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# Geostatistical features of streambed vertical hydraulic conductivities in Frenchman Creek Watershed in Western Nebraska

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## Abstract

This study evaluated the spatial variability of streambed vertical hydraulic conductivity ( $K_v$ ) in different stream morphologies in the Frenchman Creek Watershed, Western Nebraska, using different variogram models. Streambed  $K_v$  values were determined in situ using permeameter tests at 10 sites in Frenchman, Stinking Water and Spring Creeks during the dry season at baseflow conditions. Measurements were taken both in straight and meandering stream channels during a 5 day period at similar flow conditions. Each test site comprised of at least three transects and each transect comprised of at least three  $K_v$  measurements. Linear, Gaussian, exponential

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and spherical variogram models were used with Kriging gridding method for the 10 sites. As a goodness-of-fit statistic for the variogram models, cross-validation results showed differences in the median absolute deviation and the standard deviation of the cross-validation residuals. Results show that using the geometric means of the 10 sites for gridding performs better than using either all the  $K_v$  values from the 93 permeameter tests or 10  $K_v$  values from the middle transects and center permeameters. Incorporating both the spatial variability and the uncertainty involved in the measurement at a reach segment can yield more accurate grid results that can be useful in calibrating  $K_v$  at watershed or sub-watershed scales in distributed hydrological models.

**Keywords:** Frenchman Creek, geostatistical analysis, permeameter test, streambed, variogram models, vertical hydraulic conductivity

## 1 Introduction

There is a broad interest in the estimation of streambed vertical hydraulic conductivity ( $K_v$ ) due to its connection to water quality, aquatic habitat, and groundwater-surface water exchange (Cheng, Song, Chen, & Wang, 2011; Genereux, Leahy, Mitsova, Kennedy, & Corbett, 2008; Jiang et al., 2015; Wang et al., 2017). Streams are rarely isolated. The different states of connection between streams and underlying groundwater, and the water exchange pattern at the groundwater-surface water interface are mostly dependent on rainfall inputs, water head changes, and substrate permeability (Brunner, Cook, & Simmons, 2009; Castro & Hornberger, 1991). There are variations in  $K_v$  values in a watershed due to hydrologic position and scale (Katsuyama, Tani, & Nishimoto, 2010), as well as the variety of spatial and temporal factors such as the topography, the depth of streambed, bed slope, land cover, and the hydrogeological setting of the underlying aquifer (Wang et al., 2016; Woessner, 2000).

Streambed  $K_v$  is a key parameter in watershed models, so understanding its spatial variability and uncertainty is essential to accurately predict how stresses and environmental signals propagate through the hydrologic system (Abimbola, Mittelstet, Gilmore, & Korus, 2020). In situ studies have shown that streambed  $K_v$  changes significantly along the stream or river cross section (perpendicular to the stream-flow) and along the stream flow (in the downstream direction), even in a small channel segment (Chen, 2004, 2005; Cheng et al., 2011; Genereux et al., 2008; Hatch, Fisher, Ruehl, & Stemler, 2010). Since it is not practical to measure  $K_v$  at every location along a stream course,

most hydrological modelling studies usually assume homogeneity of  $K_v$  for practical reasons. Relying on literature values or limited measurements and assuming  $K_v$  is constant across a watershed may lead to more uncertainty due to the under- or over-prediction of streambed leakage and baseflow (Brunner et al., 2009; Irvine, Brunner, Franssen, & Simmons, 2012; Kurtz, Hendricks Franssen, Brunner, & Verbeecken, 2013; Leake, Greer, Watt, & Weghorst, 2008). To understand the connectivity between surface water and groundwater, it is important to estimate the spatial distribution of  $K_v$  which is one of the most important parameters controlling the movement of water from the stream to the aquifer, and vice versa (Chen & Shu, 2002; Genereux et al., 2008; Goswami, Kalita, & Mehnert, 2010; Saenger, Kitanidis, & Street, 2005; Sun & Zhan, 2007).

Several studies have found ways to measure  $K_v$  in situ (Chen, 2005, 2007; Genereux et al., 2008) mostly building on Hvorslev (1951). While some studies focused on the spatial variability of streambed  $K_v$  along transects across a channel (Cardenas & Zlotnik, 2003; Chen, 2004; Kennedy, Genereux, Mitsova, Corbett, & Leahy, 2008), others focused on both the spatial and temporal variability (Genereux et al., 2008), as well as statistical description of streambed  $K_v$  (Cardenas & Zlotnik, 2003; Song, Chen, Cheng, Summerside, & Wen, 2007). Although most of the previous studies focused on the spatial variability of streambed  $K_v$  at one or several adjacent sites in small creeks and at distant sites along large rivers (Cardenas & Zlotnik, 2003; Chen, 2005; Cheng et al., 2011; Genereux et al., 2008; Song et al., 2007), they did not develop a geostatistical distribution analysis of streambed  $K_v$  across multiple stream orders at watershed-scale. Few studies have quantified the spatial variability of  $K_v$  in different stream morphologies, and very few have considered using different variogram models (Jiang et al., 2015; Wang et al., 2017). These geostatistical methods of estimating  $K_v$  suggest challenges in determining representative samples and comparing results, considering the heterogeneity and anisotropy of streambed materials and geological conditions (Naganna, Deka, Sudheer, & Hansen, 2017). Moreover, although geostatistical methods present a wide range of interpolation procedures that can be applied to hydrological systems, there is a knowledge gap in incorporating interpolation results into hydrological modelling for better calibration.

