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Measuring and Modeling Shallow Groundwater Flow between a Semi-Karst Border Stream and Ozark Forested Riparian Zone in the Central USA

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Authors' contributions

This work was carried out in collaboration between all authors. Author JAH designed and administered the study, wrote the protocol and advised PC. Author PC developed the first draft of the manuscript. PC and JAHmanaged the literature searches, data analyses, and author PC performed the spectroscopy analysis. Authors JAH and PC managed the experimental process. All authors read and approved the final manuscript.

Original Research Article

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ABSTRACT

Aims: Quantitative information is limited pertaining to riparian forest and stream shallow groundwater interactions particularly in karst hydro-ecosystems.
Study Design, Place and Duration: Spatiotemporal variability of shallow groundwater flow was monitored along two stream reaches in a riparian Ozark border forest of central Missouri, United States. Each reach was equipped with twelve piezometers and two stream-gauging stations during the 2011 water year (WY).
Methodology: High-resolution (i.e. 15 minute) time-series data were analyzed indicating average groundwater flow per unit stream length was -3 x 10^{-5} m^3 s^{-1} m^{-1} (losing stream) for the entire study reach (total reach length = 830m) during the 2011 WY. The HYDRUS – 1D groundwater flow model was forced with observed data and outputs were assessed to improve model end user confidence in karst hydrogeologic systems.
Results and Discussion: Results indicate rapid groundwater response to rainfall events

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within two to 24 hours nine meters from the stream. Analyses indicated average stream flow loss of 28% and 7% total volume to groundwater during winter and spring seasons, respectively. During the dry season (June-September), the stream was gaining 95% of the time. During the wet season (March-June), the stream was losing 70% of the time. Based on established assessment criteria, shallow groundwater modeling performance with HYDRUS – 1D was deemed *very good* (NS = 0.95, r^2 = 0.99, RMSE = 2.38 cm and MD =1.3 cm).

Conclusion: Results supply greatly needed baseline information necessary for improved understanding of riparian forest management and shallow groundwater transport and storage processes in semi-karst forest ecosystems.

Keywords: Shallow groundwater flow; karst; Ozark border forest; Central USA; stream flow; hydraulic head; riparian; HYDRUS – 1D.

1. INTRODUCTION

Quantifying shallow groundwater flow regimes (quantity and timing) is important for effective riparian ecosystem management [1], but is often ignored due to lack of available data. Dahm et al. [2] identified that improved understanding of spatiotemporal variations in stream-groundwater exchange processes require investigation in varying geological settings, particularly in underrepresented karst geology. Shallow groundwater flow can be determined using various methods including Darcy groundwater flow calculations and tracer tests [3]. For example, Mulholland et al. [4] used seepage meters to show that groundwater flow towards the stream was 2.2 x $10^{-4}m^3$ s^{-1} at Walker Branch Creek in North Carolina, USA. While there are many methods to estimate shallow groundwater flow and quantity, a great deal of research is needed to improve confidence in investigative methods with respect to varying environmental conditions. Sophocleous [1] emphasized the need for a comprehensive interdisciplinary framework to better understand groundwater exchange in relation to land use, geology and biotic factors. Groundwater flow regimes in low order forested streams in particular are in need of investigation as the current understanding is greatly limited.

Many previous researchers used a Darcian approach to quantify saturated groundwater flow and compliment groundwater numerical modeling [5]. However, contemporary groundwater models, such as HYDRUS – 1D, use variably saturated hydraulic conductivity values [6] and can thus simulate both saturated and unsaturated groundwater flow [7,8,9]. Luo and Sophocleous [10] used HYDRUS – 1D to estimate groundwater flow values ranging from -3.5 x 10^{-8} to 3.5 x 10^{-8} m^3 s^{-1} with a coefficient of determination (r^2) value of 0.75 between simulated and measured groundwater flow in an agricultural field located in Shandong province, China. A number of other numerical groundwater models (e.g CPFLOW, MODFLOW, SUTRA, HYDRUS and FEFLOW) [11] have been shown to successfully predict groundwater flow in varying hydrogeologic settings. Despite technological progress in modeling, previous authors indicated that improved model accuracy requires proper parameterization, emphasizing the need for higher resolution (spatial and temporal) field observations [12,13]. Shallow groundwater flow studies that integrate in-situ field and modeling methodologies are necessary to improve quantitative understanding and subsequent management of groundwater resources [1,14,15].

The majority of previous surface water groundwater interaction studies in forested ecosystems were conducted in the North-Western United States [16,17,18] or outside the United States [14,19,20]. Castro and Hornberger [17] used tracers to quantify surface-subsurface water interactions in North Fork Dry Run, Shenandoah National Park, Virginia. The authors showed that physically based models need to account for interactions between the stream and the larger riparian zone (RZ) by including water table variations for numerical modeling approaches. Burt et al. [20,21] replicated experimental designs in France, the Netherlands, Poland, Romania, Spain, Switzerland and the United Kingdom by constructing piezometer grids to map water table levels in the riparian zone, including riparian woodland and upland areas. Their study results emphasized the hydrologic change in receiving water bodies due to water table fluctuations in the adjacent RZ. The observed variations in riparian zone hydraulic gradients and water table level and flow patterns were attributed to surface water –groundwater interactions associated with runoff (surface and subsurface) from surrounding hills. Their results showed a net increase in ground water level 40 m away from the stream, at the French (with an average change in groundwater table within 50 cm), UK (150 cm), Romanian (0.6 cm), Spanish (200 cm), Dutch (300 cm) and Polish (150 cm) sites during 2009, indicating greater upslope contributions to riparian zone groundwater levels, relative to localized surface water –groundwater interactions. Despite progress of previous studies in other regions, there exists a great need for shallow groundwater research in karst hydrogeological systems such as those in the Central United States where there are marked differences in riparian forest species, hydrogeology and climate.

Objectives of the following work were to a) quantify spatial and temporal variability of shallow groundwater and stream water exchange in a karst ecosystem of the central US over the period of one water year, b) validate the use of the groundwater flow model HYDRUS – 1D in a karst hydrogeologic setting, c) given results of the first two objectives, advance shallow groundwater and stream water process understanding and HYDRUS – 1D predictive confidence in hydrologically distinct Central USA riparian forests and other similar karst hydro-geological ecosystems.

2. STUDY SITE

The study took place on two reaches of Brushy Creek located within the Thomas S. Baskett Wildlife Research and Education Center (BWREC) (Fig. 1). The BWREC is located at UTM15 coordinates 569517 E and 4289338 N, 8 km east of Ashland,in the Ozark border region of South-Central Missouri, USA (48). Brushy Creek is a second order stream [22] with average slope of 0.94%, joining Cedar creek 4 km south of the BWREC, after draining a watershed of approximately 9.17 km^2. Current land use ranges from second growth forests to pastures. The watershed consists of 2.6% suburban land use, 17.9% cropland, 33% grassland, 43.2% forest, and 3.3% open water and wetlands [23].

Limestone of Ordovician and Mississippian age underlies the BWREC. Dominant soils are Weller silt loam and Clinkenbeard clay loam [24]. Streambed sediments, primarily composed of coarse gravel, cobble, and cherty fossilized materials, are on average, less than one-meter deep, overlying bedrock and layered limestone [25]. Soil within the riparian zone (RZ) consists of a mix of Cedargap and Dameron soil complexes (USDA soil map unit 66017). BWREC soils have average bulk density of 1.2 to 1.4 g cm^{-3}. Soils are well-drained and are frequently flooded soils of alluvial parent material. Vegetation consists of northern and southern division oak-hickory forest species including American Sycamore (*Platanusoccident*alisi), American Elm (*Ulmusamericana*) and Black Maple (*Acer nigrum*) dominated riparian reaches [26]. Understory vegetation is dominated by sugar maple (*Acer saccharum*), flowering dogwood (*Cornusflorida*), and black cherry (*Prunusserotina*) [27,28].

Climate in the BWREC is classified as humid - continental [29]. Mean January and July temperatures were -2.2 ℃ and 25.4 ℃ (1971-2010), respectively, while mean annual precipitation is 1037 mm, as recorded at the Columbia Regional Airport located 8km to the north of the BWREC [26].

