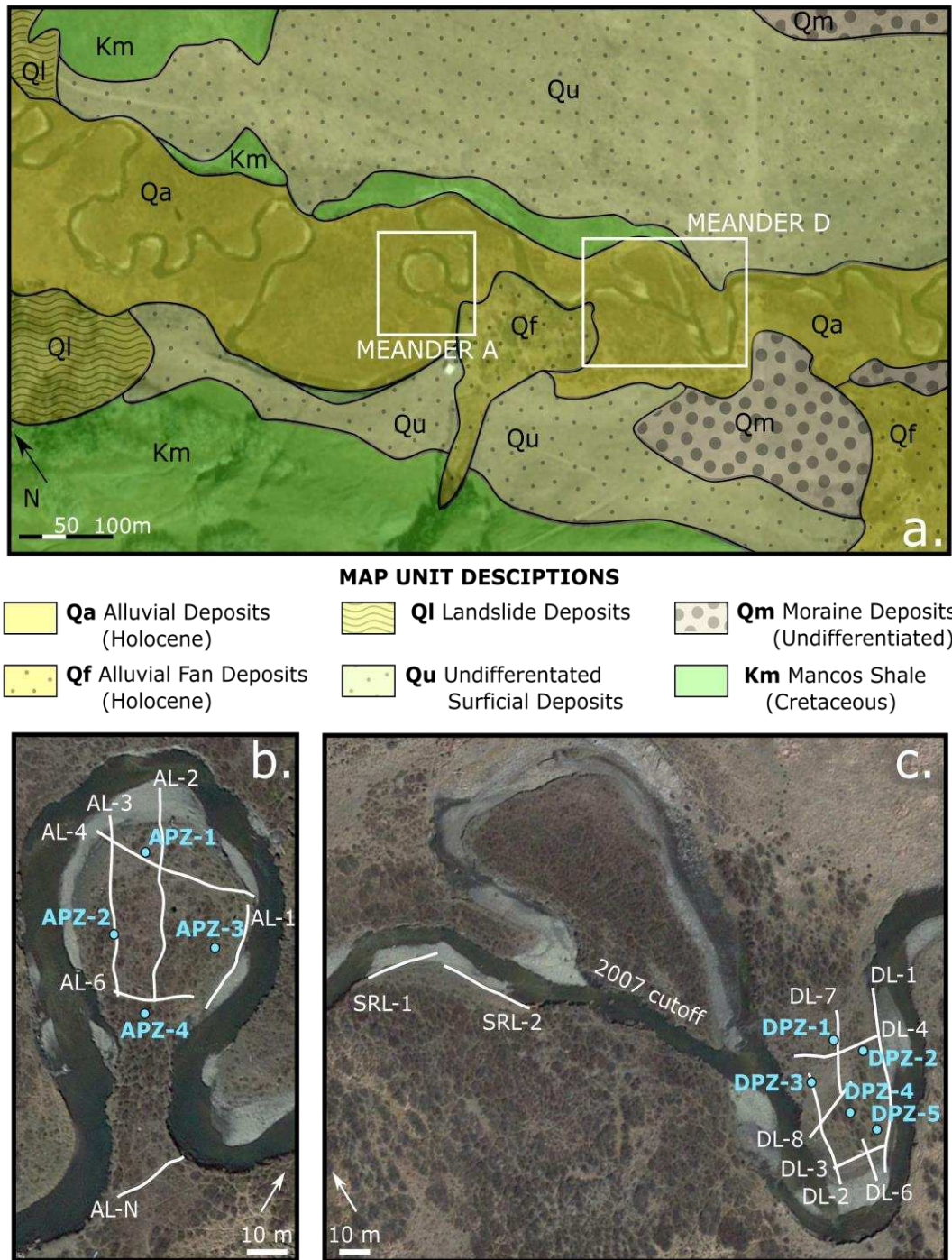


Figure 2.2: Field study area with surficial geology. a) Study site with overlay of geology. Geology modified from Gaskill et al. (1995). b) Locations of GPR transects (white lines) and piezometers (blue dots) at Meander A. c) Locations of GPR transects and piezometers near Meander D and the recent cutoff. “A”, “D”, and “SR” denote Meander A, Meander D, and straight reach, respectively. “PZ” indicates piezometer, and “L” indicates a GPR line. Sediment core descriptions and slug test data are associated with each piezometer.

portion of East River; however, for water year 2015, mean daily flows, measured with a stilling well at the study site, ranged from 1.0-14.0 m³s, (Winnick et al. 2017). Because of snowmelt-driven peak discharges, the East River, like other gravel-bed rivers in Colorado, achieve near-bankfull discharge on multiple days during the year, during which the majority of bed mobilization occurs (Andrews 1984). Downstream, near Almont, the East River was found to achieve bankfull discharge over 20 days per year (Andrews 1984).

2.2 Detailed Study Area

This study's field location is near Crested Butte Mountain Resort's pump house (38.99219 N, 106.94854 W), ~5.5 km downstream of the river's headwaters. Along the study area, the river floodplain is flanked to the northeast and southwest by lateral moraines and small outcrops of Mancos shale exposed along the channel (Figure 2.2a; Gaskill et al. 1991). The southwest valley wall is steep and includes landslide and alluvial fan deposits (Gaskill et al. 1991). The valley width along the field site is ~150 m.

Piezometer networks, sediment descriptions, water level data, and GPR surveys (described in Methods, below) and floodplain sediment descriptions were focused on two meanders along an actively migrating portion of the East River, herein referred to as "Meander A" and "Meander D", and additional GPR data was collected near a channel chute cutoff that formed ~10 years ago (Figure 2.2c). Historical aerial photographs reveal cutoff initiation in approximately 2007, with major flow diversion to the chute occurring approximately in 2012 (herein referred to as the 2007 cutoff). Piezometer networks consisted of four and five piezometers at Meander A and Meander D, respectively. The piezometers provide data on groundwater elevations and sediment

types across both meanders. We co-located GPR transects with piezometers for subsurface control on strata and water table elevations. We conducted two additional GPR surveys near the 2007 cutoff, but there are no piezometers at this location.

Meanders A and D were chosen for this study for their contrasting geometries as well as locations respectively upstream and downstream of the 2007 cutoff. Both meanders' active channels contain heterogeneous bed sediments, but meander A's range from silt- to cobble-sized while Meander D's span from silt- to boulder-sized. Graminoids, forbes, and willows (*Salix sp.*) dominate floodplain vegetation along the field location (Harte & Shaw 1995). Meander A has a wavelength of ~30 m, and a sinuosity of 6.3, while Meander D has wavelength of ~60 m and a sinuosity of 2.6. The average water slope of both locations is 0.003, as measured in late summer 2016 by both a field survey using a high precision Trimble GPS unit and from a digital elevation map created from a drone survey (Pai et al., 2017).

A key difference between the two meanders is their position within the context of the floodplain's other geomorphic features. Meander A, and the two meanders immediately downstream of Meander A, are oriented cross-valley, while Meander D's orientation is oriented down-valley (Figure 2.2a). Although both studied meanders are immediately downstream of relatively straight reaches, Meander A is located downstream of a channel reach incising valley-wall bedrock and just up-valley from an alluvial fan deposit. Meander D is located directly downstream of the 2007 cutoff, and up-valley of a meandering reach (Figure 2.2a). Flooding of the 2007 cutoff still occurs during peak discharge, and late summer monsoons can create depression storage in this low-lying feature.

CHAPTER 3

METHODS

3.1 Piezometer Installation and Sediment Descriptions

We installed piezometer networks in Meander A and Meander D during July 2016 and July 2015, respectively, and designed the configurations to create spatial coverage for water levels and sediment core data in each meander (Figure 2.2b and 2.2c). We drilled piezometers by hand auger and backpack drill combined; both drill bits were approximately 6.4 cm in diameter. We measured ground surface elevations at each piezometer using high-precision Trimble GPS units and a Topcon GPT-8200A auto-tracking pulse total station and Real Time Kinetic GPS. We recovered sediment cores using an auger, and sediment textures were described using standard National Resources Conservation Service (NRCS) soil texture classification methods (Thein

Table 3.1: Piezometer details for networks at Meanders A and D (figures 2.2b, 2.2c)

Piezometer Name	UTM-N [m]	UTM-E [m]	Ground Elevation [m]	Depth [m]	screen length [m]
APZ-1	4309873.23	330920.2301	2774.343	0.62	0.475
APZ-2	4309852.056	330922.5466	2774.153	0.43	0.35
APZ-3	4309861.628	330945.7063	2773.969	1.06	0.762
APZ-4	4309830.465	330941.6139	2774.307	0.80	0.7
DPZ-1	4309847.459	331196.257	2756.746	0.90	0.762
DPZ-2	4310060.674	330990.808	2756.381	0.60	0.762
DPZ-3	4309843.797	331185.165	2756.81	0.85	0.762
DPZ-4	4309827.884	331193.337	2756.61	1.25	0.762
DPZ-5	4309818.749	331198.819	2756.416	1.00	0.762

1979). Piezometer casing consisted of 5.08 cm inside-diameter, 6.03 cm outside-diameter PVC and had screens of 2-mm slotted PVC. Screened intervals ranged between 35 and 76.2 cm, ending at the base of the wells (Table 3.1). We developed the piezometers by flushing and pumping, and added sand packs as needed; gaps between the well casing and borehole were small (<0.5 cm).

3.2 GPR Data Acquisition and Processing

We used GPR to image the distribution of floodplain sediments and their connectivity at the two meanders of interest as well as near the 2007 cutoff (SR-1 and SR-2; Figure 2.2c). We strategically co-located GPR transect grids with shallow piezometers at meanders A and D (Figure 2.2b and 2.2c, respectively) where we recorded water-table depths and sediment core descriptions at each piezometer for subsurface control. We also positioned the straight reach GPR transects near a current point bar and the 2007 cutoff to capture radar facies of known features within the floodplain. We recorded all GPR transects with a PulseEKKO™ Pro system by Sensors & Software Inc. using 100 MHz antennae (Davis & Annan 1989). The transmitter and receiver were attached to a sled at a fixed separation of 0.50 m and dragged slowly across each transect. We collected common-offset measurements with consistent spacing of 0.250 m between antennae, activated by an odometer wheel attached to the sled. To help remove ambient electromagnetic noise, all measurements were collected using 8 stacks per sample.

We processed all data using EKKO_Project software by Sensors & Software Inc. All data were had a time-zero correction and then were 'dewowed' to remove low-frequency noise caused by inductive coupling effects or dynamic range limitations of the equipment (Annan 2009). We applied a Spherical Exponential Calibrated Compensation (SEC2) time gain to all transects to compensate for signal losses due to spherical spreading and exponential energy attenuation (Annan 2009). The SEC2 gain preserves the relative amplitude information of reflections at various depths. Transects were collected over relatively flat ground, with any irregularities in the ground surface

noted. Table A.1 presents the start gain, attenuation, and maximum gains used for the respective transects.

We combined processed GPR data with depth measurements of floodplain sediments based on the sediment cores to identify the physical origin of reflectors and change time-based information to depth. In Meander A, sedimentary data from four piezometers, ranging from 0.4-1.1 m (Table 3.1), were compared to GPR data. In Meander D, sediment data from five piezometers, ranging from 0.6-1.25 m, were used. Our radar facies assignments draw from previously documented GPR signatures (e.g. Vandenberghe & Van Overmeeren 1999; Kostic & Aigner 2007; Bridge 2009; Miall 2014; Slowil 2016), and facies configurations and dimensions are compared to current visible features along the East River, including channels and point bars.

3.3 Hydraulic Conductivity and Linear Velocity Estimates

To estimate hydraulic conductivity (K), we performed a series of falling-head slug tests at each piezometer (5 tests per well) using 1 L water slugs. Methods for processing data followed that of Hvorslev (1951) for fully-submerged well screens and of Binkhorst & Robbins (1998) for wells partially submerged well screens. Hydraulic conductivity is estimated by

$$K = \frac{r^2 \ln(L/R)}{2 L t_{37}} \quad (\text{Equation 3.1})$$

where K is the hydraulic conductivity [L^3/T], r [L] is the radius of the inside of the well casing, R [L] is the radius of the borehole, L [L] is the length of the well screen, and t_{37} [T] is the time it takes the water level to recover to 37% of the initial change (Hvorslev

1951). We only used data from the mid-time, log-linear portion of slug tests to estimate local subsurface K (Binkhorst & Robbins 1998). Additionally, for partially submerged wells, we substituted effective casing radii (r_e) and screen lengths (L_e ; the length submerged under static water table conditions) for their respective counterparts in Equation 3.1. A representative effective casing radius value for our piezometers was calculated by

$$r_e = \sqrt{r^2 + S_y(R^2 - r^2)} \quad (\text{Equation 3.2})$$

where S_y [-] is the specific yield of the sandpack (Binkhorst & Robbins 1998). We utilized a representative S_y of 0.21 for the fine-medium grained sand (Johnson 1963) to pack the wells. The resulting effective casing radius value used was 2.66 cm.

To calculate water table gradients across each meander we calculated three-point problems given water table elevations recorded at piezometers. We calculated water table gradients using multiple well configurations across each meander. Water table elevation data, code, and well configurations used to calculate water table gradients' direction and magnitude are in Appendix 6. By incorporating the hydraulic conductivity data above we estimated the average linear velocity across the meanders using Darcy's Law. We did not measure porosity directly so we used an average porosity value of 0.25. To estimate lateral hyporheic residence times across the meander, we simply divided the estimated flow path lengths across the meander, based on a straight-line distance along the maximum hydraulic gradient, by the calculated average linear velocity.

3.4 Paleochannel Mapping

We identified abandoned channels in the East River floodplain remotely using a combination of historical photography, LiDAR, and WV-2 8-band multispectral data. LiDAR and WV-2 were collected along the East River in 2015, and the WV-2 data were used to calculate NDVI. For consistency, we refer to abandoned channels identified in using LiDAR and NDVI but not in historical images as “paleochannels”. To estimate paleochannels, we used an object-oriented image analysis software package, eCognition™ by Trimble, to classify former channels using both vegetative and physical indicators, including NDVI signatures; presence of ponded water signatures (oxbows); proximity to the current river channel; and elevation relative to surrounding floodplain. We generated image-objects by merging pixels possessing similar parameter values and assigned each object to a designated class of floodplain features (such as “channel” and “surrounding floodplain”). The parameters with the greatest success of automatically identifying former channels were differences in NDVI and acute disparities in elevation captured in the LiDAR. To refine delineations of abandoned channels based on NDVI and LiDAR-based elevation, we compared our estimates to U.S. Geological Survey, U.S. Department of Agriculture Forest Service, NAIP, and historical aerial photographs of the floodplain dating back to 1955. These comparisons allowed for careful corroboration of channels abandoned after 1955, as well as an evaluation of the spatial coverage, compared to historical images alone.

CHAPTER 4

RESULTS

4.1 Sediment Characteristics

The geomorphic differences between the two studied meanders, A and D, begin at the channel geometry and continue to the subsurface with sedimentology and sediment structure. Meander A has a “textbook” sinusoidal and bilaterally symmetrical planform (Figure 2.2b). Our sediment characterization from cores indicates a consistent sediment structure across the meander: floodplain stratigraphy consists of a 0.3-0.7m layer of surficial overbank fines (silt and clay), deposits that accumulate during high flows, underlain by heterogeneous deposits containing gravels (Figure 4.1).

Alternatively, asymmetrical Meander D tapers from a wide base to an apex thinner than Meander A. Meander D’s sedimentology is composed of various types of sediment packages, including gravels, sands and pebbles, and fines; these appear in cores

Table 4.1: Summary of hydraulic conductivity estimates and screened sediment sizes. K data not available for APZ-2.

Piezometer Name	Screened Sediments	K (m/d)	Std Dev (m/d)
APZ-1	gravels, sand, and fines	0.8	4E-02
APZ-2	gravels, sand, and fines	--	--
APZ-3	gravels, sand and pebbles, and fines	2	5E-02
APZ-4	gravels, sand and pebbles, and fines	0.6	5E-02
DPZ-1	gravels, sand and pebbles, and fines	0.3	3E-02
DPZ-2	gravels and fines	0.2	4E-02
DPZ-3	sand and pebbles	4	3E-02
DPZ-4	sand and fines	0.3	5E-02
DPZ-5	sand and fines	3	3E-01

locally rather than extensively across the meander (Figure 4.1). Surficial overbank sediments are thinner (20-50 cm) in sediment cores at Meander D than Meander A.

Unlike Meander A, gravels in Meander D are encountered only in the two cores farthest from the meander apex (DPZ-1 and DPZ-2), and the remaining three sediment cores

contain only layers of fines, fine to coarse sand, and pebbles. In short, meander D is notably more heterogeneous than Meander A.

These differences in sedimentology do not appear to translate to a large range of K, measured via slug tests. The K across both meanders were similar; the geometric means of K values for Meander A and Meander D are 1 and 0.7 m/d, respectively (Table 4.1). Meander A's K values differed from one another by less than an order of magnitude, and Meander D's by just over an order of magnitude. The low variance in K values may be attributed to the fact that the piezometers' screened intervals cover multiple sediment types in each well such that the measured K values more likely reflect an effective average of the screened sediments, rather than an estimate of one discrete sediment package. Additionally, slug tests can be biased by altered, low-K near the well

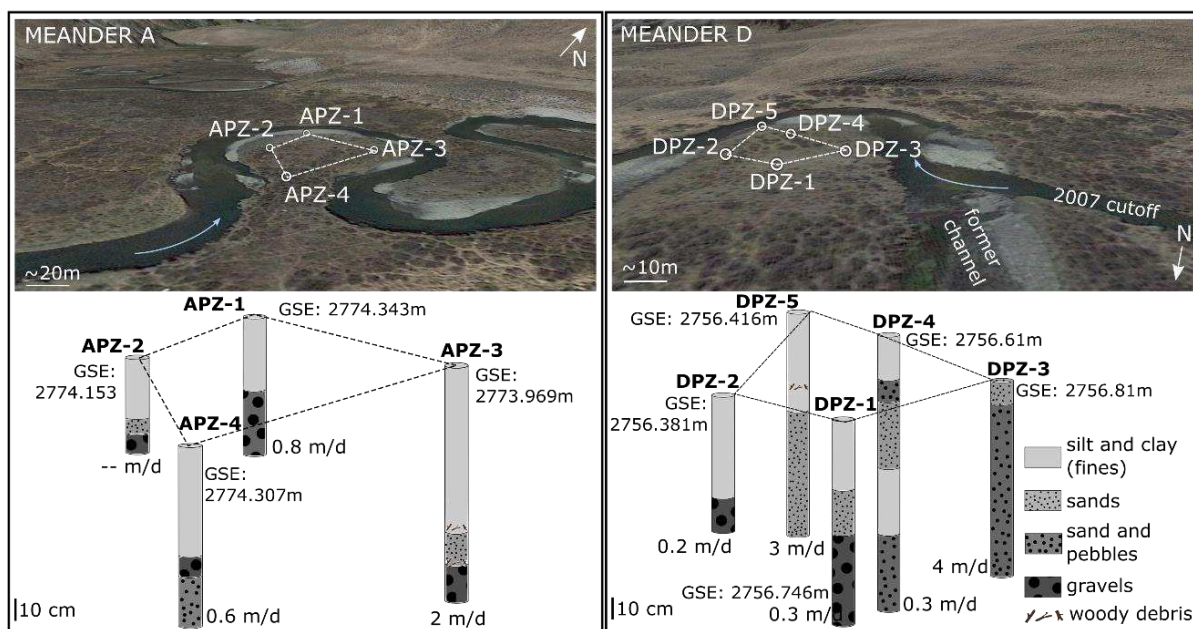


Figure 4.1: Sediment cores and hydraulic conductivity estimates. Oblique views looking down each meander. Depths are to scale. Core widths are not to scale to show details. Hydraulic conductivity is displayed in m/d. K data are not available for APZ-2 due to equipment failure. Ground surface elevation (GSE) is in meters above sea level.

(Butler & Healey 1998; Rovey & Niemann 2001), and because they do not stress the aquifer, produce smaller effective test areas with lower K estimates (Rovey & Cherkauer 1995; Rovey & Niemann 2001) and lower variance (Rovey & Cherkauer 1995; Bohling et al. 2012) than other methods, such as direct push or pumping tests. Therefore, the K values reported here may serve as a rough estimate for the East River meanders, but cannot conclusively characterize the different sediment facies without additional and alternative K measurements.

4.2 Water Table Gradients and Average Linear Velocities

The direction of the hydraulic gradients in both meanders differ between the snow melt-dominated flow regimes of late spring/early summer and the near-baseflow conditions of later summer. At Meander A, although we used multiple well configurations in our three-point gradient estimates, the gradient direction was unidirectional and of similar magnitude (Figure 4.2; Appendix 6). At Meander A, early season groundwater flow in the meander is directed cross-valley. River discharge flows cross-valley at Meander A and is greatest in early summer, along with snowmelt inputs from valley walls. Both increased discharge and snowmelt inputs potentially create the cross-valley flow across Meander A. Late in the season, concurrent with the absence of snowmelt and large amounts of precipitation, baseflow is likely a large contribution to river discharge (Winnick et al. 2017). The dominant base flow direction at this time is down-valley gradient. Across the three snapshots in time, the gradient across Meander A was unidirectional and ~ 0.01 , but flow direction varied (Figure 4.2).

At Meander D, the dominant water table gradient direction is down-valley, but there is variability in the water table across the meander (Figure 4.2). This variability is

greatest in the early season, as depicted by bidirectional flow across the meander: This may be related to the surface water flow concentrated toward the back portion of Meander D (near DPZ-3) by the 2007 cutoff. Later in the season, the largest gradient is towards the apex of the meander (0.02). The largest gradient observed across Meander D (0.03) occurs in early summer, concomitant with peak snowmelt and surface discharge.



Figure 4.2: Water table elevations (in color, meters above sea level) and gradient estimates for the two meanders during late spring (left), early summer (middle), and later summer (right), 2017. Meander A (top row) has its meander apex towards the top of the images, while Meander D has its apex towards the bottom of the image. White dashed lines are the raw calculated water table gradients. Blue lines indicate estimated flow paths across the meander. Dashed black lines are the boundaries of estimated channel sediments.

