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Wei Luo  
*Northern Illinois University, wluo@niu.edu*

Darryll T. Pederson  
*University of Nebraska-Lincoln, dpederson2@unl.edu*

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Hydraulic conductivity of the High Plains Aquifer re-evaluated using surface drainage patterns

Wei Luo\(^1\) and Darryll T. Pederson\(^2\)

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1. Introduction

Groundwater flow is governed by Darcy’s Law, which states that the groundwater flow rate is proportional to the hydraulic gradient:

\[ q = -K \frac{dh}{dl}, \]  
(1)

where \( q \) [LT\(^{-1}\)] is the volumetric groundwater flow rate per unit area (or flux); \( K \) [LT\(^{-1}\)] is the hydraulic conductivity; and \( dh/dl \) is the hydraulic gradient along the flow direction. The value of \( K \) is typically obtained by measuring discharge through a porous medium sample under different hydraulic gradients in the laboratory, by conducting pumping tests in the field and observing the effect on water levels in the subsurface, or by computer modeling coupled with field measurements of quantities such as discharge and heat flow [Daniel, 1994; Freeze and Cherry, 1979; Hornberger et al., 1998; Ingebritsen et al., 1994, 1992]. These methods are often expensive and time consuming.

2. The interaction of surface water and groundwater has long been recognized [Carlston, 1963; De Vries, 1976; Dunne, 1990; Freeze and Cherry, 1979]. Although overland flow can play a significant erosive role, the groundwater discharge and seepage induced weathering processes prepare and precondition the rocks for preferential erosion in areas weakened by weathering [Luo et al., 2011, 2010]. The erosion further concentrates groundwater flow at the points of incision due to higher and directional groundwater gradients, guiding further valley development [Dunne, 1990; Pederson, 2001]. If the aquifer is not effectively drained (i.e., recharge is greater than discharge), parts of the aquifer will experience groundwater level rises, resulting in an increase in hydraulic gradient and even saturation toward the surface in those areas, which in turn will encourage more weathering, erosion and development of additional surface drainage until an equilibrium is reached and a unique overall drainage dissection pattern is generated [Luo et al., 2011, 2010]. The drainage system resulting from such long term interplay between surface water, topography, and subsurface aquifer properties will establish a dynamic equilibrium such that the aquifer is effectively drained and recharge and discharge will be equal.

3. As shown in Figure 1, we have previously established the following quantitative relationship between surface drainage patterns and subsurface aquifer hydraulic conductivity using a derivative of Darcy’s law under the following assumptions: (1) the aquifer is effectively drained through the long term development and interplay among surface water, groundwater, and topography that has established a steady-state dynamic equilibrium; and (2) the groundwater flow is primarily horizontal such that DuPuit-Forchheimer assumptions apply [Deming, 2002; Luo et al., 2011, 2010]:

\[ K = \frac{R}{D^2[H^2 - (H - d)^2]} = \frac{4W^2R}{[H^2 - (H - d)^2]}, \]  
(2)

where \( K \) [LT\(^{-1}\)] is the hydraulic conductivity, \( W \) [L] the length of effective groundwater drainage, \( D \) [L\(^{-1}\)] drainage density, \( R \) [LT\(^{-1}\)] recharge rate, \( d \) [L] valley depth, and \( H \) [L] aquifer thickness. Note that \( D \) and \( W \) are related as

\[ D = \frac{1}{2W}. \]  
(3)

We have applied our method to the Oregon Cascades and Martin drainage systems [Luo et al., 2011, 2010] and produced results consistent with previous studies. Here we report results of applying refinements of our method of estimating \( K \) to the intensively studied High Plains Aquifer

\(^1\)Department of Geography, Northern Illinois University, De Kalb, Illinois, USA.

\(^2\)Department of Earth and Atmospheric Sciences, University of Nebraska-Lincoln, Lincoln, Nebraska, USA.

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Figure 1. Diagram illustrating the conceptual model for deriving $K$ from drainage dissection patterns [Luo et al., 2011]. $W$ is the effective drainage length and is related to drainage density $D$ as $D = 1/(2W)$. Other variables are discussed in text.

(HPA), which provided independent estimates of $K$ for comparison with our results.