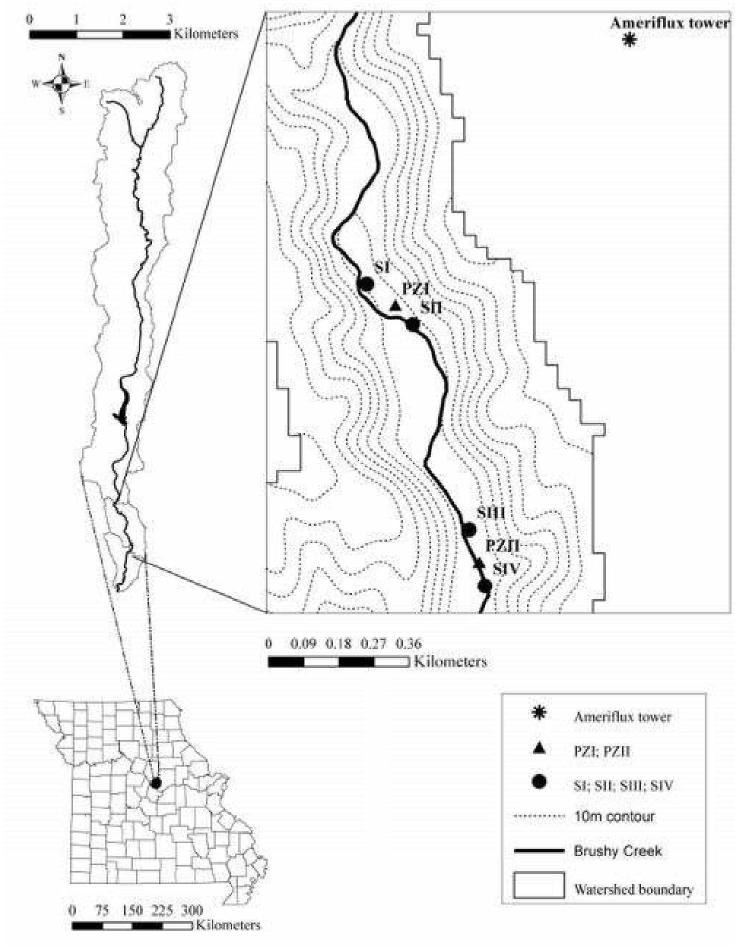

Fig. 1. Study sites and instrument locations at Baskett Wildlife Research and Education Center, Central Missouri, USA. S = stilling well sites. PZ = piezometer sites

3. METHODOLOGY

3.1 Instrumentation and Data Collection

Climate data were obtained from an AmeriFlux tower [30] installed at an elevation of 238 m (Fig. 1). Stream stage monitoring sites (hereafter referred to as SI – SIV, n=4, Fig. 2) were installed before and after each piezometer array (Fig. 1). The distance SI-SIV was 830m, while distances SI-SII, SII-SIII, SI-SIII and SIII-SIV were 160, 543, 682 and 149m respectively. Stream stage stilling wells were equipped with Solinst® Levelogger Gold pressure transducers (error ±0.003 m) and programmed to record stream stage at 15-minute intervals. To obtain high spatial resolution information, shallow groundwater levels were monitored using piezometers installed in the RZ up to 9 m perpendicular from the stream bank (Figs. 1 and 2). Between site one (SI) and site two (SII), four piezometers (Pz1, Pz2, Pz3 and Pz4) were installed in a transect (Piezometer Site I, hereafter referred to as PZI) extending from 3 m from the stream edge to 9 m into the RZ (Fig. 2). There was negligible slope and no geological differences between sites. Piezometer Site II (PZII) was located 660 m S-SE of PZI with four piezometers (Pz5, Pz6, Pz7 and Pz8). Each 3.58 m long drive-point piezometer, with 4 cm inner diameter and 76 cm slotted screen at the end, was equipped with Solinst® Levelogger Gold programmed to log water depth at 15 minute intervals. All piezometers were screened across the same elevation interval.

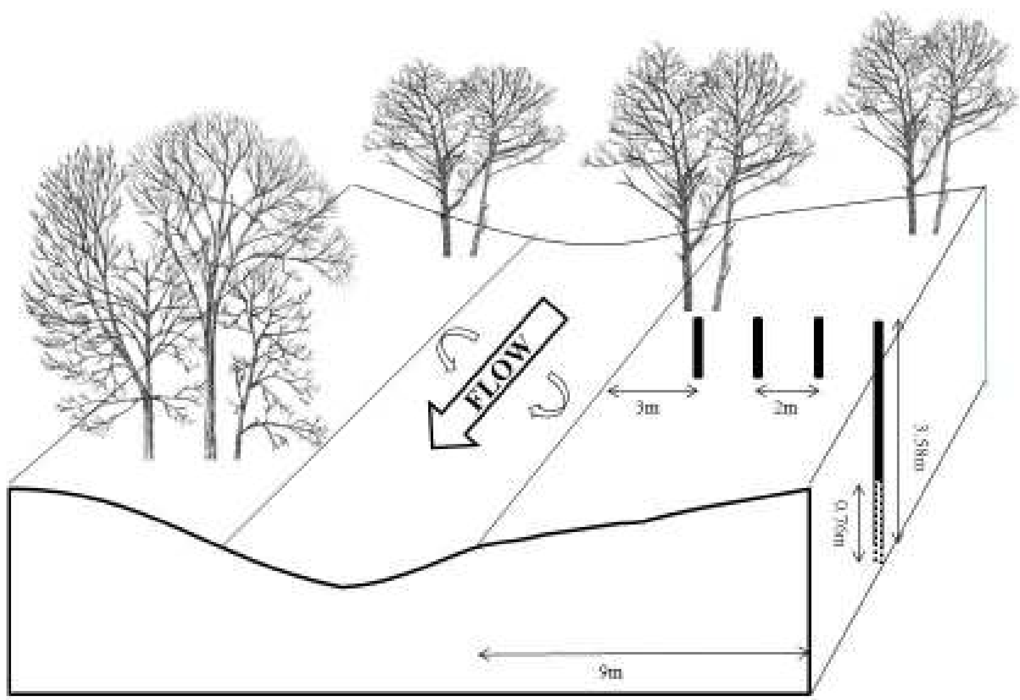

Fig. 2. Conceptual diagram of cross-section of piezometer study design at Baskett Wildlife Research and Education Center, Central Missouri, USA. Elevation of each well was measured independently and head measurements were normalized to elevation common to both piezometer sites PZI and PZII. Average depth to bedrock was approximately 3.65 m

3.2 Quantifying stream flow

Streamflow rating curves for each stage monitoring site were developed using measured stage-discharge relationships established by the stream cross section method [31] using a Marsh-McBirney® Flo-Mate flow meter (sensor error ± 2%). Stream cross section flow measurement campaigns were performed by the same personnel for various flow depths to minimize computational errors (66). Rating curves were calculated as per Dottori et al. [31]:

$$Q = a \times Z^b \qquad [1]$$

where Q is discharge in units of volume per unit time, Z is measured stream stage in units of length, and a and b are coefficients determined by stream morphology.

3.3 Quantifying Shallow Groundwater Flow

Shallow groundwater flow was calculated using Darcy's Law [5]:

$$Q_s = K_s \times \nabla h \times A \qquad [2]$$

where Q_s is shallow groundwater flow ($m^3\ s^{-1}$), K_s is hydraulic conductivity ($m\ s^{-1}$), ∇h is the hydraulic gradient ($m\ m^{-1}$), where $\nabla h = \Delta h/\Delta l$ where Δh = change in head between piezometers (m), Δl is the flowpath length between piezometers (m) and A is the cross section area (m^2). Saturated hydraulic conductivity (K_s), estimated using the piezometer method (standard slug test) [32], was $3 \times 10^{-5}\ m\ s^{-1}$ at PZI and $1 \times 10^{-5}\ m\ s^{-1}$ at PZII. Estimated K_s values corresponded to silty sand deposits [33] and agreed with results from BWREC provided by Rochow [24]. The shallow groundwater cross section area (A) was computed as product of the thickness of the saturated aquifer (i.e. average depth in piezometers) and 2 m (distance between piezometers perpendicular to flow). Since the depth to the bedrock was approximately three meters, the shallow groundwater zone was assumed primarily of alluvial composition and homogeneous (see Study Site).

Darcy velocity (*v*) for the shallow groundwater flow was calculated as per Darcy [5] and as used in Sophocleous [1], Ocampo et al. [9] and Wondzelland Swanzon [34]:

$$v = \frac{Q_s}{A} \qquad [3]$$

Darcy velocities along the piezometer transect were 4.7×10^{-7} and $1.1 \times 10^{-8}\ m\ s^{-1}$at PZI and II, respectively. Average linear velocity of shallow groundwater flow was estimated as per Freeze and Cherry [33] and as used in Levia et al. [35] and Jones and Mulholland [3]:

$$\vec{v} = \frac{Q_s}{nA} \qquad [4]$$

where $\vec{v}$ is the average linear velocity and *n* is the effective porosity. Based on porosity data summarized by Davis [36] for various geologic materials, silty sand was assumed to have an effective porosity of 0.35 to 0.50.

3.4 Quantifying Groundwater Flow per Unit Stream Length

Assuming equivalent precipitation and evapotranspiration processes along the study reaches, the groundwater flow per unit stream length was estimated using the mass balance approach:

$$\frac{dQ_f}{dx} = Q_h \qquad [5]$$

Where $Q_h$is the net groundwater flow per unit stream length (m^3 s^{-1} m^{-1}), dQ_f is the difference in stream flow (m^3 s^{-1}) measured at the upstream and downstream sampling locations of the piezometer transect and *dx* is the distance (m) between stilling wells [37,38].

3.5 Numerical Simulation with HYDRUS – 1D Model

HYDRUS – 1D characterizes infiltration, evaporation, transpiration, percolation, water flow, solute flow and heat flow through variably-saturated (vadose and saturated zone) porous soil media [7,8,28]. HYDRUS 1D models both saturated and unsaturated flow and has been tested in various media [8,39]. The model simulates lateral movement of groundwater within defined boundary conditions and thus requires less computing power relative to 2D or 3D simulations [39]. In the current work, the conceptual model representing the RZ was calibrated and validated with HYDRUS – 1D using measured groundwater head data from the piezometers located in PZI and II as per the methods of Dages et al. [39].

