Characterizing Streamflow Intermittency and Subsurface Heterogeneity in the Middle Arkansas River Basin

Kansas Geological Survey Open-File Report 2021-1

Authors: Kyle Compare[1,2], Samuel C. Zipper[1], Chi Zhang[3,4], Erin Seybold[1]

1. Kansas Geological Survey, University of Kansas, Lawrence, Kansas, USA
2. Department of Earth, Ocean & Atmospheric Science, Florida State University, Tallahassee, Florida, USA
3. Department of Geology, University of Kansas, Lawrence, Kansas, USA
4. Department of Meteorology and Geophysics, University of Vienna, Vienna, Austria

Abstract

Non-perennial streams are widespread but understudied relative to their perennial counterparts. In this study, we investigated the flow and intermittency regimes for the Arkansas River near Larned using historical streamflow, groundwater level, and climate data. We found that the river shifted between a dry regime (characterized by no flow apart from rapid responses to precipitation events) and wet regime (with near-continuous flow) several times over the past two decades. Wet and dry regimes were associated with wetter and drier than average climate conditions at the annual time scale but were not as responsive to seasonal (three-month) climate conditions. The alluvial aquifer exhibits a rapid, flashy response to precipitation, but longer-term (annual) climate appears to sustain groundwater levels in the alluvial aquifer, with wet regimes occurring when alluvial aquifer water levels rise above the streambed elevation. We sought to explore the relationship between aquifer dynamics and surface water intermittency by investigating subsurface heterogeneity. To investigate the subsurface, we conducted electrical resistivity tomography (ERT) surveys and compared ERT results to forward modeling of different subsurface hydrostratigraphic configurations. We found the best agreement between ERT and forward models for a conceptual model that included interbedded silt lenses in the alluvial aquifer, which agrees with past work, though there was substantial uncertainty in the ERT surveys due to the coarse and dry streambank sediment. Furthermore, the ERT profiles did not reach sufficient depth to characterize the confining layer separating the alluvial aquifer from the underlying High Plains aquifer, so potential exchange between these two units remains uncertain.

Funding and Acknowledgments: This work was funded by a Kansas Water Resources Institute award to Erin Seybold, Sam Zipper, and Chi Zhang. Kyle Compare was a 2020 participant in the Kansas Geological Survey's Geohydrology Internship Program (http://www.kgs.ku.edu/Hydro/gipIndex.html). We appreciate assistance and insights from Jim Butler, Ed Reboulet, Brownie Wilson, and Tom Glose.

1. Introduction

Non-perennial streams are widespread, representing more than 50% of the global stream network by some estimates (Datry et al., 2014). However, non-perennial streams are understudied relative to their perennial counterparts, particularly with regard to interactions between groundwater and surface water. With some exceptions (Batlle-Aguilar et al., 2015; Fuchs et al., 2019; Quichimbo et al., 2020; Shanafield et al., 2012), the majority of studies quantifying interactions between groundwater and surface water have focused on perennially flowing systems, in which surface water is always present. As a result, relatively less is known about the drivers of flow and groundwater-surface water interactions in non-perennial streams.

The drivers of surface flow in non-perennial streams are diverse (Shanafield et al., 2021) and can broadly be classified into storage-driven flow and precipitation-driven flow. In storage-driven flow, the water table rises to or above the bottom of the streambed, causing the streams to transition from a losing to gaining state and water to flow from the aquifer into the stream channel; these types of streams are often termed "intermittent" (Busch et al., 2020). In contrast, for precipitation-driven flow, streamflow follows precipitation events; these types of streams are often termed "ephemeral" (Busch et al., 2020). The same stream may even have different flow mechanisms at different times of year (Zimmer and McGlynn, 2017). Within Kansas, non-perennial streams have become more common over the past half-century coincident with the expansion of irrigated agriculture (fig. 1b), suggesting that this intermittency may be driven by changes in groundwater storage.

Recent work has suggested that there are frequent and potentially important groundwater-surface water interactions in non-perennial streams, even in losing streams with water tables meters below the bottom of the streambed (Quichimbo et al., 2020; Shanafield et al., 2012). Previous work around the world has documented a high degree of spatial variability in infiltration and recharge, both within and among non-perennial stream reaches (Shanafield and Cook, 2014), which theoretical work has suggested may be driven by spatial variability in streambed hydraulic properties (Noorduijn et al., 2014). However, while non-perennial streams are thought to be potentially important sources of focused recharge in Kansas and elsewhere, attempts to develop regional-scale estimates of groundwater recharge from non-perennial streams is hampered by challenges in upscaling point-based infiltration flux measurements to reach- and network-scale estimates (Cuthbert and Mackay, 2013; Rau et al., 2017).

Here, we attempt to improve understanding of the drivers of streamflow, extent of groundwater-surface water exchange, and subsurface heterogeneity at the Kansas Geological Survey (KGS) Larned Research Site, which is a non-perennial reach of the Arkansas River near Larned, Kansas (fig. 1). Specifically, this report has three goals: (i) characterize temporal patterns of streamflow intermittency, (ii) investigate potential interactions between groundwater and surface water systems during both flowing and non-flowing conditions, and (iii) understand how aquifer heterogeneity relates to stream intermittency and groundwater-surface water exchange.

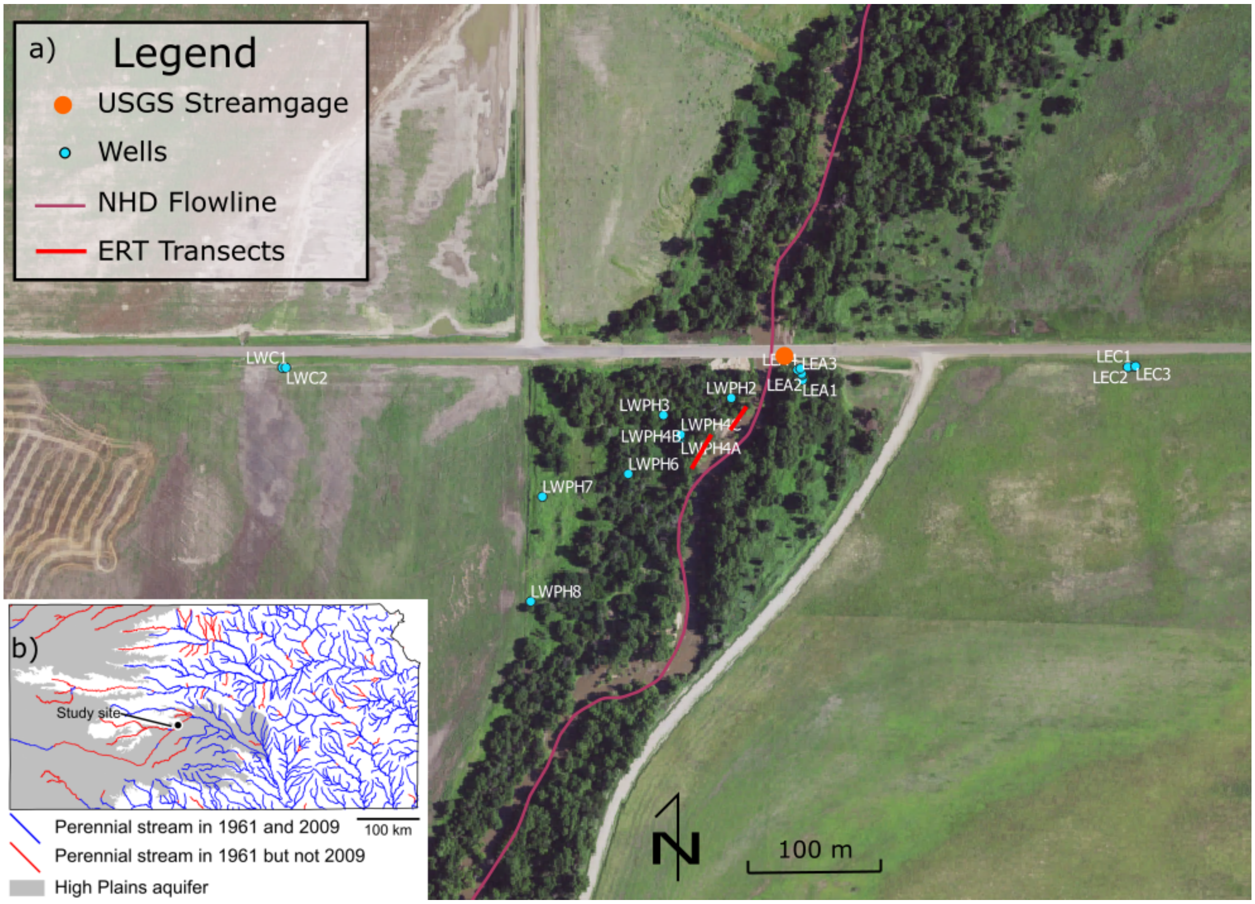

Figure 1. (a) Map of study site and (b) map showing widespread loss of perennial streams in western Kansas (modified from Zimmer et al., 2020).

2. Study Site

The KGS Larned Research Site is located about 10 kilometers northeast of Larned, Kansas, where O Road crosses the Arkansas River in Pawnee County (fig. 1). Rural farmland, including both irrigated and non-irrigated farmland, surrounds the research site. This site is co-located with a U.S. Geological Survey (USGS) gaging station (gage 07141220), which has been active since 1998. The KGS has been conducting research at this site since 2001, and past work has focused on characterizing the hydrostratigraphy of the site (Butler et al., 2004; Healey et al., 2001; Vienken et al., 2013), evapotranspiration by phreatophytic vegetation (Butler et al., 2007; Loheide et al., 2005), and aquifer responses to barometric pressure variations (Butler et al., 2011). The primary hydrostratigraphic units at the site are a surficial alluvial aquifer, a leaky confining unit, and the High Plains aquifer. The alluvial aquifer is composed of mixed gravels and sands with occasional clay layers, extending to about 10 meters below the surface (Healey et al., 2001). The leaky confining unit is composed of clay and is several meters thick, with variability across the site (Healey et al., 2001). The leaky confining unit divides the local alluvial aquifer from the regional, confined High Plains aquifer. The High Plains aquifer is composed of sand and gravel that is estimated to be about 30 meters thick at this location.

Our project builds on past work by characterizing the streamflow intermittency and investigating the hydrologic exchange among hydrostratigraphic units and the heterogeneity of sediments within the alluvial aquifer. We hope to improve the understanding of the subsurface at this site through the analysis of direct-push well data in conjunction with electrical resistivity tomography surveys to more clearly see the heterogeneity within the shallow alluvial aquifer.

3. Characterizing streamflow intermittency

3.1 The historical streamflow and intermittency regime

3.1.1 Methods:

Surficial flow of the Arkansas River at the Larned Field Site is monitored at the Arkansas River near Larned USGS gage (gage # 07141220), located at the base of the O'Rourke Bridge crossing the Arkansas River. We collected all discharge and stage data from the onset of monitoring in 1998 to the end of 2019. Since we observed a number of short (single-day) dry and wet events, we used a threshold of at least three consecutive days with a discharge measurement of zero to identify discrete dry periods in the historical flow data. The first instance of flow following three days of zero flow conditions indicated the end of that discrete dry period. We analyzed this data to determine the timing (both onset and cessation) and duration of dry periods within a year. We also performed an automated baseflow separation on the streamflow data to separate out high-frequency streamflow variation ("quickflow," inferred to be from rapid event-based flowpaths such as surface runoff) from the slowly varying component of streamflow ("baseflow," inferred to be from slowly varying sources of water such as groundwater). This is performed using the EcoHydRology package in R (Fuka et al., 2014), which uses a one-parameter digital filter (Lyne and Hollick, 1979):

$$f_k = a\,f_{k-1} + \frac{1+a}{2}\,[y_k - y_{k-1}] \qquad \text{Eq. 1}$$

where, f_k is the filtered quick response at sampling interval k, y_k is the original streamflow, and a is the filter parameter (set at a = 0.925; Nathan and McMahon, 1990). The digital filter loops over the daily streamflow data three times (forward/backward/forward) to provide a final result of estimated daily baseflow and quickflow.