2. Description of High Plains Aquifer

[4] HPA is one of the largest fresh water aquifers in the world and accounts for 30% of the groundwater used for irrigation in the United States [Dennehy, 2000]. It consists mainly of hydraulically connected geologic units of late Tertiary to Quaternary age [Gutentag et al., 1984] and represents the product of weathering, erosion, and fluvial transport processes associated with the post-Paleozoic uplift of the Rocky Mountains [Heller et al., 2003; McMillan et al., 2002] (see inset of Figure 2 for elevations). Depending on climate, rate of uplift, and continental tilting, river systems draining the Rocky Mountains transported sediments of differing particle sizes to differing distances into the high plains [Heller et al., 2003; McMillan et al., 2002]. In general, because of decreasing topographic gradients to the east, one should expect finer-grained sediment particles in alluvium associated with the lower reaches of the river systems [Frings, 2008]. Because $K$ is a function of sediment particle size, one should also expect that $K$ values decrease in alluvium with distance downstream. However, the interplay between episodic nature of uplift, continental tilting, and climatic conditions, coupled with the meandering of river systems, can produce intervals of projection of coarser sediments into lower stream reaches [Heller et al., 2003; McMillan et al., 2002], leading to dramatic changes in aquifer properties in a small area. The aforementioned interacting factors generate repeated episodes of stream incision and deposition, which cause valley systems to be filled with later floods of sediments extending sometimes to upland areas [Heller et al., 2003; McMillan et al., 2002]. Periods of incision may be limited to the valleys themselves or extend over broad areas [Heller et al., 2003; McMillan et al., 2002]. There is sufficient evidence from modern streams and ancient stream deposits to indicate that braided streams played an important role in depositing the HPA, often characterized by coarsening-upwards as well as finerning-upwards sediments [Gutentag et al., 1984]. Thus, aquifer properties (including $K$) at any point are a function of the 3-D nature of the alluvial sediments. Our calculated horizontal $K$ value represents the aggregate $K$ over the entire saturated thickness, which is composed of different sub-layers of sediments with different grain sizes.

3. Estimations of Parameters

[5] We used more sophisticated data sets and approaches in this study as compared to previous studies [Luo et al., 2011, 2010] to better quantify the relationship of $K$ and stream drainage. In previous calculation of $K$ we have used drainage density $D$, which is a half of the inverse of effective drainage length $W$ smoothed over a specified neighborhood. Here we directly use effective drainage length $W$, which is represented by the distance from any point to the nearest stream along steepest descent (downslope) averaged by watershed. We believe this is a better representation because it is not first smoothed over a specified neighborhood and thus reveals more details. Streams were extracted from DEM data using a morphology based algorithm, which calculates the curvature of the DEM by fitting a local surface to the DEM [Luo and Stepinski, 2008]. This approach has been proven to more precisely reflect the true surface dissection pattern than the traditional flow direction based D8 algorithm and its variations [Luo and Stepinski, 2008]. The DEM data used was the latest SRTM digital elevation data with 90-m resolution (http://srtm.csi.cgiar.org/). The aquifer thickness $H$ was obtained from USGS GIS data (http://co.water.usgs.gov/nawqa/hpgw/GIS.html) [Gutentag et al., 1984] by subtracting the aquifer base elevation from the topographic surface elevation. We did not use the water table elevation because $H$ as shown in Figure 1 is the maximum of the aquifer thickness. The difference between topographic surface elevation and aquifer base elevation, averaged over the watershed, offers the appropriate estimate of $H$. The valley depth $d$ was estimated using a Black Top Hat transformation [Rodriguez et al., 2002] with a moving window of ~2.2 km radius, determined by comparing results from different radii with manually measured depths [Luo et al., 2011, 2010].

[6] With the exception of areas of sandy soil, the current average annual recharge to HPA under natural conditions is only a few tenth of an inch [Gutentag et al., 1984], but there is considerable variation according to climate, soil, and topographic conditions [Gutentag et al., 1984]. Nonetheless, we can estimate the recharge rate $R$ from the long term mean annual precipitation $p$ and infiltration percentage $i$ as

$$R = p \times i.$$  

We used the 30-year (1971–2000) spatially variable average annual precipitation data from PRISM [Chris, 2006] and assigned infiltration percentage according to spatially variable soil type, referencing previous studies [Gutentag et al., 1984] (see Table 1). Using infiltration values of different soils in Table 1, the mean of calculated annual $R$ of the study area is about 3.6 cm/year (or 1.4 inch/year). This is higher than the current annual recharge of HPA under natural conditions, but it is a reasonable estimate for the region during the late Tertiary time when the HPA was deposited because the climate then was wetter and warmer than present [Gutentag et al., 1984]. By using the long term mean annual precipitation and soil type dependent infiltration percentage, the spatial variability due to climate, soil, vegetation, and topography were all implicitly taken into consideration, as soils reflect both climate and consequent vegetation as well as topography.

