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Methods Note/

Net Recharge vs. Depth to Groundwater Relationship in the Platte River Valley of Nebraska, United States

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Abstract

One-km resolution MODIS-based mean annual evapotranspiration (ET) estimates in combination with PRISM precipitation rates were correlated with depth to groundwater (d) values in the wide alluvial valley of the Platte River in Nebraska for obtaining a net recharge (Rn) vs. d relationship. MODIS cells with irrigation were excluded, yielding a mixture of predominantly range, pasture, grass, and riparian forest covers on sandy soils with a shallow groundwater table. The transition depth (d_t) between negative and positive values of the net groundwater recharge was found to be at about 2 ($\pm$1) m. Within 1 ($\pm$1) m of the surface and at a depth larger than about 7 to 8 ($\pm$1) m, the mean annual net recharge became independent of d at a level of about -4 ($\pm$12)% and 13 ($\pm$10)%, respectively, of the mean annual precipitation rate. The obtained $Rn(d)$ relationship is based on a calibration-free ET estimation method and may help in obtaining the net recharge in shallow groundwater areas of negligible surface runoff where sufficient groundwater-depth data exist.

Introduction

Regional-scale groundwater modeling is commonly performed at a linear scale of 10 to 10^3 km. Typical cell size of these models is in the order of 1 km. Groundwater recharge appears as an important source term in hydrogeological models and must be estimated from data having the same spatial resolution. With the free availability (starting in 2000) and global coverage of the Moderate Resolution Imaging Spectroradiometer (MODIS) data, evapotranspiration (ET)-estimation results based on MODIS values (e.g., Mu et al. 2011; Senay et al. 2011; Szilagyi et al. 2011a) are expected to become more frequently applied in groundwater modeling and recharge assessment (Senay et al. 2011; Szilagyi et al. 2011b, 2012).

For areas with negligible overland flow due to sandy soils and/or flat topography, such as the Nebraska Sand Hills (Figure 1) or the interfluvial sand plateau region between the Danube and Tisza Rivers in Hungary, the mean annual net groundwater recharge (Rn) can be determined (Szilagyi et al. 2011b, 2012) as

$$Rn \approx P - \mathrm{ET} \tag{1}$$

where P is the mean annual precipitation, and variations of water storage in the vadose zone are neglected over a suitably long period of typically several years. The net recharge value can be negative in areas where ET exceeds precipitation, typically in shallow groundwater areas, such

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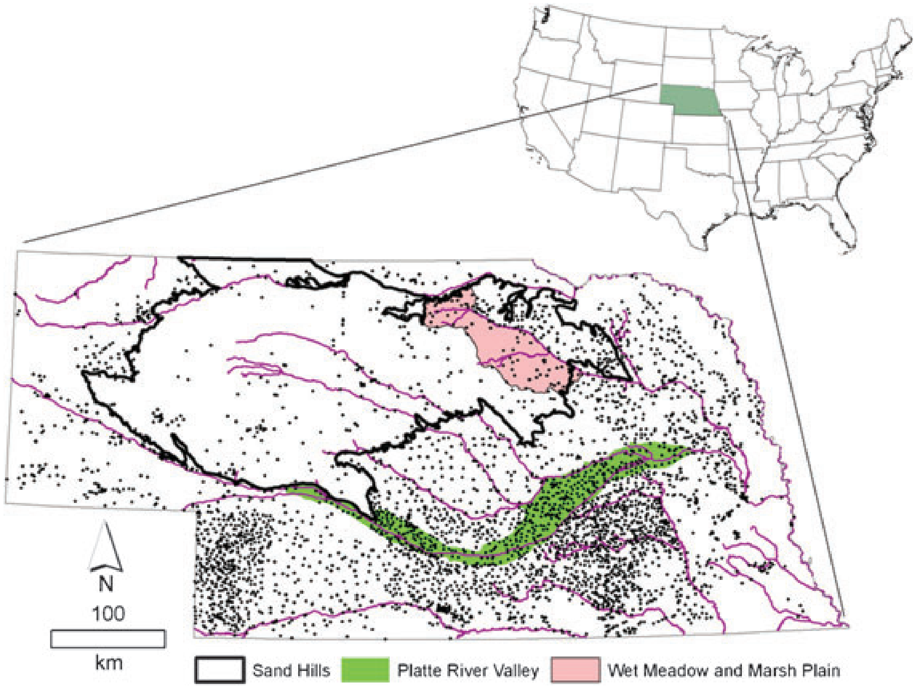

Figure 1. The two largest shallow-groundwater ecoregions (shaded) in Nebraska. The Wet Meadow and Marsh Plain region within the Sand Hills was dropped from further analysis because of poor-quality depth to groundwater grid values due to sparse well locations.