In the past 25 years, there have been significant changes in water resources in Western Nebraska. This includes increases in irrigated

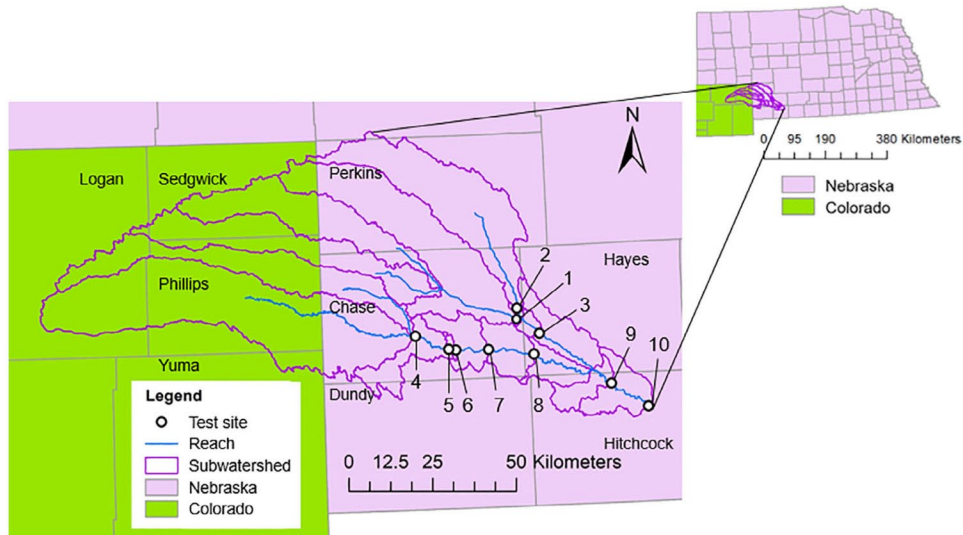
acres and irrigation wells; decreases in streamflow, groundwater levels, and groundwater allocations; conversion from flood to pivot irrigation technologies and drip irrigation; moratoriums on new irrigation wells; and the encroachment of the eastern Red Cedar and other invasive species (Twidwell et al., 2013). The Frenchman Creek Watershed in Nebraska has experienced each of these changes. In the last five decades, groundwater withdrawals in the Frenchman Creek Watershed have led to groundwater declines ranging from 2 to 17 m and decreased streamflow. In the 1960s, Frenchman Creek began several kilometers west of the Colorado border, yet today it begins 21 km east of the Colorado border (Traylor, 2012). These declines have led to reductions in groundwater allocations and a moratorium on new irrigation wells in the Upper Republican River Natural Resource District (NRD). Since  $K_v$  is a major parameter in the estimation of groundwater recharge from streams and rivers, it is therefore important to determine its spatial distribution for integrated water resource assessment and management in the watershed.

The objectives of this study were to (a) determine the statistical distribution and spatial variation of streambed  $K_v$  along different morphologies within the Frenchman Creek Watershed and (b) evaluate the accuracy and usefulness of the spatial variability of streambed  $K_v$  values as estimated from different sample sizes and using different representative samples.

## **2 Materials and methods**

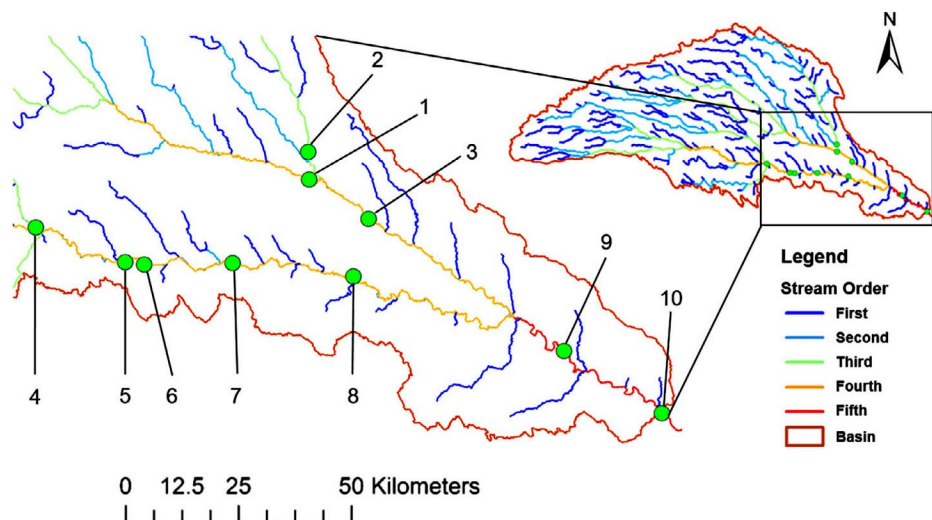
### **2.1 Study area and test sites**

The Frenchman Creek Watershed is located in two states (Nebraska and Colorado) in the United States. The watershed drains over 7,600 km<sup>2</sup> in southwest Nebraska (60%) and southeast Colorado (40%; Figure 1). It is a sub-watershed of the Republican River watershed. The primary land uses in the watershed consist of irrigated cropland, dry cropland, pasture, and rangeland, with dense vegetation including trees, shrubs, and grasses occurring in the riparian zones. The dominant soil series include Valent (70–100% sand) and Kuma (19–79% silt), and major tributaries include Stinking Water Creek, Spring Creek, and Sand Creek.



**Figure 1** Map showing the 10 sites and sub watersheds where in situ permeameter tests were performed in the Frenchman Creek Watershed.

Ten test sites (stream reaches) were selected from third-, fourth-, and fifth-order streams on the Frenchman Creek and its four main tributaries (Figure 2). The sites were selected above and below the major confluences, and were located in both meandering and straight reaches. The number of sites was constrained due to stream accessibility and streamflow. The western two thirds of the watershed was dry.