[7] All parameters (and thus result) are averaged by the watershed that drains each stream. The watershed boundaries are not shown in Figure 2a for clarity. The average size
Figure 2. Comparison of \( K \).
(a) The order of magnitude of \( K \) (in m/s) derived from our method [Luo et al., 2011, 2010];
(b) the order of magnitude of \( K \) (in m/s) derived from traditional method from well log data [Gutentag et al., 1984] (note the legend and scale bar apply to all three panels); (c) the order of magnitude of \( K \) (in m/s) derived from Nebraska Well Registry data; inset shows the location of High Plains Aquifer within US (background color from blue to orange indicates elevation from low to high).
of the watersheds is 6.2 km$^2$ with a standard deviation of 50.5 km$^2$.

4. Results and Discussion

The first comprehensive $K$ estimate of the HPA at a regional scale was reported in a USGS publication [Gutentag et al., 1984] and was based on the assumption that vertical distribution of sediments in HPA was random, typical of sediments deposited by braided streams [Gutentag et al., 1984]. Under this assumption, the areal distribution of $K$ was estimated as thickness-weighted average of different lithologic layers in drillers’ logs (using typical values of $K$ for specific lithology), supplemented with data from aquifer well pumping tests [Gutentag et al., 1984].

Figure 2 shows our result (Figure 2a) in comparison with USGS data (Figure 2b), which was interpolated from original contour data [Gutentag et al., 1984] for easy comparison. The general spatial pattern of high $K$ values ($>1 \times 10^{-3}$ m/s) in our results is consistent with that of USGS results [Gutentag et al., 1984], but our results clearly show much more detailed spatial variation. In general, the higher values are located closer to the Rocky Mountains, consistent with what one would expect in a mountain front environment. The high $K$ value near the Platte and Arkansas rivers are also prominent, reflecting the larger sediment particles in their alluvial deposits. Our $K$ values in parts of the lower reaches of the Platte and Arkansas rivers and southern HPA are 1–2 orders of magnitude higher than those of USGS data [Gutentag et al., 1984]. The reverse is true in the Sand Hills area north of the Platte River, which will be in discussed later. The comparison between our results and USGS data is also clearly shown in the two histograms in Figure 3.

Whereas the highest frequency is in the order of $10^{-3}$ m/s for both results, the overwhelming majority of USGS data fall in the order of $10^{-4}$ m/s and $10^{-3}$ m/s with very small percentage in smaller orders and virtually no values in the orders of $10^{-2}$ m/s and $10^{-1}$ m/s. We interpret this difference as due in part to far more detailed sampling with our approach, which is essentially at pixel level (~1.4 million pixels) but averaged by watershed, as compared to the USGS $K$ derived from using a limited number of drillers’ logs (several thousand) supplemented with available aquifer pump test data [Gutentag et al., 1984]. Another possible cause for the differences may be that the effect of vegetation (transpiration) was only implicitly considered with soil properties, which may result in somewhat overestimate of infiltration. However, we do not believe this is a major factor. More importantly our $K$ reflects the actual nearly horizontal groundwater flow in the HPA section representing the total thickness of saturation consisting of sediment sub-layers of different sizes resulted from the changing sediment deposition over geologic time. Our greater range of $K$ values (Figure 3) better reflects the particle size

<table>
<thead>
<tr>
<th>Soil Type</th>
<th>Assigned Infiltration Percentage $i$ (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Clay loam</td>
<td>2</td>
</tr>
<tr>
<td>Loam</td>
<td>5</td>
</tr>
<tr>
<td>Silt loam</td>
<td>8.5</td>
</tr>
<tr>
<td>Sandy loam</td>
<td>13</td>
</tr>
<tr>
<td>Sand</td>
<td>18</td>
</tr>
<tr>
<td>Caliche</td>
<td>1</td>
</tr>
</tbody>
</table>

Table 1. Infiltration Percentage of Different Soil Type

Figure 3. Comparison of histograms of our $K$ (Figure 2a) and USGS data (Figure 2b).
distribution of sediments deposited by river systems issuing from the Rocky Mountains. The effect of vertical flow that may occur in the immediate vicinity of streams can be safely ignored because (1) the maximum pre-development water table slope is only 4.1% with a mean of 0.3%, much less than 10%, the slope below which the error introduced by using the Dupuit–Forchheimer assumptions is considered negligible [Grismer and Rashmawi, 1993]; and (2) the ratio of mean vertical aquifer thickness to the mean horizontal length of the aquifer considered is only ~0.005, much less than 1, the ratio below which vertical flow can be ignored [Phillips, 2003].