Changes in the hydraulic gradient obviously affect linear velocity estimates and residence times of water flowing across the meander. The gradients' direction also influences the flow path length estimates across the meander, which vary through the summer season across both meanders (Figure 4.2; Table 4.2). The shortest lateral hyporheic residence time estimates (500 days) occur across Meander D during June and July, associated with the greatest gradient (0.09; June) and the shortest flow path length (30 m; July).

Table 4.2: Estimates of flow path lengths, average linear velocities, and residence times. Flow path lengths follow those presented in figure 4.2, with the blue lines indicating the flow path lengths. Of the blue lines in Meander D on figure 4.2, the top line is the “back” flow path, and the bottom line is the “apex” flow path. Average linear velocities are calculated using geometric means of K estimates. Geometric means for Meander A and Meander D are 1 and 0.7 m/d, respectively.

using $AVG_{geo} K$	June July A	Aug A	June D (Back)	June D (Apex)	July D (Back)	July D (Apex)	Aug D (Back)	Aug D (Apex)
flow path length [m]	50	40	45	30	45	30	40	30
v [m/d]	0.04	0.04	0.09	0.03	0.06	0.06	0.03	0.03
t_H [d]	1300	960	500	1000	800	500	1400	1000

4.3 Radar Facies

Four distinct fluvial radar facies are mapped in our datasets: a) former channels; b) lateral accretion structures; c) point bars; and d) heterogeneous gravel layers (Figure 4.3). The recognized radar facies are well-documented to be present in meandering river settings (Miall 2014; Bridge 2009) and outlined in detail here.

Former channels are indicated by a bright, concave-up bottom reflector, overlain by less-distinct, sub-horizontal layers (Figure 4.3a; Van Overmeeren 1998; Vandenberghe & Van Overmeeren 1999; Ekes & Hickin 2001; Skelly et al. 2003; Słowik 2016). The concave-up reflector demarcates the channel bottom, and the stark reflection at the channel base occurs due to the juxtaposition of coarser-grained bed

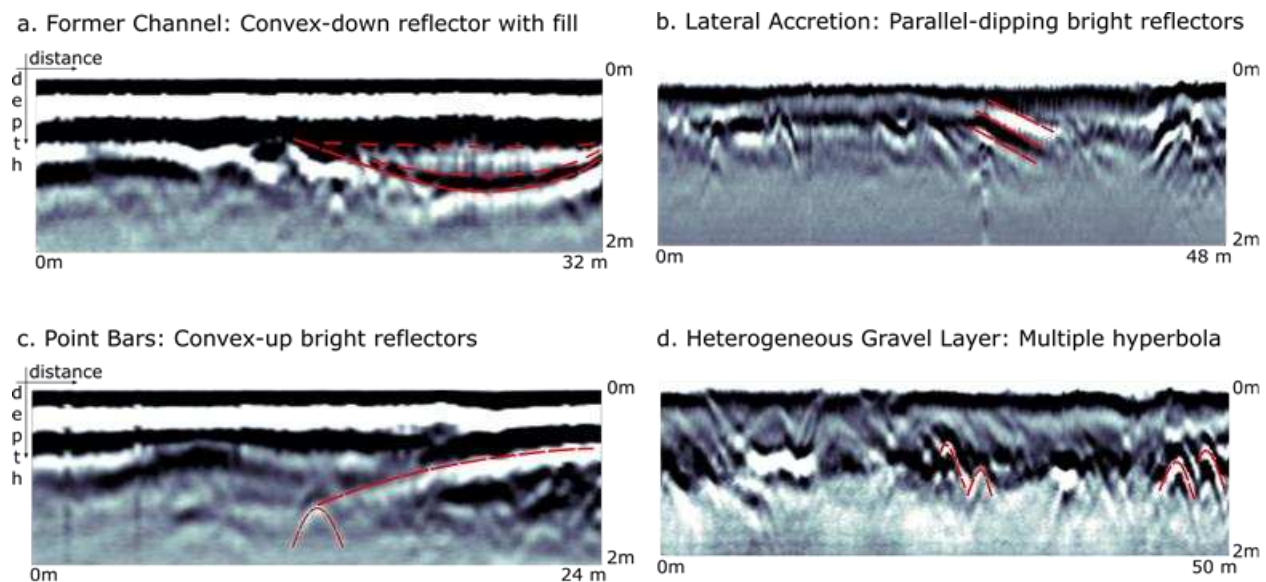


Figure 4.3 Radar facies described in the East River floodplain. a) former channel; b) lateral accretion structures; c) point bar; d) heterogeneous gravels; Key features are highlighted in red.

layer and finer-grained sediments related to channel fill, which have contrasting dielectric constants (e.g. Vandenberghe & Van Overmeeren 1999; Bridge 2009). Flooding and high flows, such as those associated with peak snow melt in this region, are a driving force behind abandoned meander infilling. It may take many flood events to infill a channel; therefore, the fine-grained fill layers may also be weakly laminated (Toonen et al. 2012). In our GPR data, we see weak, sub-horizontal, linear reflectors in channel deposits, associated with fine-grained sediments and attributed to laminated fill structures.

Lateral accretion structures are imaged by subparallel reflectors that dip in the direction of aggradation (Figure 4.3b; Vandenberghe & Van Overmeeren 1999; Kostic & Aigner 2007; Słowik 2016), a signature resulting from sediment grading of multiple aggrading layers related to the migration of the structure (Bridge 2009; Toonen et al. 2012). Lateral accretion can be present on point bars indicating the direction of point bar

migration, although they are not always present in the point bar radar signatures (Miall 2014; Bridge 2009). In meandering systems aggradation structures may be paired laterally, across the channel, with erosional features, such a basal channel scour or cutbanks (Bridge 2003; Toonen et al. 2012; Miall 2014). This lateral pairing of erosion and deposition can be noted across the current East River channel, and is imaged in GPR transects at Meander D.

The top outline of point bars are indicated by convex-up reflectors created by the contrast of coarser point-bar material mantled by finer-grained sediments (Figure 4.3c; Vandeberghe and van Overmeeren, 1999; Bridge 2003). Internal structures of point bars vary depending on bar migration processes, and can include more massive deposits of coarse sands and gravels; smaller scale cross-strata; and lateral accretion layers each with different GPR signatures (Bridge et al. 1995; Bridge 2009; Miall 2014). In the East River GPR data, the internal structures of point bars are not strongly reflected. Due to variance within point-bar deposits and the lack of clear internal structures, we utilize two key features to identify point bars: 1) the convex-up reflector shape, both laterally and longitudinally with respect to the related channel and 2) the degree of reflection, which is likely caused by contrasting sediment types between the gravel point bar and overlapping finer sediments (Bridge et al. 1995; Miall 2014; Słowik 2016) .

Heterogeneous gravel layers are characterized by discontinuous and hummocky reflectors (Figure 4.3d). Although fragmented, the reflectors can be very bright and present diffraction hyperbolae. Hyperbolae are the expression of point reflectors, where the GPR signal spreads out on either side of the object; in fluvial settings, these are

often attributed to gravel, cobbles, or boulders (Van Overmeeren 1998). At our site, hyperbola correlate to the presence of gravels seen in sediment cores. The convolution of many hyperbola creates a rugose appearance to the layer, although individual hyperbola are also visible. Similarly discontinuous and hummocky signatures have previously been attributed to poorly-sorted and heterogeneous deposits (Vandenberghe & Van Overmeeren 1999), such as alluvial fan deposits (Ekes & Hickin 2001). The lack of distinct and continuous reflectors, as well as the appearance of multiple hyperbola, characterize this facies type and indicate heterogranular deposits that contain gravels.

4.4 East River Channel Migration

An important component to our stratigraphic interpretation is the spatial context of the floodplain and location of former channels. Here, we track the path of the East River using historical images in order to evaluate GPR interpretations and their relation to the floodplains' sediment architecture and former channel locations. Historical images display three key features within the study area: 1) the development of the 2007 cutoff; 2) the progressive, sinusoidal development and increasing sinuosity of Meander A via lateral channel migration; and 3) the relative confined channel proximal to Meander D, after 1955. The combined NDVI and LiDAR-based elevations did not clearly identify abandoned channels within the floodplains of meanders A and D, but did identify some abandoned channels proximal to the study meanders (grey; Figure 4.4).

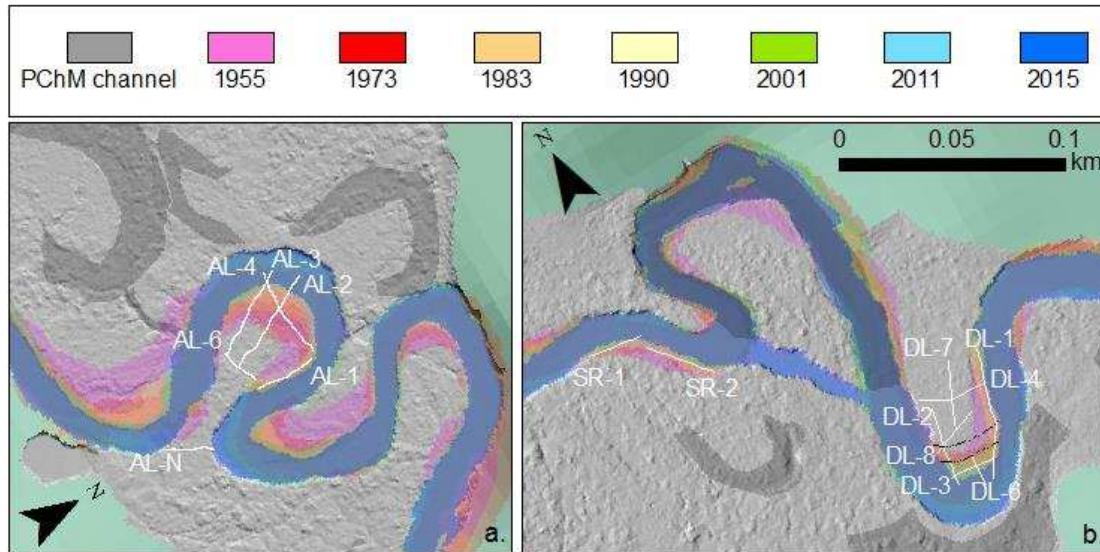


Figure 4.4: The migration of East River. Overlay of LiDAR; former channel locations, abandoned post 1955 (rainbow); paleochannels based on NDVI and elevation (grey); GPR transects (white lines); and interpreted former channel at Meander D (dashed black lines)

4.5 Integrated Stratigraphic Interpretation

Along the straight reach near the 2007 cutoff, we image the subsurface portions of a current point bar and a former channel (Figure 4.5). The imaged channel is characterized by its concave up, bright reflector overlain by sub-horizontal linear features. The imaged channel is ~1 m deep and ~15 m wide, similar to the current river's dimensions. Average channel widths along Meander A and D at the time of the study were 9 and 12m, respectively, and average bankfull widths are 18 and 36m, respectively. Overlaying the SR-2 GPR transect with channel locations from historical imagery, we corroborate the imaged channel with a former channel (Figure 4.4b). Additionally, GPR images the buried portion of the point bar (SR-1), adding to our catalog of point-bar reflections. The proximity of the coarsely grained point bar overlain by overbank fines and the concave-up channel filled with fines (~20 m; SR-1) exhibits the acute structural heterogeneity and discrete facies created by the various bedforms

preserved in the East River floodplain, created by river migration and channel abandonment.

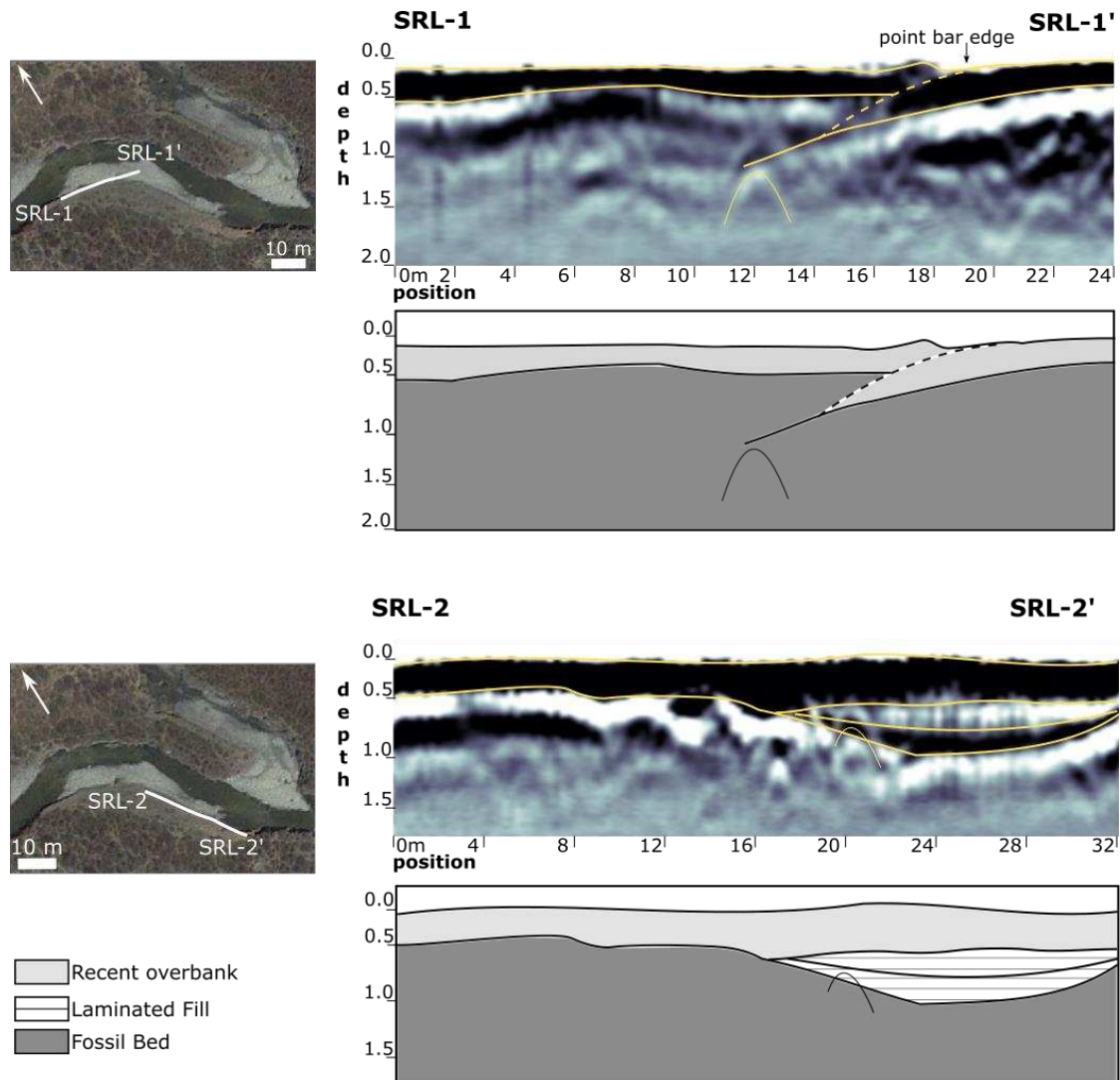


Figure 4.5: GPR transects along the straight reach. Yellow lines call out features.

As noted earlier, Meander A is simple in its stratigraphy, relative to Meander D, being composed of thick (0.3- 0.7m) overbank deposits underlain by heterogranular deposits containing gravels, identified in sediment cores and GPR signatures. Although the boreholes are relatively shallow (<1m), all contain gravels at their base, and the GPR facies indicative of gravels continue to depths up to ~1.75 m (Figure 4.6; Figure A.2;

Figure A.3). A heterogranular gravel deposit radar facies appears in the GPR transects at depth across the length and width of Meander A (Figure A.2). Gravels measured at the cutbank of the meander's neck corroborate the GPR signatures of the gravel deposits, which range from rugose reflectors to hyperbolae (Figure A.3).

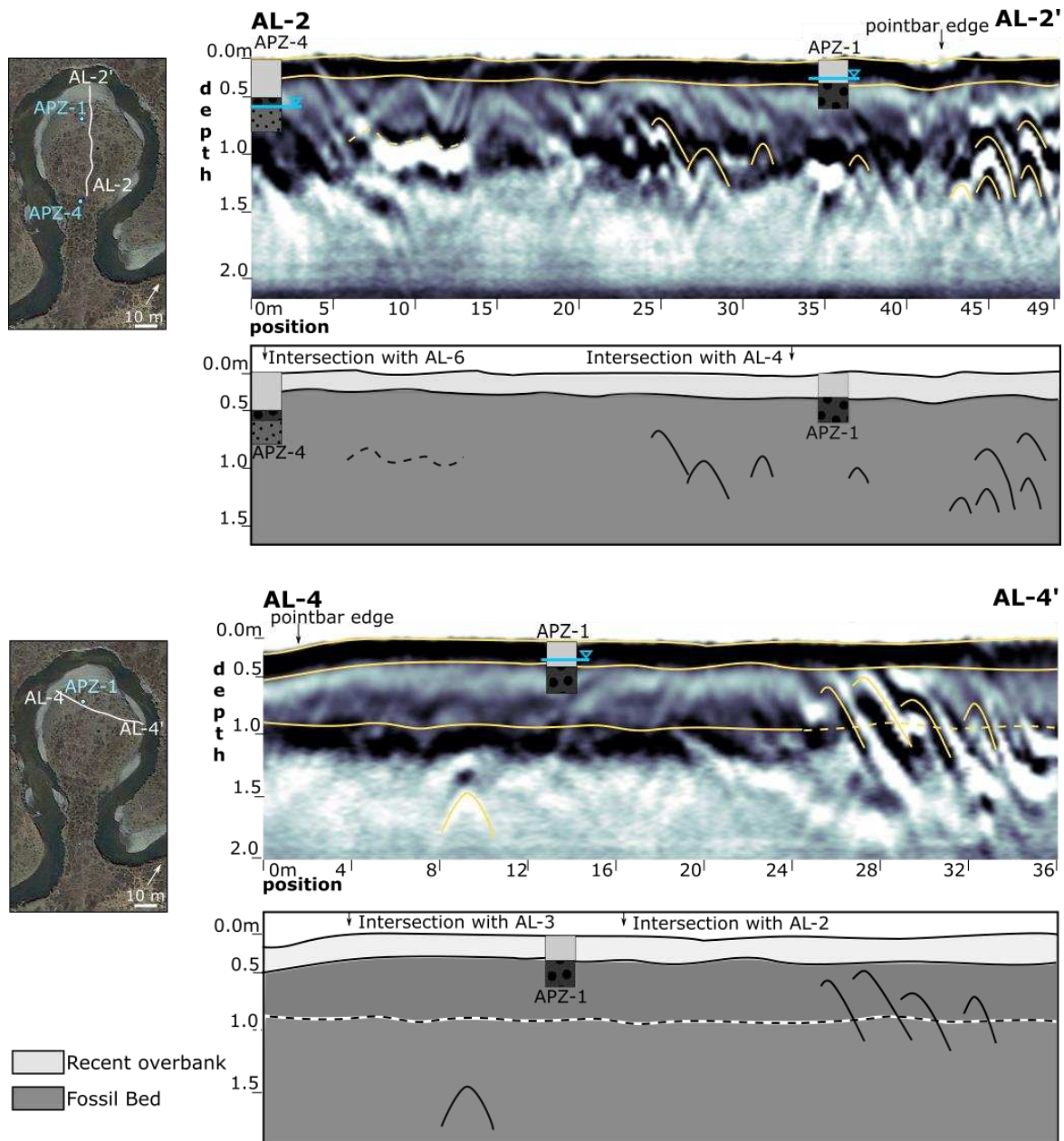


Figure 4.6: Representative GPR transects of Meander A. Blue lines indicate water table. Yellow lines mark key features. Hyperbolae related to gravels are mapped laterally and longitudinally across the meander.

The stratigraphy of Meander D is more complex than Meander A. Sedimentology differs from the base to the apex of the meander. The heterogranular radar facies (Figure 4.3d) is only located toward the back of the meander (transects DL-1, DL-2, DL-7; Figure A.3) and contains continuous heterogranular gravel layers overlain by overbank fines. This sediment package is imaged in the GPR and corroborated with the boreholes (Figure 4.7; DPZ-1 and DPZ-2). Towards the apex of the meander, gravels are absent, and the boreholes reveal interbedded layers of fines and sands. The termination of the heterogranular gravel deposit is collocated with the end of a convex-up point bar reflection, seen in GPR transects DL-1, DL-2, DL-7, and DL-8 (Figure 4.7; Figure A.3). Additionally, in transect DL-1, reflectors indicate lateral aggradation towards the nose of the meander (Figure 4.5). These bright, dipping reflectors suggest a strong contrast between the gravel deposit and finer-grained sediments towards the meander nose. Laminated channel fill is located in the middle of the transect, in-line with the current straight reach of the meander. A 10-15 m wide swath of poorly defined, sub-horizontal reflectors appears in GPR transects DL-1, DL-2, DL-6, and DL-8 (Figure 4.7; Figure A.3) and possibly represents laminated fill. Along this line sediment cores from piezometers DPZ-4 and DPZ-5 include layers of sands and fines, and pieces of organic matter and branches (Table 3.1; Figure 4.1). The final feature is an abrupt transition from the fill to a layer of gravels associated with the current point bar at the nose of Meander D (Figure 4.7). This transition can be seen in transects DL-1, DL-2, and DL-6 and may reflect an erosional feature. The sequence of point bar, accretion, fill and possible erosion radar facies down the meander indicates that the deposits are laterally discontinuous, and reflect disparate depositional regimes across the meander.