**Figure 2** Stream orders of the measurement sites.



**Figure 3** Basic design of falling-head permeameter test. *Left*: Schematic of the field permeameter. *Middle*: Photograph of permeameters with each tube installed to depth of 30 cm. *Right*: Layout of three transects along the stream for each site.

## 2.2 *In situ* permeameter tests

The *in situ* falling head permeameter test is one of several methods used to determine streambed  $K_v$  (Chen, 2000; Dong, Chen, Wang, Ou, & Liu, 2012; Genereux et al., 2008). The falling head permeameter test usually involves inserting a tube into channel sediments (Figure 3). In this study, transparent tubes were used for 10 test sites within the Frenchman Creek Watershed in the summer of 2017. Seven sites were on Frenchman Creek, two sites are on Stinking Water Creek and one site on Spring Creek. Other tributaries within the watershed were dry at the time of the study. Field measurements were carried out over a 5-day period from June 26th to 30th, 2017. The stream was at base-flow conditions with average discharge of  $0.35 \text{ m}^3 \text{ s}^{-1}$  at the USGS gage station near Palisade (06834000). The flow remained consistent throughout the study period with a flowrate on June 26th of  $0.36 \text{ m}^3 \text{ s}^{-1}$  and decreasing slightly to  $0.33 \text{ m}^3 \text{ s}^{-1}$  by June 30th. This compares to an average discharge at the Palisade gage station of  $0.65 \text{ m}^3 \text{ s}^{-1}$  from 2000 to 2020.

The duration of individual falling head tests ranged from a few minutes to a maximum of 24 hours. Permeameters left for long periods of time were covered with plastic bags to prevent evaporative losses.

Each test site comprised of at least three transects and each transect comprised of at least three streambed  $K_v$  measurements. Figure 3 is a schematic diagram showing *in situ* permeameter test installation at a test site. Transparent tubes (76 cm long and 8 cm inside diameter

or 183 cm long and 6.8 cm inside diameter) were pressed vertically into the channel sediments. The thickness of the tube wall was about 3 mm, typical of many previous studies (Genereux et al., 2008; Kennedy et al., 2008). For each  $K_v$  measurement, the tube was pressed to a depth of 30 cm, thus the lower part of the tube was filled with a sediment column of 30-cm length (Song et al., 2018). Stream water level was used as an estimate of the ambient pre-test water level, which introduces additional error in the  $K_v$  estimates (compared to Genereux et al., 2008, where groundwater head was measured) but that error is in most cases small relative to the order of magnitude differences in  $K_v$  observed between sites in this study. Clear stream water was then added slowly to fill up the tube from the top, minimizing the disturbance of sediment inside the tube. As the hydraulic head in the tube began to fall, a series of hydraulic heads at given times were recorded. Three (low conductivity) to ten (high conductivity) measurements were recorded for each test. The  $K_v$  (m/d) calculation was based on Equation (1) derived from Hvorslev's equation, similar to Genereux et al. (2008).

$$K_v = \frac{\frac{\pi D}{11m} + L_v}{(t_2 - t_1)} \ln \left( \frac{H_1}{H_2} \right)$$

where  $D$  is the inside diameter of the tube;  $L_v$  is the length of the sediment in the tube;  $t_1$  and  $t_2$  are the times between inside measurements of hydraulic heads  $H_1$  and  $H_2$ , respectively; and  $m$  is the isotropic transformation ratio  $(K_h/K_v)^{1/2}$  where  $K_h$  is the horizontal hydraulic conductivity of the sediment around the base of the tube. For this study, we used the average of the  $K_v$  values estimated with  $m = 1$  and  $m = \infty$ . If  $t_1$  is the time of the start of the permeameter test (i.e.,  $H_1 = H_0$  at  $t_1 = 0$ ), Equation (1) gives Equation (2)

$$\ln H = - \frac{K_v}{\frac{\pi D}{11m} + L_v} t + \ln H_0$$

where  $H$  is the water level inside the permeameter relative to the ambient pre-test water level, and  $t$  is time, and the slope term is set equal to the slope in the head versus time plot and solved for  $K_v$  (Genereux et al., 2008).



### 2.3 Normality test

To check whether the distributions of  $K_v$  were normal for the sites, graphical exploration (Q–Q plots and histograms) and formal tests of normality were carried out. There are quite a number of tests of normality available in the literature. D'Agostino and Stephens (1986) provided detailed descriptions of various normality tests. In this study, six normality tests were used. Anderson–Darling (AD), Cramer–von Mises (CVM), Lilliefors (LL), Pearson chi-square (CSQ), Shapiro–Francia (SF), and Shapiro–Wilk (SW) tests were applied at .05 significance level. These normality tests were categorized into tests based on correlation and regression (SW and SF tests), CSQ test, and empirical distribution test (such as LL, AD, and CVM). Some of these tests were constructed to be applied under certain conditions or assumptions. The SW (Shapiro & Wilk, 1965) test is one of the most commonly used of the six tests. According to Royston (1982), it has requirements for the sample size  $N$  ( $7 \leq N \leq 2,000$ ), while the LL (Lilliefors, 1967) test is preferable to apply for a large sample size  $N \geq 2,000$  (Cheng et al., 2011). The SF test is a simplified version of the SW test, which uses the squared normal probability plot correlation as a test statistic.

### 2.4 Gridding streambed $K_v$

For each site (stream reach), the spatial structure of  $K_v$  was analysed and different variograms were fit to the data and used as input in the software package SURFER (<http://www.GoldenSoftware.com>). Spatial interpolation of the  $K_v$  data was carried out using kriging. Kriging, which is the most commonly used geostatistical method, can estimate both the predicted values and their standard errors (Webster & Oliver, 2001). It is also an optimal interpolator which uses the spatial structure and variance of the input data points to estimate the interpolation weights and search radii to provide the best, unbiased estimate at unsampled points (Burrough & McDonnell, 1998).