3.5.1 HYDRUS – 1D computations

In HYDRUS – 1D, horizontal groundwater flow is quantified using Richards's equation [6]:

$$\frac{\partial\theta}{\partial t} = \frac{\partial}{\partial x}\left\{K(hp)\left(\frac{\partial h}{\partial x} - cos \propto\right)\right\} - S(x,t) \qquad (6)$$

where θ is volumetric soil water content (m^3 m^{-3}), *t* is time (s), *x* is the horizontal space coordinate (m) (for lateral flow), *hp* is pressure head (m), *S* is the water sink term (m^3 m^{-3} s^{-1}), α is the angle between the flow direction and the vertical axis (i.e. $\alpha = 0°$ for vertical flow, 90° for horizontal lateral flow, and $0° < \alpha < 90°$ for inclined flow) and K is unsaturated soil hydraulic conductivity (m s^{-1}) given by:

$$K(h,x) = K_s(x)\ K_r(h,x) \qquad [7]$$

where$K_r$is relative hydraulic conductivity (unitless) and K_s is saturated hydraulic conductivity (m s^{-1}). The soil hydraulic properties used in the model for the current study, θ_r – residual soil water content (m^3 m^{-3}), θ_s – saturated soil water content (m^3 m^{-3}), α –soil water retention function (m^{-1}), *n* - parameter *n* in the soil water retention function (unitless), K_s - (m s^{-1}), *l* - tortuosity (unitless), are described using a set of closed form equations developed from van Genuchten–Mualem functional relationships [40]. HYDRUS - 1D uses a Marquardt Levenberg type soil parameter estimation technique for inverse estimation of soil hydraulic parameters from measured hydraulic head data (h). A detailed description of parameter optimization and statistics of the inverse solution is provided in Šimůnek et al. [7].

3.5.2 HYDRUS – 1D conceptual model development

Given semi-karst geology and high soil permeability of the study sites, most annual rainfall (within the piezometer transect) was assumed to infiltrate to shallow groundwater. The shallow groundwater flow induced due to the presence of piezometric head differences (lateral flow in the subsurface) was assumed dominant compared to flow influenced by other mechanisms (e.g. infiltration, evapotranspiration, solute and heat flow). Change in evapotranspiration and infiltration was assumed negligible, within the piezometer transect. Observed RZ groundwater head values served as initial conditions for the simulation. Due to the availability of high frequency data, left and right boundary conditions were set as a variable pressure head type in HYDRUS – 1D, with daily time intervals. Observed hydraulic head values from the piezometers closest (Pz1 at site PZI and Pz5 at site PZII) and furthest (Pz4 at site PZI and Pz8 at site PZII) from the stream served as time dependent boundary values for the finite grid element created in HYDRUS - 1D (Fig. 3). Notably, multiple boundary layer scenarios were implemented all with excellent calibration results, presumably attributable to the close proximity of piezometers (i.e. 2 m). The observed RZ groundwater head from the remaining two piezometers at each site (Pz2, Pz3 and Pz6, Pz7 at site PZI and PZII respectively) was used for model calibration. The governing flow equation (Equation 6) was solved numerically using a standard Galerkin-type linear finite element scheme [7].

Initial soil hydraulic parameters were estimated using pedotransfer functions (PTFs), by supplying textural class and two groundwater head values as input data to ROSETTA, a built-in computer program in HYDRUS – 1D [41]. For initial estimation of soil properties, soil texture classes were identified as silt loam based on the results of previous work in the BWREC [42,43]. A 6 m horizontal soil cross-section was defined for each piezometer site. Four nodes, fixed along the soil cross section, represented each piezometer location. The initial soil water content (θ_s) was set to a uniform value of 0.43 m^3 m^{-3} (using PTFs for silty loam). Groundwater head values were simulated at each node, and compared to measured groundwater head values to obtain soil hydraulic parameters using the inverse solution method [10,39,44]. The final set of soil hydraulic parameters with the best coefficient of determination (r^2) relationship between observed and modeled hydraulic head values was used to model groundwater flow.

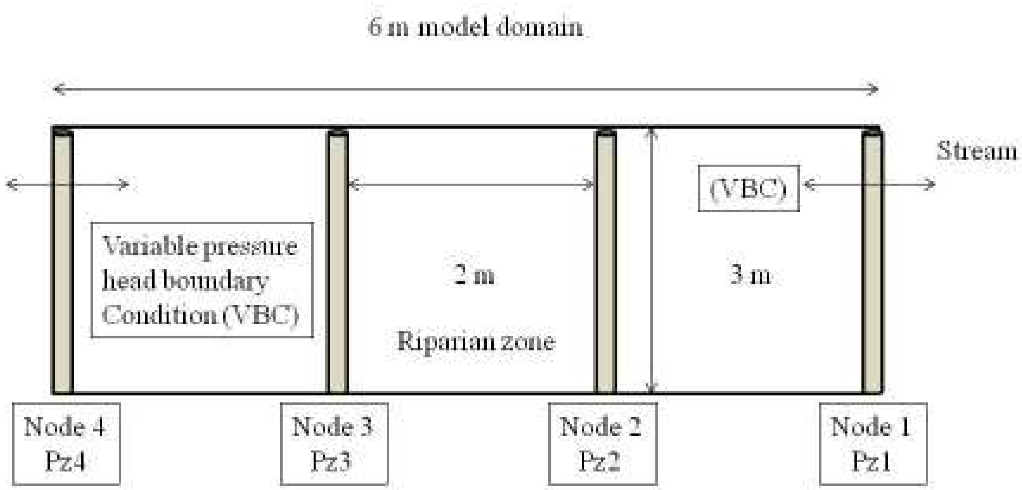

Fig. 3. HYDRUS 1D model discretization for the model area. The model domain contains four nodes of which nodes 1 and 4 are variable head boundary conditions

3.5.3 HYDRUS – 1D calibration, validation and statistical analysis

Daily averaged pressure boundary conditions were applied to the model. HYDRUS – 1Dwas calibrated for a three-month period (April 2010 to June 2010). Soil hydraulic parameters, were adjusted while calibrating the model. All the soil hydraulic parameters were consistent with those values published in the primary literature. Parameter range values were used from soil maps [23], published literature values, and correlated with measured Ks values (from pump tests) to obtain site specific soil hydraulic parameters. Final soil hydraulic parameters obtained from calibration were then used for validating the model for a three-month period (July 2010 to September 2010). To quantify model bias, simulated and observed hydraulic head values were evaluated using the Nash-Sutcliffe Efficiency parameter (NS) [45], the Root Mean Square Error (RMSE) [46], and the standard Mean Difference (MD) and regression methods. Model outputs were rated 'Very Good', 'Good', 'Satisfactory', or 'Unsatisfactory' according to the criteria recommended by Moriasi et al. [47]. The Nash-Sutcliffe (NS) efficiency parameter was used to evaluate how well HYDRUS – 1D predicted observed hydraulic head variability relative to the average observed value for the selected time period (Equation 8). The NS parameter value ranges from < 0.0 to 1.0 where 1.0 indicates the model is in perfect agreement with observed data and < 0.0 when there is a poor agreement [47]. RMSE values closer to zero indicate better model performance. Assuming that observed and simulated values are linearly related, the equation of the best-fit regression line (coefficient of determination) can indicate how well modeled values agree with observed values. Further information regarding the indices NS, RMSE, MD and standard regression is presented in Moriasi et al. [47]. The equations to calculate the aforementioned statistics are as follows:

$$NS = \frac{voN - \sum_{i=1}^{N}(x_i - yi)^2}{voN} = 1 - \frac{\sum_{i=1}^{N}(x_i - yi)^2}{\sum_{i=1}^{N}(x_i - \bar{x})^2} \qquad [8]$$

$$RMSE = \sqrt{\frac{\sum_{i=1}^{N}(x_i - yi)^2}{N}} \qquad [9]$$

$$MD = \frac{1}{N}\sum_{i=1}^{N}(x_i - yi)^2 \qquad [10]$$

where*vo*is the variance of observed values, N is the number of data points, *xi* is the observed value, *yi* is the corresponding predicted value, and $\bar{x}$ is the average observed value for the study period. The authors included NS, MD and other metrics to assess not only the PTF parameter estimations but also to assess if HYDRUS 1D can predict observed head distribution in a semi-karst hydrogeologic setting, as most previous studies only incorporated HYDRUS in alluvial floodplains.

Modeling was conducted using observed shallow groundwater data and did not include the stream water head since the hydraulic properties differ between the riparian zone and stream, and HYDRUS 1D cannot currently simulate those differences. It was assumed that shallow groundwater flow exfiltrating the piezometer near the stream exited to the stream. A common datum for all the piezometers was calculated and the model was executed to estimate the net hydraulic gradient and mass flow across the model domain. Given the shallow depth to bedrock (3 m), groundwater flow was assumed primarily horizontal due to higher hydraulic conductivity in the soil relative to bedrock. Therefore, a Dupuit flow condition (i.e. primarily horizontal flow) was assumed as the dominant flowpath which eventually converged into the stream.