Because most wells are not screened through the full saturated thickness (instead they are commonly screened in layers with high production rate), \( K \) derived from traditional well methods based on Theis Equation and its derivatives basically represents the horizontal \( K \) of the screened layer, which is only a subset of the \( K \) calculated by our method and not directly comparable to our \( K \) in terms of absolute value. However, the traditional well methods derived \( K \) can at least offer some general indication of the true spatial variation of \( K \). To have one more independent comparison with our results, we used registered well data from Nebraska and produced the map in Figure 2c following a general formula described previously based on the Thiem equation [Pederson, 1979]. A total of 142,721 wells with discharge and pumping rates were used to estimate the transmissivity at each well location, which was then converted to \( K \) by dividing by saturated thickness (pre-development water table elevation minus aquifer base elevation) and interpolated to a raster as shown for easy comparison. The general spatial pattern of \( K \) derived from the well data method corresponds very well with our result but the absolute values are different in some areas. Importantly, this overall widespread matching pattern between our method and the well data method (especially along the Platte River) reveals for the first time, to the best of our knowledge, the distinct relationship between surface drainage density and subsurface aquifer \( K \) on a regional scale.

In the Sand Hills area north of the Platte River, the general pattern of lower \( K \) value than that of the surrounding area is consistent with the well data method result and USGS data. However, our result in this area is ~2 orders of magnitude lower than that derived from the well pumping data and USGS data. Our method is not valid where there are mobile sand dunes because they are young topographic features and interdunal areas can give false positives for nonexisting drainage systems when analyzing DEM data using morphology-based algorithm. These false positives led to unusually short downslope distances to streams (low \( W \), high \( D \)) and thus low \( K \) (see equation (2)) for the Sand Hills area of Nebraska as demonstrated in Figure 2a.

## 5. Summary

Overall, we believe our method offers an effective and efficient alternative to existing methods for determining the detailed spatial variation of effective horizontal \( K \) in the HPA, a large aquifer system spanning hundreds of kilometers. Our method uses the extensive available database such as aquifer thickness, precipitation distribution, soils, DEM, which allows a detailed analysis of the aquifer (on the order of 1.4 million pixels) and thus reveals much more detailed variation of the aquifer property across the HPA, reflecting its geologic history and hydrogeologic reality. The continuous distribution of horizontal aquifer \( K \) yielded by our approach is a critical parameter in most 2-D groundwater management and transport models. In contrast, the traditional well pumping method samples limited and isolated volumes of the HPA with results better suited for development of additional wells but of questionable value for 2-D modeling and transport modeling. Our method is not appropriate for areas where the existing landscape is formed by non-fluvial geologic processes such as mobile sand dunes or for very young landscapes which have not reached a dynamic equilibrium.

Data needed for our method are widely available. Of the parameters needed as input on the right side of equation (2), effective drainage length \( W \) and valley depth \( d \) can be well constrained from DEM data, recharge rate \( R \) and aquifer thickness \( H \) can be reasonably assumed based previous literature of the study area or of similar areas. Our previous studies [Luo et al., 2011, 2010] have shown that uncertainties within \( H \) do not significantly alter the results (order of magnitude). Our method is sensitive to precipitation and infiltration rates. One order of magnitude change in \( p \) or \( i \) will result in one order of magnitude change in resultant \( K \). Our method of estimating \( R \) implicitly takes into consideration the spatial variability due to climate, soil, vegetation, and topography and represents the best we can do with available data. As long as the \( R \) (i.e., \( p \) and \( i \)) can be reasonably estimated (such as this study) the uncertainty would be less than an order of magnitude. Because DEM data are available for most areas, our method can provide reasonable estimates of the spatial distribution and order of magnitude of horizontal \( K \) under the stated assumptions, and is especially suitable for remote areas where other data are sparse and/or physical accessibility is limited.

In addition, because \( K \) is largely dependent on particle size, our method also shows the pattern of sediment deposition by rivers draining the Rocky Mountains, the aggregate of which can be considered a megafan. One can identify individual cycles of sediment deposition by ancient river systems. Similar maps can be readily prepared for other similar geologic environments using our method. Thus our method can potentially be used as tool to study patterns of past sediment movement and deposition.

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W. Luo, Department of Geography, Northern Illinois University, DeKalb, IL 60115, USA.

D. T. Pederson, Department of Earth and Atmospheric Sciences, University of Nebraska-Lincoln, Lincoln, NE 68588-0340, USA.