The performance of kriging depends on the presence of spatial autocorrelation. This implies that sites which are close together tend to be more similar than those which are further apart. Since  $K_v$  values may vary in orders of magnitude within a short distance in a heterogeneous aquifer system, the geometric mean of each site was used in

this study to capture the variations. Variograms were created from the input  $K_v$  data points by plotting the variance against the distance between pairs of points and used to optimize the interpolation weights. At channel scale, different variogram models (e.g., spherical, Gaussian, exponential, and linear) were considered and the variograms which gave the best fit were chosen for the purpose of predicting  $K_v$  at each site. At watershed scale, different models using different sample sizes were fitted to linear variograms.

The performance of the spatial interpolation models were evaluated with cross validation which can be considered an objective method of assessing the quality of a variogram, or to compare the relative quality of two or more candidate variograms. To determine the goodness-of-fit for the variograms, the two most consistently useful statistics are: the median absolute deviation of the cross-validation residuals (MAD<sub>XV</sub>) and the standard deviation of the cross-validation residuals (SD<sub>XV</sub>).

### 3 Results and discussions

#### 3.1 Vertical streambed $K_v$

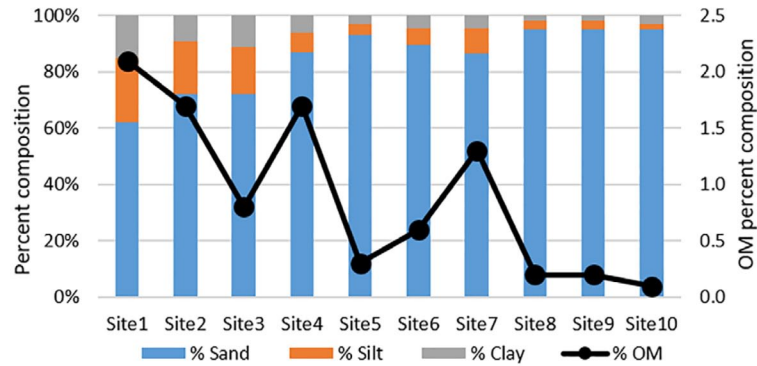
The mean  $K_v$  value varied from  $8.37 \times 10^{-3}$  to  $8.51$  m/day, about four orders of magnitude variation, indicating different types of soils with various structures across different stream orders. Stream gauge varied from 0.20 to 0.63 m while stream flow varied from 0.10 to  $0.67 \text{ m}^3 \text{ s}^{-1}$  across the 10 sites, calculated from the USGS gauge data at Palisade and the watershed area above each reach. The summary statistics of streambed  $K_v$  values and hydrological conditions at each of the 10 test sites (stream channels) are shown in Table 1.

The  $K_v$  values in Spring Creek (Site 1) and Stinking Water Creek (Sites 2 and 3) were low ( $< 4.78 \times 10^{-2}$  m/d) and consistent with a silt and clay streambed according to the grain size analysis of core samples (Figure 4). The low values are due to a large amount of small-size particles (silt and clay) that filled the pore space of the coarser sand particles. Conversely, the sites in the Frenchman Creek channel (Sites 4–10) had higher  $K_v$  values because they consisted primarily of sand which has higher hydraulic conductivity than silt or clay.

**Table 1** Summary statistics of  $K_v$  (m/day) values and hydrological conditions at 10 sites

Site	Channel shape	Sample size, n	Mean	Geometric mean	Minimum	Maximum	CV	Skewness	Kurtosis	Stream stage (m)	Streamflow ( $\text{m}^3 \text{s}^{-1}$ )
1	S	9	$8.37 \times 10^{-3}$	$7.57 \times 10^{-3}$	$3.88 \times 10^{-3}$	$1.64 \times 10^{-2}$	0.49	1.02	0.51	0.37	0.27
2	S	9	$1.67 \times 10^{-2}$	$1.25 \times 10^{-2}$	$6.11 \times 10^{-3}$	$4.81 \times 10^{-2}$	0.91	1.59	1.31	0.31	0.10
3	S	9	$4.78 \times 10^{-2}$	$4.04 \times 10^{-2}$	$1.01 \times 10^{-2}$	0.11	0.58	1.07	1.79	0.23	0.38
4	M	9	$4.22 \times 10^{-2}$	$2.23 \times 10^{-2}$	$4.87 \times 10^{-3}$	0.19	1.36	2.49	6.71	0.28	0.29
5	M	9	0.25	0.10	$4.48 \times 10^{-3}$	0.78	1.11	0.91	-0.54	0.63	0.32
6	M	12	0.27	0.16	$3.03 \times 10^{-2}$	1.13	1.12	2.27	6.03	0.46	0.32
7	M	9	0.14	$8.66 \times 10^{-2}$	$5.65 \times 10^{-3}$	0.45	0.96	1.66	2.84	0.27	0.34
8	M	9	8.51	0.62	$3.28 \times 10^{-3}$	26.8	1.32	0.86	-1.36	0.20	0.36
9	SC	9	8.02	1.42	$1.54 \times 10^{-2}$	51.8	2.06	2.91	8.60	0.46	0.65
10	M	9	3.50	1.81	$5.55 \times 10^{-2}$	9.04	0.96	1.02	-0.39	0.43	0.67

Abbreviations: CV, coefficient of variation; M, meandering (or curved) channel; S, straight channel; SC, slightly curved channel.