3.1.2 Results and Discussion

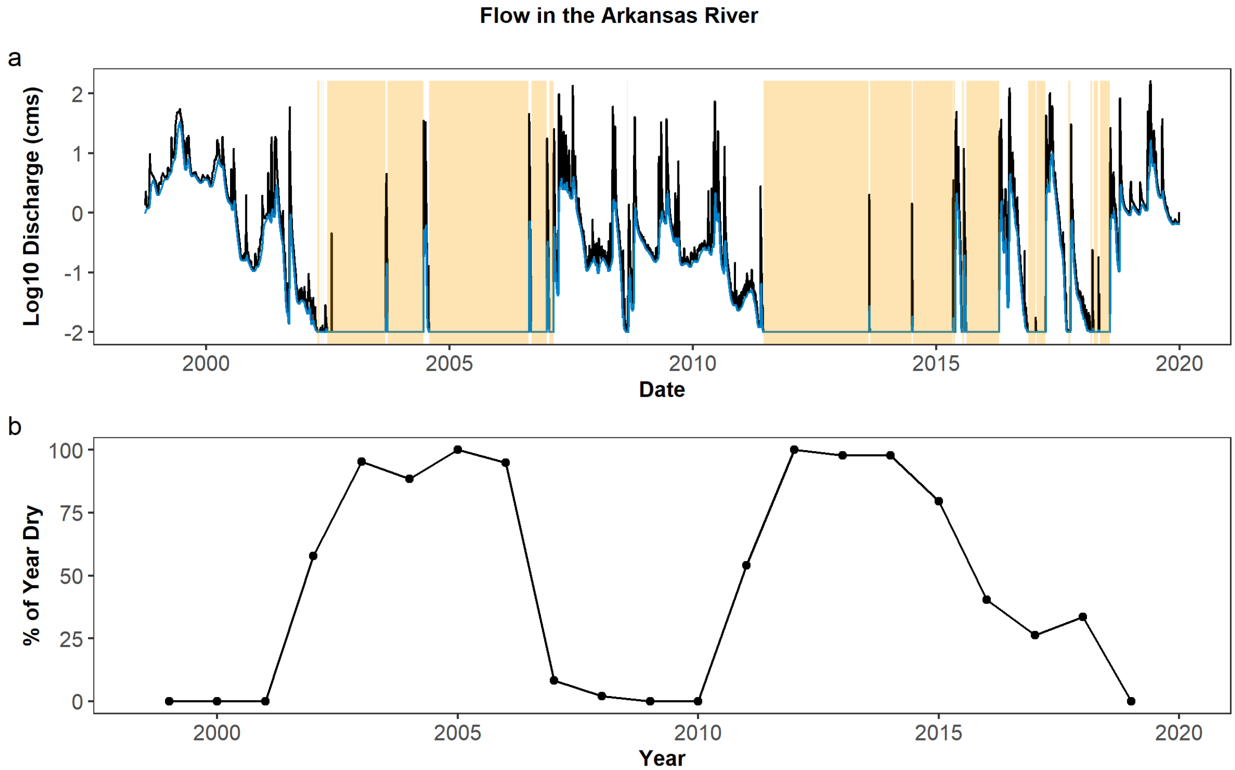

Figure 2. (a) Logarithmic hydrograph of the Arkansas River at Larned (1998–2019), with zero flow readings plotted as 10^{-2} cms. Streamflow is plotted as a black line and baseflow is plotted in blue. The highlighted sections show dry periods. (b) Annual percentage of the calendar year with zero flow.

Historically, there has been large variation in both the magnitude of streamflow and timing and duration of dry periods, with some dry periods lasting only days while others extend multiple years (fig. 2). Discrete dry periods are defined as the interval over which the stream gage records zero flow for greater than three days. There are two “dry regimes” (2003–2006, 2011–2016; fig. 2), which we defined as multi-year periods during which the river is predominantly dry with no sustained flow lasting longer than 30 days. These dry regimes are in contrast to wet regimes, in which the river is predominantly flowing (1998–2001, 2007–2010, 2019; fig. 2), though these may be broken up by the occasional dry day. There is also a transitional period between dry and wet regimes (2016–2018), where the river oscillates more rapidly between flowing and no-flow conditions.

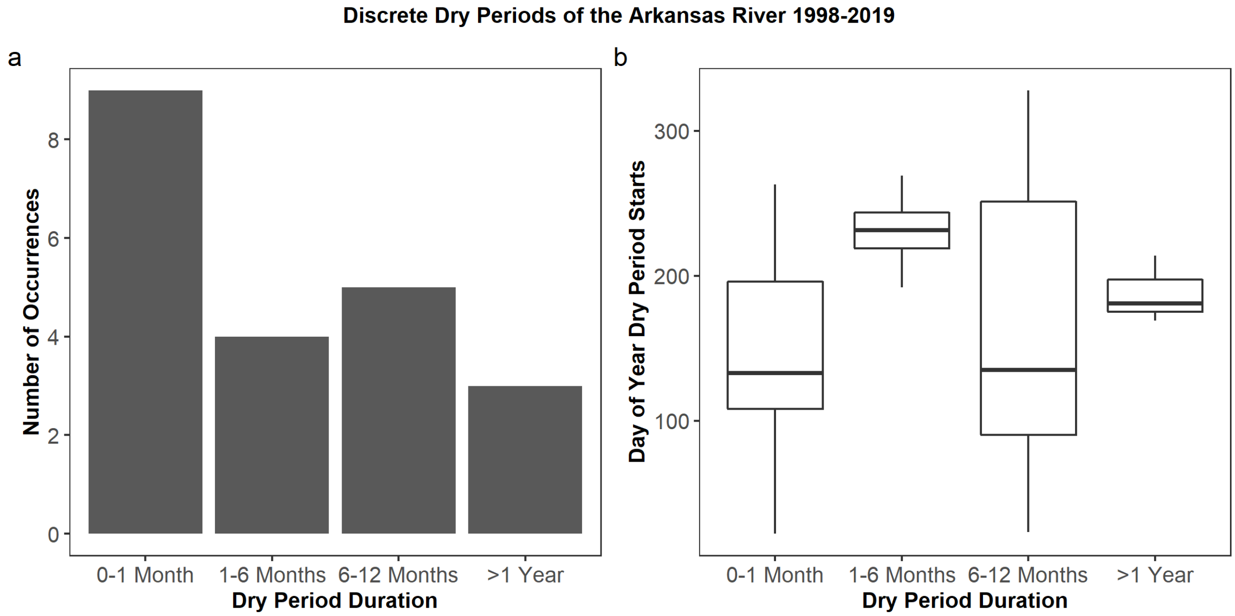

Figure 3. (a) The distribution of discrete dry periods based upon the length of the dry period, and (b) the distribution of the day of year on which dry periods begin, separated by the length of the dry period.

Although the river switches between relatively long periods dominated by wet and dry conditions, the majority of discrete dry periods are short (fig. 3b). The most common length of a dry period is less than 1 month (nine events; 43% of all dry periods), followed by 6–12 months (five events; 24%), 1–6 months (four events; 19%), and more than 1 year (three events; 14%). The shortest observed dry period was four days in length, and the longest was 789 days in length. There are also seasonal patterns in the typical onset of dry periods of different durations (fig. 3c). Extended dry periods (greater than one year), while less frequent, typically begin in the summer (late June to early August). Shorter dry periods, which are more common, have much more variability in the day of year on which they begin. On average, these shorter dry periods tend to begin in early spring (mid-May to mid-June).

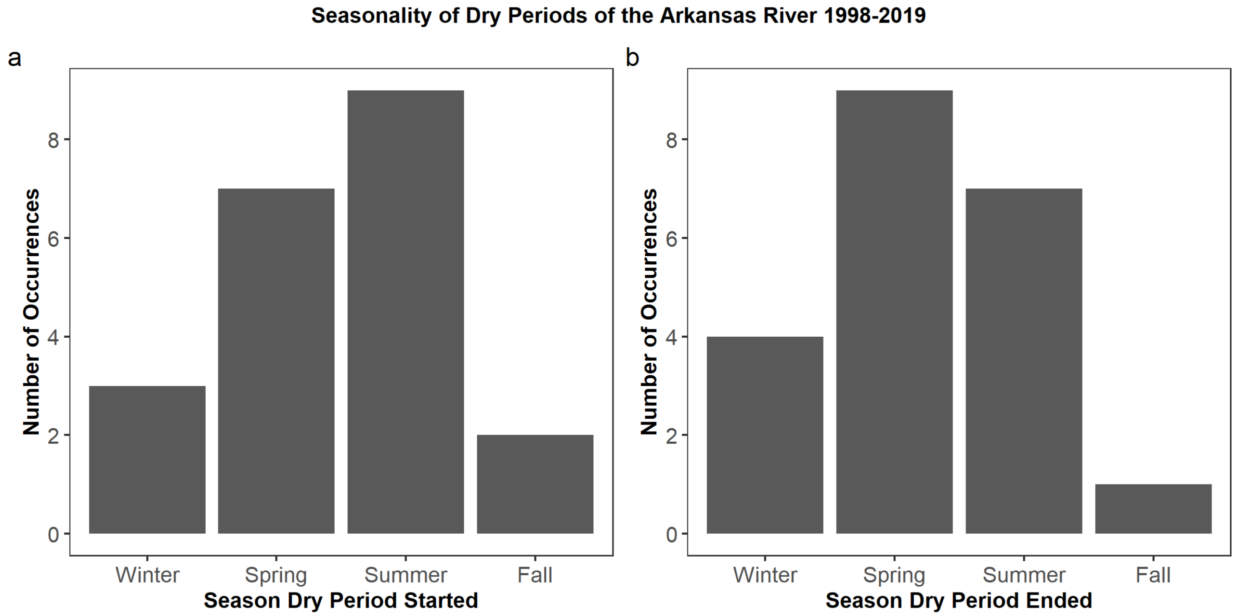

Figure 4. (a) The distribution of the seasons in which discrete dry periods begin and (b) the distribution of seasons in which discrete dry periods end.

Combining all dry periods of any length, dry periods most commonly begin and end in the summer and spring (fig. 4). The similarity between the most common start and end of dry periods is due to most dry periods having a short duration (fig. 3) and therefore ending in the same season in which they begin. The frequent onset of dry periods in the spring and summer may be explained by more evapotranspiration and crop irrigation during these seasons, both of which can reduce groundwater levels (Butler et al., 2007, 2020). The summer season also tends to have the most precipitation (fig. 6b), which may lead to short-duration flow and therefore frequent onset and cessation of dry periods.