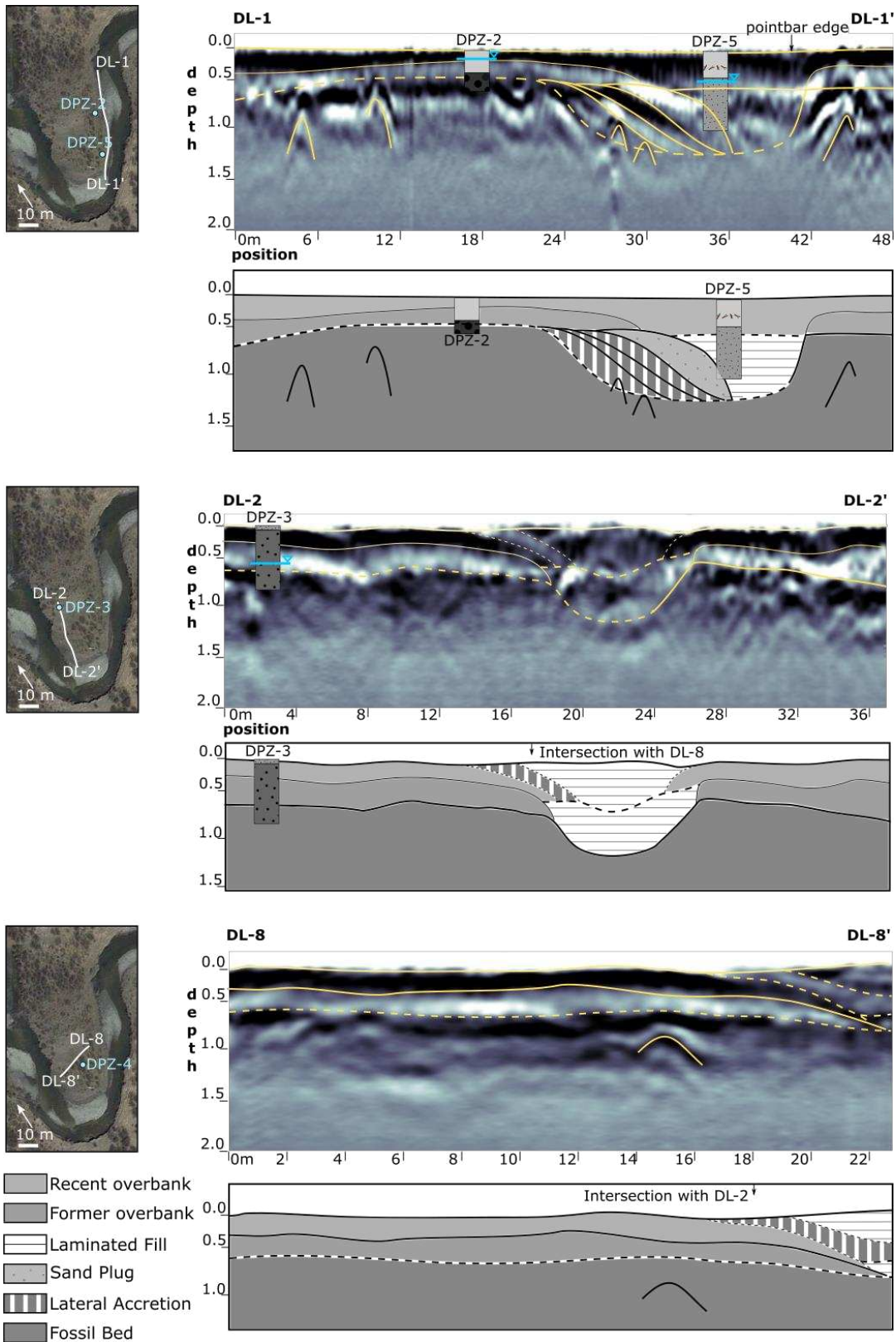


Figure 4.7: Key GPR transects of Meander D. Blue lines indicate water table. Yellow lines mark key features.

CHAPTER 5

DISCUSSION

5.1 Stratigraphy and Floodplain Evolution

The GPR-imaged floodplain stratigraphy and shape of Meander A suggest relative homogeneity in the floodplain substrate in which the meander formed. Meander A is immediately up-valley of a large alluvial fan (Figure 2.2a), which may have affected the shape of the meander by both providing a portion of material in which the meander formed, and by creating topographic barriers to down-valley river migration. Sinusoidal planforms, similar to Meander A, have been generated experimentally (e.g. Friedkin 1945; Duan & Julien 2005) and numerically (e.g. Ikeda et al. 1981; Johanneson & Parker 1989; Howard 1992; Zolezzi & Seminara 2001; Asahi et al. 2013), predicated on the assumption of a structurally homogeneous floodplain. The historical images documenting the development of Meander A display increasing sinuosity as the channel migrates towards its cutbanks (Figure 4.4a; Figure B.1). This progressive migration has been observed in experimental and numerical studies of river meandering in homogenous materials (reviewed by Camporeale et al. 2007). The meander's relatively simple stratigraphy of heterogranular gravel deposits overlain by overbank deposits are seen in the sediment cores and imaged in the GPR transects across the meander. Progressive lateral migration of the channel and point-bar aggradation (Figure 4.4a) may have erased evidence of earlier channel locations, leaving coarsely grained point-bar deposits overlain by overbank fines observed across Meander A (Figure 5.1). Additionally, active migration of Meander A is possibly linked to its orientation relative to the valley and the down-valley alluvial fan deposit. The cross-valley orientation of the meander results in a lower downstream floodplain slope relative to meanders oriented

down-valley. Reduction in slope generally creates conditions suitable for neck cutoffs, which occur when a meander becomes so sinuous that the neck is breached (e.g. Miall 2014).

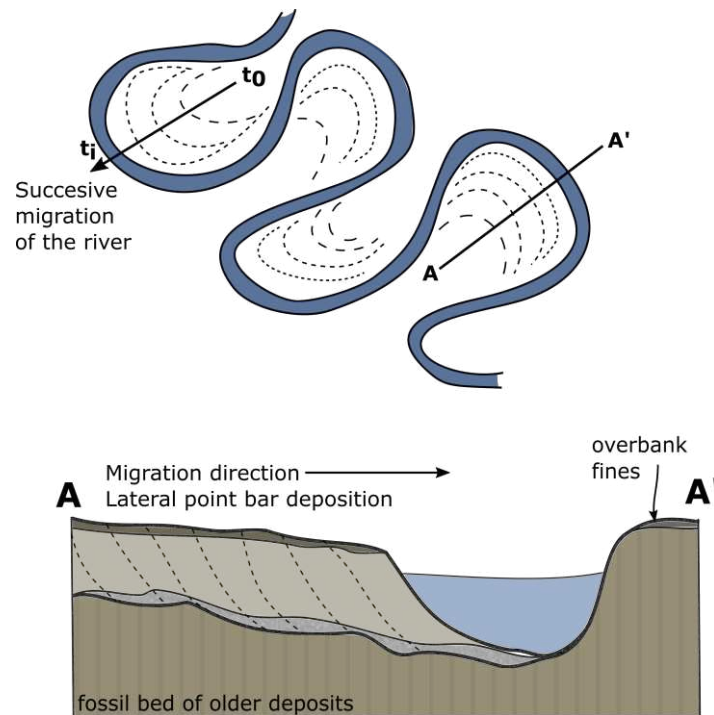


Figure 5.1: Sketch of progressive lateral migration of meanders and corresponding sediment deposits. Top: Planview of lateral migration of the river and associated increase in sinuosity. Bottom: Cross section of the sediment deposits associated with channel migration, relative to the current channel.

Although the East River continues to meander, it has not created a neck cutoff at Meander A. This is possibly due to a small topographic barrier, detected in LiDAR at the meander neck, impeding river migration down-valley. The alluvial fan just down-valley of Meander A is a likely source for material and increased elevation of Meander A's neck.

The stratigraphic interpretation for Meander D is a preserved channel, possibly a chute related to a former channel. Support for a preserved chute is 1) the stratigraphy, indicative of a former channel; 2) the preservation of the channel stratigraphy captured

by the GPR survey and 3) the relative instability of the East River channel near Meander D, indicated by recent local cutoffs. The stratigraphy across the interpreted cutoff at Meander D is similar to other documented cutoff configurations (Toonen et al. 2012; Miall 2014; Figure 5.2): plugs of fines (Toonen et al. 2012; Miall 2014; Słowik 2016) and organic debris (Toonen et al. 2012; Słowik 2016) associated with low discharges during channel abandonment and disconnection with the main channel (Toonen et al. 2012). The interpreted former channel from our data (dashed black lines; Figure 4.4b) is collocated with the 1955 channel (Figure 4.4b; pink), but unlike Meander A, the channel deposit is still preserved as seen in GPR signatures and sediment core data. Abandonment of the chute preserves the channel structures. The diversion of flow from the chute reduces erosive power and allows for sediment deposition and the preservation of channel features in the abandoned chute. If the former chute is only occupied by the river during overbank flow, flow in the abandoned channel will be lower and slower, and not capable of moving the coarser channel bed sediment. Instead, lower flows in abandoned chutes are associated with deposition of finer sediments, and deposition over the former channel shape preserves the channel (Toonen et al. 2012). If the river laterally migrated towards the cutbank, like at Meander A, point bar aggradation would have progressively erased the former channel position, which is not supported by our data. Instead, our data at Meander D support the preservation of a channel bottom and channel fill of finger-grained sediments.

The preserved channel facies at Meander D and the laminated, fine-grained fill (~30-42m along transect DL-1; Figures 4.7 and 5.2) indicate that the channel disconnected from the main channel at some point and was only active during high

flows. Unsteady flow dynamics, such as the snowmelt- versus baseflow-dominated flow regime in the East River, can promote chute formation (Asahi et al. 2013; Van Dijk et al. 2014) and control chute infilling (Van Dijk et al. 2012). The reach of river near Meander D has experienced multiple cutoff events, both upstream and downstream of Meander D, documented by historical photographs (Figure B.2).

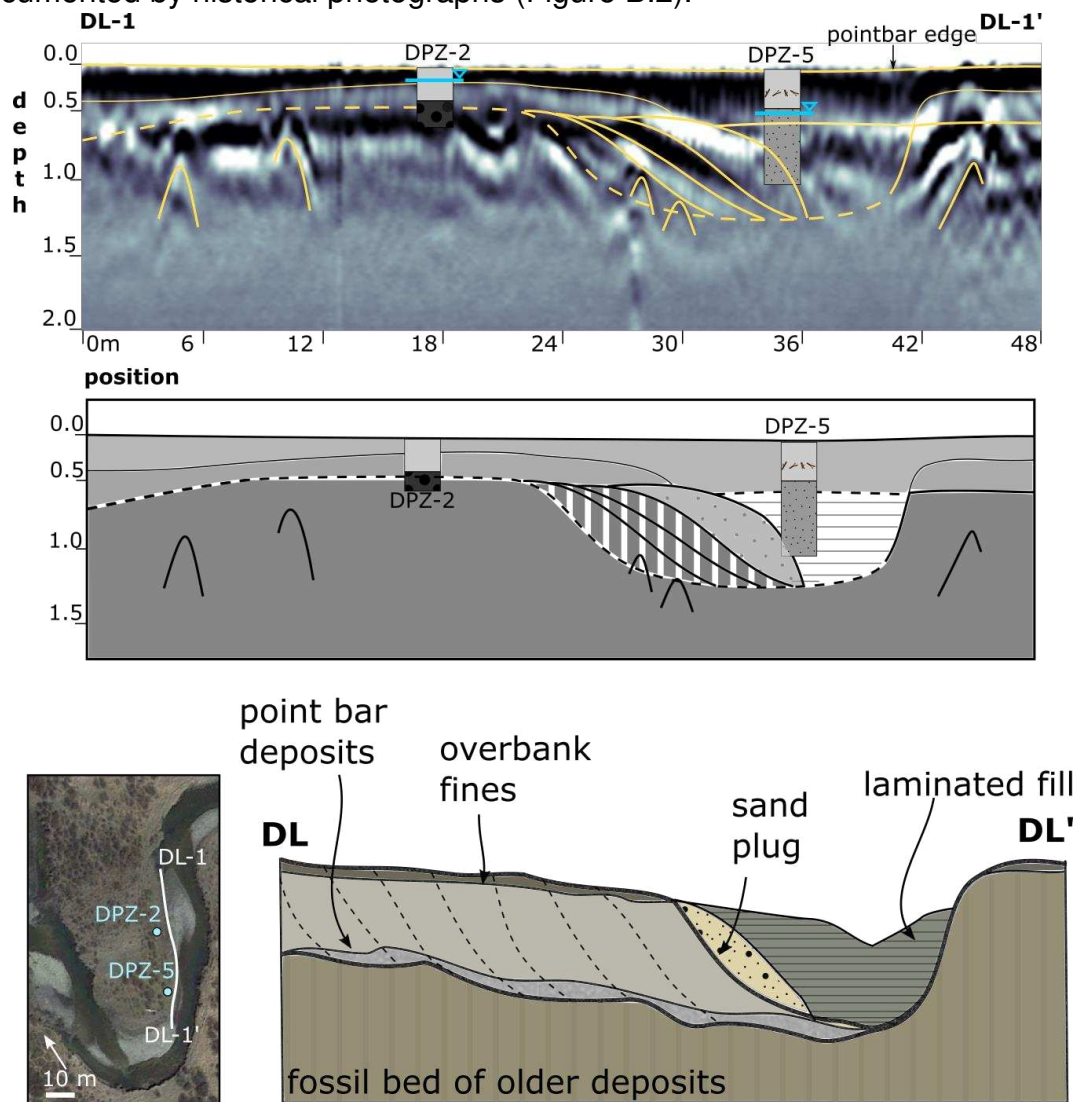


Figure 5.2: Stratigraphic interpretation of Meander D as a preserved channel deposit. Top: Disparate sediment types and GPR signatures across Meander D suggest variable depositional settings across the meander. Middle: Sketch of the various deposits and their relation to each other. Bottom: Stratigraphic interpretation of meander of a preserved channel deposit (modified from Toonen et al. 2012).

Cutoff events can occur in succession (e.g. Słowik 2016), as rivers adjust to local energy imbalance created by cutoffs (Lanzoni & Seminara 2006; Van Dijk et al. 2014). Chute cutoffs, in particular, are attributed to ratios of meander radius of curvature (R_c) to channel width (W) between 1.8 and 3.7 (Harvey 1989). The current $R_c:W$ of Meander D is 2.5 (R_c of ~45m; width of 20 m), well within the range suitable to generate a cutoff. The $R_c:W$ ratio at Meander D would indicate that this reach may still be unstable and facilitative of cutoffs.

Additionally, unstable flow regimes, unsteady sediment inputs, and bend geometry can divide river discharge, leading to channel bifurcation and chutes forming near meander apexes (Van Dijk et al. 2014). The preserved channel at meander D may

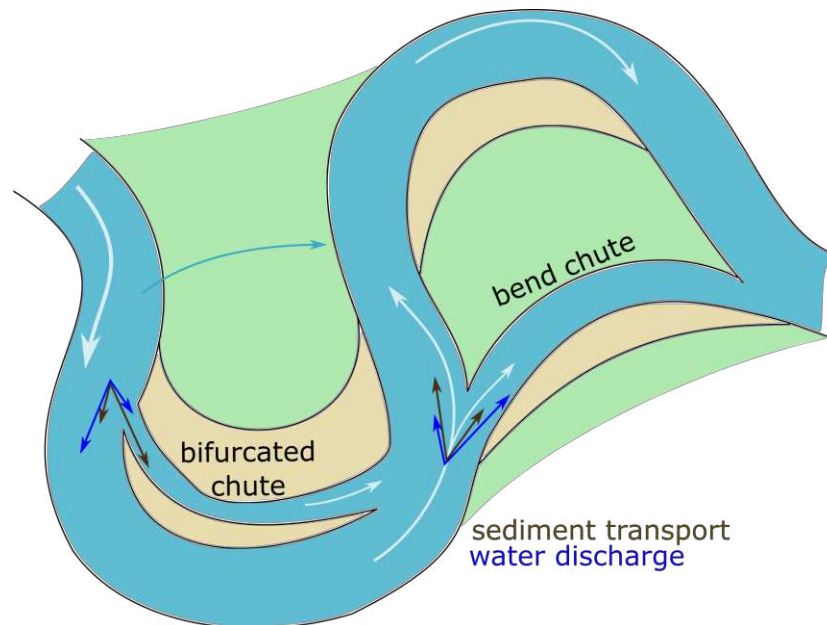


Figure 5.3: Comparison sketch of discharge and sediment transport capabilities of bifurcated and bend chute cutoffs (modified from Van Dijk et al. 2014). Light blue arrows indicates fluid flow, and dark blue and brown arrows indicate river discharge and sediment transport, respectively. The low water discharge through the bifurcated chute, relative to larger amounts of sediment, increases the sedimentation rate and subsequently, failure rate of bifurcated chutes when compared to bend or neck cutoffs.

have been a former bifurcated chute, supported by the preserved channel's location relative to the current channel. Bifurcated chutes have a high rate of failure relative to bend and neck cutoffs due to sediment-discharge dynamics and gradient advantages (Figure 5.3; Van Dijk et al. 2017). The high amounts of sediment transport through the chute relative to low river discharge, increase deposition through the chute. Meander D's stratigraphy, its preservation, and its context within an unstable reach of the river, would support a stratigraphic interpretation of a former bifurcated chute, infilling, and abandonment.

5.2 Floodplain Heterogeneity and Strata Connectivity

K estimates across the study site are relatively consistent, with approximately an order of magnitude difference across both meanders and all sediment types. However, when combined with water table gradients, flow path lengths, and strata connectivity as

Table 5.1: Comparison of linear velocity, and residence time estimates of groundwater flowing laterally across Meander D when using geometric (top) and arithmetic (bottom) means of K values. Flow path length estimates remain the same between comparisons.

using $AVG_{geo} K$	June D (Back)	June D (Apex)	July D (Back)	July D (Apex)	Aug D (Back)	Aug D (Apex)
flow path length [m]	45	30	45	30	40	30
v [m/d]	0.09	0.03	0.06	0.06	0.03	0.03
t_H [d]	500	1000	800	500	1400	1000
using $AVG_{arithm.} K$	June D (Back)	June D (Apex)	July D (Back)	July D (Apex)	Aug D (Back)	Aug D (Apex)
flow path length [m]	45	30	45	30	40	30
v [m/d]	0.2	0.06	0.1	0.1	0.06	0.06
t_H [d]	240	500	400	240	640	500

estimated by GPR and remote sensing data, linear transport velocities and lateral hyporheic residence times across the meanders may be quite different. We estimated the linear velocities and residence times at Meander D using a geometric mean of K

values (0.7 m/d). However, depending on orientation of fluvial packages, higher K strata can create preferential flow paths, which would increase linear flow velocities. Across Meander D, the largest K values are located near the preserved former channel (DP-3 and DP-5; Figure 4.1), which is oriented down-valley, parallel to the groundwater gradient (Figure 4.2). Considering an arithmetic mean of Meander D's K values (1.7 m/d), perhaps appropriate given the orientation of packages with respect to the gradient, linear velocity estimates double, while residence times are halved (Table 5.1). Despite the narrow range of estimated K values at this site, there may be a significant difference in estimated linear velocity when strata orientation is considered. This finding illustrates the importance of capturing both the range of magnitude of hydraulic characteristics related to fluvial strata, as well as strata connectivity and orientation in physical hydrologic models. Hyporheic residence times are often compared to biogeochemical reaction times (e.g. Bardini et al. 2012; Gomez-velez & Harvey 2014; Gomez-Velez et al. 2015), and doubling of residence time can have a significant impact on estimates of hyporheic efficacy in transforming solutes.

Across Meander A, hydraulic gradient magnitudes do not change greatly throughout the late spring and summer 2017, and GPR and sediment cores support a simplified stratigraphy devoid of structures with disparate sedimentologies. Despite these facts, residence times across the meander likely widely vary due to the meander's high sinuosity. Lateral hyporheic residence times reflect transport across the widest part of the meander (Table 4.2); if we take into consideration flow paths across Meander A's neck, (~20 m wide) instead, estimated hyporheic residence times in early season drop from 1300 to 500 days.

5.3 Mapping Abandoned Channels and Strata at the Floodplain Scale

The location, extent, and orientation, of hydrofacies have a large impact on physical and chemical hydrologic processes. In this study, we map former channels to try to identify and simplify major floodplain strata types, such as channel fill, channel bottoms, and point bars and relate these features to current and former channel locations. The goal of this work is to assess the possibility of simplifying floodplain stratigraphy into hydrostratigraphic packages that are identifiable remotely.

In addition to historical imagery, we augmented the body of recognizable former channels (paleochannels) by combining LiDAR-based elevation and NDVI data. Former channels are identified via elevation differences and land cover disparities related to gradual infilling and vegetative succession of abandoned channels (Figure A 2.2). Although the technique is efficient and portable, it has limitations, including the need for significant human input. Initial maps generated by the eCognition software required hand delineation following software classification. This method is predicated on well-documented attributes of abandoned channels, including topographical, hydrological, and vegetative differences between former channels and floodplains related to repeated disturbance via flooding, gradual infilling, and vegetative succession. When combined with 3-D fluvial sediment architecture, elevation and NDVI data may indicate the location and extent of fluvial deposits. Our data support the use of surficial features to map fine-grained fill in mapped channels and coarse-grained point bars at the inside bend of mapped channel meanders, but does not provide enough evidence to support the use of meander geometry as a proxy for meander sediment heterogeneity.

CHAPTER 6

CONCLUSIONS

In this study, we interpret the stratigraphy and hydrology of two meanders with contrasting geometries in a montane floodplain using sediment cores, slug test data, historical photography, LiDAR, remote sensing, and GPR data. We compare our interpretations to former and current channel locations to evaluate heterogeneity of floodplain stratigraphy and its effect on linear groundwater transport velocity and lateral residence times of lateral across the meander. This study also explores how surficial floodplain features, such as former channel locations and channel geometry can assist in the mapping of hydrofacies' extent and orientation. Findings indicate that meander stratigraphy is the result of many geomorphic factors, such as channel geometry (both past and present), channel migration, and geomorphic context of the channel relative to other landforms. Our results suggest that the more sinuous and symmetrical meander (Meander A) has a more simplified stratigraphy, and that the asymmetric Meander D has a more heterogeneous stratigraphy. Our results also highlight risks related to physically representing an alluvial aquifer by a homogeneous grainsize distribution and hydraulic conductivity and effects on estimated lateral hyporheic residence times. In particular, capturing floodplain strata extent and orientation in Meander D, and river meander geometry and dynamic hydrology related to groundwater gradients' magnitude and orientation relative to meander shape in Meander A are important to quantifying water transport processes. Ideally, this study serves as a bridge between small-scale, detailed, field studies and basin-scale modeling of hydrological processes and provides a pathway for mapping the extent and orientation of representative fluvial deposits using surficial floodplain characteristics that should be useful for hydrological modeling.