**Figure 4** Soil textural compositions based on sieve analysis of 30 cm soil cores from the center permeameters in the middle transects at the 10 sites. Core recovery was typically greater than 95%. OM is the organic matter.

Meandering channels showed a more heterogeneous distribution of streambed  $K_v$  when compared to straight channels as shown by the higher coefficients of variation (Table 1). This difference in average  $K_v$  values shows that morphologic changes in meandering reaches caused changes in the streambed leading to erosion and deposition of sediments (Song et al., 2017; Zhang et al., 2017). For meandering stream channels, higher streambed  $K_v$  values were observed at the erosional outer bends and the middle of the channels than at the depositional inner bends due to the differences in particle size (Table 2). As a result of morphologic changes in meandering reaches, relatively larger particles were observed at the erosional channel banks, and the lower streambed  $K_v$  observed at the depositional channel banks can be attributed to finer particles that can cause streambed sediment

**Table 2** Summary statistics of average  $K_v$  (m/day) values at meandering sites

Site	Channel shape	Erosional outer bend and middle of channel	Depositional inner bend
4	M	0.05	0.02
5	M	0.35	0.05
6	M	0.35	0.12
7	M	0.16	0.10
8	M	9.29	6.94
9	SC	11.82	0.43
10	M	4.27	1.96

Abbreviations: M, meandering (or curved) channel; SC, slightly curved channel.

clogging (Song et al., 2017). Moreover, even though our study was carried out in losing streams, higher streambed  $K_v$  were observed near the apex of reach bends. This agrees with the findings of Zhang et al. (2017) who observed that higher vertical water exchange fluxes significantly occurred near the apex of bends of gaining streams. Although streambed  $K_v$  along any side (left or right bank) of meandering channels showed some heterogeneity, in general, higher spatial variability was observed across stream channels than along stream channels.

### 3.2 Statistical distribution of streambed $K_v$

Owing to the fact that there are contradicting results as to which test is the optimal or best test (Yap & Sim, 2011), the aforementioned six normality tests were compared in order to see how they performed for both non-transformed and log-transformed  $K_v$  values at the 10 sites (Tables 3 and 4). A  $p$ -value less than .05 implies nonnormality of a distribution. The Q–Q plots for all combined log-transformed  $K_v$  values are shown in Figure 5. For each Q–Q plot, since both sets of quantiles come from the same distribution, the points form a fairly straight line, thus indicating that both sets of quantiles come from normal distributions.

**Table 3** Test for normality for non-transformed  $K_v$  values ( $p$ -value)

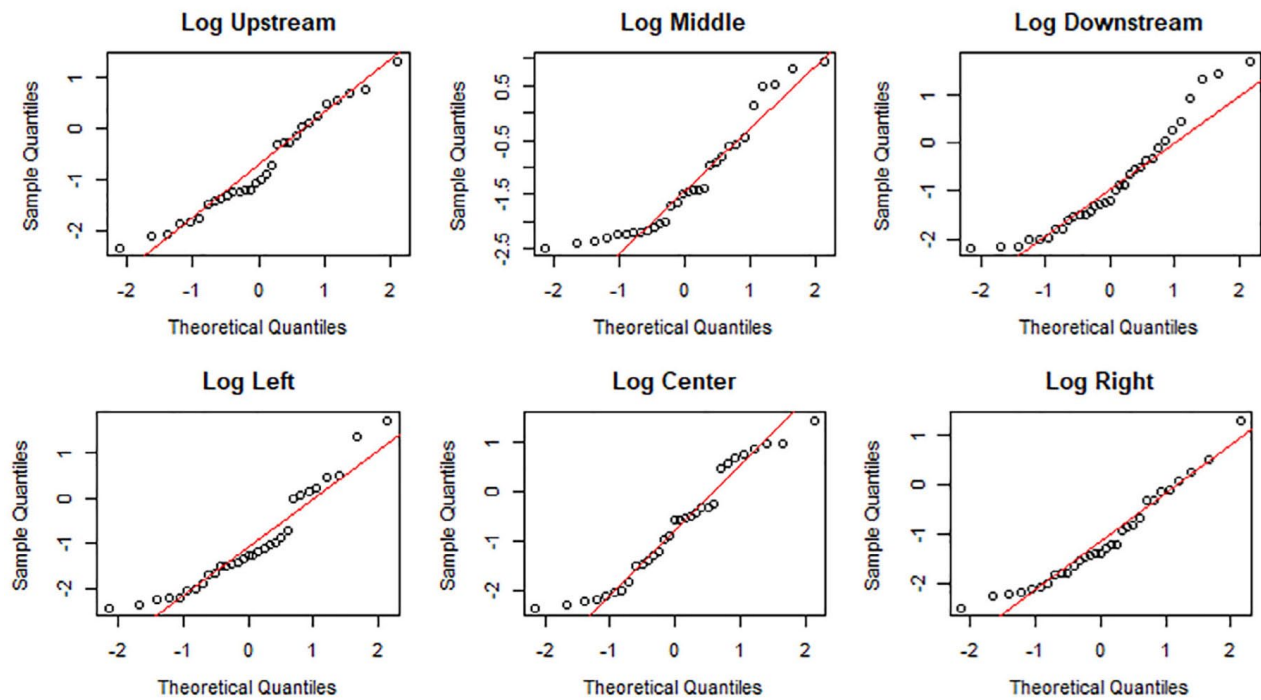
Site no.	Anderson–Darling	Cramer–von Mises	Lilliefors (Kolmogorov–Smirnov)	Pearson chi-square	Shapiro–Francia	Shapiro–Wilk
1	0.38	0.39	0.54	0.26	0.32	0.35
2	0.00*	0.00*	0.00*	0.00*	0.01*	0.00*
3	0.19	0.21	0.37	0.05	0.15	0.27
4	0.00*	0.00*	0.02*	0.01*	0.00*	0.00*
5	0.06	0.07	0.21	0.05	0.08	0.06
6	0.01*	0.01*	0.07	0.05	0.00*	0.00*
7	0.05	0.05	0.02*	0.15	0.03*	0.04*
8	0.00*	0.00*	0.02*	0.00*	0.01*	0.01*
9	0.00*	0.00*	0.00*	0.00*	0.00*	0.00*
10	0.06	0.08	0.12	0.15	0.07	0.06

\* $p$ -Value less than .05 implies non-normality of distribution.