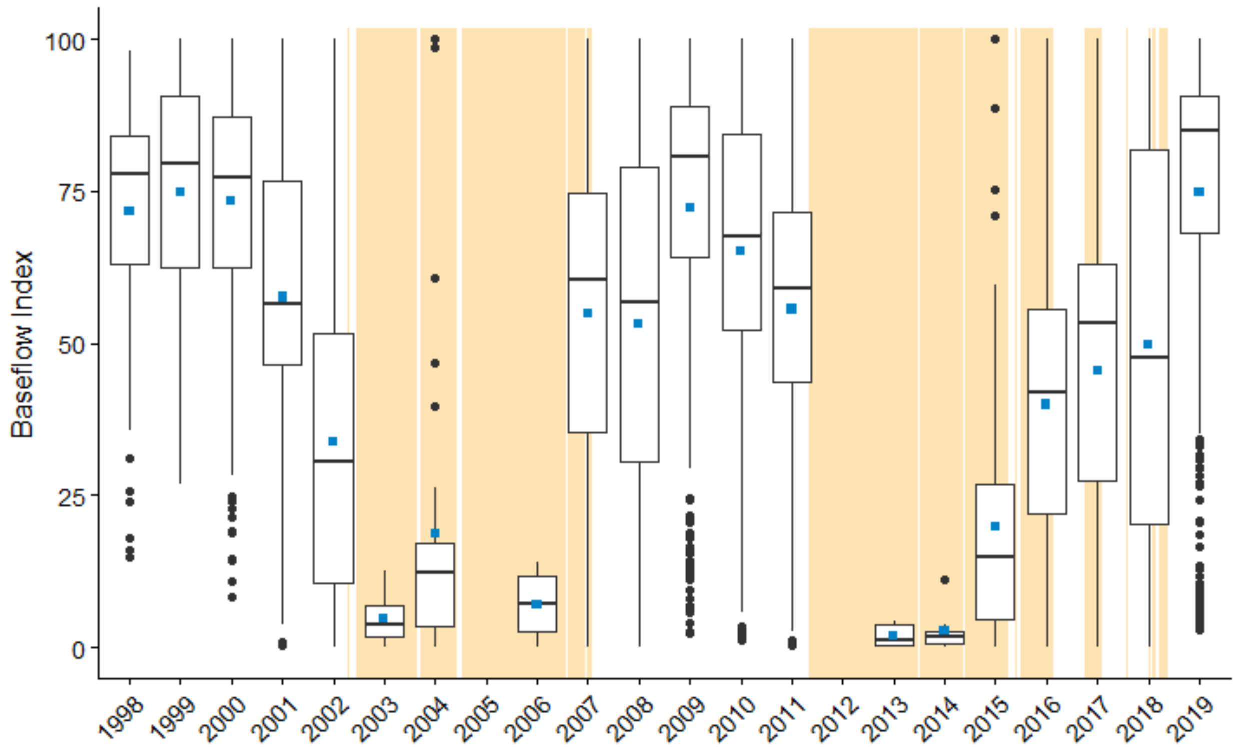

Figure 5. Baseflow index by year (boxplot of daily baseflow index with colored dot representing annual mean baseflow index).

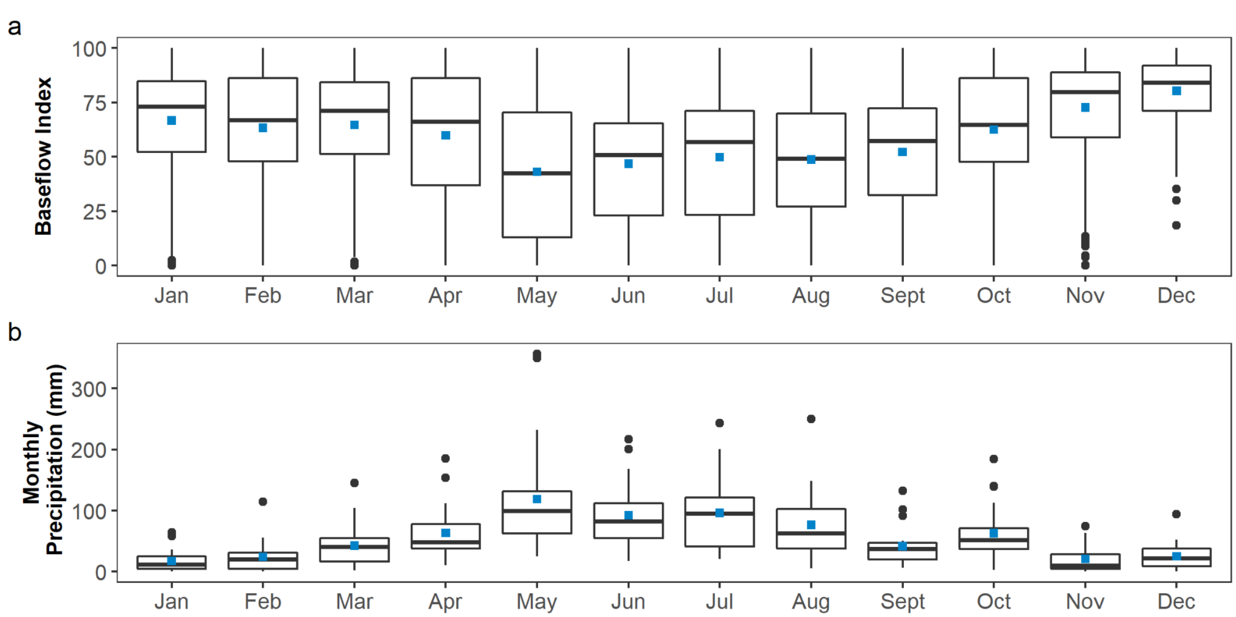

Figure 6. (a) Baseflow index by month across all years (colored dot representing monthly mean) and (b) monthly precipitation across all years (colored dot representing monthly mean).

To specifically examine the contribution of groundwater to streamflow, we use the baseflow as a proxy for the magnitude of groundwater contributions to streamflow. Multi-year dry regimes exhibited lower average baseflow indices (2001–2006, 2011–2015; fig. 5). This may be the result of either higher quickflow volumes (indicating that the short-duration flow is driven by precipitation events) and/or lower baseflow (indicating groundwater does not contribute substantially to streamflow). Inversely, the years during a wet regime exhibit higher baseflow indices, showing that groundwater has an important role in sustaining continuous flow at this site. This relationship seems to indicate that a higher baseflow index correlates with a flowing regime and lower baseflow index corresponds to a dry regime.

Relating this phenomenon to the season that initiates a dry regime, the winter and fall months have higher baseflow indices than the spring and summer months (fig. 6a) and also initiate the least number of dry regimes (fig. 4a). This reinforces the claim that baseflow helps sustain the river and prevent complete drying. This may indicate that the river system has a storage-dominated flow regime at this site, in which flow can be sustained during climatically dry periods if groundwater levels are sufficiently high.

3.2 Potential drivers of intermittency dynamics

3.2.1 Methods

To better understand the streamflow and drying dynamics described in Section 3.1, we explored their relationship with indicators of climatic conditions and human impacts. For climatic data, we compiled daily minimum temperature, maximum temperature, and precipitation data from four nearby weather stations for the same time period that streamflow data were available. The weather stations used were NOAA Global Historical Climatology Network Daily (GHCND) stations USC00141530, USC00141531, USC00147192, and USC00143218, all within a 20-mile radius of the field site.

To evaluate the climatic conditions through time, we used the Standardized Precipitation Evapotranspiration Index (Vicente-Serrano et al., 2009), which is based on the monthly water balance between precipitation and potential evapotranspiration:

$$D_i = P_i - PET_i \qquad \text{Eq. 2}$$

where D_i is a surplus ($D > 0$) or deficit ($D < 0$) of water, P_i is precipitation, and PET_i is calculated potential evapotranspiration. In this case, the Hargreaves equation was used to estimate PET_i with the SPEI R package (Beguería and Vicente-Serrano, 2013). The SPEI normalizes monthly anomalies using a log-logistic transformation to develop a unitless drought indicator that can be compared across space and time. A value of zero indicates that a month is at approximately average conditions relative to that month in all other years, negative values indicate dry conditions, and positive values indicate wet conditions. Since droughts of different durations may have different hydrological impacts, D can be calculated across different timescales. In the case of a three-month timescale, for instance, the cumulative D for the month of interest and the preceding two months are combined prior to normalization. Here, we calculated 3-month and 12-month SPEI to reflect shorter-term (seasonal) and longer-term (annual) conditions, respectively. It is important to note that, while SPEI is commonly referred to as a “drought index,” it can be used to calculate both dryness (SPEI < 0) and wetness (SPEI > 0) relative to average conditions.

The two potential human impacts we considered were local irrigation and upstream reservoir construction. The area surrounding the Larned Research Site is largely agricultural. As a result, groundwater extraction for irrigation has a major influence on groundwater levels in the High Plains aquifer at this location (Butler et al., 2020), but it is less clear how pumping affects the alluvial aquifer. To quantify pumped groundwater volumes, we compiled data for water wells within a 10 km radius of the study site from the KGS Water Information Management and Analysis System (WIMAS) database, and we compared yearly cumulative pumping to intermittency dynamics.

Upstream, the construction of Horsethief Reservoir is the main change over the time interval of interest. This reservoir was constructed in 2009 by damming Buckner Creek, west of Jetmore, Kansas. Buckner Creek flows into the Pawnee River, which in turn flows into the Arkansas River. To determine whether the construction of this reservoir impacted flow at the Larned Research Site, we used double-mass curves to compare flow at the Arkansas River near the Larned gage to two USGS gages along the Pawnee River: the Pawnee River near Burdett (gage # 07140850) and the Pawnee River near Rozel (gage # 07141200). The Burdett gage is upstream of the confluence with Buckner Creek and therefore unaffected by Horsethief Reservoir, while the Rozel gage is downstream of the confluence and therefore is affected by Horsethief Reservoir. Both gages are more than 35 km upstream (as the crow flies) of the junction between the Pawnee and the Arkansas.

3.2.2 Results and Discussion

Disentangling the impacts of pumping and climate is challenging because these two potential drivers are related to each other — there is a clear inverse relationship between annual precipitation (fig. 8a) and annual groundwater withdrawals (fig. 8b). Overall, this relationship has an R^2 of 0.34 (fig. 9). This makes logical sense as most of the pumping is for agriculture and therefore pumping needs are greatest during dry climatic periods.

Surprisingly, there is not a strong correlation between either annual precipitation or annual groundwater withdrawals and percent of the year dry. There is almost no linear correlation between these parameters (fig. 9a and 9c). Closer inspection of fig. 8, however, reveals a potential indirect relationship where years with combined low precipitation and high pumping precede the initiation of a dry regime. For example, in 2002 there was both low precipitation and high groundwater pumping followed by the start of the first dry regime the next year. The same occurs in 2011 with the second dry regime. This pattern supports the inference that dry regimes are preceded by years with combined low precipitation and high pumping.

Multi-year time lags also may be important for transitions from dry regimes to wet regimes. The year 2004 has the highest amount of precipitation and low pumping, but the Arkansas River does not transition into a flowing regime until 2007. Similar to what was seen in the dry regime, it seems that it takes multiple years of high precipitation and low pumping (2004–2006, 2013–2016) to transition from a dry regime to a wet regime.