4. RESULTS AND DISCUSSION

4.1 Hydroclimate During Study

Climate at the BWREC during WY2011 (October 2010 to September 2011) was characteristically variable with mean air temperature of 12.5 ˚C and total precipitation of 647 mm. It was on average cooler and drier during the study relative to average temperature and precipitation (13˚C and 930 mm, respectively) recorded at the Ameriflux tower from 2005 to 2011. Seasonal precipitation during WY 2011 (winter, spring, summer, fall) was 170 mm (December - March), 250 mm (March – June), 135 mm (June – September) and 94 mm (September to December). During the current study, stream flow was ephemeral, exhibiting characteristic high flows in spring and summer, and no flow by mid-October. Carter and Anderson [48] estimated that flow meter velocity observation error, measured at 45-second intervals, at 0.2, 0.6 and 0.8 depths was less than 2.3 %. Thus, in the current work, error associated with streamflow measurements and stream stage errors was estimated to be $\pm 1.05 \times 10^{-4}$ m^3 s^{-1} and therefore assumed negligible. Annual average stream flow at SII was 0.06 m^3 s^{-1} greater than streamflow at SI, indicating that the stream reach (between SI and SII) was, on average, a gaining stream (Fig. 4, Table 1). Average stream flow at SIV was more than twice that of SIII indicating that the stream reach (between SIII and SIV) was also gaining. There was negligible precipitation (27 mm) and thus surface flow from October 15 to November 23 2010,during which time stream-shallow groundwater flow could not be quantified using Equation 5 (Figs. 5 and 6). Average daily stream flow during the study period was 0.22 m^3 s^{-1} at SII followed by SI (0.16 m^3 s^{-1}), SIV (0.13 m^3 s^{-1}) and SIII (0.04 m^3 s^{-1}), indicating that the stream was intermittently gaining and losing along the entire reach.

4.2 Groundwater Flow per Unit Stream Length

Applying methods and calculations (as presented earlier in text) of high-resolution (15 min) stream stage and groundwater level data showed that average annual groundwater flow per unit flow length was -3×10^{-5} m^3 s^{-1} m^{-1} (thus a losing stream) for the entire study reach (SI to SIV, total reach length = 830.8 m), and was 4.2×10^{-4} and 5.8×10^{-4} m^3 s^{-1} m^{-1} for the stream reaches SI to SII and SIII to SIV, respectively (Table 2). Figs. 5 and 6 show groundwater flow relationships between stilling wells SI-SII and SIII-SIV. It is arguable that the semi-karst geology of the BWREC may increase groundwater flow values. Direct karst geological influence of results was beyond the scope of the current work, but supplies impetus for future investigations. In the current work, average groundwater flow towards the stream was 0.07 and 0.09 m^3 s^{-1} from SI to SII and SIII to SIV, respectively. Maximum daily groundwater flow was 0.27 and 0.51 m^3 s^{-1} (thus a gaining stream) while the minimum groundwater flow was -2.07 and -0.001 m^3 s^{-1} (losing stream) at SI-SII and SIII-SIV, respectively. Estimated groundwater flow of -2.07 $m^3\ s^{-1}$ at SI (losing stream) was consistent with observed decrease in depth to groundwater of 75.32 cm (from 105.79 cm on March 15) at PZI. Fig. 4 shows the relationships between depth to groundwater and stream stage, illustrating a high degree of shallow groundwater connectivity between the stream and adjacent RZ. Maximum daily groundwater flow (0.51 $m^3\ s^{-1}$) coincided with minimum depth to groundwater from the surface of the soil (101.93 cm on February 17) at PZII. During WY2011, groundwater flow accounted for approximately 0.07 and 0.09 $m^3\ s^{-1}$ of the mean daily stream discharge of 0.22 $m^3\ s^{-1}$ for SII and 0.13 $m^3\ s^{-1}$ for SIV respectively. However, for the entire length of the study reach, SI to SIV, (830 m), a mean daily discharge of -0.03 $m^3\ s^{-1}$ was lost to the aquifer during WY2011 (Tables 1, 2 and 3).

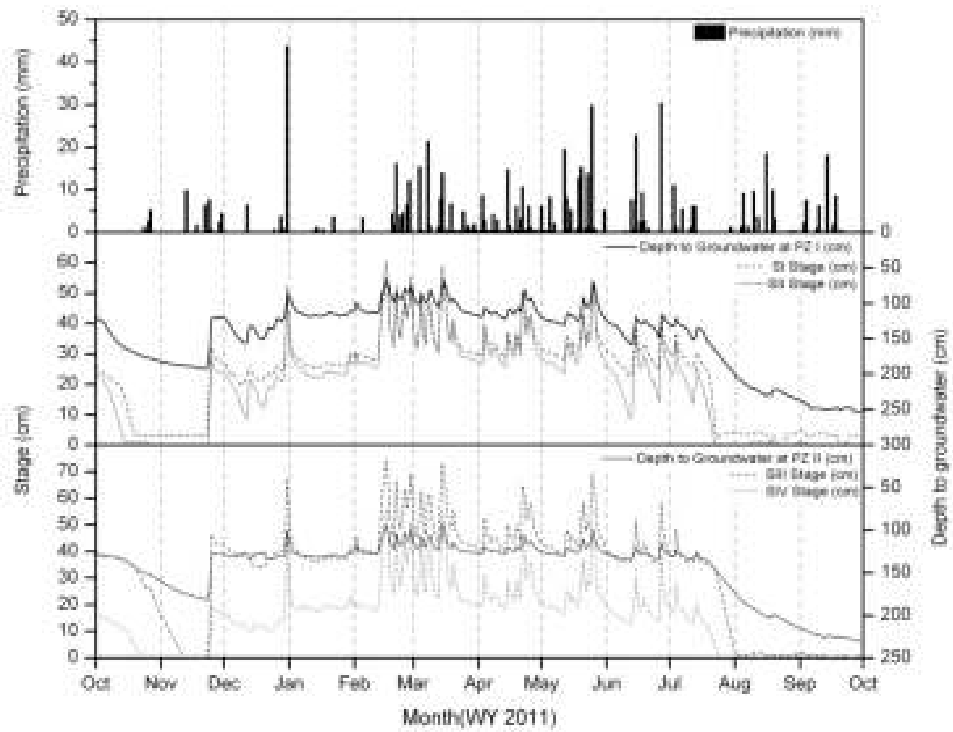

Fig. 4. Measured rainfall (mm), stream stage (cm) and average depth to groundwater (cm) at piezometersites during WY 2010 at Baskett Wildlife Research and Education Center, Central Missouri, USA

Table 1. Stream discharge ($m^3\ s^{-1}$) descriptive statistics for WY 2011 of Brushy Creek flow monitoring sites at Baskett Wildlife Research and Education Center, Central Missouri, USA

Descriptive Statistics	SI ($m^3\ s^{-1}$)	SII ($m^3\ s^{-1}$)	SIII ($m^3\ s^{-1}$)	SIV ($m^3\ s^{-1}$)
Mean	0.16	0.22	0.04	0.13
Std Deviation	0.45	0.29	0.13	0.23
Minimum	0.00	0.00	0.00	0.00
Maximum	3.80	1.78	1.22	1.59

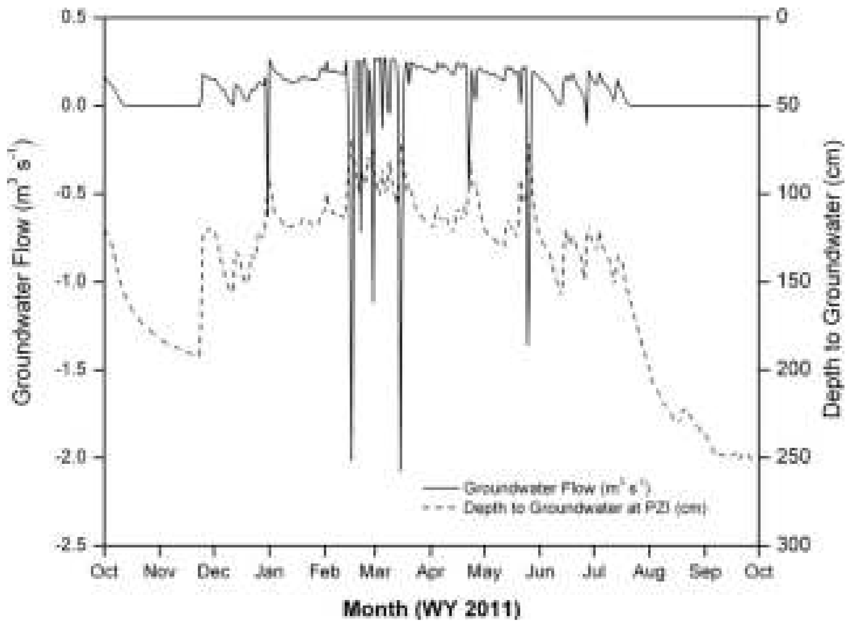

Fig. 5. Average groundwater flow ($m^3\ s^{-1}$) and depth to groundwater (cm) at SI-SII (PZI), with average of four wells, for water year 2011 at Baskett Wildlife Research and Education Center, Central Missouri, USA