**Table 4** Test for normality for log-transformed  $K_v$  values ( $p$ -value)

Site no.	Anderson-Darling	Cramer-von Mises	Lilliefors (Kolmogorov-Smirnov)	Pearson chi-square	Shapiro-Francia	Shapiro-Wilk
1	0.95	0.98	1.00	0.80	0.99	0.93
2	0.07	0.07	0.04*	0.09	0.08	0.06
3	0.23	0.24	0.12	0.26	0.18	0.34
4	0.77	0.82	0.84	0.26	0.79	0.75
5	0.47	0.47	0.74	0.46	0.55	0.46
6	0.91	0.89	0.93	0.34	0.96	0.96
7	0.24	0.28	0.33	0.46	0.15	0.28
8	0.20	0.26	0.62	0.15	0.23	0.14
9	0.05	0.03*	0.05	0.02*	0.09	0.12
10	0.19	0.27	0.28	0.46	0.09	0.13

\* $p$  Value less than .05 implies non-normality of distribution.

**Figure 5** Q-Q plots of all streambed  $K_v$  values. *Top*: Log-transformed values for all upstream, middle and downstream transects; *Bottom*: Log-transformed values for all left, center, and right permeameters.

**Table 5** Geostatistical summary of  $K_v$  (m/day) at 10 sites

Site	Model	MAD <sub>XV</sub>	SD <sub>XV</sub>
1	Gaussian	0.0016	0.0045
2	Linear	0.0024	0.0111
3	Gaussian	0.0197	0.0322
4	Spherical	0.0109	0.0774
5	Linear	0.1309	0.3176
6	Exponential	0.1116	0.3744
7	Spherical	0.0522	0.1564
8	Linear	2.6062	11.893
9	Exponential	1.9349	18.594
10	Gaussian	1.5081	3.3649

### 3.3 Spatial distribution of streambed $K_v$

The goodness-of-fit statistics for the best variogram models for predicting streambed  $K_v$  at the 10 sites (channels) are presented in Table 5. For a heterogeneous aquifer, the effective hydraulic conductivity is known to be the geometric mean since samples of hydraulic conductivity in most cases follow a lognormal distribution (Dagan, 1981; Desbarats & Srivastava, 1991; Gómez-Hernández & Gorelick, 1989; Madden, 1976; Warren & Price, 1961). Since the  $p$ -values from at least four of the six tests are greater than .05 for all the 10 sites (Table 4), it can be stated with 95% confidence that the  $K_v$  data for the sites tended to follow a lognormal distribution. The spatial structure of the 10 geometric means (one value for each site) was also analysed and a linear variogram was fit to the data. In addition, 10  $K_v$  values from middle transects and center permeameters were also used in gridding. Table 6 shows the nugget effect which is a reflection of measurement errors and variations that occur over distances smaller than the spacing of permeameter tests. The nugget effect was highest when all 93 tests were used (2.77) due to its highest variability of  $K_v$  values as indicated by its standard deviation and the coefficient of variation.

Results in Table 6 also show that the goodness-of-fit statistics are better when the geometric means for the 10 sites are used instead of all 93  $K_v$  values. This is because the geometric mean gives a better summary of values at each site. When all values are used, there are a

**Table 6** Statistics, parameters, and cross-validation goodness of- fit for linear variograms using different sample sizes

	N = 93*	N = 10**	N = 10***
Standard deviation	6.80	0.66	3.33
Coeff. of variation	3.36	1.54	1.98
Nugget effect	2.77	2.156E-02	0
Anisotropy angle	30.8°	37.9°	38.1°
Variogram slope	8.75E-05	1.53E-06	8.818E-05
MADXV	0.18	0.08	0.16
SDXV	7.21	0.26	0.87

\* Statistics calculated using all permeameter tests.