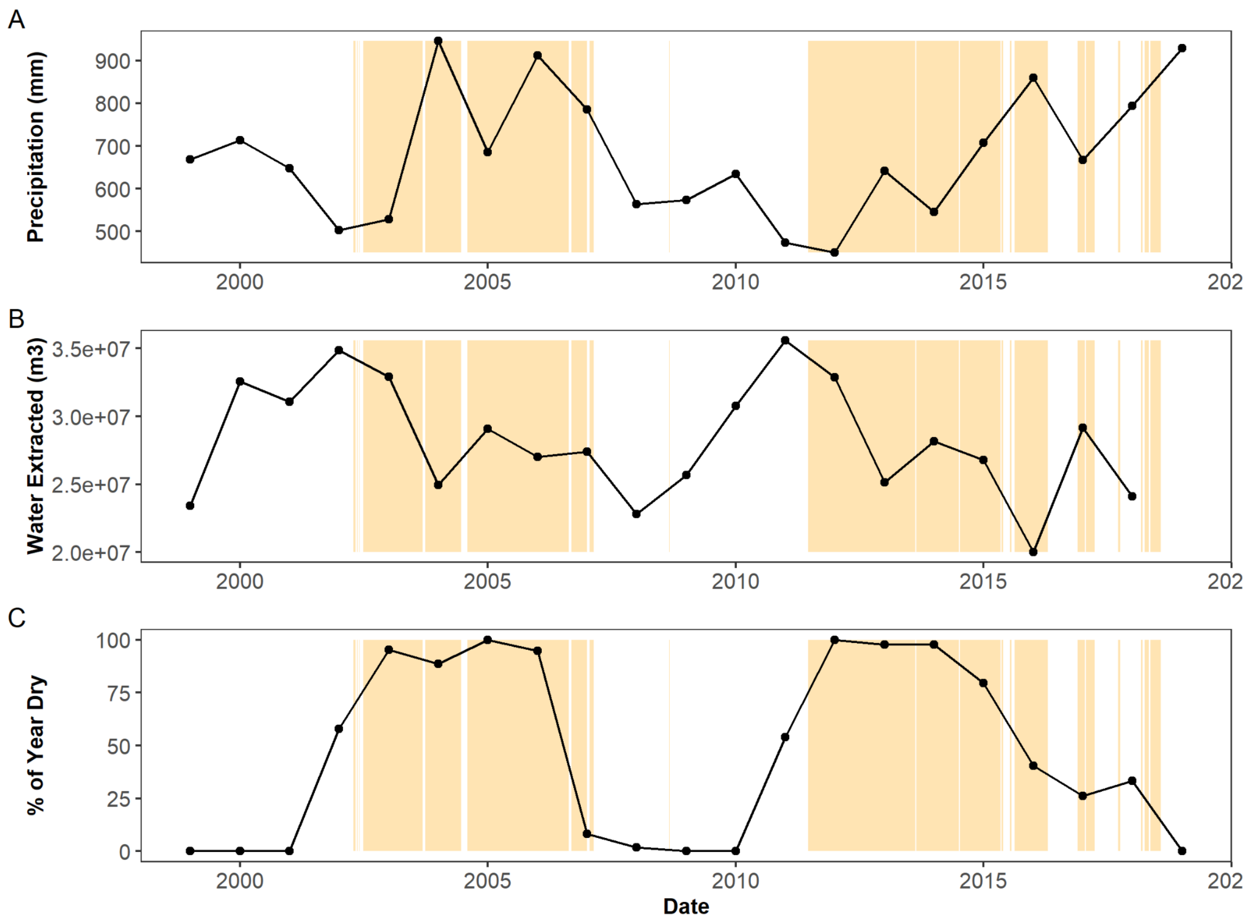

Figure 8. (a) Annual precipitation; (b) annual water use from WIMAS; and (c) annual percent no-flow days. No-flow periods of the Arkansas River are highlighted

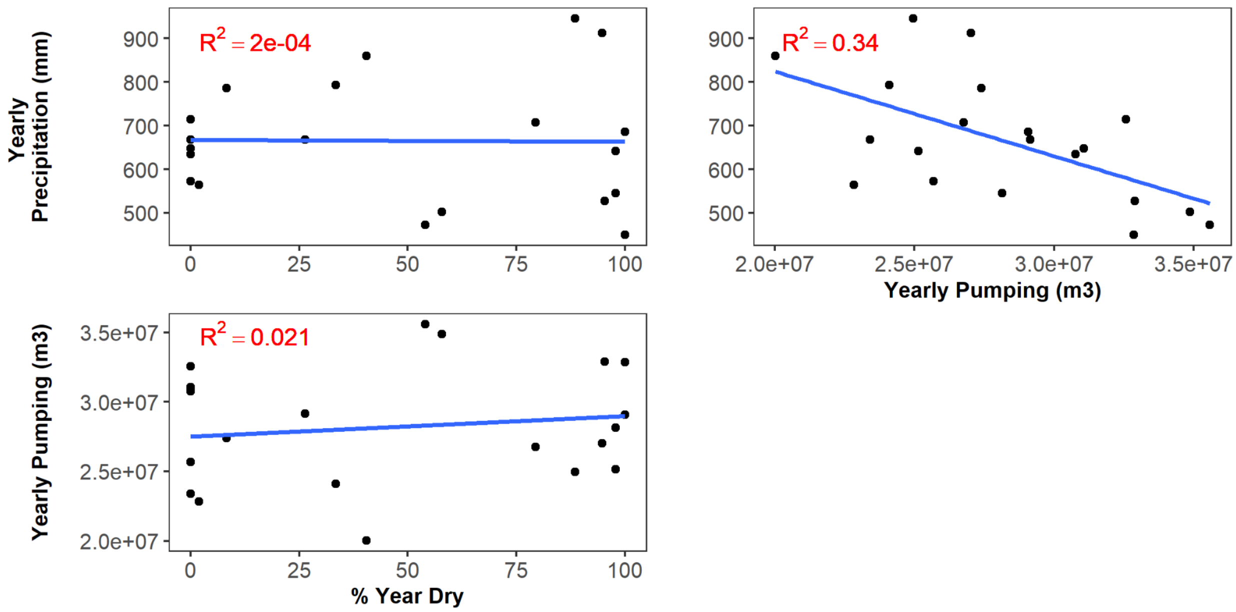

Figure 9. (a) Relationship between annual cumulative precipitation and percentage of the year the Arkansas River is dry; (b) relationship between annual cumulative precipitation and annual cumulative groundwater pumping within a 10 km radius of the study site; and (c) relationship between annual cumulative groundwater pumping and percentage of the year in which the Arkansas River is dry.

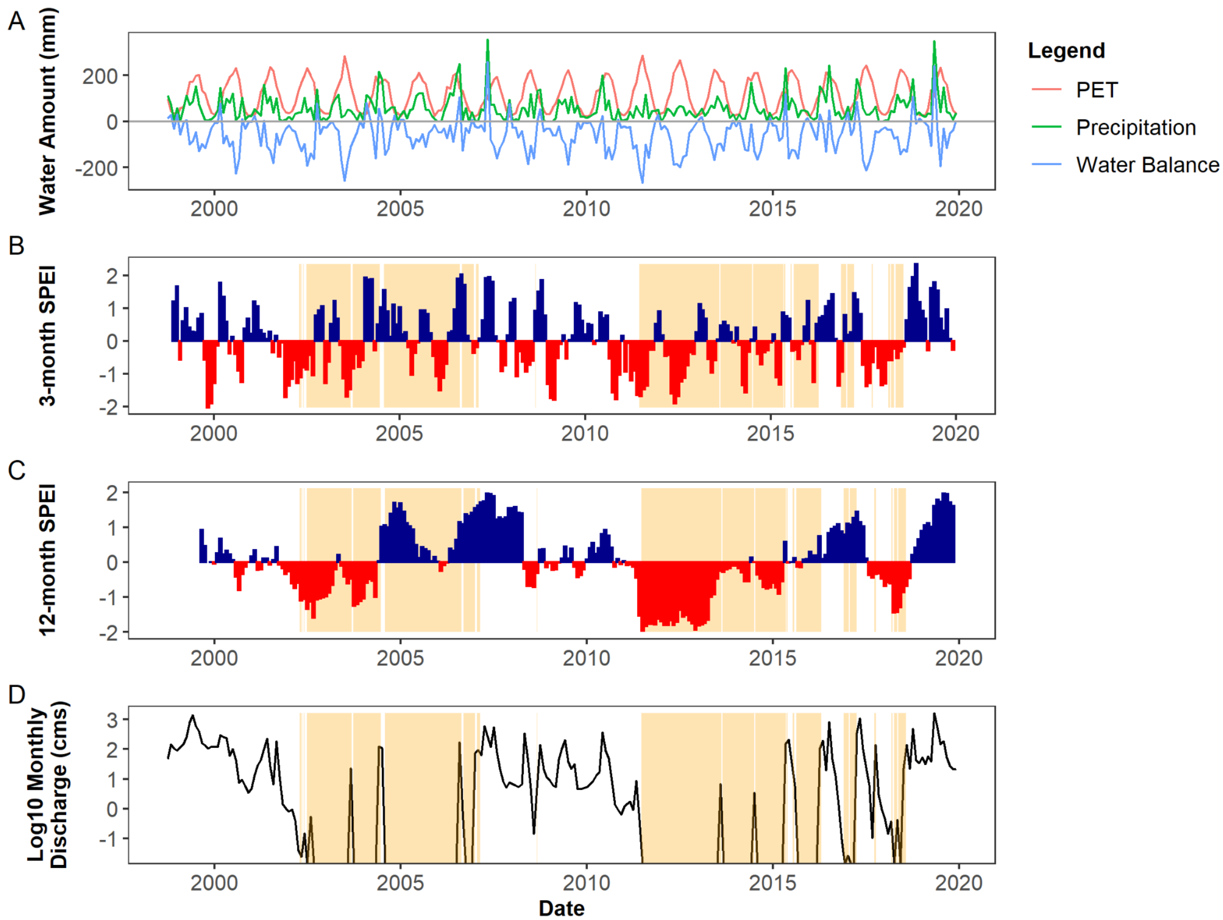

Figure 10. (a) Meteorological input data at monthly resolution (precipitation, potential evapotranspiration [PET], water balance), where water balance is defined as (precipitation - PET); (b) 3-month SPEI; (c) 12-month SPEI; and (d) mean monthly discharge. For all panels, the dry periods are shaded with an orange background.

The time lags between climate and stream intermittency can be seen in more detail in the SPEI data. Precipitation and temperature show a strong seasonal pattern (fig. 10a), with higher PET and precipitation and more negative water-balance estimates during summer months. There does not seem to be an obvious relationship between no-flow conditions and either the climatic water balance (fig. 8a) or the 3-month SPEI (fig. 10b). Although the no-flow periods tend to start during negative 3-month SPEI conditions, many negative 3-month SPEI values are not accompanied by no-flow periods (fig. 10b).

However, the 12-month SPEI time series has a much clearer relationship with no-flow conditions (fig. 10c). The extended dry regimes are preceded by and coincide with periods characterized by large, persistent negative SPEI values (i.e., 2002, 2011). Notably, flow does not begin immediately after the 12-month SPEI returns to positive values (i.e., 2005); only after prolonged wetter-than-average conditions does flow return (i.e., 2007). This lends credibility to the idea that flow at this site is storage-driven. Prolonged periods of negative SPEI deplete this storage, and prolonged wet conditions are needed to recharge the system, resulting in a lag time between climatically wet periods and the return of river flow.

In addition to groundwater pumping, we hypothesized that Horsethief Reservoir may be another source of anthropogenic influence on the intermittency of the Arkansas River. However, our comparison with the gages on the Pawnee River indicated that the creation of Horsethief Reservoir had little effect on flow at the Larned Research Site. A change in flow coincident with the completion of the reservoir would be indicated by a change in slope of the relationship between the Arkansas River gage and the Pawnee at Burdett gage. We found a clear change in slope of the relationship between the two gages on the Pawnee River (fig. 11c) indicating that the Pawnee River at Rozel (which is downstream of the reservoir) begins to accumulate flow at a more rapid rate than the Pawnee River at Burdett (which is not affected by the reservoir) following the construction of the reservoir. In contrast, there is no clear shift in the relationship between the Arkansas River near Larned and either of the two Pawnee River gages (fig. 11a and 11b) after the completion of Horsethief Reservoir, indicating that there was not a substantial shift in Arkansas River flow associated with the construction of Horsethief Reservoir.