CHAPTER 7

FUTURE WORK

A key source of uncertainty in this study is the K estimates of different sediment packages observed in the East River floodplain. K estimates across the study site are relatively consistent, with approximately an order of magnitude difference across both meanders and all sediment types. However, we believe that this similarity may be due in part to the method used and/or the fact that each piezometers were screened over intervals that contained various sediment facies and that specific sedimentologies were not targeted. The variety of floodplain sediments observed across the floodplain would indicate a greater range and spatial heterogeneity of hydraulic conductivities. For contrast, Anderson et al. (1999) conducted a detailed study of hydrofacies and their spatial distribution within braided stream deposits developed in glacial outwash sediments. Fluvial organization of glacial sediments observed in the their study proffers a point of comparison for the hydraulic conductivities of East River sedimentological units, such as channel fill, overbank deposits, and bedload-transported gravel deposits. Anderson et al. (1999) found many orders of magnitude between the various hydrofacies, and that spatial heterogeneity significantly impacted modeled preferential flow paths. We relate specific deposits documented in the Anderson et al. (1999) study to deposits in the East River by their sedimentology and interpreted depositional mechanism (Table 7.1).

Table 7.1: Comparison of fluvial hydrofacies documented in Anderson et al. (1999) and the East River. Characteristics include sediment description, interpretation of depositional mechanisms, and measured hydraulic conductivities. Documented deposits are located within a braided fluvial system developed in glacial outwash sediments, similar in range and depositional setting to the East River. Interpretative connection to the East River deposits facilitates assignment of K estimates to the different hydrofacies. ¹Data and interpretations from Anderson et al. (1999) study.

Lithofacies reference ¹	Description ¹	Interpretation ¹	K [m/d] ¹	Connection to East River
Fm	Massive very fine- to fine-grained sand and silt	Deposited as overbank or channel fill	0.60	Top sediment layer on all meanders and fill at meander D
St	Fine to medium sand	Deposited during lower flow regime	26	Fill in former channel at meander D
Gs	Stratified fine to medium sandy gravel	Deposited during sustained, relatively low-flow conditions on flanks of point bars	78	Various layers in meander D
Gm	Matrix-supported medium to coarse gravel	Deposited as bedload during waning high flows	76	Heterogeneous gravel layers across meander A and towards the back of meander D
Gow	Moderately well-sorted, open-work gravel with silt-filled upper surfaces	Deposited during conditions of fluctuating discharge; gravel aggraded during periods of high discharge and then became filled with suspended material when discharge decreased	1040	Former channel bottom and top layer of point bars

In the East River, despite the range of sediment packages, all K estimates resulting from slug tests are most similar to the overbank fines of the Anderson et al. (1999) study. If we were to model the East River as a homogeneous, low conductivity floodplain, we may also overlook preferential flow paths related to gravel deposits, which had K estimates four orders of magnitude greater than the overbank fines in the Anderson et al. study. In the case of Meander A, the extensive gravels across the meander may facilitate significant lateral surface-groundwater exchange, particularly across the neck and apex of the meander (Boano et al. 2006). Alternatively, at Meander D, the deposit of fines across the meander would inhibit the movement of water across the meander. Impacts of alluvial stratigraphy on water movement in the river corridor depend both on the contrast of hydraulic conductivities within deposits, as well as the deposits' orientation relative to the channel and valley (Tonina & Buffington 2007).

Future work should focus on quantifying detailed K estimates of stratigraphic units and observing their control on water movement through the floodplain. An alternative method of measuring K in the East River floodplain would be the collection of disparate sediment packages, followed by grainsize analysis and estimation of hydraulic conductivities using representative grainsize distributions for each stratigraphic unit. This approach of K estimation has been shown to be comparable to direct laboratory and field measurements (e.g. Lu et al. 2012). To observe water movement in the floodplain, a down-well tracer test could be employed at the study site. In Spring 2017, inclement weather prevented a down-well salt and electrical resistivity tracer test that was designed to capture surface water exchange across Meander D. Connection between the injection well and the river was confirmed using physical parameters, such as temperature and fluid conductivity, which are commonly different between surface water and groundwater. An electrical resistivity survey was designed to capture 3-D movement of the salt tracer across the meander. The observed stratigraphy of Meander D make it a superb location for the future K measurements and tracer tests and facilitates hypothesis testing of stratigraphic controls on lateral hyporheic exchange.

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APPENDIX A

ADDITIONAL GPR INFORMATION

This appendix includes details on GPR processing and additional GPR figures and interpretation that support the main manuscript.

Table A.1: Processing Notes for GPR. PZ = piezometer associated with each transect.

Transect	Gain Type	Attenuation	Start Gain	Maximum Gain	Velocity (m/ns)	Associated PZ	Water Level (m)
AL1	SEC2	10	3	130	0.05	APZ-3	0.56
AL2	SEC2	10	3	130	0.05	APZ-1	0.34
AL3	SEC2	10	3	130	0.05	APZ-2	0.34
AL4	SEC2	10	3	130	0.05	APZ-1	0.34
AL6	SEC2	10	8	133	0.05	AP-4	0.57
neck	SEC2	10	8	133	0.05	Gravels; 0.6 and 0.25 m	
DL1	SEC2	20	5	50	0.05	DPZ-2; DPZ-5	0.26; 0.4
DL2	SEC2	20	5	50	0.05	DPZ-3	0.53
DL3	SEC2	20	5	50	0.05	DPZ-5	0.41
DL4	SEC2	20	5	50	0.05	DPZ-1	0.61
DL5	SEC2	20	5	50	0.05		
DL6	SEC2	20	5	50	0.05		
DL7	SEC2	20	5	50	0.05	DPZ-1	0.61
DL8	SEC2	20	5	50	0.05		
SR1	SEC2	5	5	50	0.06		
SR2	SEC2	20	5	50	0.06		

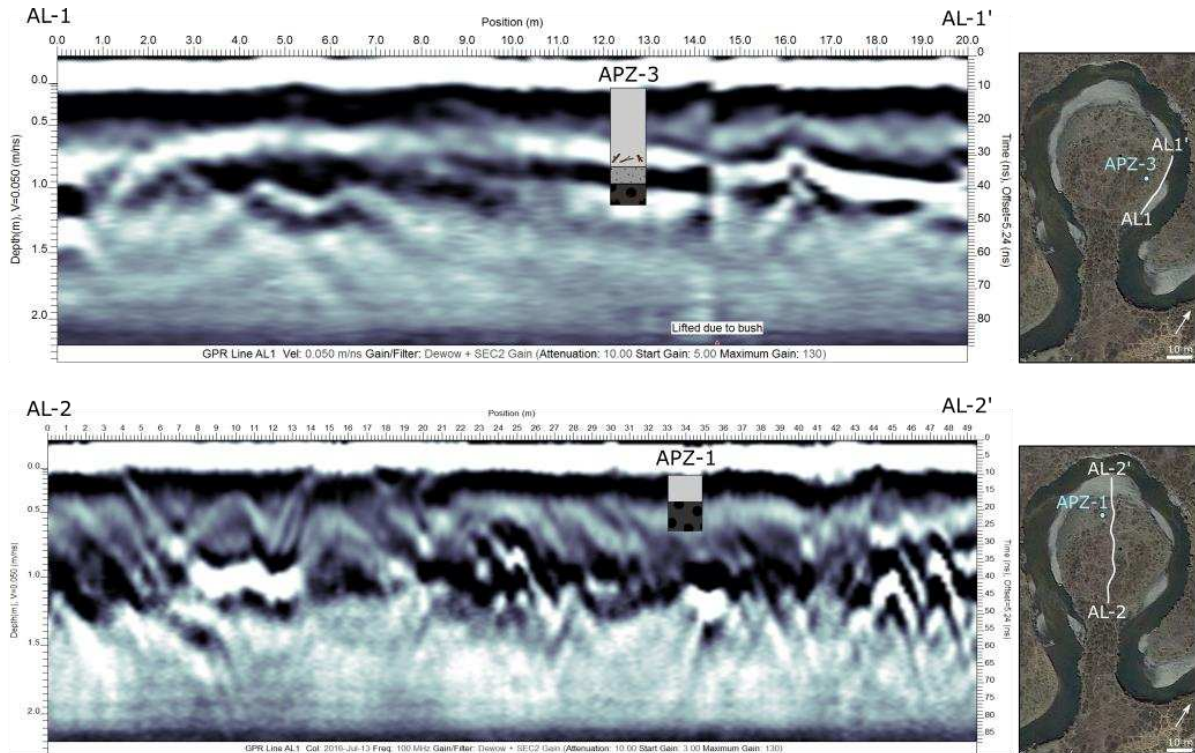


Figure A.1: Non-annotated GPR transects collected at Meander A.

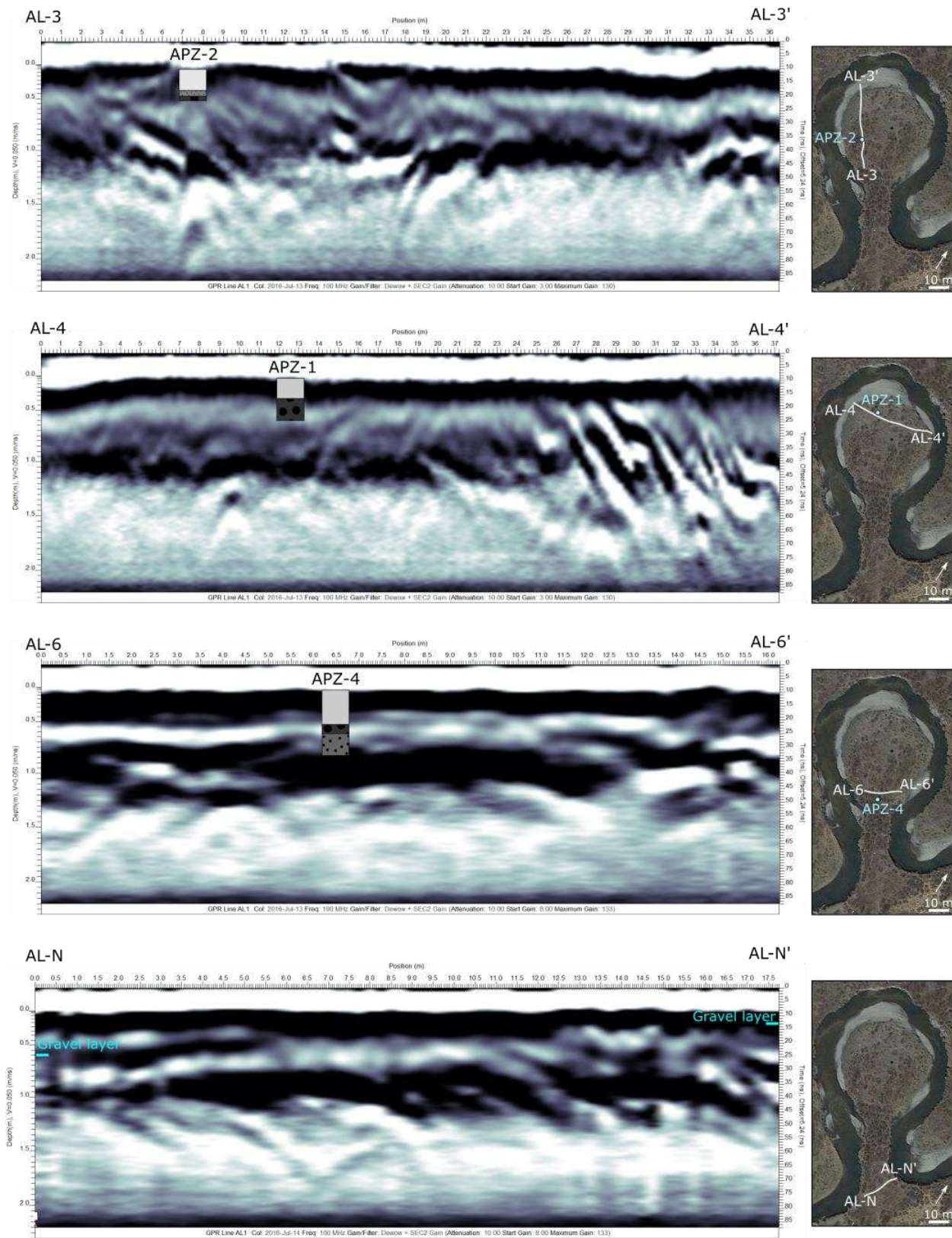


Figure A.1 (continued)

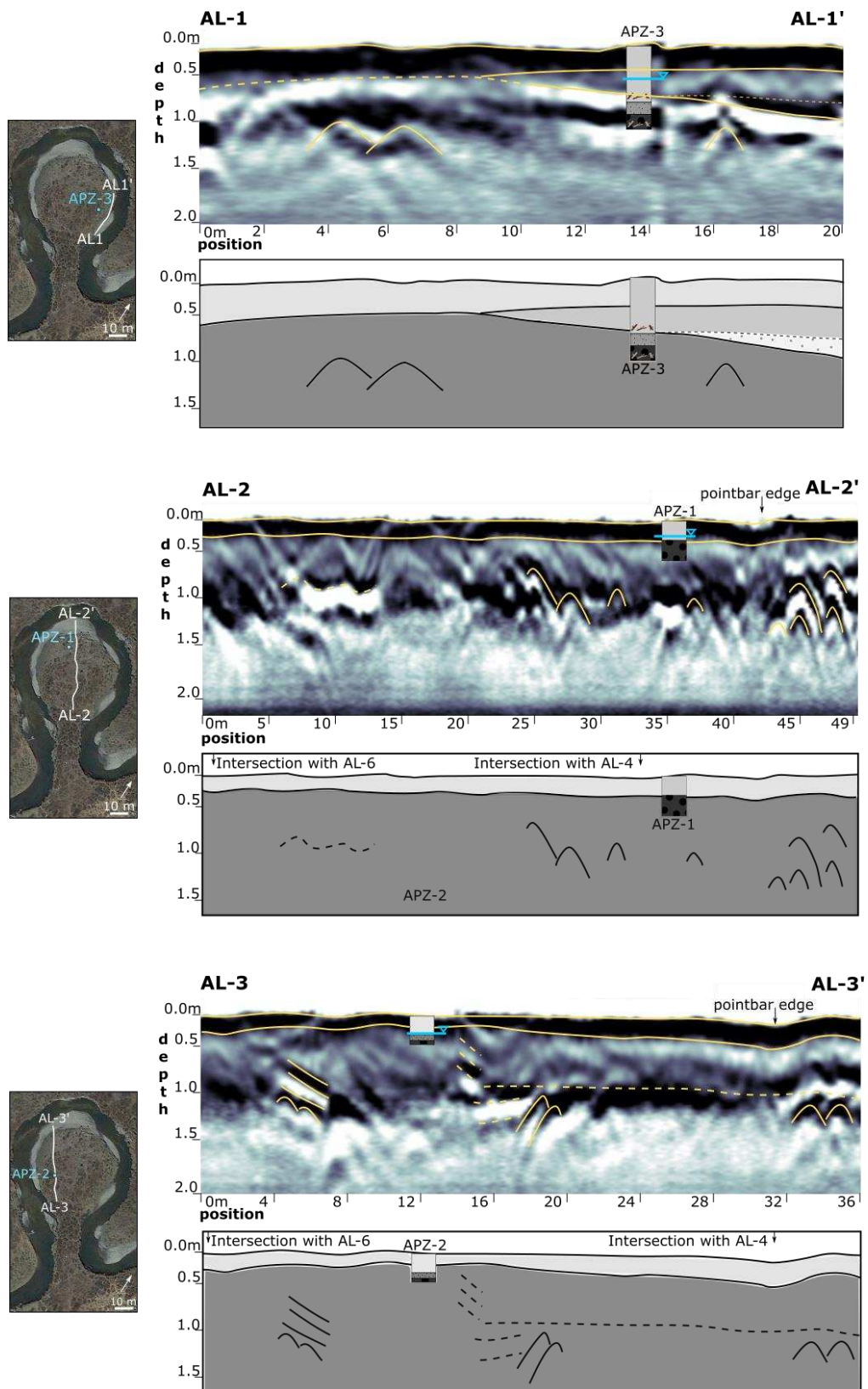


Figure A.2: Stratigraphic interpretation of GPR transects along Meander A

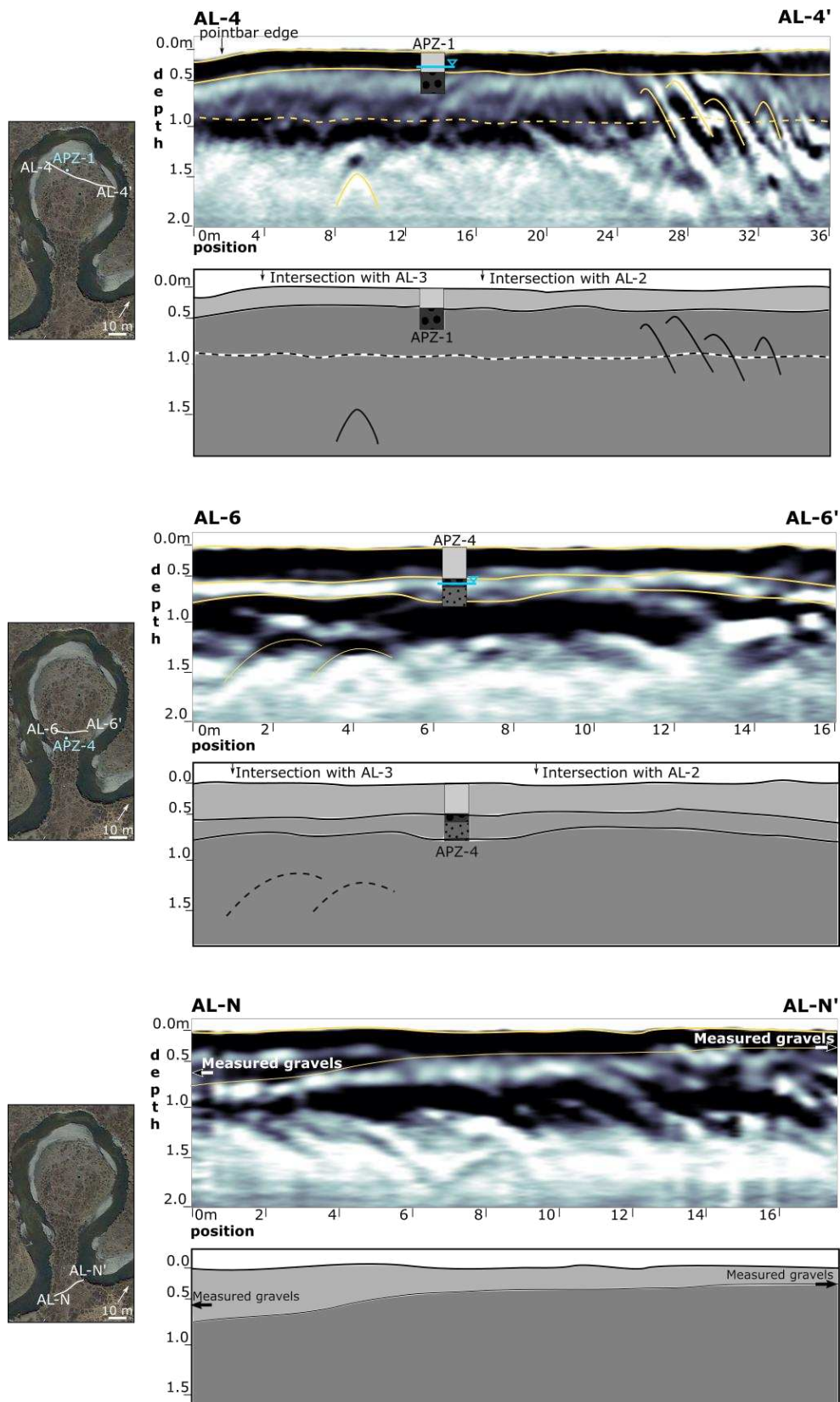


Figure A.2 (Continued)

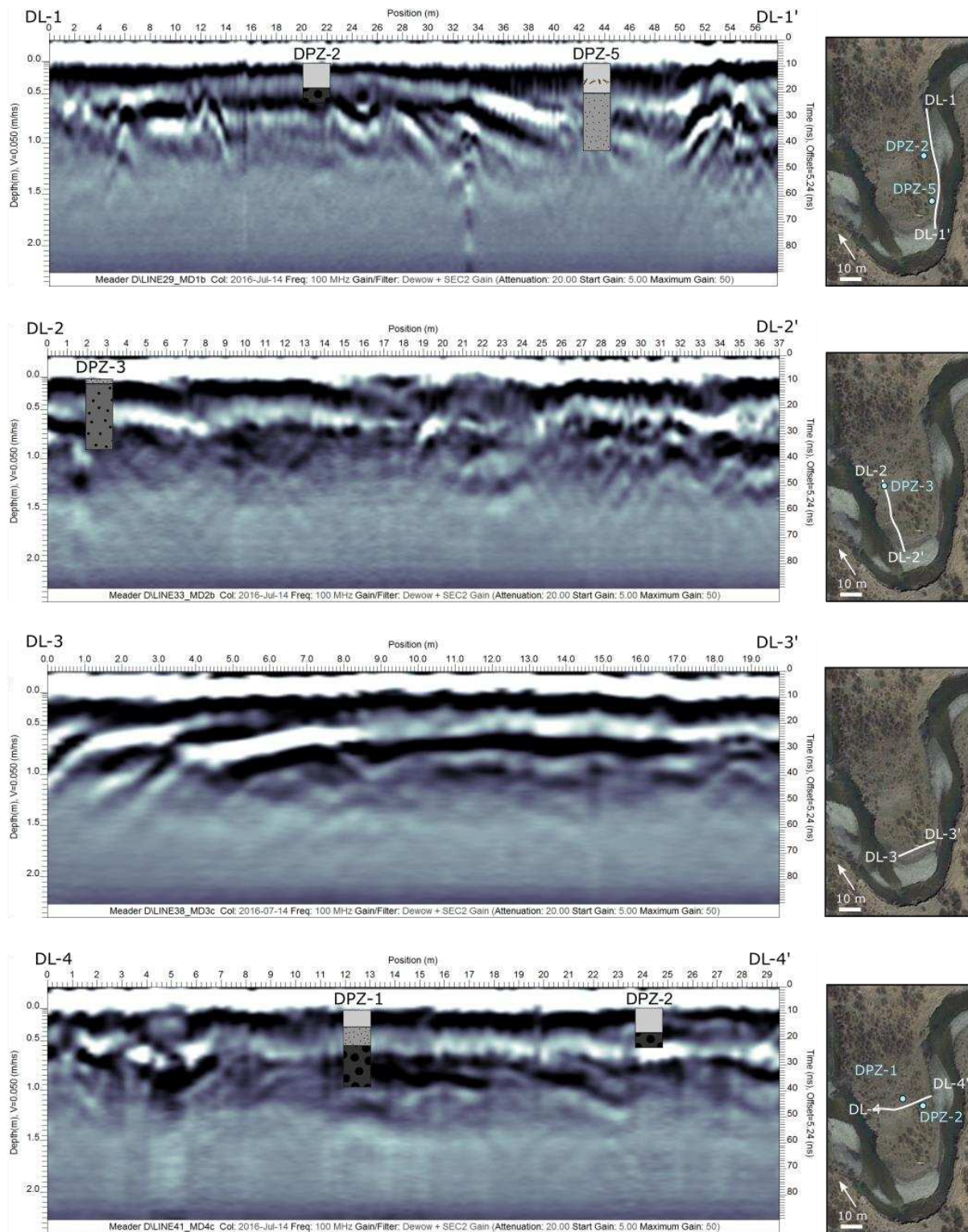


Figure A.3: Non-annotated GPR transects collected at Meander D. (continues on next page)