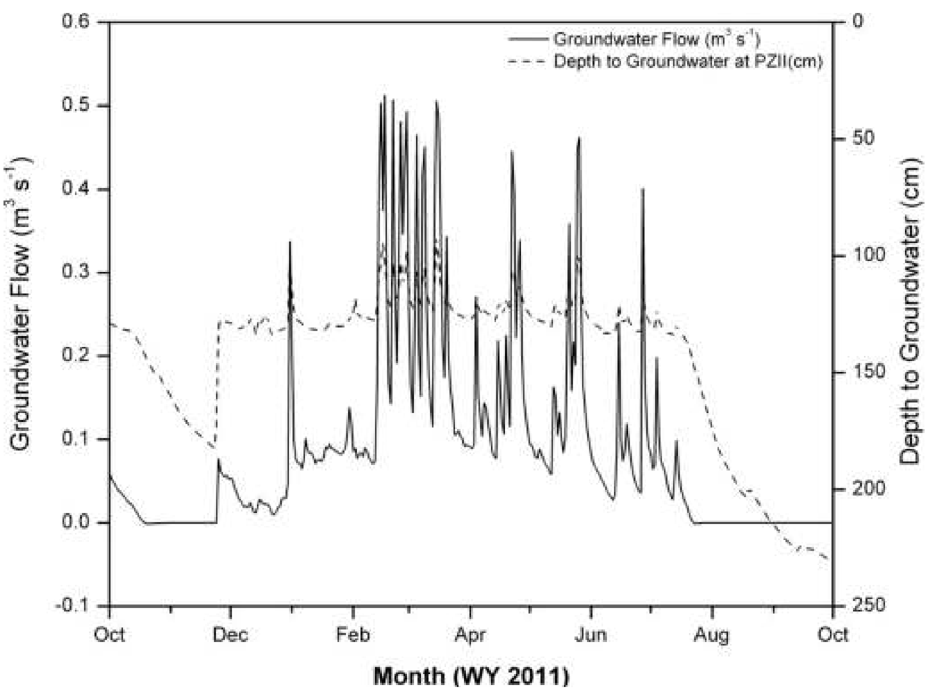

Fig. 6. Average groundwater flow ($m^3\ s^{-1}$) and depth to groundwater (cm) at SIII-SIV (PZII), (n = 4 wells), for water year 2011 at Baskett Wildlife Research and Education Center, Central Missouri, USA

Table 2. Average stream discharge difference (∂Q in $m^3 s^{-1}$) and average groundwater flow per unit stream length (∂Q/∂x in $m^3 s^{-1} m^{-1}$) at four monitoring sites (negative sign indicates a losing flow condition) during water year 2011 at Baskett Wildlife Research and Education Center, Central Missouri, USA

	Groundwater flow between gauging sites ($m^3 s^{-1}$)			Groundwater flow per unit stream length between gauging sites ($m^3 s^{-1} m^{-1}$)		
Site	**SI**	**SII**	**SIII**	**SI**	**SII**	**SIII**
SI	-			-		
SII	0.07	-		4.2×10^{-4}	-	
SIII	-0.12	-0.18	-	-1.7×10^{-4}	-3.4×10^{-4}	-
SIV	-0.03	-0.1	0.09	-3.4×10^{-5}	-1.4×10^{-4}	5.8×10^{-4}

Table 3. Descriptive statistics of groundwater flow ($m^3 s^{-1}$) between three study site locations at Baskett Wildlife Research and Education Center, Central Missouri, USA

Descriptive statistics	Groundwater flow ($m^3 s^{-1}$)		
	Entire study Reach SI -SIV	**Between SI-SII**	**Between SIII-SIV**
Mean	-0.03	0.07	0.09
Standard deviation	0.08	0.27	0.51
Minimum	-2.23	-2.07	0.00
Maximum	0.24	0.23	0.11

Daily average stream discharge at SIV was higher during winter and spring (0.25 and 0.20 $m^3 s^{-1}$) seasons, relative to fall and summer (0.04 and 0.02 $m^3 s^{-1}$). During the brief period when streamflow was negligible, average groundwater flow towards the stream was two orders of magnitude greater at SIII-SIV (5×10^{-4} $m^3 s^{-1}$) relative to SI-SII (7×10^{-6} $m^3 s^{-1}$). During the winter season, SIII-SIV had 0.10 $m^3 s^{-1}$ more water flow from the RZ relative to SI-SII. Ultimately, groundwater input to the stream accounted for approximately 27% of the total stream discharge volume at stream reach one (PZI) and 69% at stream reach two (PZII) during WY 2011. This finding corroborates the results of previous authors who showed that shallow groundwater flow directions near the stream are highly spatially variable and bidirectional with shallow groundwater flowing intermittently towards and away from the stream [34,49]. Marzolf et al. [49] reported average stream flow of 0.003 and 0.002 $m^3 s^{-1}$ during summer and fall seasons (i.e. one-tenth the flow of Brushy Creek, with reach length = 830m) with groundwater flow of 1×10^{-4} and 2×10^{-4} $m^3 s^{-1}$ (i.e. one-hundredth the groundwater flow observed at SIV-SIII, Brushy creek) in Walker Branch Creek in Tennessee (reach length = 62m). The higher flow in Brushy Creek relative to the Walker Branch Creek study is explained in part by larger drainage area and study reach length. The groundwater flow per unit stream length at Walker Branch Creek (1×10^{-5} $m^3 s^{-1} m^{-1}$) was 10% that of Brushy Creek (4×10^{-4} $m^3 s^{-1} m^{-1}$) thus proportionally corroborating similar drainage area-stream-groundwater exchange patterns between the two studies.

4.3 Shallow Groundwater Depth

Average depth to groundwater was 69.70 cm at PZI and 92.32 cm at PZII during spring months (February to June), with an average 32% difference between sites, and 253.41cm at PZI and 231.30 cm at PZII during fall months (September to December), with an average 8% difference between sites (Fig. 4 and 8). During the dry season (October – November) depth to groundwater was 214.9 cm and 197.61 cm at PZI and PZII with an average 8% difference between sites, respectively, and water level in the piezometers dropped below average level (126.62 and 150. 93 cm at PZI and PZII). Unsurprisingly, under negligible stream surface flow conditions, there was decreased variability of groundwater level across the RZ at both PZI (< 1% change) and PZII (< 1% change). This is typical for streams of arid regions or streamflow in dry seasons [50]. Hydraulic gradient analysis descriptive statistics are provided in Table 4.

Implications for these results are important for riparian management. For example, shallow groundwater flow depends on a number of additional hydroclimatic factors including, but not limited to, air temperature, evapotranspiration, soil water status and unsaturated zone depth [51]. It is also worth noting that both exfiltration and infiltration processes have ecological importance, as the amount of water stored in the RZ and how far surface water infiltrates controls transport of key nutrients such as nitrate, phosphorous and potassium [3,19,52]. While quantifying many of these factors was beyond the scope of the current work, the current study supplies very important baseline data that could benefit future studies.

Table 4. Descriptive statistics of hydraulic gradient (cm) for shallow groundwater monitoring sites (PZI and PZII) for water year 2011 at Baskett Wildlife Research and Education Center, Central Missouri, USA

Descriptive statistics	Hydraulicgradient (cm)	
	PZI	PZII
Mean	-10.05	5.65
Standard Deviation	6.27	3.30
Minimum	-32.47	-5.90
Maximum	0.90	10.21

4.4 Modeling with HYDRUS - 1D

4.4.1 Calibration of HYDRUS – 1D

As per calibration outcomes (April to June 2010) the following soil hydraulic parameters were used in HYDRUS-1D 0.065 m^3 m^{-3}, 0.41 m^3 m^{-3}, 0.075 m^{-1}, 1.89, 1. 2 x 10^{-5}m s^{-}1 and 0.5 for θ_r, θ_s, α, n, K_s and l respectively (see Methods) with corresponding r^2 values of 0.98 and 0.90 for PZI and PZII, respectively. Simulated hydraulic head values were compared to observed hydraulic head values to validate (July 2010 to September 2010) the model as per the methods of Dages et al. [39]. Model validation resulted in r^2 values ranging from 0.98 to 0.99 (Table 5).

Table 5. Descriptive statistics of modeled groundwater flow (cm d^{-1}) for sites PZI and PZII for the water year 2011 at Baskett Wildlife Research and Education Center, Central Missouri, USA

Descriptive Statistics	Modeled groundwater flow (cm d^{-1})	
	PZI	PZII
Mean	105.93	106.25
Standard Deviation	0.07	0.06
Minimum	105.74	106.03
Maximum	106.09	106.32

The coarse four grid node discretization was sufficient for the model setup, as noted in the calibration results, to capture the effect of the daily pressure head boundary conditions. This is due to the semi-karst hydraulic conductivity between wells. Given that the model domain remained within six meters between the piezometers, a nodal spacing of two meters was sufficient. However, the model was forced in finer grid spacings (up to 100 grids at 10 by 10 cm each) and minimal (2%) change in the predicted head values was observed. Model results based on the coarser grid spatial scale is encouraging for land managers who wish to use the freely available HYDRUS 1 D to estimate groundwater flow rates, but not have access to high power computing.