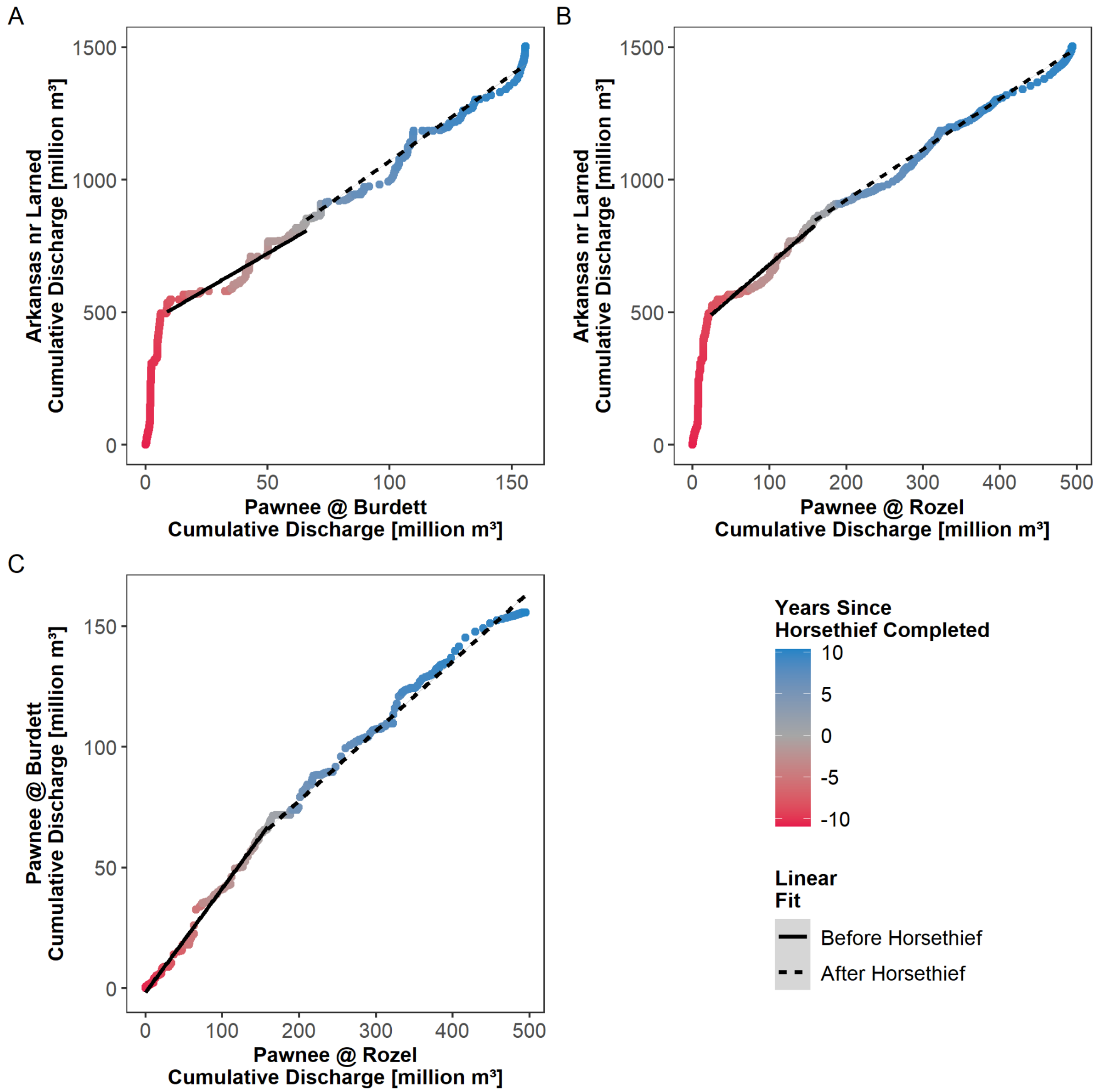

Figure 11. Double mass curves comparing (a) the Arkansas River near Larned and the Pawnee River at Burdett (which is not influenced by Horsethief Reservoir), (b) the Arkansas River near Larned and the Pawnee River at Rozel (which is influenced by Horsethief Reservoir), and (c) the Pawnee River at Burdette and the Pawnee River at Rozel. The black lines are linear best-fits before and after the completion of the Horsethief Reservoir in September 2009. Data in all plots begin on October 1, 1998, and end on December 31, 2019. For plots with the Arkansas River near Larned (a and b), the "Before Horsethief" fit does not include data before March 1999 because the period of record begins with a high-flow condition.

3.3 Groundwater-surface water interactions and no flow

3.3.1 Methods

The KGS has compiled hourly water-level data from the Arkansas River (USGS) and subhourly water-level data for the alluvial aquifer (well LEC1) and the High Plains aquifer (LEC2) from 2001 onward; well LEC2 is part of the High Plains aquifer index well network (Butler et al., 2020). We compared river stage, alluvial aquifer head, and High Plains aquifer head using both time series and correlation plots. Daily USGS stage data were not available at the beginning of the record, so we calculated a rating curve using data from 2007 to 2011 to fill in missing data (fig. 12):

$$Stage\ [m] = 0.312 * \sqrt{Flow\ [cms]} + 593.2$$

Eq. 3

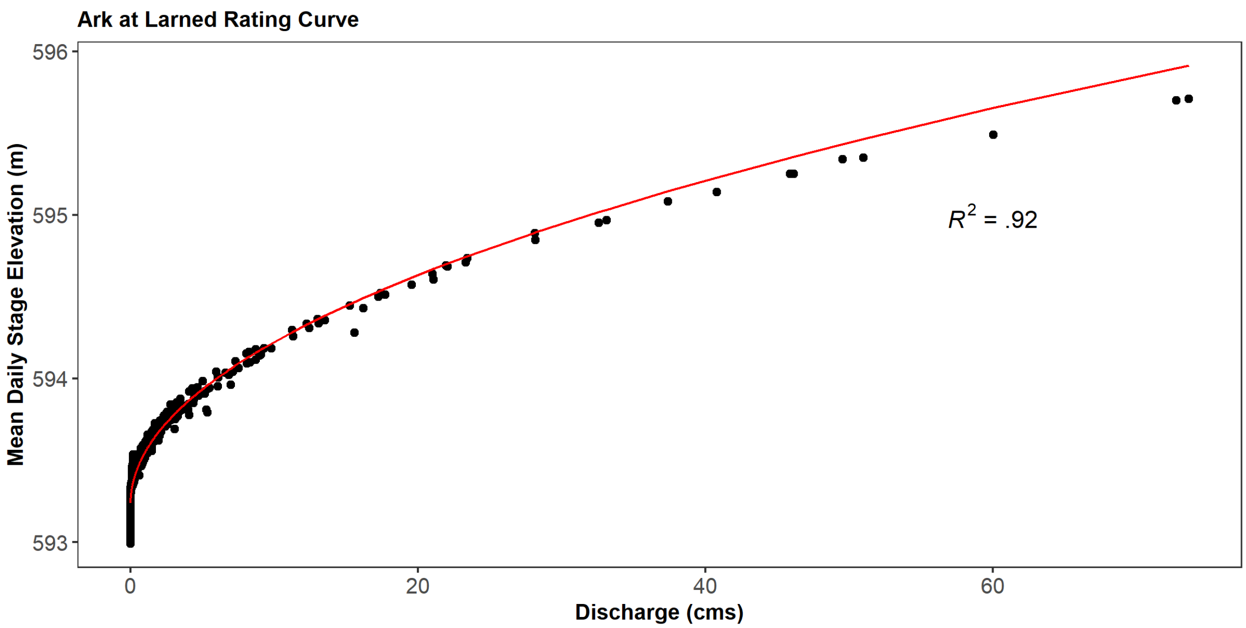

Figure 12. Rating curve used to back-fill historical stage data based on discharge.

3.3.2 Results and Discussion

The time series of the Arkansas River stage, Arkansas River alluvial aquifer head, and the High Plains aquifer head show a dynamic relationship between these two stores of water (fig. 13). The river regimes can be seen clearly, as stage values are only present when the river is flowing. Thus, the large gaps in stage data represent the dry regimes and positive stage values represent the flowing regimes. The first dry regime, which begins around 2003, is accompanied by a general downward trend in both the alluvial aquifer and High Plains aquifer until winter 2006–2007. The river transitions to a flowing regime in 2007, which correlates with a large increase in both High Plains and alluvial aquifer heads. Although there are gaps in aquifer data, we see a similar pattern associated with the end of the second dry regime in 2018–2019. Based on these two observations, it seems that there is a more rapid and dynamic relationship between the alluvial aquifer and the river stage with a strong temporal match between changes in water levels in these two units, compared to the relationship between the High Plains aquifer head and the alluvial aquifer, where the relationships among water-level changes in the different units are less tightly coupled.

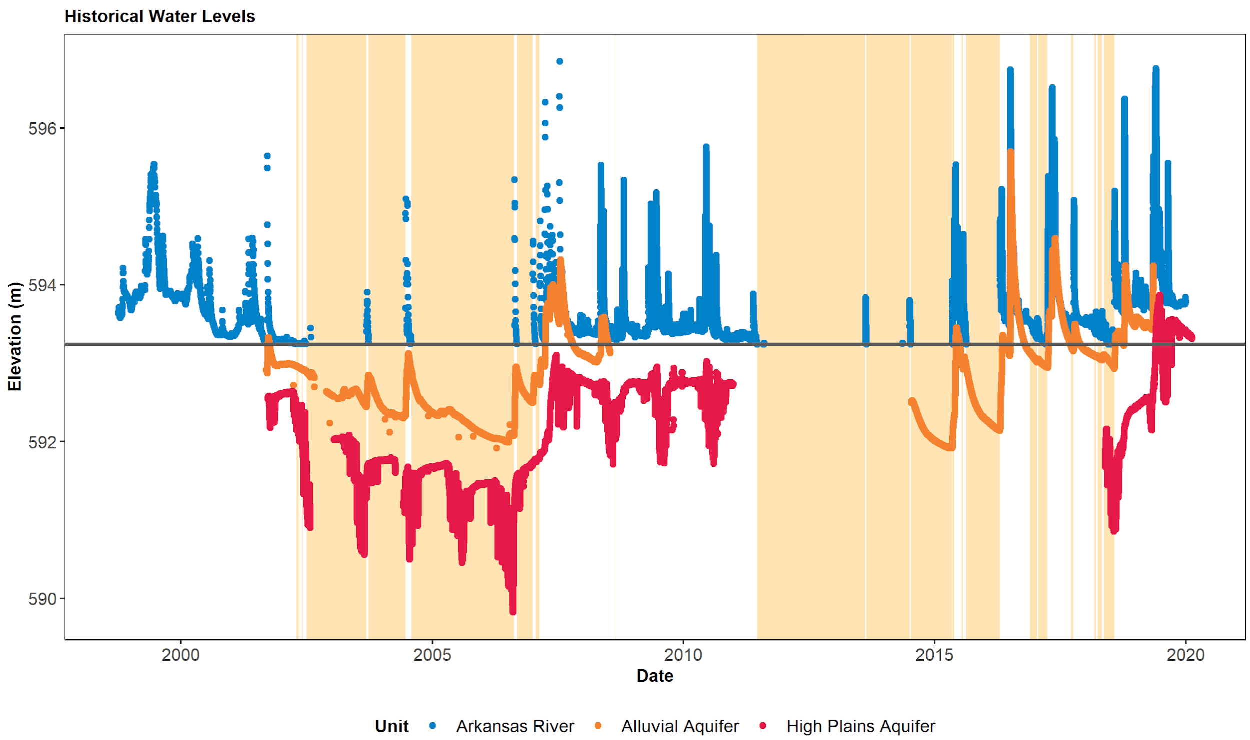

Figure 13. Time series plot of river elevation, alluvial aquifer head, and High Plains aquifer head. The grey line signifies the elevation of the riverbed.