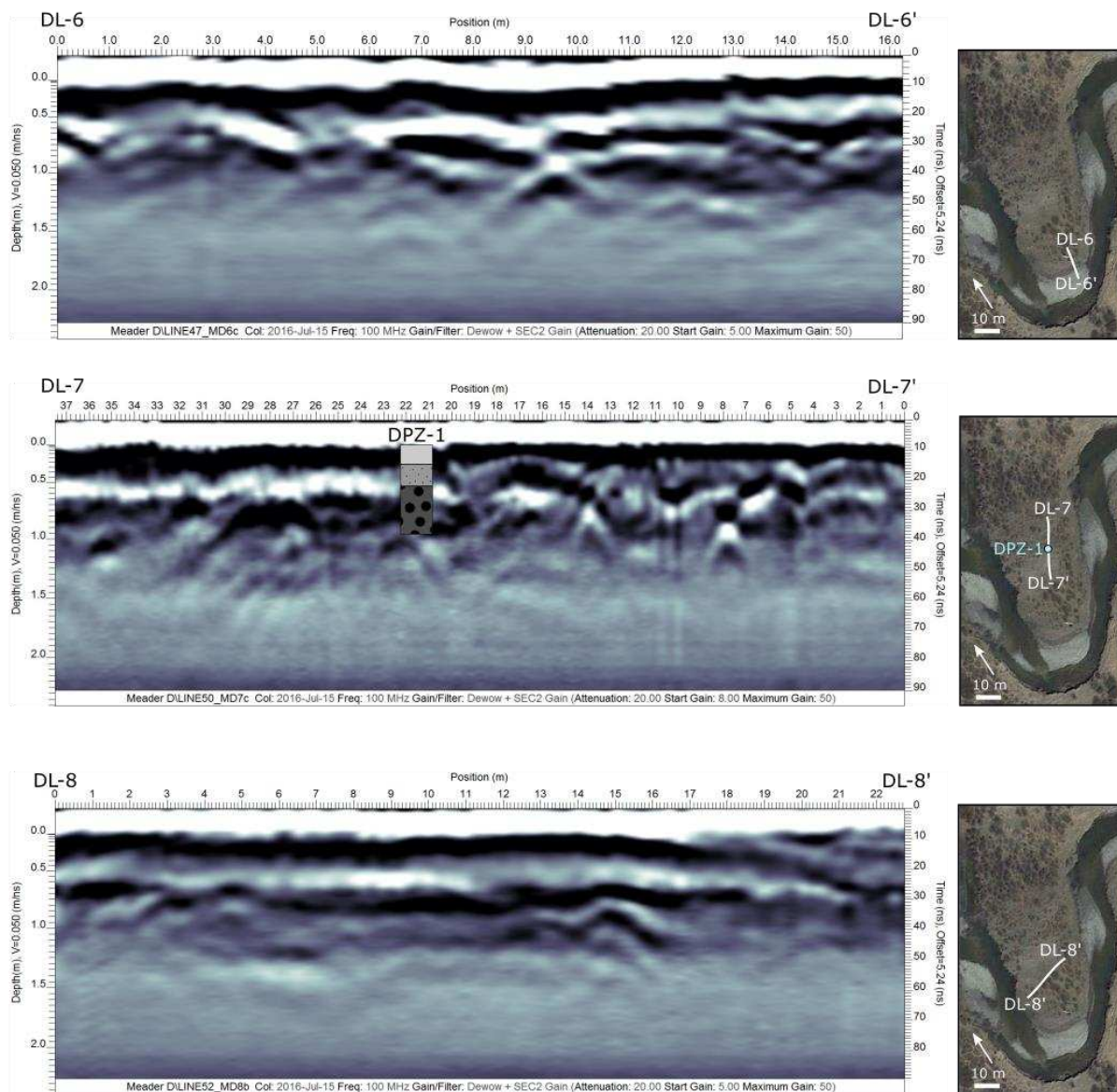


Figure A.3: (continued)

APPENDIX B

HISTORIC CHANNEL LOCATIONS OF THE EAST RIVER

This appendix includes maps of former channel locations based off of historical photography and NDVI/elevation data.

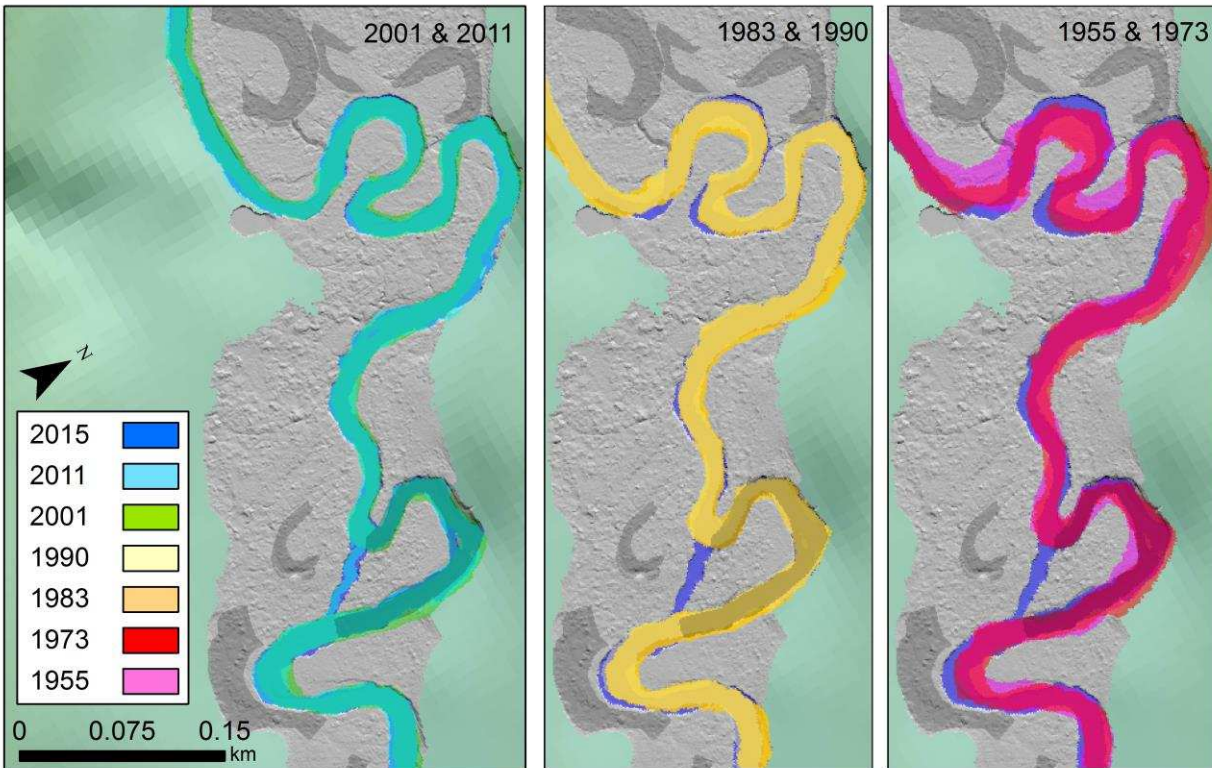


Figure B.1: Former channel locations as shown in historical images. The most recent channel configuration (2015) is plotted below all the other years in each panel for spatial reference. Elevation/NDVI mapped channels are in grey.

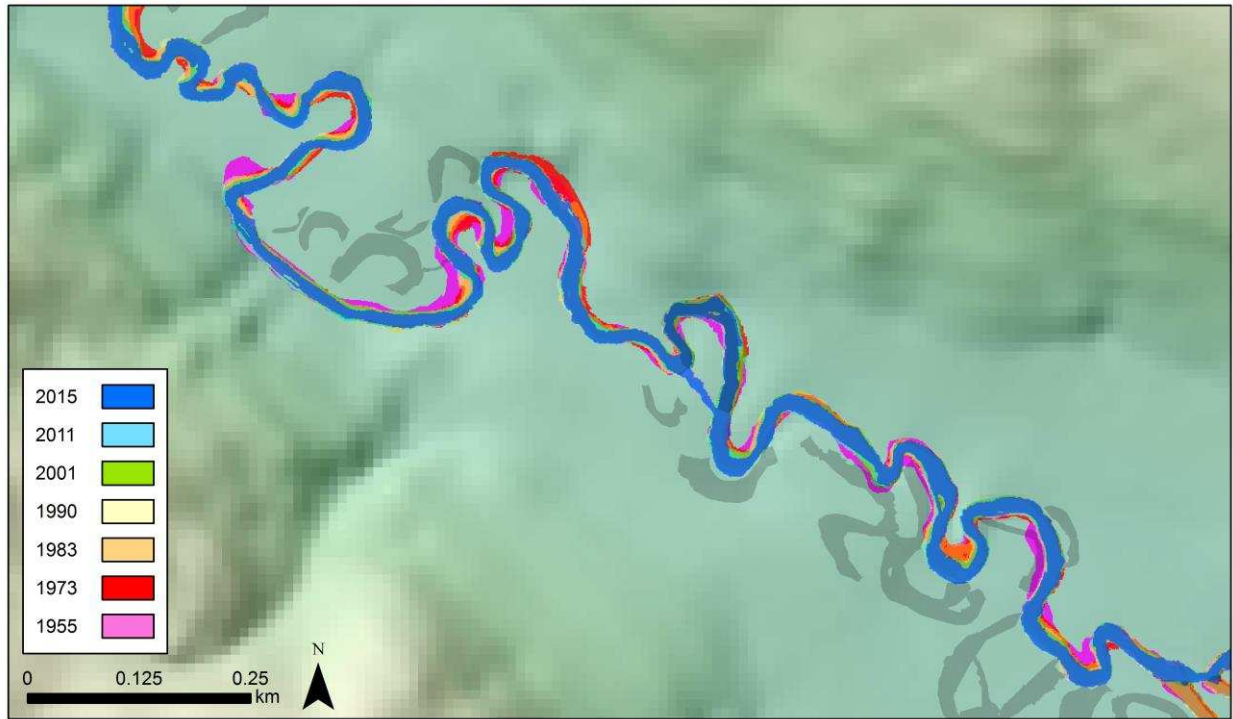


Figure B.2: Channel reconstructions (1955-2015) based on historical photography (rainbow) and elevation/NDVI (grey). Note multiple historical cutoffs up- and downstream of Meander D.

APPENDIX C

GRAVEL DEPOSIT DEPTHS

Gravel depths are inferred from tile-probe transect. Rebar was pushed into floodplain sediments at the given locations. Depth of refusal is assumed to be depth of gravels

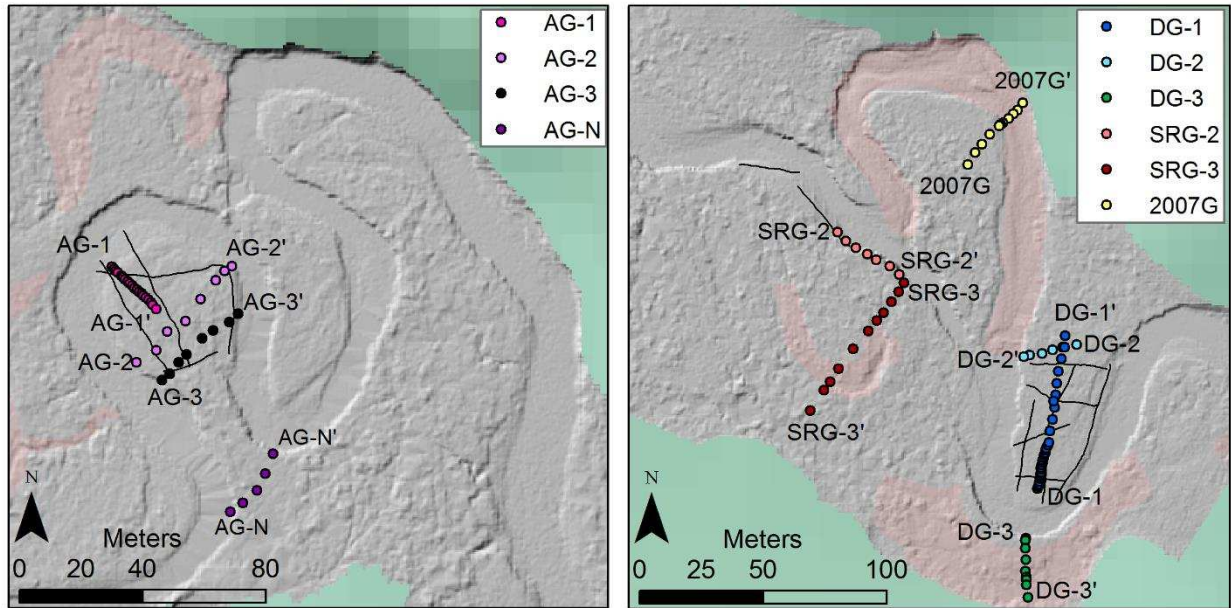


Figure C.1 Locations of tile-probe transects. C.1a (left) is along Meander A and C.1b (right) is along Meander D. Mapped abandoned channels are colored in light orange.

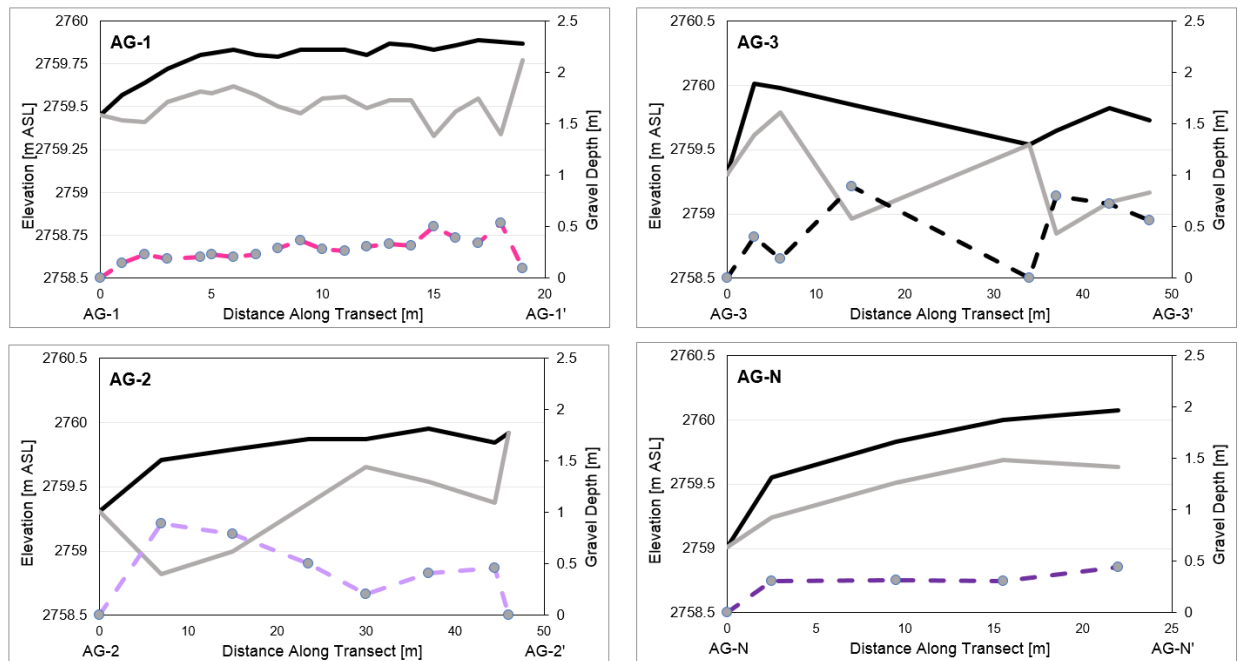


Figure C.2 Graphs of ground surface elevations (black solid) and inferred gravel depths based on tile probe measurements (dotted), and calculated gravel elevations (grey solid) at Meander A.

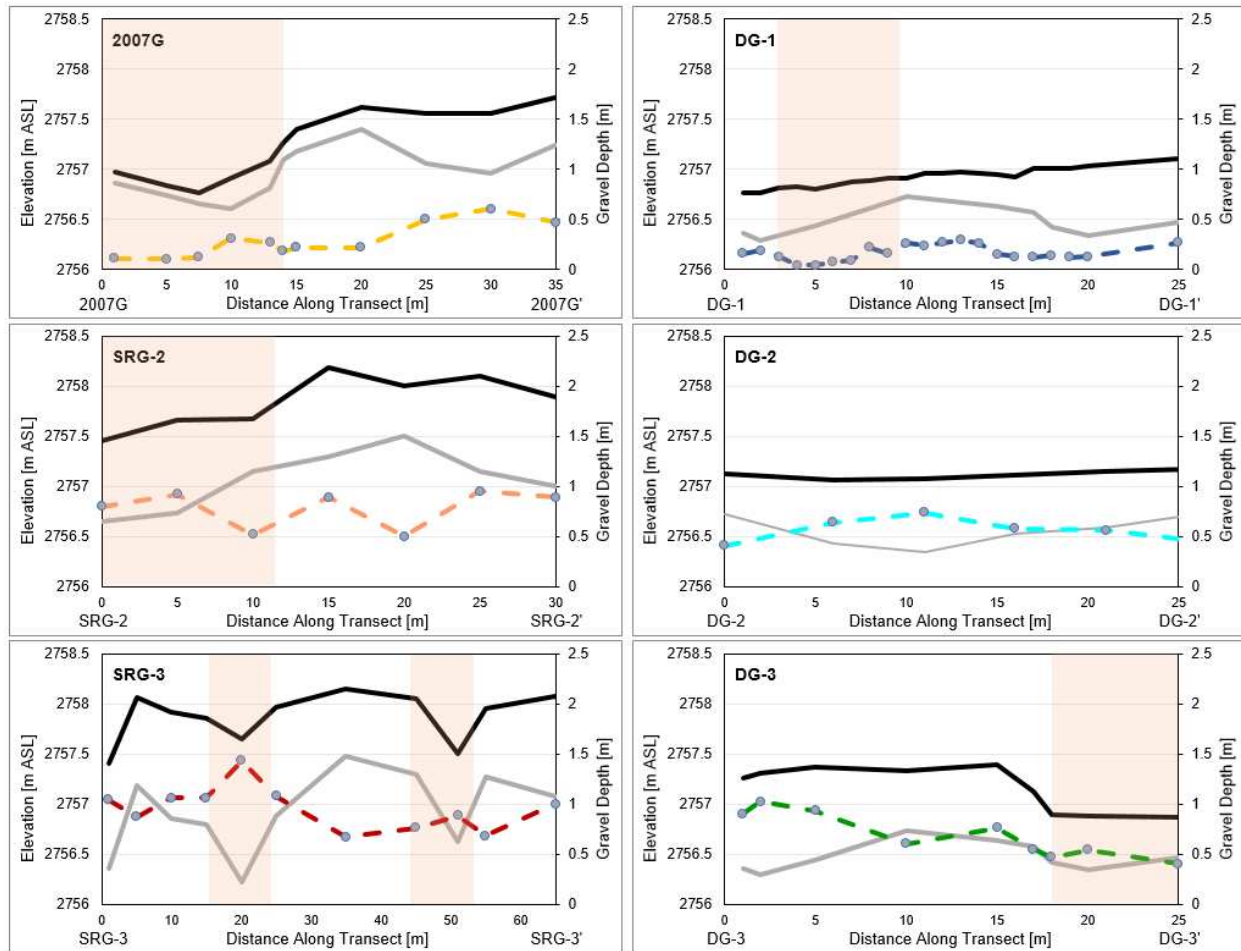


Figure C.3 Graphs of ground surface elevations (black solid), inferred gravel depths (dotted), and calculated gravel elevations (grey solid) at Meander D, based on tile probe data. Light orange shading is the location of mapped abandoned channels.

APPENDIX D

RESULTS OF AN IN-STREAM TRACER TEST

D.1: Summary of August 2016 pulse injection instream tracer test

On August 10, 2016, we performed a pulse tracer injection using conservative tracers, Rhodamine water tracer (RWT) and NaCl. We instantaneously injected 1.3 L of RWT and 22.7 kg of NaCl into the East River. RWT and NaCl was measured instream only. RWT can be measured at very small concentrations, but is photosensitive and can degrade in sunlight. Because of this, we used a companion “true” conservative tracer of NaCl. For this test, NaCl measurements appeared more reliable than the RWT measurements and are presented below. A key take away from this test is that pulse tracer injections record advection through the river channel more than dispersion and hyporheic exchange. Because of this, a continuous injection test was performed in September 2016. All instream mass estimates presented here integrate the measured concentration $[M/L^3]$ through time and multiply it by the total discharge $[L^3]$ for the time of measurement.