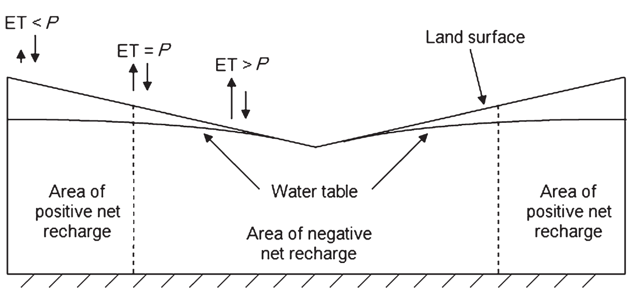

Figure 2. Schematic cross-section of an alluvial valley. Net recharge is positive where the water table is deep and the corresponding ET is lower than precipitation (*P*). Negative net recharge occurs near stream channels where ET enhanced by the shallow water table exceeds precipitation. In energy-limited environments, negative net recharge may never occur.

as wetlands and river valleys (Figure 2). Negative *Rn* means borrowing water for ET from the groundwater due to deficit of precipitation in a given location. Availability of model-specific correlation between *Rn* and depth to water table (*d*) would be a substantial asset prior to the development of regional groundwater-flow models. Inference of this relationship requires the availability of three maps including ET, *P*, and depth to the water table (or in lieu of the latter, a digital elevation model [DEM], and an elevation of the water table map).

In the groundwater literature, a term called groundwater ET (*ETg*) is frequently employed (e.g., McDonald and Harbaugh 1988; Harbaugh and McDonald 1996), which relates to the part of ET that originates from the groundwater. Groundwater models, such as MODFLOW, often use a linear *ETg* vs. depth function (McDonald and Harbaugh 1988; Harbaugh and McDonald 1996). In several recent applications, the linear *ETg* function has been replaced by exponentially decaying ones (Shah et al. 2007; Luo et al. 2009) or by segmented linear functions that generally reflect an exponential decay (Banta 2000). Although for the original linear function only the extinction depth needs to be specified (beside the actual ET rate), the exponential functions (including the segmented linear approach) need extra parameters, such as a decay coefficient.

For estimating groundwater source/sink terms in groundwater models, recharge to the groundwater must also be estimated concurrently with and separately from the groundwater ET term. The objective of this study is to obtain the net recharge as a function of depth to the groundwater. Note that the aforementioned formulation (Equation 1) of the net recharge term already yields the sum of the previous two terms (i.e., recharge to the groundwater and groundwater ET). With the help of a 1-km resolution DEM, precipitation (PRISM Climate Group 2004) and MODIS-based ET data (i.e., CREMAP [Complementary-Relationship-based Evapotranspiration MAPping] by Szilagyi et al. 2011a), our objective of determining the net recharge rate as a function of depth to the groundwater can be obtained without any sort of calibration, as demonstrated in the following.

Although subgrid scale sampling of the land surface-elevation data may improve groundwater ET (Li et al. 2008; Kambhammettu et al. 2012) and therefore the present *Rn* estimates, depth to groundwater or elevation of the groundwater-table values are typically not available at the subgrid or block scale (which can be smaller than 1 km) for regional modeling due to the high associated cost of obtaining this kind of information. Even in Nebraska (land area of 200,345 km^2), where the economy substantially relies on groundwater use for irrigation, a state-wide groundwater-elevation map is interpolated from a total of about 5000 measurement locations (i.e., one point for each 40 km^2 on average), and the latest such map is available (CSD 2001) only for 1995.