To better understand these relationships, we compared water levels across these three datasets (fig. 14). The relationship between the Arkansas River stage and the alluvial aquifer head (fig. 14a) is not a simple linear relationship. The time element is identified with the color scale, and we see that there are event-scale hysteresis loops in the aquifer and river stage relationship. With the large loop in fig. 14a and the upper loop seen in fig. 14c, time progresses in the counterclockwise direction, indicating that river stage increased more rapidly than aquifer stage and enhanced losing conditions from the river into the alluvial aquifer. This relationship may be explained with infiltration of river flood waters into the alluvial aquifer and subsequently the High Plains aquifer. The relationship between the river stage and High Plains aquifer is less clear (fig. 14c) and suggests that the water-level variations in the river have a less direct influence on HPA head due to the presence of the confining layer. There is a more linear relationship between the High Plains aquifer and alluvial aquifer (fig. 14b) water levels despite the confining unit separating the two. This suggests some degree of leakage across the confining unit or a potential common driver, such as climate.

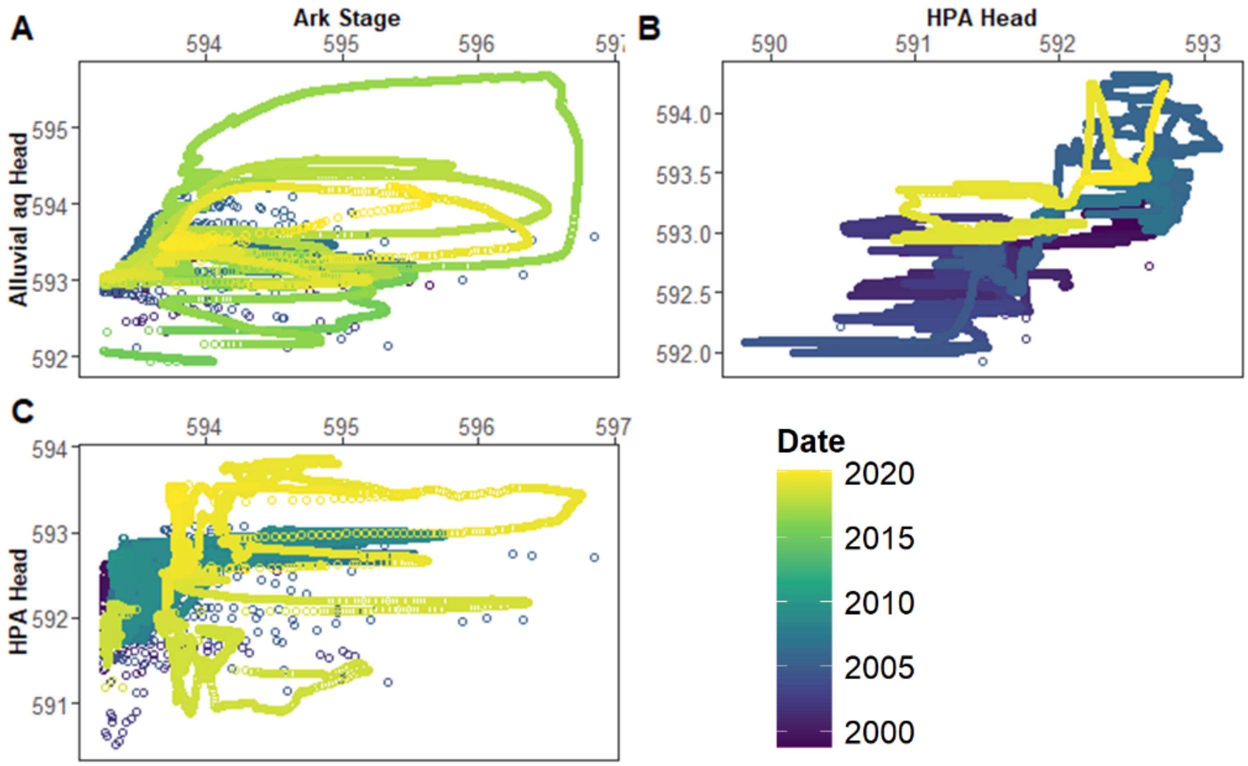

Figure 14. Plots of (a) river stage vs. alluvial aquifer head; (b) alluvial aquifer head vs. HPA head; and (c) HPA head vs. river stage. River stage values that are equal to the riverbed elevation (i.e., no flow) were removed before plotting.

4. Characterizing subsurface heterogeneity

Section 3 of this report showed a complex relationship between climate, surface flow, and water levels in the alluvial and High Plains aquifers. Because past work has shown that subsurface heterogeneity can affect flow and surface water-groundwater interactions in non-perennial streams (Noorduijn et al., 2014), we wanted to evaluate subsurface heterogeneity at this field site. We used three data sources: existing logs of electrical conductivity collected during well installation, electrical resistivity tomography (ERT) transects collected at the field site, and forward modeling of different hydrostratigraphic conceptual models.

4.1 Electrical conductivity profiles

4.1.1 Methods

Previous work has been performed to characterize the subsurface heterogeneity of the site. Direct-push profiles of electrical conductivity were collected during well installation at the site (Butler et al., 2004; Healey et al., 2001). Here, we review profiles from wells that are approximately 0.25 km from the channel on both the east (wells LEC1–3) and west (wells LWC1–2) sides of the river. Using these well logs, we developed a conceptual model of the subsurface beneath the Arkansas River (fig. 15).

4.1.2 Results and Discussion

Based on the electrical conductivity profiles, we developed a conceptual model that includes three major hydrogeologic units of the area from top to bottom: 1) the Arkansas River alluvial aquifer, 2) a clay confining unit, and 3) the High Plains aquifer. The profiles provide electrical conductivity as a function of depth, which can vary as a function of sedimentary properties (i.e., sand/silt/clay content) and water content. We based our description of the subsurface geology on the interpretations of the electrical conductivity data from past work (Butler et al., 2004; Healey et al., 2001). The alluvial aquifer is interpreted to be primarily made of sands with some silt and clay layers. This interpreted silt layer was present in the electrical conductivity profile collected on the western side of the river but not on the eastern side. In addition, there are differences in the thickness of the interpreted clay confining unit, ranging from 15 feet thick on the western side to 8 feet thick on the eastern side. Beneath this is the High Plains aquifer, which is not fully penetrated by these profiles. The lateral variability in clay and silt layers within the alluvial aquifer supports the hypothesis that subsurface heterogeneity at the site may impact water movement in the shallow subsurface.

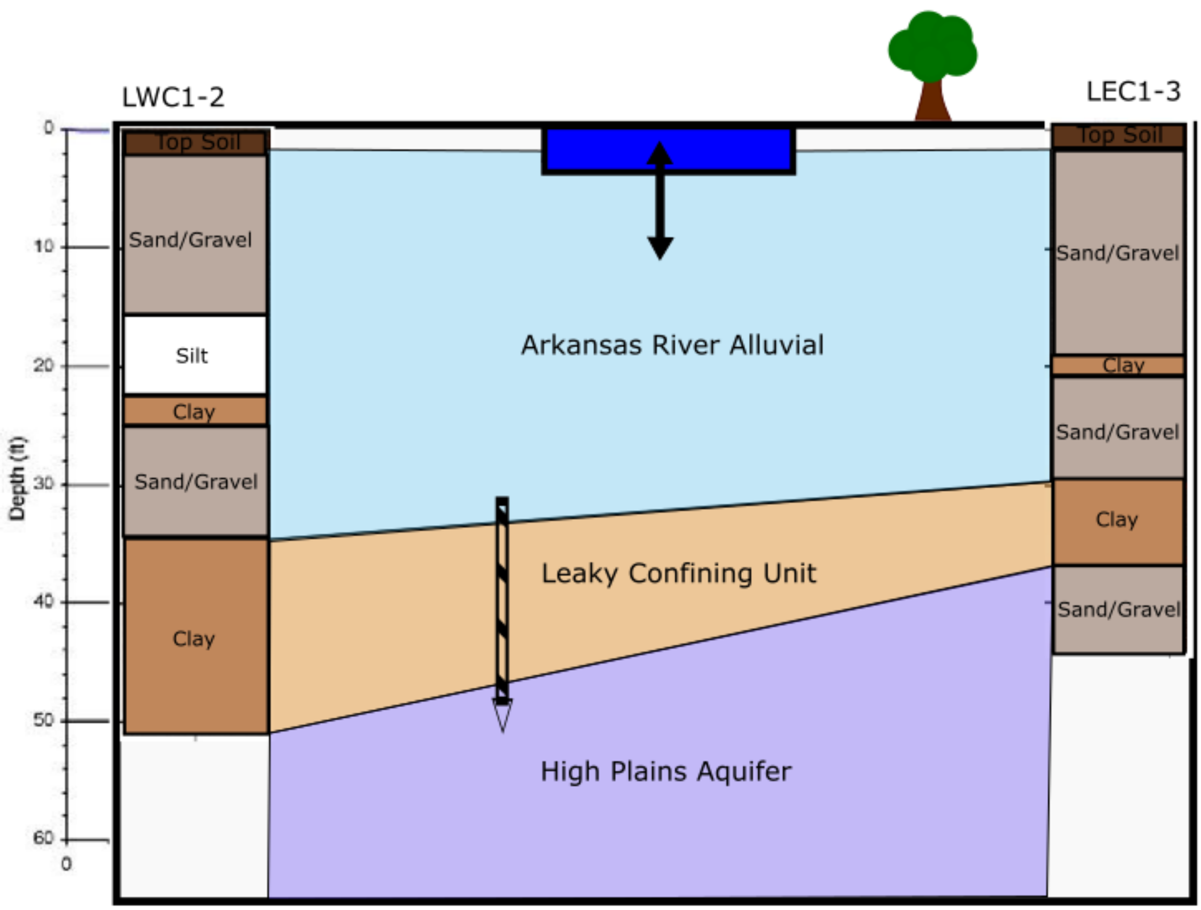

Figure 15. Fence diagrams based on electrical conductivity profiles of Healey et al. (2001).

4.2 Electrical resistivity tomography (ERT)

4.2.1 Methods

To better understand the subsurface heterogeneity, we collected two electrical resistivity tomography transects on June 4, 2020, the first along the riverbank and the second along a sandbar within the river (fig. 1). Transects were conducted with the ARES-II system using stainless steel electrodes with a 1-meter spacing. All the profiles were measured with both Wenner and Dipole-dipole configuration. As current is run through the electrodes, primarily porewater and groundwater act as conductors, whereas sediment and rock generally act as resistors. Five main properties affect the electrical conductivity (inverse of resistivity) of a terrain: porosity, saturation, moisture salinity, moisture temperature, and colloidal amount/composition (McNeil, 1980). The main factors we expect to vary are saturation and sediment type (a combination of porosity and colloidal amount/composition). High porosity allows for more water to act as a conductor and leads to lower resistivity. Colloidal amounts/composition can be thought of as the amount of ions from soil matrix that can be associated into the pore water. Clays may have adsorbed cations on particle surfaces, which may dissociate when immersed with water (McNeil., 1980). This increases electrical conductivity and lowers resistivity. This property is not seen in quartz, leading to a higher resistivity for quartz sands. Using these properties,we can estimate both the location of the water table and the composition of subsurface sediments using ERT transects.

4.2.2 Results and Discussion

The ERT transects were conducted with mixed success. We used the raw data to create a pseudosection of the subsurface, which we inverted to reduce the noise of the data. ResiPy, an open source Python program, was used to invert the data (fig. 16). Because of the extremely coarse and dry conditions of the sandy substrate, the electrodes did not have great contact and therefore the ERT data were noisy and only penetrated to a depth of ~10 m. Despite the noise present in these data, they do show a significant amount of lateral and vertical heterogeneity in resistivity, which may be attributable to differences in sediment composition.