Calibration period model runs were initially forced using high temporal resolution (15 min) pressure head data. No significant difference between observed and predicted heads was detected. However, the computing time and power demand was high (one to two days). To test the model for more practical use by land managers, temporal resolution was reduced to daily time steps, thus reducing the computing time 2-3 hours. It is worth mentioning that the predicted head values remained consistent with the observed head values with both a coarser grid and daily time steps.

4.4.2 HYDRUS – 1D simulated groundwater flux

HYDRUS – 1D predicted hydraulic conductivity (Ks) to be 1.2×10^{-5} m s^{-1} using pedotransfer functions and inverse modeling. Descriptive statistics for groundwater flow are shown in Table 5. The Ks value is the same as that reported by Valett et al. [18] for a study conducted in Rio Calveras, New Mexico, USA, in an alluvial sediment RZ. In another study conducted by Fellows et al. [53] at Rio Caleveras, the average groundwater velocity was reported to be 7×10^{-7} m s^{-1} when the summer stream discharge was 0.0003 m^3 s^{-1}. Compared to those results, the difference in groundwater velocity estimated at Brushy Creek (1×10^{-5} m s^{-1}) may be due to higher stream discharge (0.04 m^3 s^{-1}) relative to that of Rio Caleveras. The ratios between groundwater velocity and stream discharge were 0.07 and 0.007 m^{-2} between Rio Caleveras and Brushy Creek, respectively, indicating that the stream discharge at Brushy Creek could be influenced more by shallow groundwater flow, which given the karst geology of the BWREC may not be surprising. Fluctuations in groundwater flow were instantaneous relative to rising limb of the stream stage hydrograph at PZI, but exhibited a lag time (approximately one day) at PZII (Figs. 7 and 8). This result could be attributable to greater groundwater flow towards the stream at PZII relative to PZI. This finding is corroborated in multiple previous studies [34,54,55] that reported that ground water flowed on average towards the stream from the RZ, with only slight changes in net groundwater flow direction between wet and dry months. Fig. 8 shows that during July through October, the net change in shallow groundwater flow was zero due to lack of stream flow during that period. However, groundwater flow was still observed clearly indicating presence of a substantial subsurface flow regime below the streambed. This observation supplies basis for future mechanistic investigations in semi-karst geological regions of the central USA and elsewhere.

Net groundwater flow varied spatially and temporally depending on stream discharge, precipitation and evapotranspiration. These results are consistent with findings of Wondzell and Swanzon [34] who advocated the use of more complex models (e.g. 3-D) to better characterize subtle groundwater mechanistic relationships. The average linear velocity (using Equation 8 and n = 0.50) did not vary between sites PZI and PZII (2.45×10^{-5} m s^{-1}), and there was limited variation between seasons indicating relatively consistent streambed conductivity (i.e. microscopic flowpaths) to shallow groundwater over time.

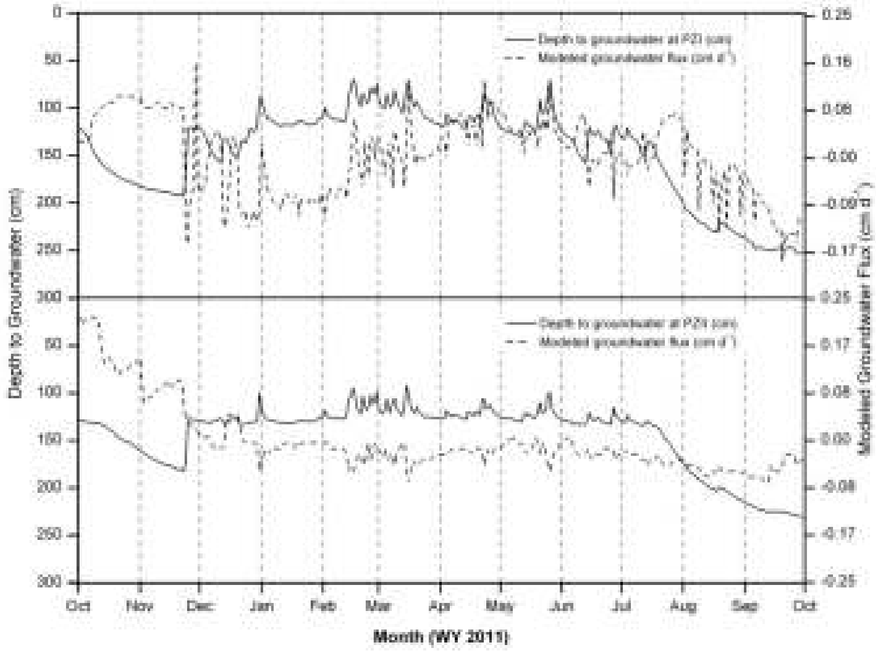

Fig. 7. Depth to groundwater (cm) and simulated groundwater flow (cm d^{-1}) at PZI (top) and PZII (bottom) for water year 2011 at Baskett Wildlife Research and Education Center, Central Missouri, USA

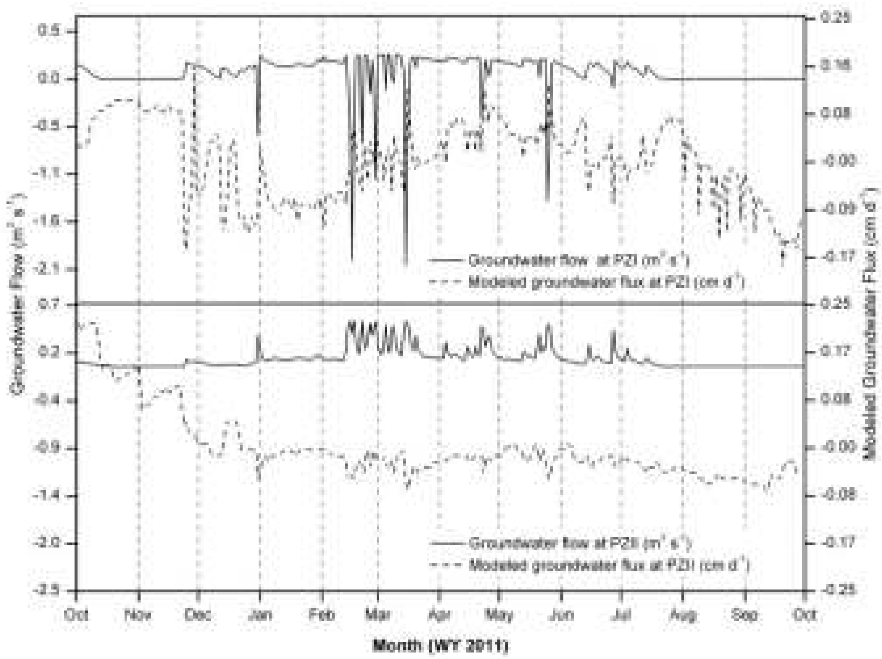

Fig. 8. Depth to ground water flow (m^{3} s^{-1}) and simulated groundwater flow (cm d^{-1}) at PZI (top) and PZII (bottom) for water year 2011 at Baskett Wildlife Research and Education Center, Central Missouri, USA

4.4.3 Validation of HYDRUS – 1D

HYDRUS – 1D simulations were generated for the 2011 WY for both sites (PZI and PZII) using validated soil hydraulic parameters with r^2 values ranging from 0.99 to 0.98 at PZI and 0.98 to 0.96 at PZII. Model validation Nash-Sutcliffe values ranged from 1.00 to 0.99 at PZI, with the former for the hydraulic head in the piezometer most adjacent to the stream, indicating an excellent fit of the modeled hydraulic head to observed hydraulic head. For PZII, NS values ranged from 0.90 to 0.99, indicating a very good fit of the modeled hydraulic head. The RMSE ranged from 2.38 cm to 3.51 cm, while the MD ranged from 1.30 cm to 2.36 cm between the stream and PZI. For PZII, RMSE ranged from 11.16 cm to 2.92 cm, while the MD ranged from 2.24 cm to 10.08 cm. The coefficient of determination (r^2) between observed and modeled hydraulic head was 0.99 for both PZI and PZII, respectively (Table 6). Differences between RMSE values between piezometers may be due to subsurface process heterogeneity, or perhaps model conceptual error (beyond the scope of the current work). Model statistics (Table 6) indicated that HYDRUS – 1D, was accurate in predicting the hydraulic head measurements at PZI and PZII for the study period and was thus rated 'Very Good' in all cases according to the criteria set by Moriasi et al. [47]. HYDRUS - 1D predicted Ks values of 1.2×10^{-5} m s^{-1} which is in close agreement with the average Ks measured from field measurements (1.5×10^{-5} m s^{-1}) and also in agreement with Ks predicted from USDA - National Resources Conservation Service (NRCS) Web Soil Survey (WSS) [23] (1.25×10^{-5} m s^{-1}). Fig. 9 compares modeled hydraulic heads against observed heads for the entire study period.