Estimation of the *Rn*(*d*) Function

Two large, shallow groundwater regions were selected within Nebraska for obtaining the $Rn(d)$ function: the Platte River Valley, and the Wet Meadows and Marsh Plain ecoregion (Chapman et al. 2001), north of the Platte River (Figure 1). After inspection of the depth to groundwater values, the latter region was dropped from further analysis because a large percentage (20%) of the values yielded negative depth (i.e., groundwater elevation was higher than the land surface) as the result of subtractions. Figure 1 displays about two-third (3640 points) of the original well locations used for the 1995 groundwater-elevation map, that is, only those places where groundwater-elevation data are collected by the Conservation and Survey Division of the University of Nebraska. Clearly, a low density of well locations is the cause of the large number of negative depth values in the Wet Meadows and Marsh Plain ecoregion.

The Platte River Valley, however, with its dense irrigation-well system (in excess of 420 well locations) for groundwater-elevation readings in the spring when pumping effects are minimal, proved to yield more accurate depth to groundwater values (i.e., only 4% of the cells had negative values). The wide alluvial valley has an area of about 8000 km^2, enjoys a continental climate with a mean depth to the groundwater of 4.36 m, and a mean annual precipitation of 639 mm (PRISM Climate Group 2004), changing from 500 mm in the west to 680 mm in the east. The valley is covered with cropland, predominantly irrigated corn and soybean, giving way to riparian forest of primarily cottonwood (*Populus*) species mixing with range, pasture, and grassland close to the braided channel.

A recently revised version (Szilagyi 2013) of the original, state-wide CREMAP-estimated mean annual ET rates of 2000 to 2009 (Szilagyi et al. 2011a) were correlated with the depth to groundwater values at the MODIS-grid scale of about 1 km. This included the following.

A 30-m DEM was resampled at the MODIS cell size with assigning the mean value of the corresponding 30-m cell values. The 1995 well readings of groundwater elevation were hand-interpolated (Summerside et al. 2001) into approximately 15 m (50 ft) and approximately 3 m (10 ft) increment contours, the latter for the Platte River Valley, with the help of additional 7.5-min maps of perennial stream, spring, and lake-elevation data. The resulting 1 × 2 degree (1:250,000) digitized contour maps (altogether 11) were mosaicked and transformed into a 1-km resolution grid by the ArcGIS (ESRI, Redlands, California). The groundwater-elevation data of 1995 were corrected by the change in groundwater-elevation values between the springs of 2000 and 2010 (Korus et al. 2010) by first interpolating the groundwater-elevation change values obtained at the well locations of Figure 1 onto a 1-km grid with a spline method. Note that here it is implicitly assumed that the 2000 to 2010 changes were the same as the 1995 to 2005 ones. Unfortunately, no change values are available for this latter interval. The corrected groundwater-elevation values were finally subtracted from the 1-km DEM to obtain a depth to groundwater map (Figure 3). Cells with irrigation (50% of the total number of cells within the Platte River Valley) were excluded from the present analysis (Figure 4), as irrigation distorts the sought-for $Rn(d)$ relationship through artificially inflated ET rates. Irrigation information came from the 2005 statewide land use maps (CALMIT–NDNR 2006; Dappen et al. 2007).

CREMAP-ET rates in the Platte River Valley were verified by flux-tower measurements at two different locations, employing eddy-covariance systems (Landon et al. 2009). The estimated mean annual ET rates were within 3 and 6% of the flux-tower-derived values (Szilagyi et al. 2011a). CREMAP requires the following information: air temperature and humidity, incident global radiation (in lieu of it, sunshine duration or possible percent sunshine), and daytime surface temperatures from MODIS. A detailed description of the CREMAP method is found in Szilagyi et al. (2011a) and Szilagyi (2013).

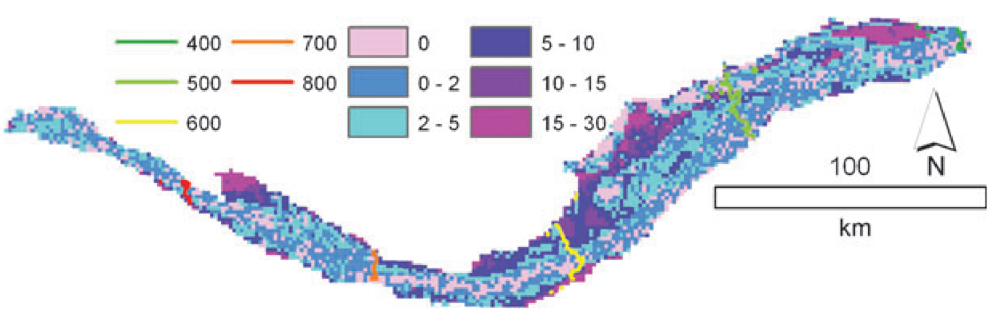

Figure 3. Depth to the ground water (m) and elevation contours (m) of the groundwater table within the Platte River Valley of Nebraska.