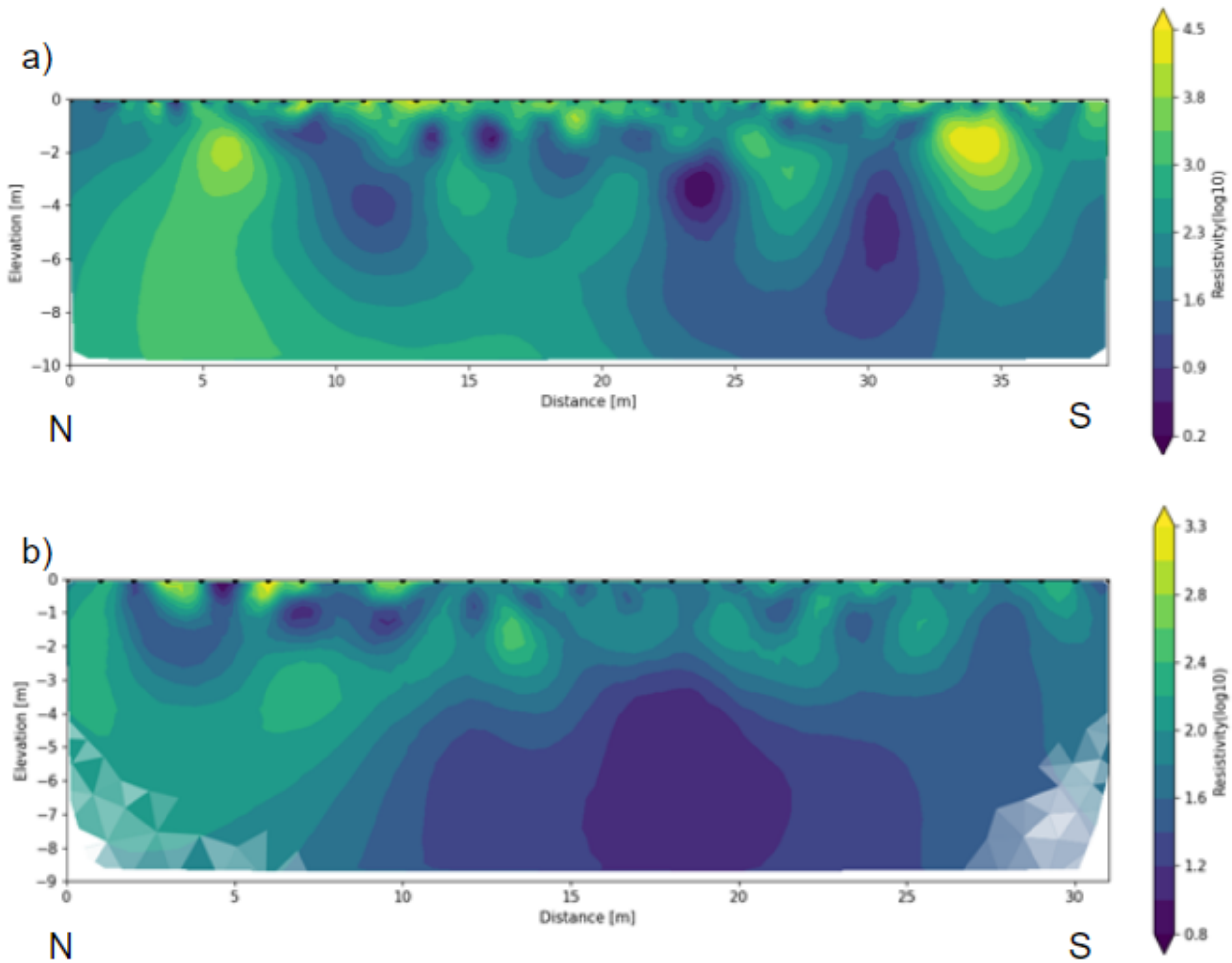

Figure 16. (a) Inverted ERT profile of the western bank of the Arkansas River, root mean squared (RMS) error = 1.06%, and (b) inverted ERT profile of the sandbar, RMS = 1.13%. N and S indicate north and south ends of transect, respectively. Profile locations are shown in fig. 1.

4.3 Forward modeling of different subsurface conditions

4.3.1 Methods

To broaden the conceptual understanding of the area, we attempted to undertake forward modeling. To do this, we created three possible conceptual models with increasing heterogeneity, based on the direct-push electrical conductivity and ERT data described in the preceding sections. We developed three conceptual models with increasing complexity (fig. 17). The simplest model is a relatively homogeneous, three-layer model consisting of a thin layer of unsaturated sand and gravel at the top, a thick layer of saturated sand and gravel, and a thin layer of silt (fig. 17a). The second conceptual model has increased vertical heterogeneity consisting of an unsaturated sand and gravel layer at the top, four layers of alternating saturated sand/gravel and silt, and a clay layer at the bottom (fig. 17b). The third model is vertically and horizontally heterogeneous, consisting of an unsaturated sand and gravel layer on the top, a thick saturated sand and gravel layer containing silt lenses, and a clay layer on the bottom (fig. 17c). We input these conceptual models into AGI EarthImager software using estimated resistivity for each unit and simulated estimated apparent resistivity using the software's forward modeling capabilities. The following values were used for each layer: 3000 Ωm for unsaturated sands and gravel, 1000 Ωm for saturated sands and gravel, 200 Ωm for silt, and 10 Ωm for clay (McNeil, 1980). In addition, we added 1.0% background noise.

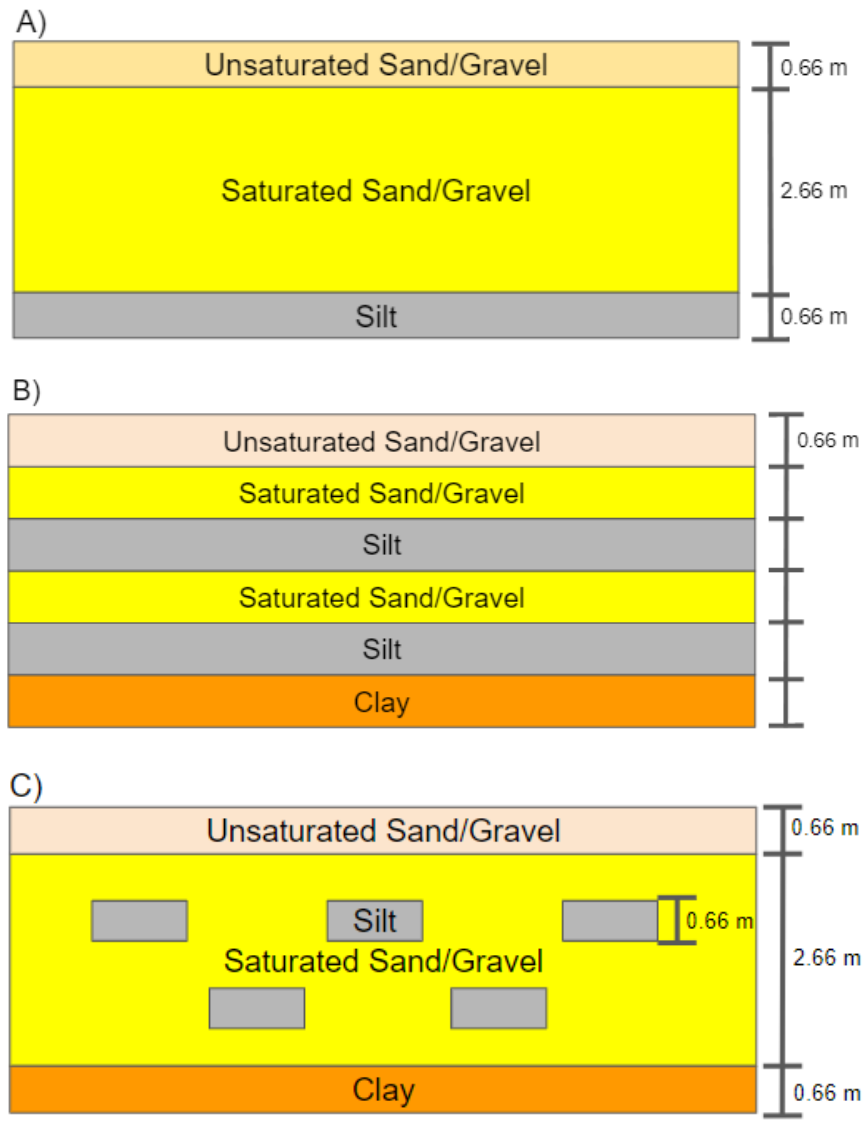

Figure 17. Three conceptual models that were simulated using forward models. All models are 4 meters thick and 31 meters long: (a) simplest model, (b) vertical heterogeneous model, and (c) vertically and laterally heterogeneous model.

4.3.1 Results and Discussion

The three subsurface models were input into EarthImager, run through forward modeling functions to create a resistivity pseudosection, and then inverted. The results were compared to the actual inverted data for the Arkansas River sandbar ERT transect (fig. 18). Note that the EarthImager software uses a different color scheme and may slightly differ from ResiPy inversion, but this is the same transect as fig. 16a.

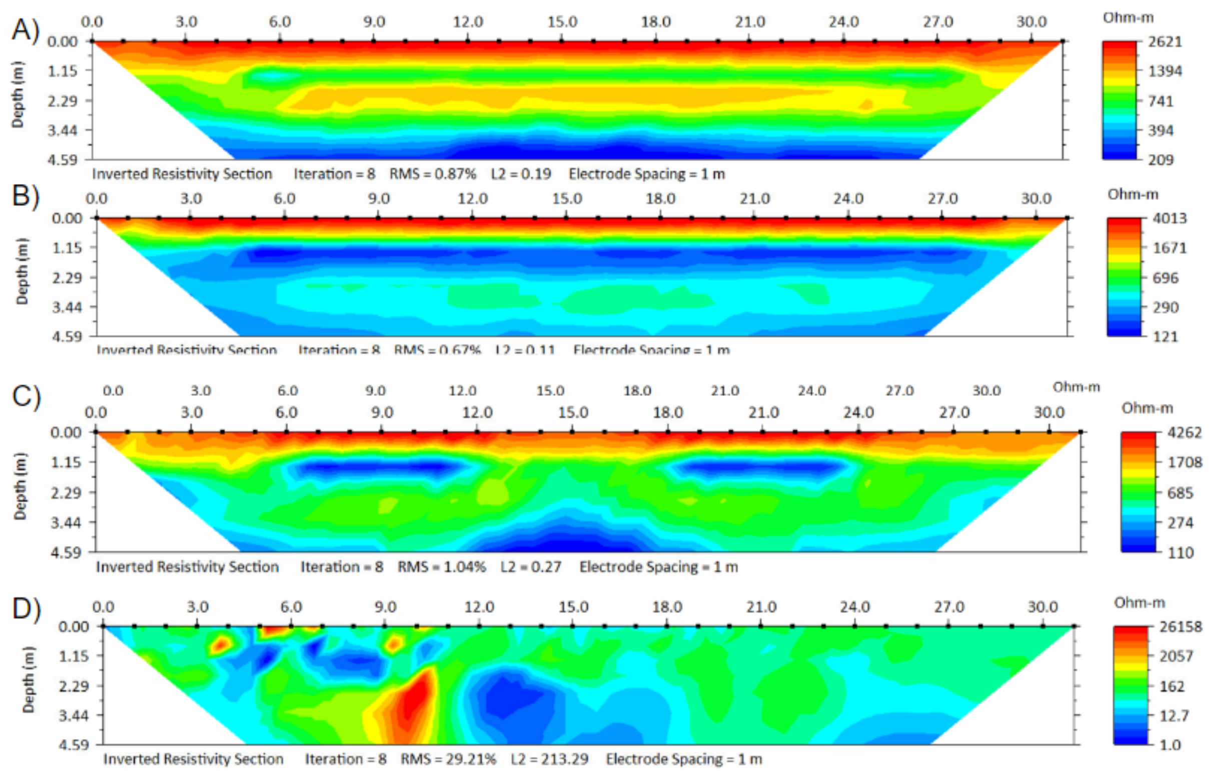

Figure 18. Forward model inversions using (a) simplest model (fig. 17a), (b) vertical heterogeneous model (fig. 17b), (c) vertically and laterally heterogeneous model (fig. 17c), and (d) inverted data from the Arkansas River sandbar ERT transect for comparison. Note that colorbar limits vary among plots.