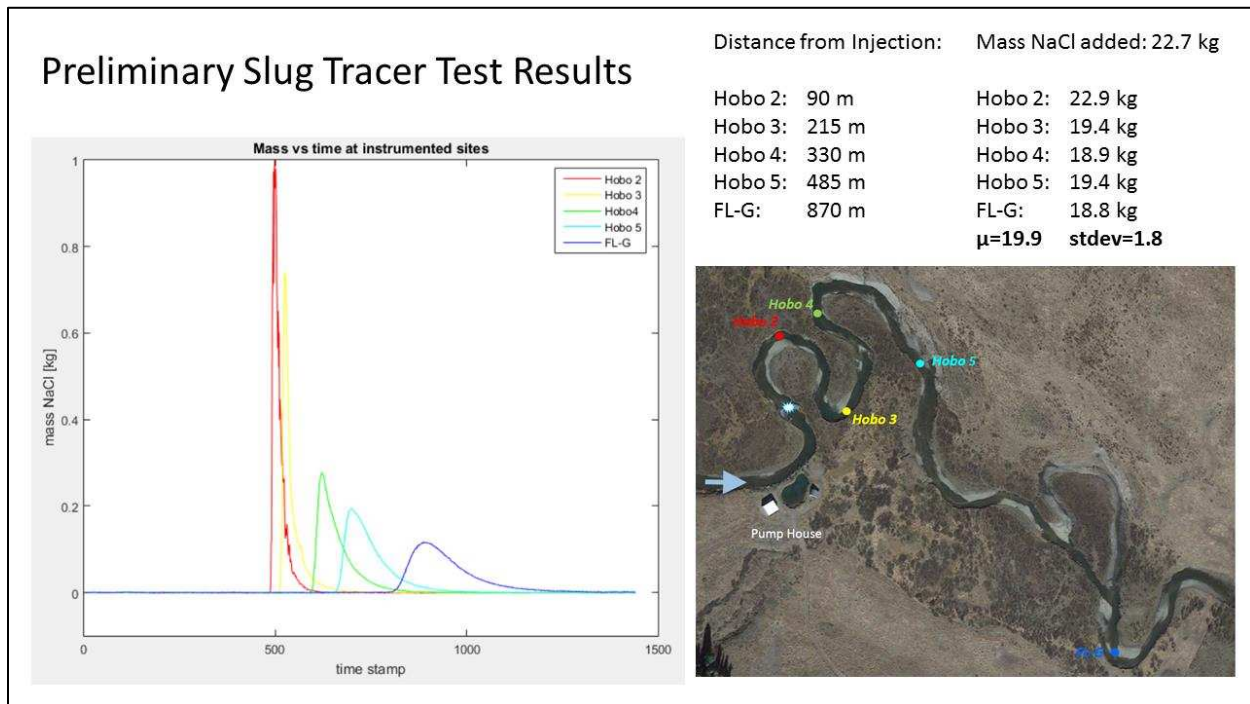


Figure D.1: Summary of NaCl concentration measurements and mass estimates instream during the August 2016 pulse tracer injection.

D.2: Summary of September 2016 continuous injection instream tracer test

September 10, 2016, we conducted a 4 hour instream, continuous tracer test along the 2007 cutoff and Meander D. We used NaCl and Rhodamine water tracer (RWT) as conservative tracers. During and after the test, RWT and NaCl concentrations were measured both instream, along the river corridor, and in piezometers located at Meander D. The following figures describe the test set up, measurements and summary of our findings is below.

Major findings and suggestions:

- 1) Although we injected tracer for 4 hours, no tracer breakthrough was observed in the wells at Meander D. The absence of tracer recorded in the well network (measured during and hours/days/weeks after the test) and the minimal in-stream loss may suggest that little tracer entered the meander during these near base-flow conditions.
- 2) The estimated total mass of RWT measured at Location 3 is greater than the mass of RWT injected (442g). We attribute this overestimation to poor mixing/too short of mixing length between our injection site and our instrument placement. The length for adequate mixing was ~ 0.3 km, the distance between the injection site (location 1 in Figure D.2) and the second instrument (location 4).
- 3) The absence of tracer in wells may be the result of the relatively short tracer injection duration and base flow conditions.

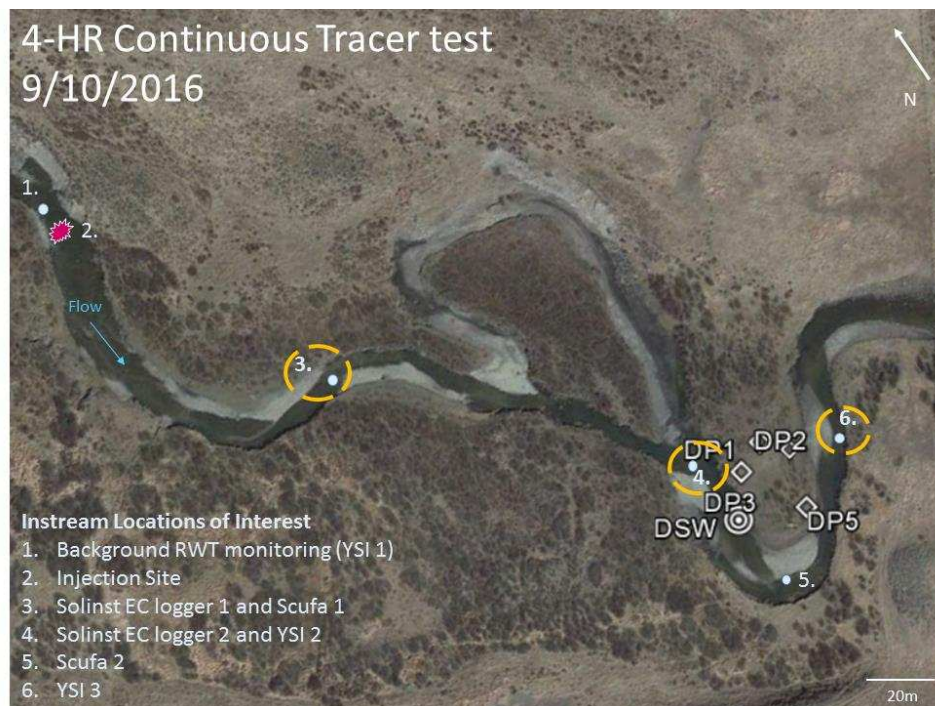


Figure D.2: Summary of tracer injection and measurement locations.

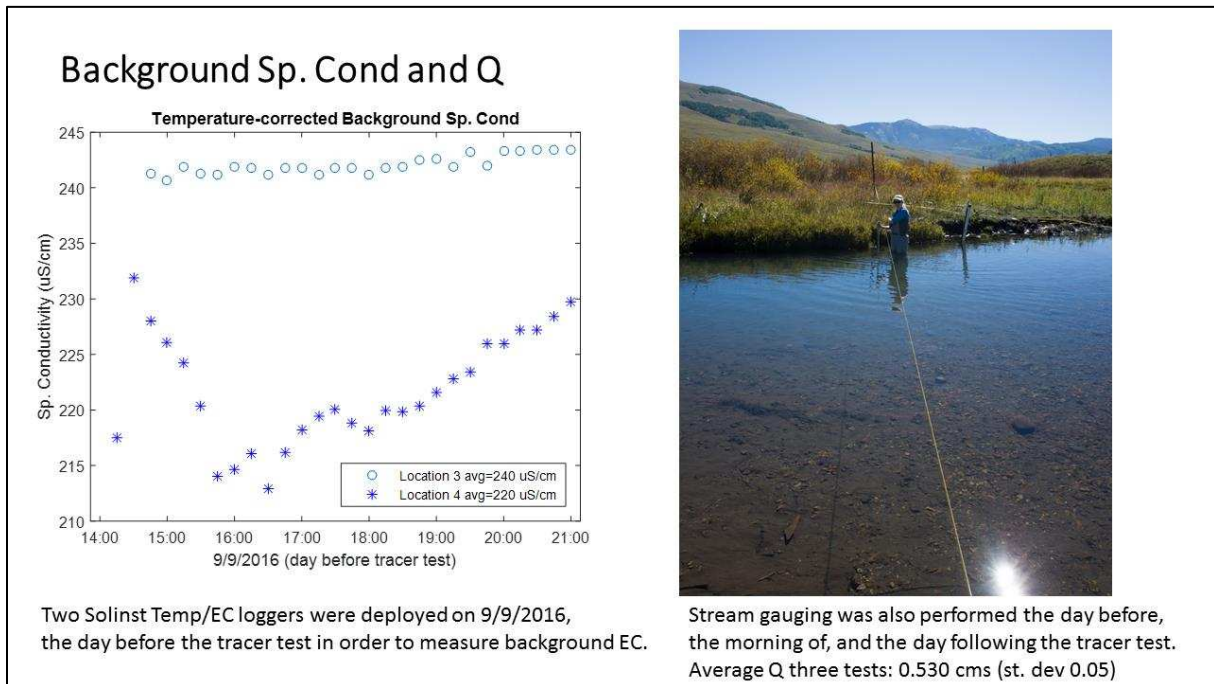


Figure D.3 Background measurements of specific conductivity. All specific conductivity measurements presented were temperature corrected via:
 $EC_{corr} = Sp.Cond / (1 + ((Temp - 25) * 2.1 / 100))$. Right panel shows stream gauging that used to estimate discharge.

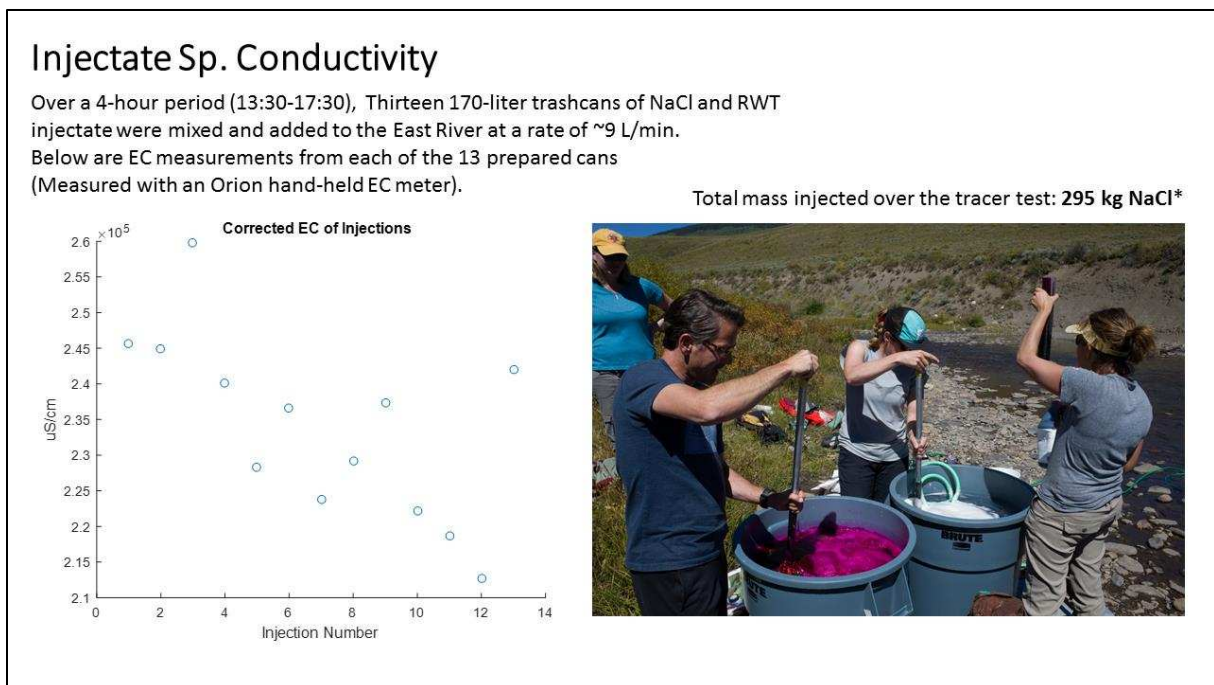


Figure D.4: Tracer injection details.

Instream [RWT] and Sp Cond Plots for all locations

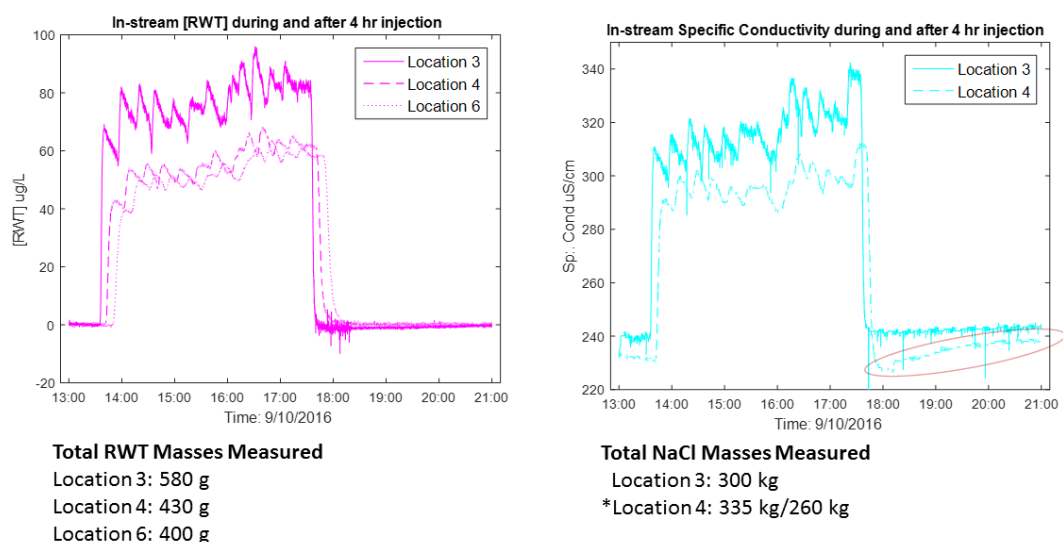


Figure D.5: Summary of [RWT] and specific conductivity (Sp Cond) measurement. We used the conversion: $0.5 \text{ uS/cm} = 1 \text{ mg/l NaCl}$ to estimate the concentration of NaCl for a give timestep.

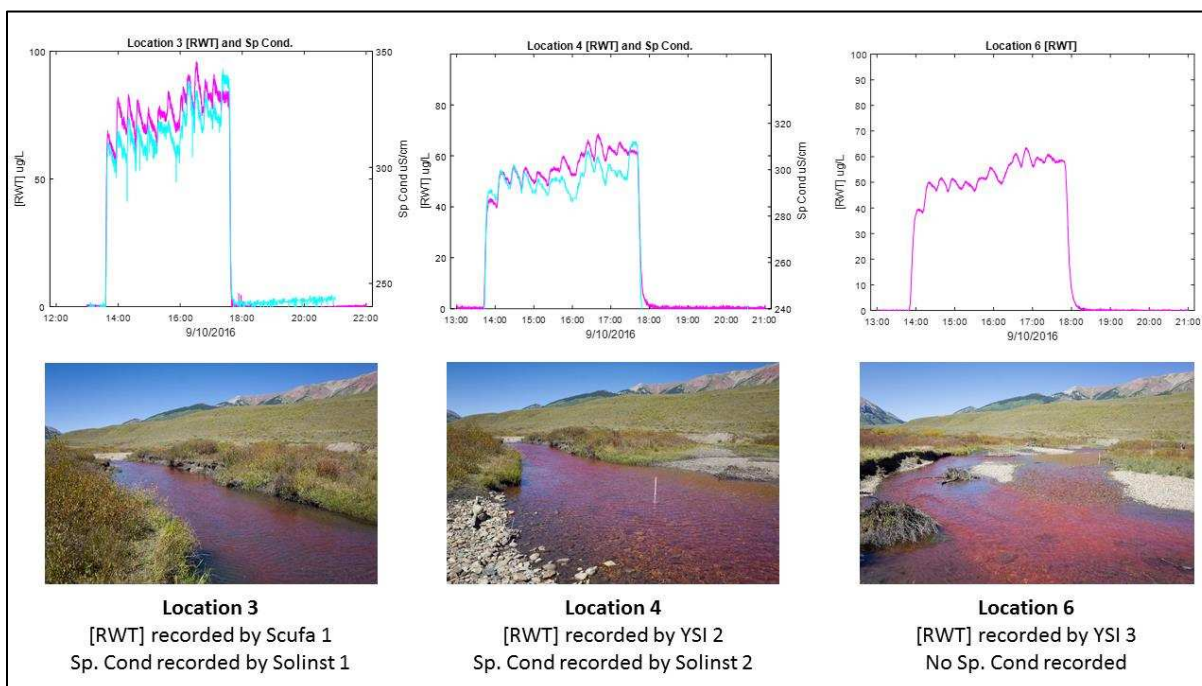


Figure D.6: Comparison of [RWT] and specific conductivity along the East River corridor during and after the tracer test.

Log-log [RWT] and log-log Sp. Cond at each location

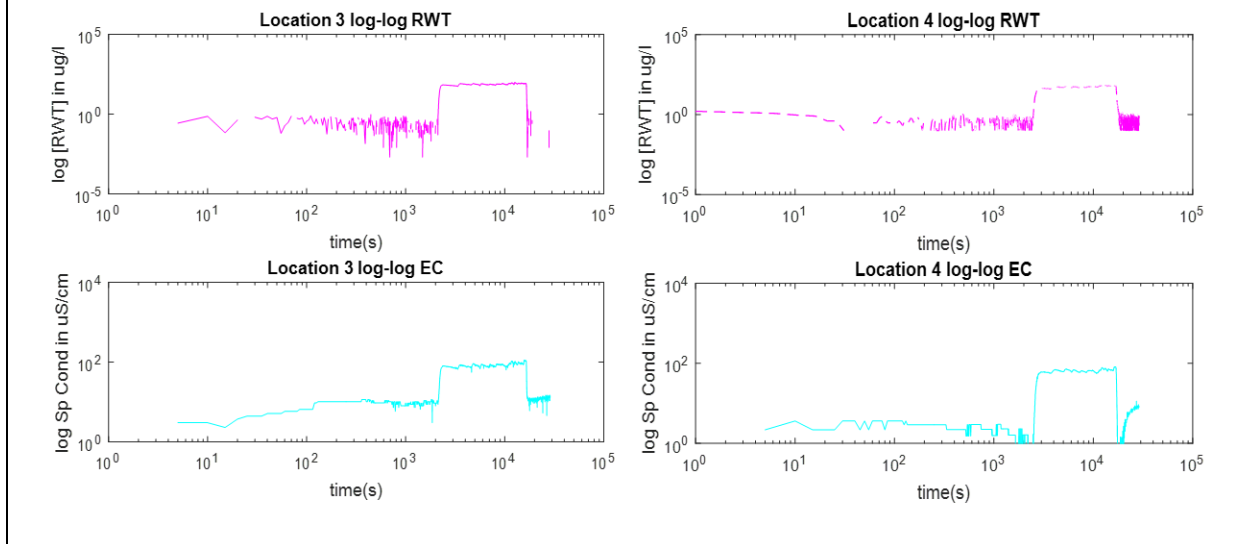


Figure D.7: Log-log plots of [RWT] and specific conductivity. Log-log plots are helpful in identifying log tails.

D.3: Summary of May 2017 down-well injection test at Meander D

In May 2017 a down-well injection test was planned and prepared for, but inclement weather prevented the full test. The goal of the project was to track tracer through Meander D, from injection down a well that was connected hydrologically to the river. To track the tracer's movement, we designed an electrical resistivity (ER) geophysical survey, coupled with additional down-well fluid electrical conductivity meters. ER methods are sensitive to the presence of the added salt tracer, and measurements will be collected both during and after the injection. The circular ER survey configuration is made up of 76 electrodes (Figure D.8) and combines transect and square arrays, facilitating the measurement of tracer location with depth across, as well as direction of tracer movement, across the survey. By directly injecting the tracer down a well that is hydrologically connected to the river, rather than in-stream, we greatly reduce the amount of salt needed.

Below is a figure of the test's layout and the test's planned components, along with notes on their success.