Table 6. Model performance statistics comparing observed versus modeled Hydraulic Head (Hp) (cm) between piezometer site PZI and PZII, for the calibration period (April to June 2010) at Baskett Wildlife Research and Education Center, Central Missouri, USA

Model Node	**r^2**	**NS**	**RMSE(cm)**	**MD (cm)**
Piezometer	**Piezometer site PZI**			
Pz2	0.99	0.99	2.38	1.3
Pz3	0.99	0.99	3.51	2.36
	Piezometer site PZII			
Pz6	0.99	0.99	2.92	2.24
Pz7	0.98	0.9	11.16	10.08

NS=Nash-Sutcliffe; RMSE=Root Mean Squared Error; MD=Mean Difference

4.5 Study Limitations

HYDRUS - 1D was used in this work to improve manager confidence in the model in semi-karst watersheds. Stream flow and groundwater interactions below and within the streambed were not addressed, as data on surficial streambed geology were not available. User-friendly groundwater models like HYDRUS - 1D should be developed for practitioner uses that simulate three dimensional processes (e.g. simplified MODFLOW, HydroGeosphere, Parflow, GSFlow, and others). HYDRUS - 1D can be parameterized with finer mesh size than used in the current work. However, associated computational power, run time, and labor costs can become prohibitive for use by land managers. It is worth mentioning that calibration runs showed that the difference in results between a coarser and finer mesh size was negligible (<2%). In the current study, piezometer spacing (2 m) was helpful to improve confidence between observed and modeled hydraulic head values. Future studies using HYDRUS – 1D with increased piezometer spacing in karst hydrosystems is warranted.

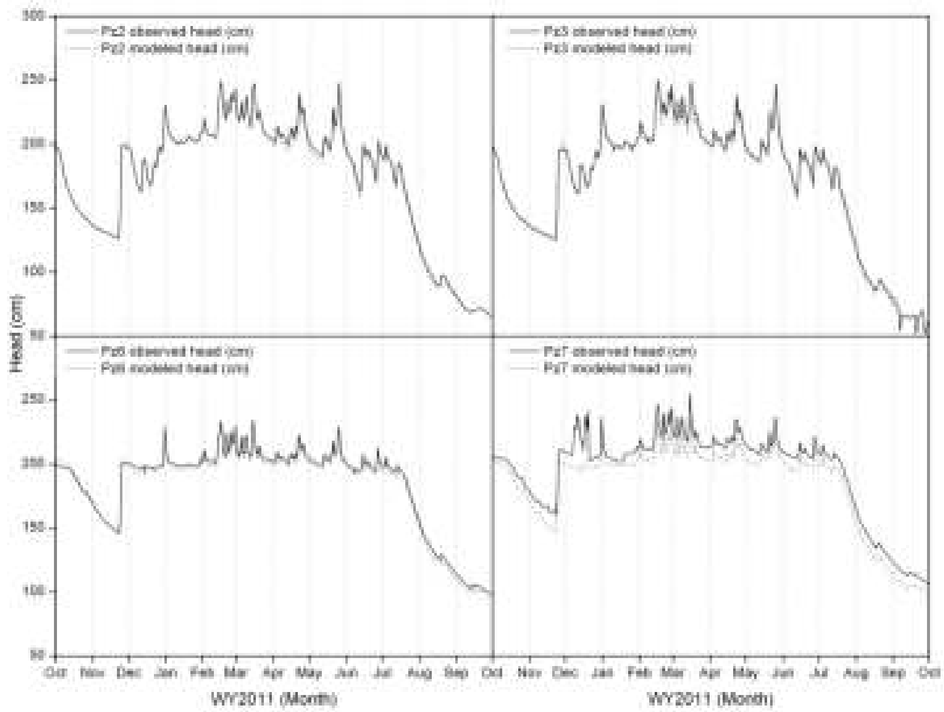

Fig. 9. Observed versus HYDRUS – 1D modeled hydraulic head (Hp) for piezometers Pz2 and Pz3 (located in the piezometer site - PZI) and Pz6 and Pz7 (located in the piezometer site -PZII) over the WY 2011 at Baskett Wildlife Research and Education Center, Central Missouri, USA

4.6 Future Directions

HYDRUS – 1D successfully predicted hydraulic head values, thus illustrating the model's ability to accurately predict shallow groundwater level and flow in semi-karst hydrogeology. Site-specific soil hydraulic parameters are necessary for accurate model runs [8]. Therefore, future studies that model a larger area should include additional observation nodes and finer discretization of model domains. In the current study, HYDRUS – 1D was able to accurately predict groundwater flow using a minimum number of key soil and physical parameters including Ks, soil texture, soil depth, precipitation and hydraulic head. Quantification of those key parameters in the current study showed that the method is easily transferrable to other karst systems of the central US given sufficient key forcings are known.

5. SUMMARY AND CONCLUSIONS

This work quantified shallow groundwater connectivity between a semi-karst Ozark forested riparian zone (RZ) and a second order stream in Central Missouri, USA. Improved understanding of shallow groundwater flow regimes are needed to improve understanding of water availability in many regions including the Central USA. The current work showed that on average, the entire study reach was a losing stream with approximately 19% of streamflow lost to groundwater. However, at individual study sites, the groundwater flow that accounted for stream discharge was 27% at PZI (with 37% increase in total stream flow), and therefore a gaining stream, relative to 69% at PZII (with 218% increase in total stream flow), indicating a gaining stream reach. During late summer and early fall, stream flow was greatly influenced by groundwater flow (70 to 50% of the stream flow was groundwater input). Conversely, during high precipitation events, stream water infiltrated the RZ and increased groundwater storage, as shown by a decrease in depth to groundwater by 41%. Due to silty sand deposits and semi-karst geology present in the study sites, shallow groundwater response was rapid (within a couple of hours). HYDRUS – 1D results were "Very Good" (NS = 0.95, r^2 = 0.99, RMSE = 2.38 cm and MD =1.3 cm) in terms of estimating groundwater depth and flow in the RZ.

This work provides distinct baseline hydrologic and groundwater information that will guide land managers and supplement future investigations in karst hydrogeological forested riparian zones of the Central USA. Models are increasingly used to predict the hydrology of small-scale watersheds and their potential hydrologic response to disturbance. However, field (i.e. observed) data and digital topographic data necessary for modeling are often costly and labor intensive. Results from this work show that with limited input parameters net groundwater flux can be accurately predicted leading to reasonable computations of groundwater storage. Results indicate that HYDRUS – 1D should be considered a reliable management tool (following proper calibration and validation) for establishing groundwater resources management practices for forested RZs in Missouri, the central US, and similar semi-karst hydrogeological regions globally.

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COMPETING INTERESTS

The authors declare no competing interests.

REFERENCES

1. Sophocleous M. Interactions between groundwater and surface water: the state of the science. Journal of Hydrogeology. 2002;10:52-67, doi: 10.1007/s10040-001-0170-8.
2. Dahm CN, Grimm NB, Marmonier P, Valett HM, Vervier P. Nutrient dynamics at the interface between surface waters and groundwaters. Freshwater Biology. 1968;40:427-451.doi: 10.1046/j.1365-2427.1998.00367.
3. Jones JB, Mulholland PJ. Streams and Groundwater. Elsevier Publishing Co., San Diego, CA. 2000a;120-190.
4. Mulholland PJ, Marzolf ER, Webster JR, Hart DR, Hendricks SP. Evidence that hyporheic zones increase heterotrophic metabolism and phosphorus uptake in forest streams. Limnology and Oceanography. 1997;443-451.
5. Darcy H. Principles to follow and formulas to be used in the question of the distribution of water. Public foundations of the city of Dijon; 1856.