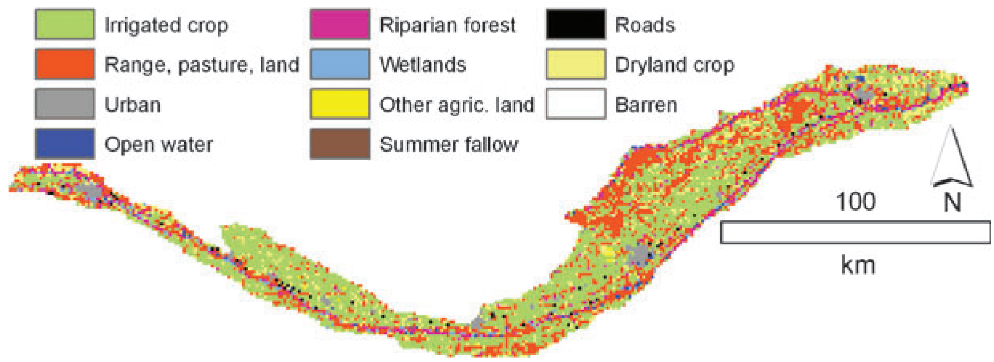

Figure 4. Land use, land cover distribution at 1-km resolution within the Platte River Valley in 2005 (after Dappen et al. 2007).

Figure 5a and 5c plots the resulting ET and $P-$ ET values against the depth to groundwater for each MODIS cell where irrigation is absent. The correlation of the ET and $P-$ ET values against groundwater depth were repeated (panels "b" and "d") with groundwater-elevation values obtained by universal kriging of ArcGIS using a subset (seen in Figure 1) of the original well locations employed for the 1995 groundwater-elevation map. The resulting depth values are inferior to the original groundwater-depth values because of the (1) smaller number of available wells, and (2) lack of additional high-resolution stream- and lake-elevation data that were utilized for the derivation of the 1995 groundwater-elevation contour lines. As the 1995 groundwater-elevation map does not contain information about the accuracy of the contouring (i.e., about the error in the interpolated values), this accuracy can be contained by the 3.7-m value of the mean prediction error of kriging. The true accuracy of the 1995 groundwater-elevation map is certainly smaller than 3.7 m. As a 3-m contouring was possible in the Platte River Valley, the error is probably in the order of 1 m or less. The shapes of the resulting polynomials fitted to the krigged values are similar to the ones obtained from the original hand-interpolated depth values (panels "a" and "c"). The curves level off (i.e., ET and net recharge become constant) at a depth less than 1 ($\pm$1) m and again at a depth greater than 7 to 8 ($\pm$1) m, which can be best seen in panels "a" and "c." This is so because the root system of the vegetation can easily access the shallow groundwater when it is within 1 m of the ground (the central part of Figure 2), and conversely, when the groundwater is below a certain depth, the roots cannot tap it any longer so they must rely solely on the moisture retained in the vadose zone (the laterals of Figure 2).

The shape of the fitted polynomial curve, utilizing the more accurate groundwater-elevation data of Figure 5c, is similar to the numerical model result of Maxwell and Kollet (2008) that employed a fully integrated, coupled watershed and land-surface model over the Little Washita catchment in the southern Great Plains of Oklahoma and to the modeling result of Soylu et al. (2011) employing a sandy soil under a climate characteristic of southern Nebraska. Saturation (leveling out) of the net recharge curve in Figure 5c takes place above approximately 1 ($\pm$1) m (it is $\sim$2 m for Maxwell and Kollet [2008] and $\sim$0.3 m for Soylu et al. [2011]) and below approximately 7 to 8 ($\pm$1) m, almost exactly as that of Maxwell and Kollet (2008) while Soylu et al. (2011) have a depth of 12.5 m. As seen in Figure 5c, the net groundwater recharge can be positive or negative, depending on the depth to the groundwater. The depth where this transition occurs (d_t) is at approximately 2 ($\pm$1) m. As the mean annual precipitation changes by more than 30% going west-to-east in the Platte River Valley, it may be necessary to normalize the net recharge rates by the corresponding precipitation, as the net recharge may depend on the actual precipitation rate. As a result, when the groundwater is within 1 ($\pm$1) m of the surface, the net (negative) recharge stabilizes at -4 ($\pm$12)% of P (Figure 6), and similarly, when the groundwater is below approximately 7 to 8 ($\pm$1) m, the net (positive) recharge again levels off at about 13 ($\pm$10)% of precipitation (the value after the plus/minus sign corresponds to the mean length of the whiskers of Figure 6 within the respective depth range). Szilagyi et al. (2011b) obtained a regionally representative net recharge value of 14 ($\pm$14)% of P for the Sand Hills of Nebraska (Figure 1) where the mean depth to the groundwater is larger than 10 m.