Qualitatively, a few observations can be made. The structure of the three conceptual models can still be seen in the inversions, with some slight differences, such as alternating layers being seen in the model A. Of these three models, the third conceptual model with silt lenses scattered throughout (fig. 18c) is visually the closest match to the true data (fig. 18d). They both contain pockets of different resistivities throughout, though the true data have higher heterogeneity than any of the conceptual models.

Quantitatively, our ability to interpret these results is constrained by several limitations. The true sandbar ERT transect inversion has not fully converged at a solution, as the final iteration had a root mean squared data misfit of 29.21%. We were not able to successfully improve this value despite substantial exploration of different parameter options within the software. Thus, this inversion is not the true inverted image of the subsurface, and consequently we are primarily making qualitative remarks. The poor model fit may be due to the noisy ERT data mentioned in Section 4.2. Second, the range of resistivities is an order of magnitude greater in the real transect than in any of the conceptual models. It appears that there is a very highly resistive material (shown as the red patch in fig. 18d) that may be skewing the scale and making comparison more difficult. Due to the Arkansas River's history of flooding and this site being known as a local recreational site, the values of this highly resistive patch (more than 20,000 Ωm) may indicate a transported boulder (McNeil, 1980) or an unexpected anthropogenic material present in the subsurface below this reach.

5. Conclusions and Future Research Needs

This report aimed to improve the understanding of river intermittency, groundwater-surface water exchange, and subsurface heterogeneity on a reach of the Arkansas River near Larned, Kansas. Analysis of streamflow time series revealed very dynamic flow and intermittency over the past two decades. During this time, the Arkansas River experienced both wet regimes, characterized by persistent flow, and dry regimes, characterized by a persistent lack of flow. While wet and dry cycles can occur over short timescales (days or weeks), transitions between dry and wet regimes occurred at the scale of years.

We identified and evaluated several potential factors controlling the intermittency of this reach. The onset of dry periods has a weak correlation with seasonality, with more dry periods starting in the summer and more likely to end in the spring. However, flow regimes can span multiple years and do not seem to be affected by short-term (seasonal) variability in weather but are more associated with long-term (annual) dynamics. Although it is challenging to separate the influence of weather and groundwater pumping due to their correlation, there may be a delayed effect on flow regime. This is suggested by the consistent temporal relationship between lowering of the High Plains aquifer levels and a decline in the alluvial aquifer levels and streamflow.

Subsurface heterogeneity may affect river intermittency. Well logs and electrical resistivity tomography surveys show that the subsurface is laterally and vertically heterogeneous. As a result, there may be preferential flow paths between the hydrostratigraphic units beneath the river reach as well as exchange between the alluvial aquifer and the confined High Plains aquifer. However, geophysical investigations we conducted at the site using ERT were inconclusive. Further research is needed on subsurface heterogeneity in both the alluvial aquifer and confining unit to determine how pumping in the High Plains aquifer and exchange across the confining layer may affect river intermittency.

6. References

Batlle-Aguilar, J., Xie, Y., & Cook, P. G. (2015). Importance of stream infiltration data for modelling surface water–groundwater interactions. *Journal of Hydrology*, *528*, 683–693. https://doi.org/10.1016/j.jhydrol.2015.07.012

Beguería, S., & Vicente-Serrano, S. M. (2013). SPEI: Calculation of the Standardised Precipitation-Evapotranspiration Index (Version R package version 1.6). Retrieved from https://CRAN.R-project.org/package=SPEI

Busch, M. H., Costigan, K. H., Fritz, K. M., Datry, T., Krabbenhoft, C. A., Hammond, J. C., Zimmer, M., Olden, J. D., Burrows, R. M., Dodds, W. K., Boersma, K. S., Shanafield, M., Kampf, S. K., Mims, M. C., Bogan, M. T., Ward, A. S., Perez Rocha, M., Godsey, S., … Allen, D. C. (2020). What's in a name? Patterns, trends, and suggestions for defining non-perennial rivers and streams. *Water*, *12*(7), 1980. https://doi.org/10.3390/w12071980 Butler, J. J., Jin, W., Mohammed, G. A., & Reboulet, E. C. (2011). New iInsights from wWell rResponses to fFluctuations in bBarometric pPressure. *Groundwater*, *49*(4), 525–533. https://doi.org/10.1111/j.1745-6584.2010.00768.x

Butler, J. J., Kluitenberg, G. J., Whittemore, D. O., Loheide, S. P., Jin, W., Billinger, M. A., & Zhan, X. (2007). A field investigation of phreatophyte-induced fluctuations in the water table. *Water Resources Research*, *43*(2). https://doi.org/10.1029/2005WR004627

Butler, J. J., Whittemore, D. O., Reboulet, E. C., Knobbe, S., Wilson, B. B., & Bohling, G. C. (2020). High Plains Aquifer Index Well Program: 2019 Annual Report (Open-File Report No. 2020-2). Lawrence KS: Kansas Geological Survey. Retrieved from http://www.kgs.ku.edu/Publications/OFR/2020/OFR2020-2.pdf

Butler, J. J., Whittemore, D. O., Zhan, X., & Healey, J. M. (2004). Analysis of two pumping tests at the O'Rourke Bridge site on the Arkansas River in Pawnee County, Kansas (KGS Open-File Report No. 2004-32). Lawrence KS: Kansas Geological Survey. Retrieved from http://www.kgs.ku.edu/Hydro/Publications/2004/OFR04_32/index.html

Cuthbert, M. O., & Mackay, R. (2013). Impacts of nonuniform flow on estimates of vertical streambed flux. *Water Resources Research*, *49*(1), 19–28. https://doi.org/10.1029/2011WR011587

Datry, T., Larned, S. T., & Klement, T. (2014). Intermittent rivers: A challenge for freshwater ecology. *BioScience*, *64*(3), 229–235.

Fuchs, E. H., King, J. P., & Carroll, K. C. (2019). Quantifying disconnection of groundwater from managed-ephemeral surface water during drought and conjunctive agricultural use. *Water Resources Research*, *55*(7), 5871–5890. https://doi.org/10.1029/2019WR024941

Fuka, D., Walter, M., Archibald, J., Steenhuis, J., & Easton, Z. (2014). EcoHydRology: A community modeling foundation for Eco-Hydrology (Version 0.4.12). Retrieved from https://CRAN.R-project.org/package=EcoHydRology

Healey, J. M., Butler Jr, J. J., & Whittemore, D. O. (2001). Stream-aquifer interaction investigations on the middle Arkansas River: Construction of groundwater observation wells in Edwards, Pawnee, and Barton County, Kansas (KGS Open-File Report No. 2001-48). Lawrence KS: Kansas Geological Survey. Retrieved from http://www.kgs.ku.edu/Hydro/Publications/2001/OFR01_48/index.html

Loheide, S. P., Butler, J. J., & Gorelick, S. M. (2005). Estimation of groundwater consumption by phreatophytes using diurnal water table fluctuations: A saturated-unsaturated flow assessment. *Water Resources Research*, *41*(7). https://doi.org/10.1029/2005WR003942

Lyne, V., & Hollick, M. (1979). Stochastic time-variable rainfall-runoff modelling. In *Institute of*

Engineers Australia National Conference (pp. 89–93).
McNeil, J. D. (1980). Electrical Conductivity of Soils and Rocks, Technical note TN-5. *GEONICS Limited, Ontario, Canada, 22 p.*
Nathan, R. J., & McMahon, T. A. (1990). Evaluation of automated techniques for base flow and recession analyses. *Water Resources Research*, *26*(7), 1465–1473. https://doi.org/10.1029/WR026i007p01465
Noorduijn, S. L., Shanafield, M., Trigg, M. A., Harrington, G. A., Cook, P. G., & Peeters, L. (2014). Estimating seepage flux from ephemeral stream channels using surface water and groundwater level data. *Water Resources Research*, *50*(2), 1474–1489. https://doi.org/10.1002/2012WR013424
Quichimbo, E. A., Singer, M. B., & Cuthbert, M. O. (2020). Characterising groundwater–surface water interactions in idealised ephemeral stream systems. *Hydrological Processes*, *34*(18), 3792–3806. https://doi.org/10.1002/hyp.13847
Rau, G. C., Halloran, L. J. S., Cuthbert, M. O., Andersen, M. S., Acworth, R. I., & Tellam, J. H. (2017). Characterising the dynamics of surface water-groundwater interactions in intermittent and ephemeral streams using streambed thermal signatures. *Advances in Water Resources*, *107*, 354–369. https://doi.org/10.1016/j.advwatres.2017.07.005 Shanafield, M., Bourke, S. A., Zimmer, M. A., & Costigan, K. H. (2021). An overview of the hydrology of non-perennial rivers and streams. *WIREs Water*, *8*(2), e1504. https://doi.org/10.1002/wat2.1504
Shanafield, M., & Cook, P. G. (2014). Transmission losses, infiltration and groundwater recharge through ephemeral and intermittent streambeds: A review of applied methods. *Journal of Hydrology*, *511*, 518–529. https://doi.org/10.1016/j.jhydrol.2014.01.068
Shanafield, M., Cook, P. G., Brunner, P., McCallum, J., & Simmons, C. T. (2012). Aquifer response to surface water transience in disconnected streams. *Water Resources Research*, *48*(11). https://doi.org/10.1029/2012WR012103
Vicente-Serrano, S. M., Beguería, S., & López-Moreno, J. I. (2009). A multiscalar drought index sensitive to global warming: The Standardized Precipitation Evapotranspiration Index. *Journal of Climate*, *23*(7), 1696–1718. https://doi.org/10.1175/2009JCLI2909.1
Vienken, T., Reboulet, E., Leven, C., Kreck, M., Zschornack, L., & Dietrich, P. (2013). Field comparison of selected methods for vertical soil water content profiling. *Journal of Hydrology*, *501*, 205–212. https://doi.org/10.1016/j.jhydrol.2013.08.004
Zimmer, M. A., Kaiser, K. E., Blaszczak, J. R., Zipper, S. C., Hammond, J. C., Fritz, K. M., Costigan, K. H., Hosen, J., Godsey, S. E., Allen, G. H., Kampf, S., Burrows, R. M., Krabbenhoft, C. A., Dodds, W., Hale, R., Olden, J. D., Shanafield, M., DelVecchia, A. G., … Allen, D. C. (2020). Zero or not? Causes and consequences of zero-flow stream gage readings. *WIREs Water*, e1436. https://doi.org/10.1002/wat2.1436
Zimmer, M. A., & McGlynn, B. L. (2017). Ephemeral and intermittent runoff generation processes in a low relief, highly weathered catchment. *Water Resources Research*, *53*(8), 7055–7077. https://doi.org/10.1002/2016WR019742