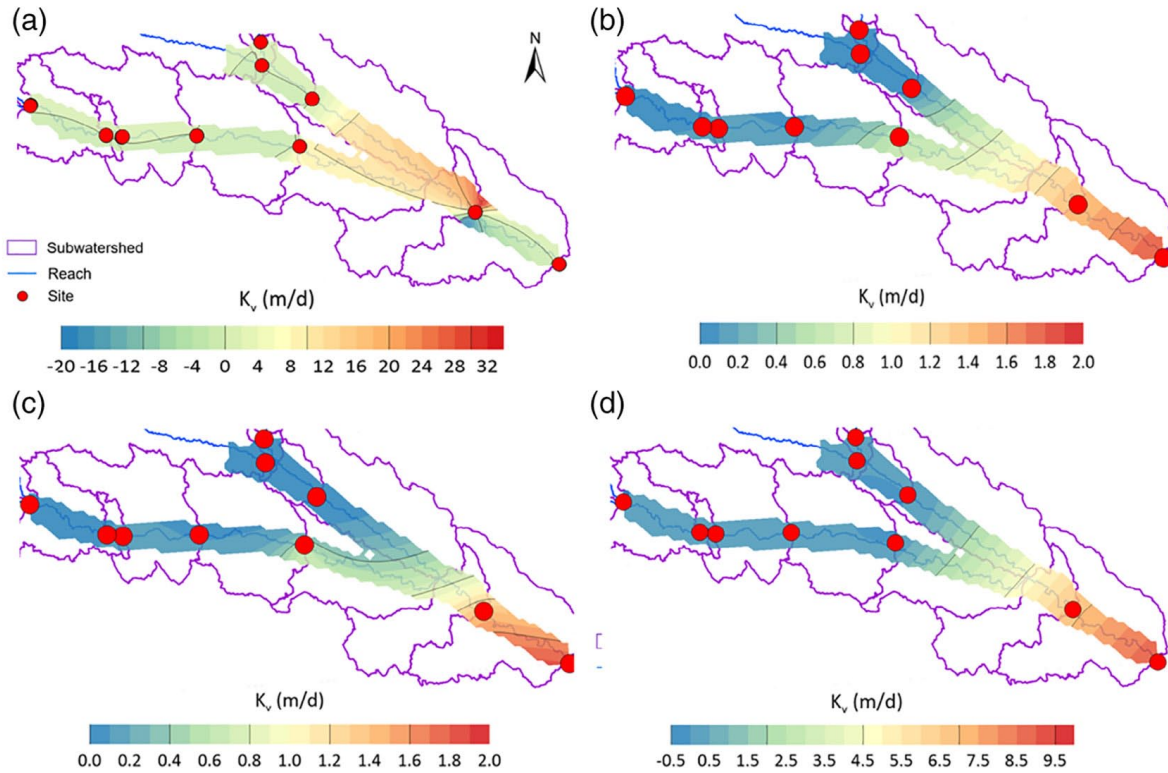
\*\* Statistics calculated using the geometric means of the 10 sites.

\*\*\* Statistics calculated using the values from middle transects and center permeameters.

few extreme  $K_v$  values that tend to affect spatial interpolation, leading to negative  $K_v$  values downstream (Figure 6a). Negative  $K_v$  are due to Site 9 having the highest heterogeneity (0.015–51.8 m/d), coefficient of variation and skewness of the 10 sites (Table 1). This supports the argument that using the 93  $K_v$  values directly (without using geometric mean per site) can lead to spatial interpolation errors. Furthermore, using only 10 values from the middle transects and center permeameters gives MADXV value of 0.16 which is similar to the MADXV estimated when all 93  $K_v$  values are used (MADXV = 0.18 m/d) although the SDXV value of 0.87 m/d is much lower than 7.2 m/d.

The resulting contours are plotted using non-transformed  $K_v$  values gridded as linear (Figure 6a–d). Since using geometric means produced the best grid, log-transformed  $K_v$  values were also gridded as linear (Figure 6c). It is important to note that the contours do not start from the headwater subwatersheds because, at the time of study, the channels of the headwater subwatersheds were dry. Thus, extrapolation beyond the range of the data was treated sceptically since it was based on untestable assumptions about the behaviour of the data beyond their observed support, and there were no points around to constrain the models. Future work will focus on using a different approach for  $K_v$  (or saturated hydraulic conductivity,  $K_{sat}$ ) estimation in dry streambeds. This will be significant for sites in headwater subwatersheds where streams are normally dry but recharge the groundwater system seasonally or intermittently.





**Figure 6** Spatial pattern based on (a) all 93 permeameters at the 10 sites using non-transformed values; (b) 10 geometric means using non-transformed values; (c) 10 geometric means using log-transformed values; (d) 10 values from center permeameters in the middle transects using non-transformed values. The boundary of spatial interpolation has been increased for better visualization of contour lines.

Figure 6b, c shows the similarity between the contours for non-transformed and log-transformed geometric mean values when both are gridded as linear. Although using 10  $K_v$  values from center permeameters in the middle transects for gridding produced a relatively similar spatial pattern (Figure 6d) when compared to using geometric means (Figure 6b,c), there tends to be an over-prediction of  $K_v$  values across the watershed. This is because of the higher SDXV value of 0.87 m/d when compared to a value of 0.26 m/d (Table 6).

### 3.4 Effect of sedimentation processes on spatial distribution of $K_v$

In order to better understand the sedimentation processes, this study used the SSURGO database which consists of information about soil as collected by the National Cooperative Soil Survey over the