Conclusions

For shallow groundwater regions of negligible surface runoff, the net recharge vs. depth to groundwater relationship, $Rn(d)$, can be obtained with the help of spatially distributed mean annual ET and measured precipitation rates, provided sufficient depth-to-groundwater data exist. The present study employed a revised version of the original CREMAP method (Szilagyi et al. 2011a), which is a calibration-free ET estimation approach. The resulting transition depth of 2 ($\pm$1) m between negative and positive net recharge areas fits well within the typical maximum extinction depth value of 5 m, which groundwater modelers consider as an upper limit (Shah et al. 2007; Luo et al. 2009; Ajami et al. 2012) for groundwater ET. The longest roots of the vegetation may reach the groundwater (or its capillary zone, which for sandy soils is thin) at a depth between 2 to 5 m, producing groundwater ET with the total ET remaining below the precipitation rate.

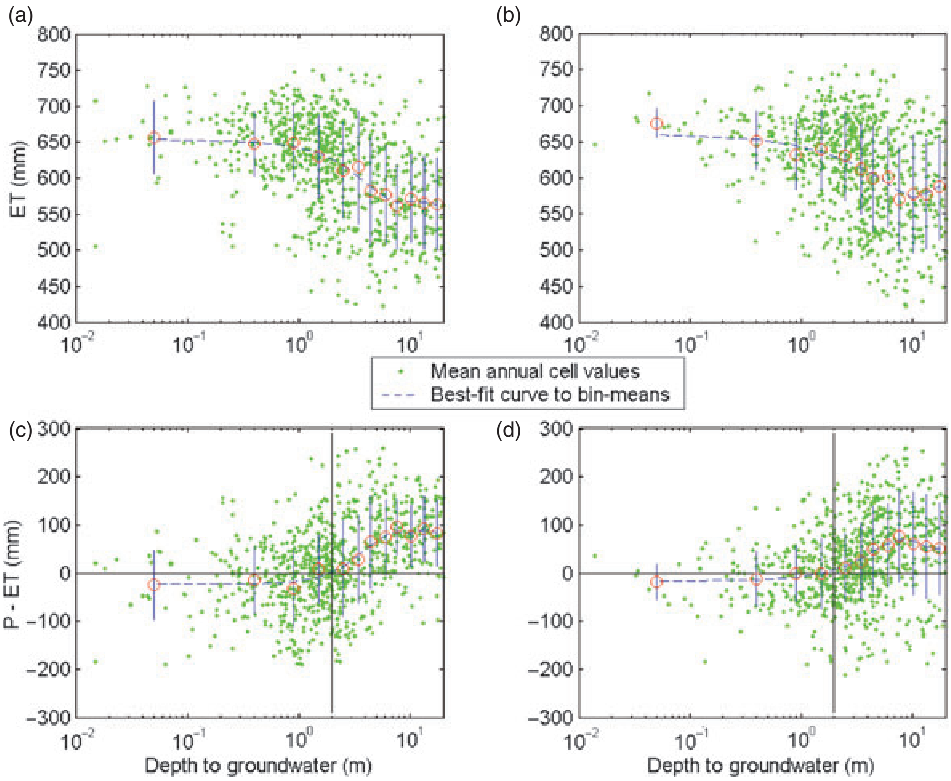

Figure 5. Mean annual ET (top panels) and $P-$ ET rates (bottom panels) vs. depth to groundwater (a, c: hand-contoured depth values from a full set of well locations; b, d: kriging-interpolated depth values from a reduced number of well locations) for the Platte River Valley in Nebraska. The length of each whisker denotes the standard deviation of the binned data around the bin mean (circle). Total numbers of data points displayed are 1169 (a, c) and 1119 (b, d). A large number of cells with irrigation (50% of total) were omitted. See Table 1 for the coefficients of the fifth-order best-fit polynomials to bin means.