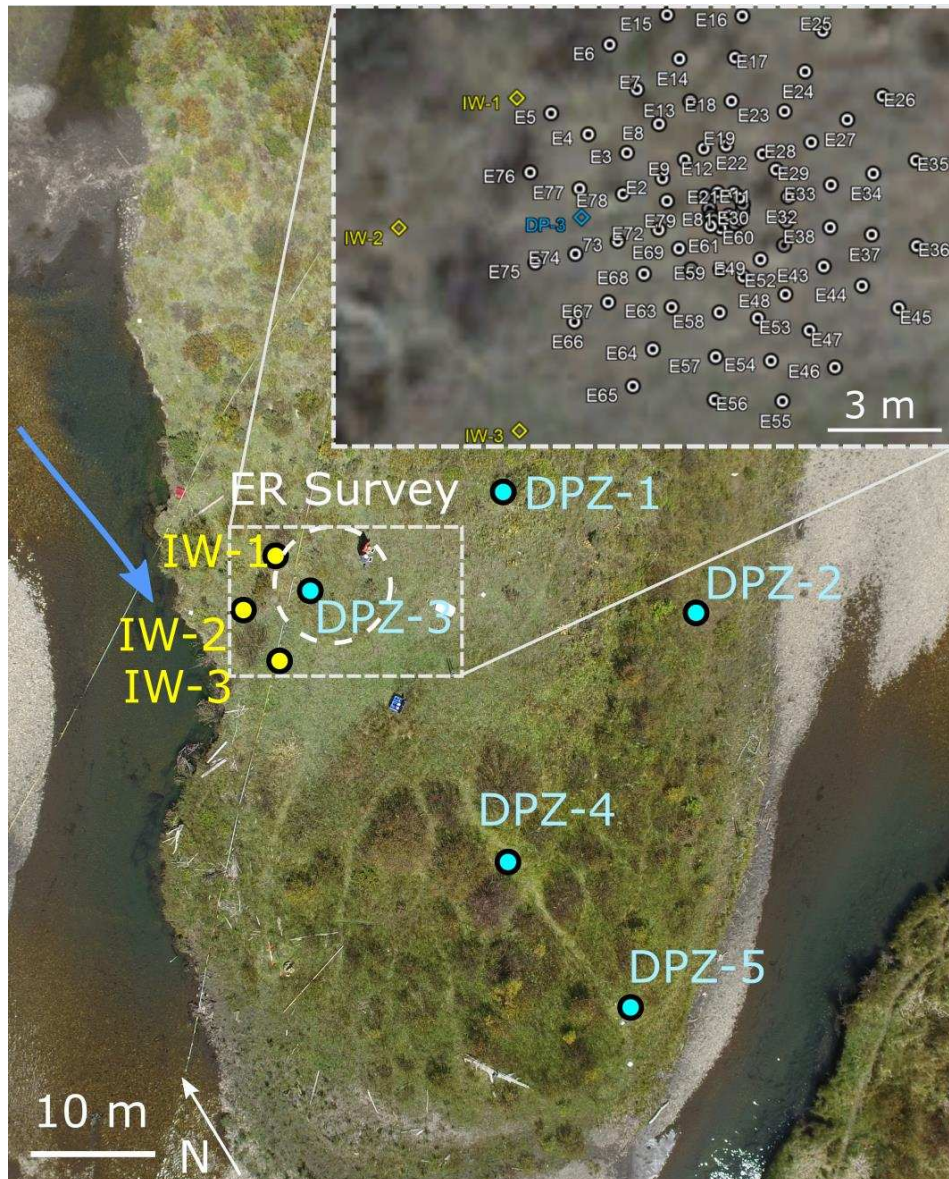


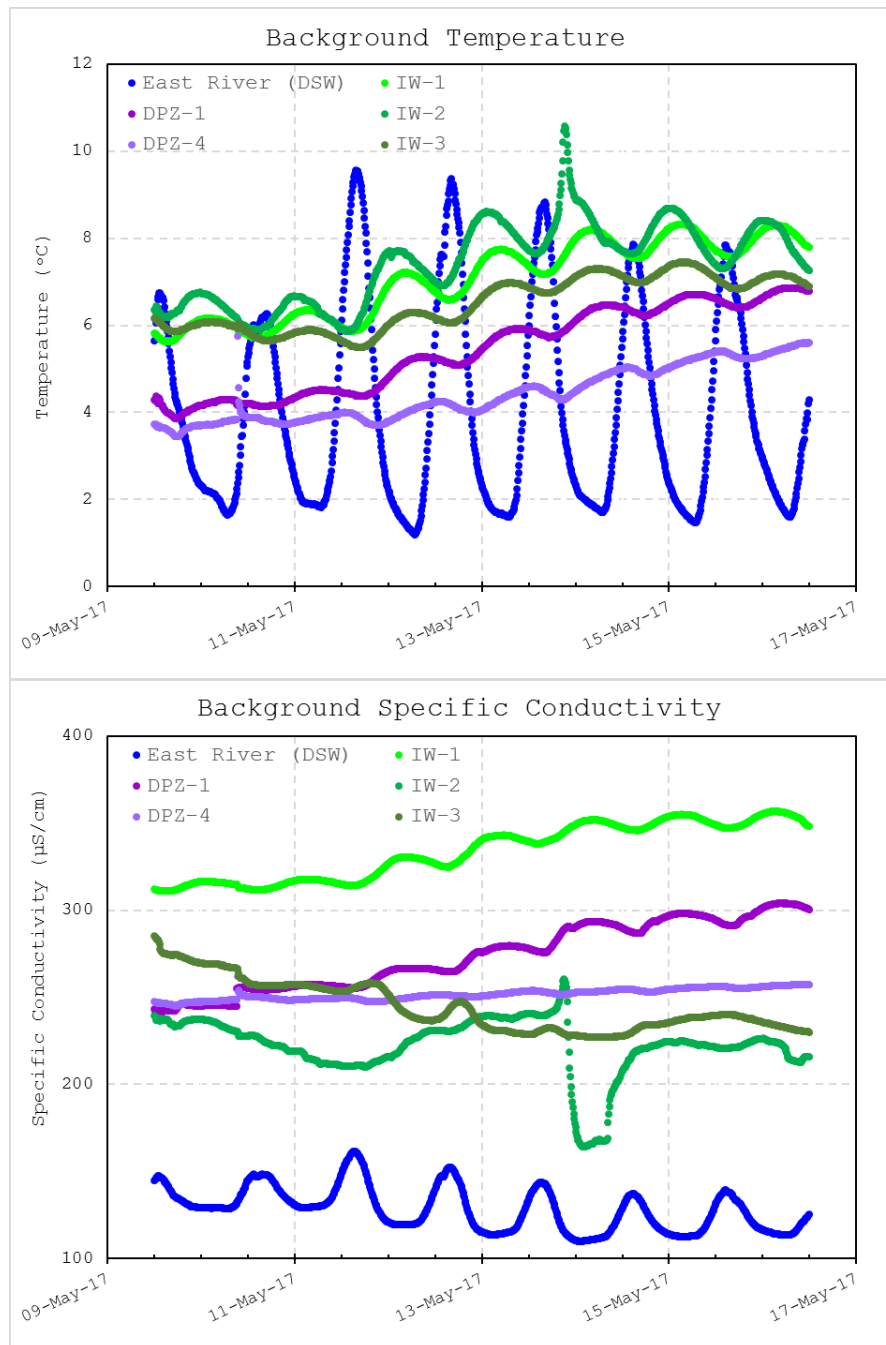
Figure D.8 Map of tracer test design at Meander D. Blue dots indicate groundwater piezometers, yellow dots indicate newly constructed wells for tracer injection, and white dotted circle indicates the extent of the ER/IP survey. Inset is the electrode configuration relative to DP-3 and the injection wells.

Test Components:

- 1) **Establish connection between surface (river) water and the groundwater in the meander subsurface: (Completed; Figure D.9)** Before beginning the tracer test, we collected ~7 days of background conductivity and temperature data of river water and groundwater using the shallow well network. Generally, surface and groundwater display differences in physical characteristics, such as temperature, conductivity, and pH. To confirm connection with the river, we installed 3 new wells (~1-2" in diameter) on the upstream side of the meander

(yellow dots on Figure D.8). In the background data the river water appears to have a signature distinct from groundwater in the piezometers, and the new injection wells (particularly IW-1) exhibit patterns indicative of mixing of the two waters.

Figure D.9 Background temperature (top) and specific conductivity (bottom) for the East



River (DSW; blue), piezometers (purples) and injection wells (greens).

- 2) Down-well tracer injection: (*Not completed*)** A salt-water mixture will be injected down the well with the strongest connection to the river. Tracer will be injected at 200 mg/L concentration, which is below the federal and state MCLs for NaCl (250 mg/l). We are not raising the entire river or groundwater concentration up to 200 mg/l, but rather introducing a small amount (1-2 gal/min) of water with a salt concentration below the MCL. (Historically, river discharge alone is typically ~200,000 to 250,000 gal/min during this time of year). Injection will be conducted continuously for approximately 4 hours. Continuous injection allows for significantly lower concentrations of salt needed because the method helps mitigate effects of dilution in the groundwater. Depending on tracer spreading or slow movement of the tracer in the geophysical survey, the length of injection, however, may be modified as needed, up to ~15 hours.
- 3) Tracer measurement: (*Not completed*)** We will track tracer movement using point measurements by conductivity meters placed in piezometers, and a geophysical survey comprised of ER/IP transects will be positioned in a circular pattern across the meander (dashed lines on Figure D.9). The ER/IP methods are sensitive to the presence of the added salt tracer, and measurements will be collected both during and after the injection. The location of injection, as well as the locations of instrumentation will be recorded using a high-precision GPS unit.

APPENDIX E

SUMMARY OF HYDROLOGIC MONITORING NETWORK

As part of my work in the East River SFA, I am currently creating data packages for these monitoring networks. Data will be available for other scientists' use. Data include well metadata, including well dimensions, locations and manual water level measurements; raw transducer and barometric pressure data, and calculated water elevations (sub-hourly and daily averages). This includes data at 5 meanders (A-E), not just the two comparison meanders outlined in this thesis.

Table E.1 Locations and notes on piezometers and stilling wells in the hydrologic monitoring networks established in the East River Watershed Function Scientific Focus area. "PZ" denotes piezometer and "baro Logger" for barometric pressure logger.

Well	Latitude	Longitude	Record of data	Notes	Well	Latitude	Longitude	Record of data	Notes
GP1	38.97795	-107.00315	Oct 2015-	PZ	CP1	38.92322	-106.95006	Oct 2015-	PZ
GP1-Air	38.97795	-107.00315	Oct 2015-	baro Logger	CP1-Air	38.92322	-106.95006	Oct 2015-	baro Logger
GP2	38.97808	-107.00303	Oct 2015-	PZ	CP2	38.92326	-106.94986	Oct 2015-Sept 2017	PZ
GP3	38.978	-107.00275	Oct 2015-	PZ	CP3	38.92321	-106.94962	Oct 2015-	PZ
GP4	38.97792	-107.00293	Oct 2015-Sept 2017	PZ	CP4	38.92314	-106.94981	Oct 2015-Sept 2017	PZ
GP5	38.97783	-107.00282	Oct 2015-	PZ	CP5	38.92304	-106.94996	Oct 2015-Sept 2017	PZ
GSW	38.97807	-107.00317	Oct 2015-	stilling well	CP6	38.92299	-106.94977	Oct 2015-	PZ
EP1	38.88702	-106.90887	Oct 2015-Sept 2017	PZ	DP1	38.92135	-106.9472	Oct 2015-Sept 2017	PZ
EP2	38.8867	-106.90874	Oct 2015-Sept 2017	PZ	DP2	38.92128	-106.9471	Oct 2015-Sept 2017	PZ
EP3	38.88674	-106.90901	Oct 2015-June 2016	PZ	DP3	38.92129	-106.94733	Oct 2015-Sept 2017	PZ
EP3-Air	38.88674	-106.90901	Oct 2015-	PZ	DP4	38.92115	-106.94722	Oct 2015-Sept 2017	PZ
EP4	38.88681	-106.90891	Oct 2015-June 2016	PZ	DP5	38.92108	-106.94715	Oct 2015-Sept 2017	PZ
EP5	38.88689	-106.90905	Oct 2015-Sept 2017	PZ	DSW	38.92115	-106.94744	Oct 2015-Dec 2016	stilling well
ESW	38.88612	-106.90874	Oct 2015-	stilling well					
FP-1	38.8861	-106.90807	Oct 2015-	PZ	AP-1	38.92334	-106.95101	June 2016-	PZ
FP-2	38.88591	-106.908	Oct 2015-	PZ	AP-2	38.92315	-106.95096	June 2016-Sept 2017	PZ
FP-3	38.88572	-106.90839	Oct 2015-	PZ	AP-3	38.92324	-106.95068	June 2016-	PZ
FP-4	38.88591	-106.90829	Oct 2015-	PZ	AP-4	38.92296	-106.95074	June 2016-	PZ
FP-5	38.88601	-106.9088	Oct 2015-	PZ					

APPENDIX F

WATER TABLE ELEVATIONS AND GRADIENTS: SUMMARIES AND MATLAB CODE

F.1: Summary of data for water table elevation plotting and gradient calculations

Table F.1: Data inputs for our water table elevation mapping and water table gradients. This table was saved as A_Gradient_Summer17.CSV and read into Matlab to run “Thesis_Gradient_Fig” and “Thesis_threePt” Matlab code (below).

A_UTM_Lat	A_UTM_Long	A_May_15_17	A_June_15_17	A_Aug_15_17
4309873.23	330920.2301	2773.733938	2773.796938	2773.560938
4309852.056	330920.2301	2773.954601	2774.017601	2773.731601
4309861.628	330945.7063	2773.789383	2773.839383	2773.377383
4309830.465	330941.6139	2774.103165	2774.173165	2773.707165

Table F.2: Data inputs for our water table elevation mapping and water table gradients. This table was saved as D_Gradient_Summer17.CSV and read into Matlab to run “Thesis_Gradient_Fig” and “Thesis_threePt” Matlab code (below).

D_UTM_Lat	D_UTM_Long	D_May_15_17	D_June_15_17	D_Aug_15_17
4309847.459	331196.257	2756.315	2756.453	2756.125
4309839.583	331205.7	2756.346	2756.416	2756.015
4309843.797	331185.165	2756.642	2756.707	2756.271
4309827.884	331193.337	2756.439	2756.565	2756.136
4309818.749	331198.819	2756.351	2756.374	2755.987

Table F.3: Summary table of water table gradients calculated using different three-point configurations at Meander A and Meander D. Although different well configurations were used at Meander A, the gradient directions were across the meander and were of similar magnitude.

3-pt Gradient calculations	Meander A		Meander D					
	1,3,4	2,3,4	1,3,5	2,3,5	1,3,4	1,2,4	2,4,5	1,4,5
May	0.01	0.01	0.028	0.014	0.029	0.006	0.011	0.22
June	0.011	0.011	0.023	0.014	0.022	0.009	0.022	0.36
Aug	0.012	0.012	0.015	0.013	0.014	0.011	0.017	0.23

Table F.4 Summary of flow path lengths (lengths correspond to blue dashed lines in Figure 4.2); average pore velocity estimates (v); and estimated residence time of lateral hyporheic exchange (t_H) and Meander A and Meander D. At meander A June and July estimates are lumped due to similarity between the two months' data.

using AVG _{geo} K	June July A	Aug A	June D (Back)	June D (Apex)	July D (Back)	July D (Apex)	Aug D (Back)	Aug D (Apex)
flow path length [m]	50	40	45	30	45	30	40	30
v [m/d]	0.04	0.04	0.09	0.03	0.06	0.06	0.03	0.03
t_H [d]	1300	960	500	1000	800	500	1400	1000

F.2: Matlab code for plotting water table elevations:

```
%Plotting Water Table Elevations for meander A and Meander D; Summer 2017
%Updated by HFM 16 Dec 2017
```

```
% Import Arrays from "A_Gradient_Summer17.csv" and
% D_Gradient_Summer17.csv".
%These will contain both meanders' UTM coordinates and water levels for
%three dates in May, June, and August 2017.
```

```
%%Meander A: =====
%May-----
```

```
cy=linspace(min(A_UTM_Lat),max(A_UTM_Lat));
cx=linspace(min(A_UTM_Long),max(A_UTM_Long));
```

```
dy=linspace(min(A_UTM_Lat),max(A_UTM_Lat));
dxx=linspace(min(A_UTM_Long),max(A_UTM_Long));
```

```
[CX,CY]=meshgrid(cx,cy);
[DX,DY]=meshgrid(dxx,dy);
```

```
CP_Grid=griddata(A_UTM_Long,A_UTM_Lat,A_May_15_17,CX,CY);
DP_Grid=griddata(A_UTM_Long,A_UTM_Lat,A_May_15_17,DX,DY);
```

```
figure; contourf(cx,cy,CP_Grid); caxis([2773,2774.5]);
title('Meander A Water Elevations May 15 2017');
colorbar;xlabel('long UTM'); ylabel('lat UTM');
```

```
%June-----
```

```
cy=linspace(min(A_UTM_Lat),max(A_UTM_Lat));
cx=linspace(min(A_UTM_Long),max(A_UTM_Long));
```

```
dy=linspace(min(A_UTM_Lat),max(A_UTM_Lat));
dxx=linspace(min(A_UTM_Long),max(A_UTM_Long));
```

```
[CX,CY]=meshgrid(cx,cy);
[DX,DY]=meshgrid(dxx,dy);
```

```
CP_Grid=griddata(A_UTM_Long,A_UTM_Lat,A_June_15_17,CX,CY);
DP_Grid=griddata(A_UTM_Long,A_UTM_Lat,A_June_15_17,DX,DY);
```

```
figure; contourf(cx,cy,CP_Grid); caxis([2773,2774.5]);
title('Meander A Water Elevations June 15 2017');
colorbar;xlabel('long UTM'); ylabel('lat UTM');
```

```
%Aug-----
```

```
cy=linspace(min(A_UTM_Lat),max(A_UTM_Lat));
cx=linspace(min(A_UTM_Long),max(A_UTM_Long));
```

```

dy=linspace(min(A_UTM_Lat),max(A_UTM_Lat));
dxx=linspace(min(A_UTM_Long),max(A_UTM_Long));

[CX,CY]=meshgrid(cx,cy);
[DX,DY]=meshgrid(dxx,dy);

CP_Grid=griddata(A_UTM_Long,A_UTM_Lat,A_Aug_15_17,CX,CY);
DP_Grid=griddata(A_UTM_Long,A_UTM_Lat,A_Aug_15_17,DX,DY);

figure; contourf(cx,cy,CP_Grid); caxis([2773,2774.5]);
title('Meander A Water Elevations August 15 2017');
colorbar;xlabel('long UTM'); ylabel('lat UTM');

%%Meander D=====
%May-----

cy=linspace(min(D_UTM_Lat),max(D_UTM_Lat));
cx=linspace(min(D_UTM_Long),max(D_UTM_Long));

dy=linspace(min(D_UTM_Lat),max(D_UTM_Lat));
dxx=linspace(min(D_UTM_Long),max(D_UTM_Long));

[CX,CY]=meshgrid(cx,cy);
[DX,DY]=meshgrid(dxx,dy);

CP_Grid=griddata(D_UTM_Long,D_UTM_Lat,D_May_15_17,CX,CY);
DP_Grid=griddata(D_UTM_Long,D_UTM_Lat,D_May_15_17,DX,DY);

figure; contourf(cx,cy,CP_Grid); caxis([2755.5,2757]);
title('Meander D Water Elevations May 15 2017');
colorbar;xlabel('long UTM'); ylabel('lat UTM');

%June-----

cy=linspace(min(D_UTM_Lat),max(D_UTM_Lat));
cx=linspace(min(D_UTM_Long),max(D_UTM_Long));

dy=linspace(min(D_UTM_Lat),max(D_UTM_Lat));
dxx=linspace(min(D_UTM_Long),max(D_UTM_Long));

[CX,CY]=meshgrid(cx,cy);
[DX,DY]=meshgrid(dxx,dy);

CP_Grid=griddata(D_UTM_Long,D_UTM_Lat,D_June_15_17,CX,CY);
DP_Grid=griddata(D_UTM_Long,D_UTM_Lat,D_June_15_17,DX,DY);

figure; contourf(cx,cy,CP_Grid); caxis([2755.5,2757]);
title('Meander D Water Elevations June 15 2017');
colorbar;xlabel('long UTM'); ylabel('lat UTM');

%Aug-----

```

```

cy=linspace(min(D_UTM_Lat),max(D_UTM_Lat));
cx=linspace(min(D_UTM_Long),max(D_UTM_Long));

dy=linspace(min(D_UTM_Lat),max(D_UTM_Lat));
dxx=linspace(min(D_UTM_Long),max(D_UTM_Long));

[CX,CY]=meshgrid(cx,cy);
[DX,DY]=meshgrid(dxx,dy);

CP_Grid=griddata(D_UTM_Long,D_UTM_Lat,D_Aug_15_17,CX,CY);
DP_Grid=griddata(D_UTM_Long,D_UTM_Lat,D_Aug_15_17,DX,DY);

figure; contourf(cx,cy,CP_Grid); caxis([2755.5,2757]);
title('Meander D Water Elevations August 15 2017');
colorbar;xlabel('long UTM'); ylabel('lat UTM');

```

F.3: Matlab code for calculating and plotting water table gradients using the three-point approach

```

%Arrow plotting function needed for plotting gradient direction
%Written by K. Singha; given to HFM June 2016
%-----
function handles = plot_arrow( x1,y1,x2,y2,varargin )
%
% plot_arrow - plots an arrow to the current plot
%
% format:   handles = plot_arrow( x1,y1,x2,y2 [,options...] )
%
% input:    x1,y1   - starting point
%           x2,y2   - end point
%Options - come as pairs of "property","value" as defined for "line" and
%"patch" controls, see matlab help for listing of these properties.
% note that not all properties were added, one might add them at the end of
%this file.
%
% additional options are:
% 'headwidth': relative to complete arrow size, default value is 0.07
% 'headheight': relative to complete arrow size, default value is 0.15
% (encoded are maximal values in pixels, for the case that the arrow is very
%long)
%
% output:   handles - handles of the graphical elements building the arrow
%
% Example:  plot_arrow( -1,-1,15,12,'linewidth',2,'color',[0.5 0.5
0.5],'facecolor',[0.5 0.5 0.5] );
% plot_arrow( 0,0,5,4,'linewidth',2,'headwidth',0.25,'headheight',0.33 );
% plot_arrow; % will launch demo

% =====
% for debug - demo - can be erased
% =====
if (nargin==0)
    figure;
    axis;

```

```

        set( gca,'nextplot','add' );
        for x = 0:0.3:2*pi
            color = [rand rand rand];
            h = plot_arrow( 1,1,50*rand*cos(x),50*rand*sin(x),...
                'color',color,'facecolor',color,'edgecolor',color );
            set( h,'linewidth',2 );
        end
        hold off;
        return
    end
% =====
% end of for debug
% =====

% =====
% constants (can be edited)
% =====
alpha      = 0.15;    % head length
beta       = 0.07;    % head width
max_length = 22;
max_width  = 10;

% =====
% check if head properties are given
% =====
% if ratio is always fixed, this section can be removed!
if ~isempty( varargin )
    for c = 1:floor(length(varargin)/2)
        try
            switch lower(varargin{c*2-1})
                % head properties - do nothing, since handled above already
                case 'headheight', alpha = max( min( varargin{c*2},1 ),0.01 );
                case 'headwidth',  beta = max( min( varargin{c*2},1 ),0.01 );
                end
            catch
                fprintf( 'unrecognized property or value for: %s\n',varargin{c*2-
1} );
            end
        end
    end
end

% =====
% calculate the arrow head coordinates
% =====
den      = x2 - x1 + eps;
% make sure no division by zero occurs
teta     = atan( (y2-y1)/den ) + pi*(x2<x1) - pi/2;
% angle of arrow
cs       = cos(teta);
% rotation matrix
ss       = sin(teta);
R        = [cs -ss;ss cs];
line_length = sqrt( (y2-y1)^2 + (x2-x1)^2 );
% sizes
head_length = min( line_length*alpha,max_length );

```

```

head_width = min( line_length*beta,max_length );

x0          = x2*cs + y2*ss;
% build head coordinates
y0          = -x2*ss + y2*cs;
coords      = R*[x0 x0+head_width/2 x0-head_width/2; y0 y0-head_length y0-
head_length];

% =====
% plot arrow (= line + patch of a triangle)
% =====
h1          = plot( [x1,x2],[y1,y2],'k' );
h2          = patch( coords(1,:),coords(2,:),[0 0 0] );

% =====
% return handles
% =====
handles = [h1 h2];

% =====
% check if styling is required
% =====
% if no styling, this section can be removed!
if ~isempty( varargin )
    for c = 1:floor(length(varargin)/2)
        try
            switch lower(varargin{c*2-1})

                % only patch properties
                case 'edgecolor', set( h2,'EdgeColor',varargin{c*2} );
                case 'facecolor', set( h2,'FaceColor',varargin{c*2} );
                case 'facelighting',set( h2,'FaceLighting',varargin{c*2} );
                case 'edgelighting',set( h2,'EdgeLighting',varargin{c*2} );

                % only line properties
                case 'color'      , set( h1,'Color',varargin{c*2} );

                % shared properties
                case 'linestyle', set( handles,'LineStyle',varargin{c*2} );
                case 'linewidth', set( handles,'LineWidth',varargin{c*2} );
                case 'parent',    set( handles,'parent',varargin{c*2} );

                % head properties - do nothing, since handled above already
                case 'headwidth',;
                case 'headheight',;

            end
        catch
            fprintf( 'unrecognized property or value for: %s\n',varargin{c*2-
1} );
        end
    end
end

%Plotting Water Table Gradient via 3-point method
%Meander A and Meander D; Summer 2017