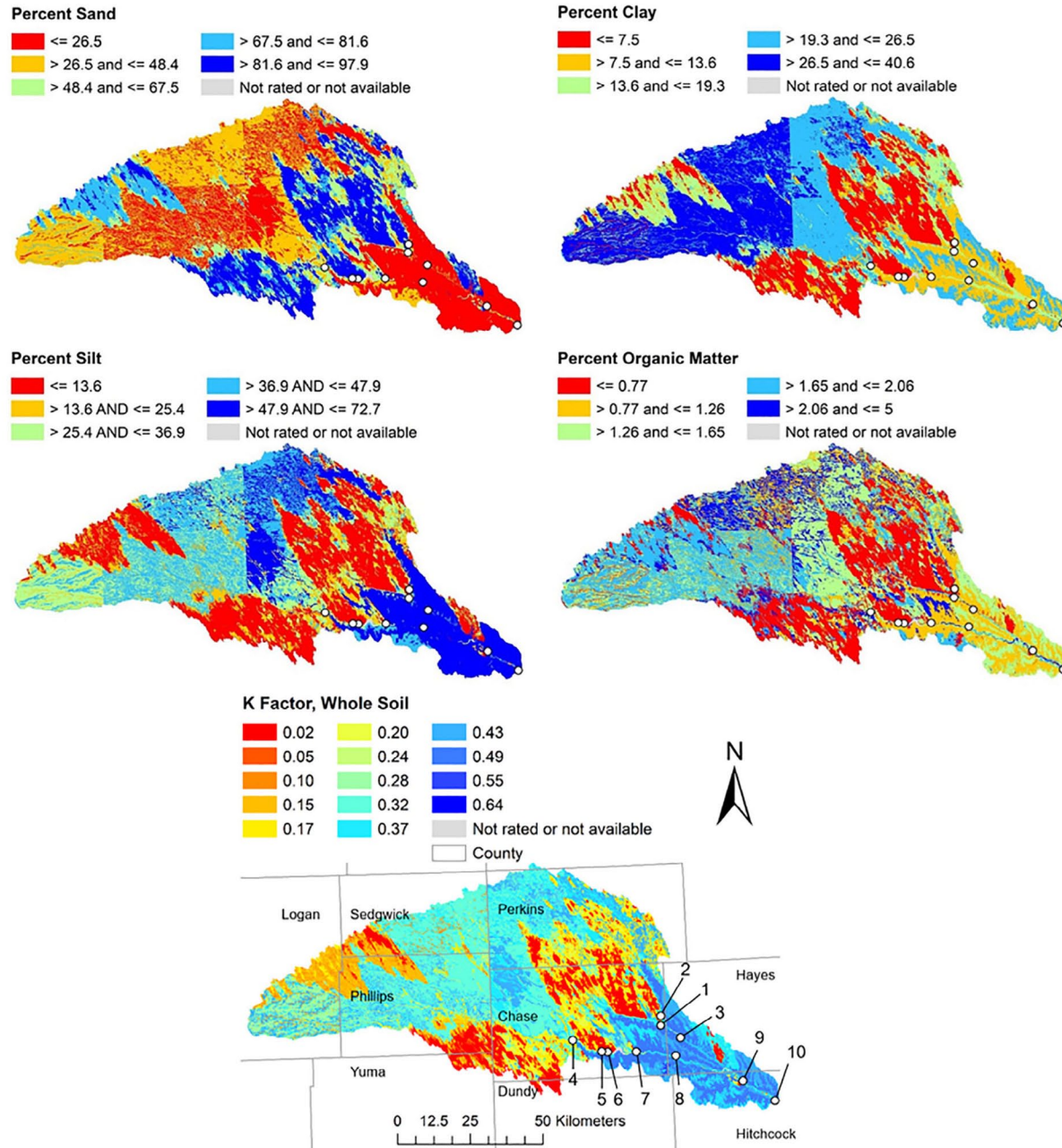
course of a century (USDA-NRCS, 2018) in the United States. We extracted the textural data (i.e., organic matter, sand, silt, and clay contents) and erodibility index for each county covered by the Frenchman Creek Watershed in Nebraska and Colorado (Figure 7; Abimbola et al., 2020). Although sieve analysis (Figure 4) of permeameter cores show that the streambed is mostly sandy (>95%) in the downstream areas with higher  $K_v$  values, the spatial distributions of the texture of topsoil (0–50 cm) show that there is about twice more silt than sand in the downstream areas (Sites 9 and 10) of the watershed compared to the upland areas (Figure 7). The higher  $K_v$  values in downstream areas is a result of downstream transition which occurs in stream channels as a result of the assortment of sediments coming from all points in a subwatershed or watershed and the spatial variation of soil textural properties of the sediment. Moreover, the sediment source of the tributaries plays a major role in controlling the grain-size distribution for streambed sediments (Singer, 2008).

In various applications of distributed hydrological models, spatial interpolation results of  $K_v$  can be useful in hydrological modelling using tools such as Soil and Water Assessment Tool (SWAT) and MODFLOW. Instead of assuming homogeneity of streambed  $K_v$  across a watershed, using grid results as input will help in calibrating this important parameter at reach or sub-watershed scale. Consequently, this will reduce prediction uncertainty and improve hydrological modelling results.

## 4 Conclusions

Spatial variability in the magnitudes and spatial patterns of streambed  $K_v$  will enhance the understanding of water and solute fluxes between groundwater and surface water in different morphologies, thus reducing uncertainties on streamflow prediction.

Although streambed  $K_v$  along any side (left or right bank) of meandering channels showed some heterogeneity, in general, higher spatial variability was observed across stream channels than along stream channels. Higher streambed  $K_v$  values were observed at the erosional outer bends and the middle of the channels than at the depositional inner bends. Also, higher streambed  $K_v$  was observed near the apex of reach bends.



**Figure 7** Spatial distributions of the soil properties for 0–50 cm depth. The sharp vertical and horizontal boundaries between classes in some maps are county boundaries which are effects of differences in how county soil surveys were conducted.

Geostatistical analysis shows that using the geometric means of the 10 sites performs better than using either all the  $K_v$  values from the 93 permeameter tests or 10  $K_v$  values from the middle transects and center permeameters. In general, understanding the spatial variability of  $K_v$  and the uncertainty involved in both measurement and gridding is crucial to accurately predicting how stresses and environmental signals propagate through the hydrologic system. Grid results can be useful in calibrating streambed  $K_v$  at reach or sub-watershed scale in distributed hydrological models.

Future work will focus on using a different approach for  $K_v$  (or saturated hydraulic conductivity,  $K_{sat}$ ) estimation in dry streambeds. This will be significant for sites in headwater subwatersheds where streams are normally dry but recharge the groundwater system seasonally or intermittently. More work is also needed to understand the spatio-temporal variations in streambed  $K_v$  across various natural systems, and to develop fast and accurate methods that will allow the surveying and gridding of larger watersheds, and also allow the representation of the dynamic behaviour of streambed  $K_v$  in distributed hydrological models.

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**Data availability** — The data that support the findings of this study are available from the corresponding author upon reasonable request.

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