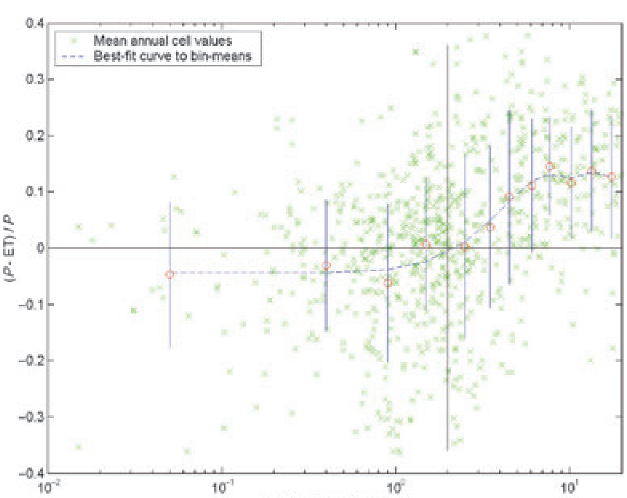

Figure 6. Mean annual ($P-$ ET)/P ratios vs. depth to groundwater (from Figure 5c) for the Platte River Valley in Nebraska. The length of each whisker denotes the standard deviation of the binned data around the bin mean (circle). Total number of data points is 1169. See Table 1 for the coefficients of the fifth-order best-fit polynomial to bin means.

Two saturation depth values (above or below net recharge becomes independent of the depth to the groundwater) were obtained, 1 ($\pm$1) m and 7 to 8 ($\pm$1) m, the latter saturation depth being almost identical to the modeled value of Maxwell and Kollet (2008) but is considerably smaller than that of Soylu et al. (2011). The resulting $Rn(d)$ values contain a significant uncertainty of about 10% to 15% of the corresponding mean annual precipitation value, reflecting the accumulated uncertainties in the ET, P, and d values (Szilagyi et al. 2011b).

In summary, the proposed method provides a simple, data-driven and calibration-free approach for the estimation of the net groundwater recharge vs. depth function for modeling of shallow groundwater systems of regional extent where runoff is negligible. It requires the following information: CREMAP-derived ET, precipitation, and depth to the groundwater. Note that the fitted $Rn(d)$ curves are not directly transferable to arbitrary locations of negligible runoff, even under the same soil and vegetation conditions, because, for example, with increasing humidity of the environment, the $P-$ ET term may always be positive and therefore the resulting curve may completely lack its negative-value segment. At the same time, one could expect that the saturation depths (i.e., 1 [$\pm$1] m and 7 to 8 [$\pm$1] m) would not change with climate, provided vegetation and soil types are the same, as the model results of Maxwell and Kollet (2008) suggest. The $Rn(d)$ curves, obtained with the help of MODIS data, may become useful in modeling of shallow groundwater systems for

Table 1
Polynomial Coefficient Values of Figures 5 ($\times 10^{-4}$) and 6 ($\times 10^{-7}$) in Decreasing Order of the Exponent

Figures	Fifth	Fourth	Third	Second	First	Constant	RMSE	ΔRMSE
5a	13	−567	8498	−40, 634	−90,079	6,545,225	5.57	4
5b	−8	336	−4733	34, 404	−211,960	6,610,963	7.97	5
5c	−34	1486	−22,340	124, 303	−50,347	−224,876	7.84	5
5d	−10	513	−8779	50, 768	42,068	−161,660	4.52	26
6	−52	2334	−35,600	202, 600	−102,300	−429,700	0.013	4

RMSE is the root-mean-square error (mm for Figure 5) and ΔRMSE is the decrease in the RMSE value (in percentages) when the order of the polynomial is increased by 1.

pre-2000 time periods (or future scenarios) when MODIS data are not available. The $Rn(d)$ curves then can assist with specifying external forcing to these shallow groundwater systems to be modeled.

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Disclaimer

The views, conclusions, and opinions expressed in this study are solely those of the writers and not the University of Nebraska, state of Nebraska, or any political subdivision thereof.

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