```

```

%modified by HFM 16 Dec 2017 (see below)

% Import Arrays from "A_Gradient_Summer17.csv" and
% D_Gradient_Summer17.csv".
%These will contain both meanders' UTM coordinates and water levels for
%three dates in May, June, and August 2017.

%%IMPORTANT%% To run this file, you must first run the plot arrow.m script

%%Meander A: =====
%May-----

% function threepoint
% written by K. Singha, last edit 6/2016
% modified by E.B. Voytek, 6/2016
% modified by H.F. Malenda for East River data 12/2017
% insert x,z locations and head values of the three wells
xloc=[A_UTM_Long(1), A_UTM_Long(3), A_UTM_Long(4)];
zloc=[A_UTM_Lat(1), A_UTM_Lat(3), A_UTM_Lat(4)];
heads=[A_May_15_17(1), A_May_15_17(3), A_May_15_17(4)];

% plot well locations
figure; scatter(xloc,zloc,200,'filled')
axis equal, box on

% find change in head and distances between all pairs of wells
for i=1:length(xloc)-1;
    dx(i)=sqrt((xloc(i)-xloc(i+1)).^2+(zloc(i)-zloc(i+1)).^2);
    dh(i)=heads(i)-heads(i+1);
    theta(i)=atan((xloc(i)-xloc(i+1))./(zloc(i)-zloc(i+1)));
end
dx(i+1)=sqrt((xloc(1)-xloc(i+1)).^2+(zloc(1)-zloc(i+1)).^2);
dh(i+1)=heads(1)-heads(i+1);
theta(i+1)=atan((xloc(1)-xloc(i+1))./(zloc(1)-zloc(i+1)));

% calculate the apparent gradient between wells
gradient=abs(dh./dx);

% find the wells with highest and lowest head, and the intermediate point
lowhead=find(heads==min(heads));
hihead=find(heads==max(heads));
midhead=setxor([hihead,lowhead],[1:3]);

if hihead~=3 & lowhead~=3;
    ind=1;
elseif hihead~=2 & lowhead~=2;
    ind=3;
elseif hihead~=1 & lowhead~=1;
    ind=2;
end

% find the distance along the line between the lowest and highest head
% where the mid head is, and plot that location
finddist=(heads(midhead)-heads(lowhead))./gradient(ind);
xmid=sign(zloc(hihead)-zloc(lowhead))*sin(theta(ind))*finddist;

```

```

zmid=sign(zloc(hihead)-zloc(lowhead))*cos(theta(ind))*finddist;
xlocxpt=xloc(lowhead)+xmid;
zlocxpt=zloc(lowhead)+zmid;
hold on
plot(xlocxpt,zlocxpt,'x');

% draw the equipotential (dotted line), calculate the flow direction
plot([xloc(midhead) xlocxpt],[zloc(midhead) zlocxpt],'--k');
newangle=atan((xlocxpt-xloc(midhead))./(zlocxpt-zloc(midhead)));

if sign(zloc(hihead)-zloc(lowhead)) == 1;
direction=rad2deg(newangle)+90;
elseif sign(zloc(hihead)-zloc(lowhead)) == -1;
direction=rad2deg(newangle)+270;
end

% calculate the true gradient and plot the direction of flow
finddist2=cos(newangle-(pi/2+theta(ind)))*sqrt((xloc(hihead)-
xlocxpt).^2+(zloc(hihead)-zlocxpt).^2);
dhdx=(heads(hihead)-heads(midhead))/finddist2;
xflow=sign(zloc(hihead)-zloc(lowhead))*sin(pi/2-newangle)*finddist2;
zflow=sign(zloc(hihead)-zloc(lowhead))*cos(pi/2-newangle)*finddist2;
plot([xloc(hihead)+xflow],[zloc(hihead)-zflow],'x');
plot_arrow(xloc(hihead),zloc(hihead),xloc(hihead)+xflow,zloc(hihead)-
zflow,...
'linewidth',3,'color','r','edgecolor','r','facecolor','r');
% label gradient and direction of flow angle in title
title(['Gradient: ',num2str(dhdx,'%4.3f'),' ; Angle: ',
num2str(direction,'%3.0f')])

%June-----

% insert x,z locations and head values of the three wells
xloc=[A_UTM_Long(1), A_UTM_Long(3), A_UTM_Long(4)];
zloc=[A_UTM_Lat(1), A_UTM_Lat(3), A_UTM_Lat(4)];
heads=[A_June_15_17(1), A_June_15_17(3), A_June_15_17(4)];

% plot well locations
figure;
scatter(xloc,zloc,200,'filled')
axis equal, box on

% find change in head and distances between all pairs of wells
for i=1:length(xloc)-1;
dx(i)=sqrt((xloc(i)-xloc(i+1)).^2+(zloc(i)-zloc(i+1)).^2);
dh(i)=heads(i)-heads(i+1);
theta(i)=atan((xloc(i)-xloc(i+1))./(zloc(i)-zloc(i+1)));
end
dx(i+1)=sqrt((xloc(1)-xloc(i+1)).^2+(zloc(1)-zloc(i+1)).^2);
dh(i+1)=heads(1)-heads(i+1);
theta(i+1)=atan((xloc(1)-xloc(i+1))./(zloc(1)-zloc(i+1)));

% calculate the apparent gradient between wells
gradient=abs(dh./dx);

```

```

% find the wells with highest and lowest head, and the intermediate point
lowhead=find(heads==min(heads));
hihead=find(heads==max(heads));
midhead=setxor([hihead,lowhead],[1:3]);

if hihead~=3 & lowhead~=3;
    ind=1;
elseif hihead~=2 & lowhead~=2;
    ind=3;
elseif hihead~=1 & lowhead~=1;
    ind=2;
end

% find the distance along the line between the lowest and highest head
% where the mid head is, and plot that location
finddist=(heads(midhead)-heads(lowhead))./gradient(ind);
xmid=sign(zloc(hihead)-zloc(lowhead))*sin(theta(ind))*finddist;
zmid=sign(zloc(hihead)-zloc(lowhead))*cos(theta(ind))*finddist;
xlocxpt=xloc(lowhead)+xmid;
zlocxpt=zloc(lowhead)+zmid;
hold on
plot(xlocxpt,zlocxpt,'x');

% draw the equipotential (dotted line), calculate the flow direction
plot([xloc(midhead) xlocxpt],[zloc(midhead) zlocxpt],'--k');
newangle=atan((xlocxpt-xloc(midhead))./(zlocxpt-zloc(midhead)));

if sign(zloc(hihead)-zloc(lowhead)) == 1;
    direction=rad2deg(newangle)+90;
elseif sign(zloc(hihead)-zloc(lowhead)) == -1;
    direction=rad2deg(newangle)+270;
end

% calculate the true gradient and plot the direction of flow
finddist2=cos(newangle-(pi/2+theta(ind)))*sqrt((xloc(hihead)-
xlocxpt).^2+(zloc(hihead)-zlocxpt).^2);
dhdx=(heads(hihead)-heads(midhead))/finddist2;
xflow=sign(zloc(hihead)-zloc(lowhead))*sin(pi/2-newangle)*finddist2;
zflow=sign(zloc(hihead)-zloc(lowhead))*cos(pi/2-newangle)*finddist2;
plot([xloc(hihead)+xflow],[zloc(hihead)-zflow],'x');
plot_arrow(xloc(hihead),zloc(hihead),xloc(hihead)+xflow,zloc(hihead)-
zflow,...
    'linewidth',3,'color','r','edgecolor','r','facecolor','r');
% label gradient and direction of flow angle in title
title(['Gradient: ',num2str(dhdx,'%4.3f'),''; Angle:
',num2str(direction,'%3.0f')])

%August-----

% insert x,z locations and head values of the three wells
xloc=[A_UTM_Long(1), A_UTM_Long(3), A_UTM_Long(4)];
zloc=[A_UTM_Lat(1), A_UTM_Lat(3), A_UTM_Lat(4)];
heads=[A_Aug_15_17(1), A_Aug_15_17(3), A_Aug_15_17(4)];

```

```

% plot well locations
figure;
scatter(xloc,zloc,200,'filled')
axis equal, box on

% find change in head and distances between all pairs of wells
for i=1:length(xloc)-1;
    dx(i)=sqrt((xloc(i)-xloc(i+1)).^2+(zloc(i)-zloc(i+1)).^2);
    dh(i)=heads(i)-heads(i+1);
    theta(i)=atan((xloc(i)-xloc(i+1))./(zloc(i)-zloc(i+1)));
end
dx(i+1)=sqrt((xloc(1)-xloc(i+1)).^2+(zloc(1)-zloc(i+1)).^2);
dh(i+1)=heads(1)-heads(i+1);
theta(i+1)=atan((xloc(1)-xloc(i+1))./(zloc(1)-zloc(i+1)));

% calculate the apparent gradient between wells
gradient=abs(dh./dx);

% find the wells with highest and lowest head, and the intermediate point
lowhead=find(heads==min(heads));
hihead=find(heads==max(heads));
midhead=setxor([hihead,lowhead],[1:3]);

if hihead~=3 & lowhead~=3;
    ind=1;
elseif hihead~=2 & lowhead~=2;
    ind=3;
elseif hihead~=1 & lowhead~=1;
    ind=2;
end

% find the distance along the line between the lowest and highest head
% where the mid head is, and plot that location
finddist=(heads(midhead)-heads(lowhead))./gradient(ind);
xmid=sign(zloc(hihead)-zloc(lowhead))*sin(theta(ind))*finddist;
zmid=sign(zloc(hihead)-zloc(lowhead))*cos(theta(ind))*finddist;
xlocxpt=xloc(lowhead)+xmid;
zlocxpt=zloc(lowhead)+zmid;
hold on
plot(xlocxpt,zlocxpt,'x');

% draw the equipotential (dotted line), calculate the flow direction
plot([xloc(midhead) xlocxpt],[zloc(midhead) zlocxpt],'--k');
newangle=atan((xlocxpt-xloc(midhead))./(zlocxpt-zloc(midhead)));

if sign(zloc(hihead)-zloc(lowhead)) == 1;
    direction=rad2deg(newangle)+90;
elseif sign(zloc(hihead)-zloc(lowhead)) == -1;
    direction=rad2deg(newangle)+270;
end

% calculate the true gradient and plot the direction of flow

```

```

finddist2=cos(newangle-(pi/2+theta(ind)))*sqrt((xloc(hihead)-
xlocxpt).^2+(zloc(hihead)-zlocxpt).^2);
dhdx=(heads(hihead)-heads(midhead))/finddist2;
xflow=sign(zloc(hihead)-zloc(lowhead))*sin(pi/2-newangle)*finddist2;
zflow=sign(zloc(hihead)-zloc(lowhead))*cos(pi/2-newangle)*finddist2;
plot([xloc(hihead)+xflow],[zloc(hihead)-zflow],'x');
plot_arrow(xloc(hihead),zloc(hihead),xloc(hihead)+xflow,zloc(hihead)-
zflow,...
    'linewidth',3,'color','r','edgecolor','r','facecolor','r');
% label gradient and direction of flow angle in title
title(['Gradient: ',num2str(dhdx,'%4.3f'),''; Angle:
',num2str(direction,'%3.0f')])

%%Meander D: =====
%May-----

% function threepoint
% written by K. Singha, last edit 6/2016
% modified by E.B. Voytek, 6/2016
% insert x,z locations and head values of the three wells
xloc=[D_UTM_Long(1), D_UTM_Long(5), D_UTM_Long(2)];
zloc=[D_UTM_Lat(1), D_UTM_Lat(5),D_UTM_Lat(2)];
heads=[D_May_15_17(1),D_May_15_17(5),D_May_15_17(2)];

% plot well locations
figure; scatter(xloc,zloc,200,'filled')
axis equal, box on

% find change in head and distances between all pairs of wells
for i=1:length(xloc)-1;
    dx(i)=sqrt((xloc(i)-xloc(i+1)).^2+(zloc(i)-zloc(i+1)).^2);
    dh(i)=heads(i)-heads(i+1);
    theta(i)=atan((xloc(i)-xloc(i+1))./(zloc(i)-zloc(i+1)));
end
dx(i+1)=sqrt((xloc(1)-xloc(i+1)).^2+(zloc(1)-zloc(i+1)).^2);
dh(i+1)=heads(1)-heads(i+1);
theta(i+1)=atan((xloc(1)-xloc(i+1))./(zloc(1)-zloc(i+1)));

% calculate the apparent gradient between wells
gradient=abs(dh./dx);

% find the wells with highest and lowest head, and the intermediate point
lowhead=find(heads==min(heads));
hihead=find(heads==max(heads));
midhead=setxor([hihead,lowhead],[1:3]);

if hihead~=3 & lowhead~=3;
    ind=1;
elseif hihead~=2 & lowhead~=2;
    ind=3;
elseif hihead~=1 & lowhead~=1;
    ind=2;
end

% find the distance along the line between the lowest and highest head

```

```

% where the mid head is, and plot that location
finddist=(heads(midhead)-heads(lowhead))./gradient(ind);
xmid=sign(zloc(hihead)-zloc(lowhead))*sin(theta(ind))*finddist;
zmid=sign(zloc(hihead)-zloc(lowhead))*cos(theta(ind))*finddist;
xlocxpt=xloc(lowhead)+xmid;
zlocxpt=zloc(lowhead)+zmid;
hold on
plot(xlocxpt,zlocxpt,'x');

% draw the equipotential (dotted line), calculate the flow direction
plot([xloc(midhead) xlocxpt],[zloc(midhead) zlocxpt],'--k');
newangle=atan((xlocxpt-xloc(midhead))./(zlocxpt-zloc(midhead)));

if sign(zloc(hihead)-zloc(lowhead)) == 1;
direction=rad2deg(newangle)+90;
elseif sign(zloc(hihead)-zloc(lowhead)) == -1;
direction=rad2deg(newangle)+270;
end

% calculate the true gradient and plot the direction of flow
finddist2=cos(newangle-(pi/2+theta(ind)))*sqrt((xloc(hihead)-
xlocxpt).^2+(zloc(hihead)-zlocxpt).^2);
dhdx=(heads(hihead)-heads(midhead))/finddist2;
xflow=sign(zloc(hihead)-zloc(lowhead))*sin(pi/2-newangle)*finddist2;
zflow=sign(zloc(hihead)-zloc(lowhead))*cos(pi/2-newangle)*finddist2;
plot([xloc(hihead)+xflow],[zloc(hihead)-zflow],'x');
plot_arrow(xloc(hihead),zloc(hihead),xloc(hihead)+xflow,zloc(hihead)-
zflow,...
'linewidth',3,'color','r','edgecolor','r','facecolor','r');
% label gradient and direction of flow angle in title
title(['Gradient: ',num2str(dhdx,'%4.3f'),' ; Angle:
',num2str(direction,'%3.0f')])

%June-----

% insert x,z locations and head values of the three wells
xloc=[D_UTM_Long(1), D_UTM_Long(5), D_UTM_Long(2)];
zloc=[D_UTM_Lat(1), D_UTM_Lat(5), D_UTM_Lat(2)];
heads=[D_June_15_17(1), D_June_15_17(5), D_June_15_17(2)];

% plot well locations
figure;
scatter(xloc,zloc,200,'filled')
axis equal, box on

% find change in head and distances between all pairs of wells
for i=1:length(xloc)-1;
dx(i)=sqrt((xloc(i)-xloc(i+1)).^2+(zloc(i)-zloc(i+1)).^2);
dh(i)=heads(i)-heads(i+1);
theta(i)=atan((xloc(i)-xloc(i+1))./(zloc(i)-zloc(i+1)));
end
dx(i+1)=sqrt((xloc(1)-xloc(i+1)).^2+(zloc(1)-zloc(i+1)).^2);
dh(i+1)=heads(1)-heads(i+1);
theta(i+1)=atan((xloc(1)-xloc(i+1))./(zloc(1)-zloc(i+1)));

```

```

% calculate the apparent gradient between wells
gradient=abs(dh./dx);

% find the wells with highest and lowest head, and the intermediate point
lowhead=find(heads==min(heads));
hihead=find(heads==max(heads));
midhead=setxor([hihead,lowhead],[1:3]);

if hihead~=3 & lowhead~=3;
    ind=1;
elseif hihead~=2 & lowhead~=2;
    ind=3;
elseif hihead~=1 & lowhead~=1;
    ind=2;
end

% find the distance along the line between the lowest and highest head
% where the mid head is, and plot that location
finddist=(heads(midhead)-heads(lowhead))./gradient(ind);
xmid=sign(zloc(hihead)-zloc(lowhead))*sin(theta(ind))*finddist;
zmid=sign(zloc(hihead)-zloc(lowhead))*cos(theta(ind))*finddist;
xlocxpt=xloc(lowhead)+xmid;
zlocxpt=zloc(lowhead)+zmid;
hold on
plot(xlocxpt,zlocxpt,'x');

% draw the equipotential (dotted line), calculate the flow direction
plot([xloc(midhead) xlocxpt],[zloc(midhead) zlocxpt],'--k');
newangle=atan((xlocxpt-xloc(midhead))./(zlocxpt-zloc(midhead)));

if sign(zloc(hihead)-zloc(lowhead)) == 1;
    direction=rad2deg(newangle)+90;
elseif sign(zloc(hihead)-zloc(lowhead)) == -1;
    direction=rad2deg(newangle)+270;
end

% calculate the true gradient and plot the direction of flow
finddist2=cos(newangle-(pi/2+theta(ind)))*sqrt((xloc(hihead)-
xlocxpt).^2+(zloc(hihead)-zlocxpt).^2);
dhdx=(heads(hihead)-heads(midhead))/finddist2;
xflow=sign(zloc(hihead)-zloc(lowhead))*sin(pi/2-newangle)*finddist2;
zflow=sign(zloc(hihead)-zloc(lowhead))*cos(pi/2-newangle)*finddist2;
plot([xloc(hihead)+xflow],[zloc(hihead)-zflow],'x');
plot_arrow(xloc(hihead),zloc(hihead),xloc(hihead)+xflow,zloc(hihead)-
zflow,...
    'linewidth',3,'color','r','edgecolor','r','facecolor','r');
% label gradient and direction of flow angle in title
title(['Gradient: ',num2str(dhdx,'%4.3f'),''; Angle:
',num2str(direction,'%3.0f')])

%August-----

% insert x,z locations and head values of the three wells

```

```

xloc=[D_UTM_Long(1), D_UTM_Long(5), D_UTM_Long(2)];
zloc=[D_UTM_Lat(1), D_UTM_Lat(5), D_UTM_Lat(2)];
heads=[D_Aug_15_17(1), D_Aug_15_17(5), D_Aug_15_17(2)];

% plot well locations
figure;
scatter(xloc,zloc,200,'filled')
axis equal, box on

% find change in head and distances between all pairs of wells
for i=1:length(xloc)-1;
    dx(i)=sqrt((xloc(i)-xloc(i+1)).^2+(zloc(i)-zloc(i+1)).^2);
    dh(i)=heads(i)-heads(i+1);
    theta(i)=atan((xloc(i)-xloc(i+1))./(zloc(i)-zloc(i+1)));
end
dx(i+1)=sqrt((xloc(1)-xloc(i+1)).^2+(zloc(1)-zloc(i+1)).^2);
dh(i+1)=heads(1)-heads(i+1);
theta(i+1)=atan((xloc(1)-xloc(i+1))./(zloc(1)-zloc(i+1)));

% calculate the apparent gradient between wells
gradient=abs(dh./dx);

% find the wells with highest and lowest head, and the intermediate point
lowhead=find(heads==min(heads));
hihead=find(heads==max(heads));
midhead=setxor([hihead,lowhead],[1:3]);

if hihead~=3 & lowhead~=3;
    ind=1;
elseif hihead~=2 & lowhead~=2;
    ind=3;
elseif hihead~=1 & lowhead~=1;
    ind=2;
end

% find the distance along the line between the lowest and highest head
% where the mid head is, and plot that location
finddist=(heads(midhead)-heads(lowhead))./gradient(ind);
xmid=sign(zloc(hihead)-zloc(lowhead))*sin(theta(ind))*finddist;
zmid=sign(zloc(hihead)-zloc(lowhead))*cos(theta(ind))*finddist;
xlocxpt=xloc(lowhead)+xmid;
zlocxpt=zloc(lowhead)+zmid;
hold on
plot(xlocxpt,zlocxpt,'x');

% draw the equipotential (dotted line), calculate the flow direction
plot([xloc(midhead) xlocxpt],[zloc(midhead) zlocxpt],'--k');
newangle=atan((xlocxpt-xloc(midhead))./(zlocxpt-zloc(midhead)));

if sign(zloc(hihead)-zloc(lowhead)) == 1;
    direction=rad2deg(newangle)+90;
elseif sign(zloc(hihead)-zloc(lowhead)) == -1;
    direction=rad2deg(newangle)+270;
end

```

```

% calculate the true gradient and plot the direction of flow
finddist2=cos(newangle-(pi/2+theta(ind)))*sqrt((xloc(hihead)-
xlocxpt).^2+(zloc(hihead)-zlocxpt).^2);
dhdx=(heads(hihead)-heads(midhead))/finddist2;
xflow=sign(zloc(hihead)-zloc(lowhead))*sin(pi/2-newangle)*finddist2;
zflow=sign(zloc(hihead)-zloc(lowhead))*cos(pi/2-newangle)*finddist2;
plot([xloc(hihead)+xflow],[zloc(hihead)-zflow],'x');
plot_arrow(xloc(hihead),zloc(hihead),xloc(hihead)+xflow,zloc(hihead)-
zflow,...
    'linewidth',3,'color','r','edgecolor','r','facecolor','r');
% label gradient and direction of flow angle in title
title(['Gradient: ',num2str(dhdx,'%4.3f'),' ; Angle:
',num2str(direction,'%3.0f')])

```