6. Richards LA. Capillary conduction of liquids through porous mediums. Physics. 1931;1:318-333.
7. Šimůnek J, Šejna M, Saito H, Sakai M, van Genuchten MT. The HYDRUS-1D Software Package for Simulating the Movement of Water, Heat, and Multiple Solutes in Variably-saturated Media. Version 4.0. HYDRUS Software Series 3, Department of Environmental Sciences, University of California Riverside, Riverside, CA, USA. 315. 2008.
8. Ramos T, Šimůnek J, Gonçalves M, Martins J, Prazeres A, Castanheira N, Pereira L. Field evaluation of a multicomponent solute transport model in soils irrigated with saline waters. Journal of Hydrology. doi: 2011;10.1016/j.jhydrol.2011.07.016.
9. Ocampo CJ, Sivapalan M, Oldham C. Hydrological connectivity of upland-riparian zones in agricultural catchments: Implications for runoff generation and nitrate transport. Journal of Hydrology. 2007;331:643-658, doi: 10.1016/j.jhydrol.2006.06.010
10. Luo Y, Sophocleous M. Seasonal groundwater contribution to crop-water use assessed with lysimeter observations and model simulations. Journal of Hydrology. 2010;389:325-335., doi: 10.1016/j.jhydrol.2010.06.011.
11. Maest AS, Kuipers JR, Travers CL, Atkins DA. Predicting Water Quality at Hardrock Mines: Methods and Models, Uncertainties, and State-of-the-Art, Kupiers and Associates, Montana; 2005.
12. Hurst CA, Thorburn PJ, Lockington D, Bristow KL. Sugarcane water use from shallow water tables: implications for improving irrigation water use efficiency. Agriculture and Water Management. 2004;65:1-19, doi: 10.1016/S0378-3774(03)00207-5.
13. Katsuyama M, Shibata H, Yoshioka T, Yoshida T, Ogawa A, Ohte N. Applications of a hydro-biogeochemical model and long-term simulations of the effects of logging in forested watersheds. Sustainability Science. 2009;4:179-188, doi: 10.1007/s11625-009-0079.
14. Abesser C, Wagener T, Nuetzmann G. Groundwater-Surface Water Interaction: Process Understanding, Conceptualization and Modeling; 2008; IAHS Publication, Oxfordshire, UK.
15. Akerman D, Stein DE. Evaluating the Effectiveness of Best Management Practices Using Dynamic Modeling. Journal of Environmental Engineering. 2008;134:628-639, doi: 10.1061/(ASCE)0733-9372.
16. Tabacchi E, Lambs L, Guilloy H, Planty-Tabacchi AM, Muller E, Decamps H. Impacts of riparian vegetation on hydrological processes. Hydrological Processes. 2000;14(16-17):2959-2976.
17. Castro NM, Hornberger GM. Surface-subsurface water interactions in an alluviated mountain stream channel. Water Resources Research. 1991;27:1613-1621.
18. Valett HM, Fisher SG, Stanley EH. Physical and chemical characteristics of the hyporheic zone of a Sonoran Desert stream. Journal of the North American Benthological Society. 1990;9:201-215.
19. Burt TP, Matchett LS, Goulding KWT, Webster CP, Haycock NE. Denitrification in riparian buffer zones: the role of floodplain hydrology. Hydrological Processes. 1999;13:1451-1463.
20. Burt T, Pinay G, Matheson F, Haycock N, Butturini A, Clement J, Danielescu S, Dowrick D, Hefting M, Hillbricht-Ilkowska A. Water table fluctuations in the riparian zone: comparative results from a pan-European experiment. Journal of Hydrology. 2002A;265:129-148, doi: 10.1016/S0022-1694(02)00102-6.
21. Burt T, Bates P, Stewart M, Claxton A, Anderson M, Price D. Water table fluctuations within the floodplain of the River Severn, England. Journal of Hydrology. 2002B;262:1-20, doi: 10.1016/S0022-1694(01)00567-4.

22. Strahler A. Hypsometric (area-altitude) analysis of erosional topography. Bulletin of the Geological Society of America. 1952;63:1117-1142.
23. US Department of Agriculture (USDA), Natural Resources Conservation Service (NRCS). Web Soil Survey; 2009.
24. Rochow JJ. A vegetational description of a mid-Missouri forest using gradient analysis techniques. American Midland Naturalist. 1972;377-396.
25. Keller WD. The common rocks and minerals of Missouri, University of Missouri, Columbia; 1961.
26. Belden AC, Pallardy SG. Successional trends and apparent Acer saccharum regeneration failure in an oak-hickory forest in central Missouri, USA, Plant Ecology. 2009;204:305-322, doi: 10.1007/s11258-009-9593-4.
27. Reed SE. Ambrosia Beetle Habitat Use, Host Use, and Influence on Early Wood Colonizing Microbes in an Oak-Hickory Forest In Forestry. University of Missouri: Columbia, Missouri. Dissertation; 2010.
28. Radcliffe DE, Simůnek J. Soil Physics with HYRDUS: Modeling Applications. 1st Edition. CRC Press, New York. 2010;51-99.
29. Critchfield HJ. General climatology. Prentice-Hall, Englewood Cliffs, New Jersey; 1966.
30. Gu L, Meyers T, Pallardy SG, Hanson PJ, Yang B, Heuer M, Hosman KP, Liu Q, Riggs SJ, Sluss D. Influences of biomass heat and biochemical energy storages on the land surface fluxes and radiative temperature. Jouranal of Geophysical Research Letters. 2007;112:1-11, doi: 10.1029/2006JD007425.
31. Dottori F, Martina MLV, Todini E. A Dynamic Rating Curve Approach to Indirect Discharge Measurement. Hydrologic Earth Systems Science. 2009;13:847-863.
32. Amoozegar A. Piezometer Method (Saturated Zone): Methods of Soil Analysis, Part 4-Physical Methods; 2002. Soil Science Society of America Inc. Madison, Wisconsin.
33. Freeze RA, Cherry JA. Groundwater. Prentice-Hall, Inc., Englewood Cliffs, New Jersey; 1979.
34. Wondzell SM. Swanson EJ. Seasonal and storm dynamics of the hyporheic zone of a 4th-order mountain stream I: Hydrologic processes. Journal of the North American Benthological Society. 1996;15:3-19.
35. Levia DF, Carlyle-Moses DE, Tanaka T. Reflections on the State of Forest Hydrology and Biogeochemistry, Springer, London. 2011;729-733.
36. Davis SN. Porosity and permeability of natural materials, in R. J. M. De Wiest (Eds.), Flow through Porous Media, Academic Press, New York. 1969;53-89.
37. Harvey J, Wagner B. The physical template: Hydrology, Hydraulics, and Physical structure, in Jones JJ and Mulholland PJ. (Eds.), Streams and Ground waters, Elsevier Publishing Co, San Diego, CA.; 2000B.
38. Scordo EB, Moore RD. Transient storage processes in a steep headwater stream. Hydrological Processes. 2009;23:2671-2685, doi: 10.1002/hyp.7345.
39. Dages C, Voltz M, Ackerer P. Parameterization and evaluation of a three-dimensional modeling approach to water table recharge from seepage losses in a ditch. Journal of Hydrology. 2008;348:350-362, doi: 10.1016/j.jhydrol.2007.10.004.
40. Van Genuchten MT. A closed-form equation for pre- dicting the hydraulic conductivity of unsaturated soils. Journal of the Soil Science Society of America. 1980;44:892-898.
41. Schaap MG, Leij FJ, van Genuchten MT. Rosetta: a computer program for estimating soil hydraulic parameters with hierarchical pedotransfer functions. Journal of Hydrology. 2001;251:163-176., doi: 10.1016/S0022-1694(01)00466-8.
42. Krusekopf HH, Scrivner CL. Soil survey of Boone County, Missouri. U.S. Department of Agriculture and Missouri Agricultural Experiment Station; 1962.

43. Garrett HE, Cox GS. Carbon dioxide evolution from the floor of an oak-hickory forest. Soil Science Society of America. 1973;37:641-644.
44. Šimůnek J, Wendroth O, Van Genuchten MT. Parameter estimation analysis of the evaporation method for determining soil hydraulic properties. Journal of the Soil Science Society of America. 2008;62:894-905.
45. Nash JE, Sutcliffe J. River flow forecasting through conceptual models part I—A discussion of principles. Journal of Hydrology. 1970;10:282-290, doi: 10.1016/0022-1694(70)90255-6.
46. Willmott CJ. On the validation of models. Physical Geography. 1981;2:184-194.
47. Moriasi DN, Arnold JG, Liew MW, Bingner RL, Harmel RD, Veith TL. Model evaluation guidelines for systematic quantification of accuracy in watershed simulations. Transactions of the American Society of Agricultural and Biological Engineers. 2007;50:885-900.
48. Carter RW, Anderson IE. Accuracy of current meter measurements. Journal of the American Society of Civil Engineers. 1963;89:105-115.
49. Marzolf ER, Mulholland PJ, Steinman AD. Improvements to the diurnal upstream-downstream dissolved oxygen change technique for determining whole-stream metabolism in small streams. Canadian Journal of Fisheries and Aquatic Sciences. 1994;51:1591-1599, doi: 10.1139/f94-158.
50. Hughes FMR. The influence of flooding regimes on forest distribution and composition in the Tana River floodplain, Kenya. Journal of Applied Ecology. 1990;27:475-491.
51. Lewandowski J, Lischeid G, Nützmann G. Drivers of water level fluctuations and hydrological exchange between groundwater and surface water at the lowland River Spree (Germany): field study and statistical analyses. Hydrological Processes. 2009;23:2117-2128, doi: 10.1002/hyp.7277.
52. Lewandowski J, Nutzmann G. Surface water-groundwater interactions: hydrological and biogeochemical processes at the lowland River Spree (Germany), IAHS Publications-Series of Proceedings and Reports, Oxfordshire, UK, Report -321. 2007;30-38.
53. Fellows CS, Valett HM, Dahm CN. Whole-stream metabolism in two montane streams: Contribution of the hyporheic zone. Limnology and Oceanography. 2001;46:523-531.
54. Wroblicky GJ, Campana ME, Valett HM, Dahm CN. Seasonal variation in surface-subsurface water exchange and lateral hyporheic area of two stream-aquifer systems. Water Resources Research. 1998;34:317-328, doi: 10.1029/97WR03285.
55. Harvey J, Wagner B. Interactions between Streams and their subsurface Hyporheic Zone, in Jones JJ, and Mulholland PJ. (Eds.), Streams and Ground waters. Elsevier Publishing Co., San Diego, CA.; 2000A.

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