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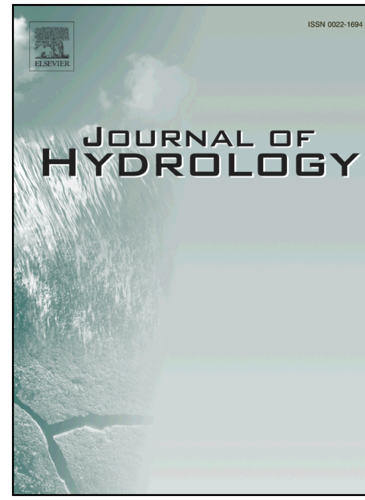

Research papers

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Groundwater Flux Estimation in Streams: A Thermal Equilibrium Approach

Yan Zhou, Garey A. Fox*, Ron B. Miller, Robert Mollenhauer, Shannon Brewer

Yan Zhou, Ph.D., and Ron Miller, Ph.D., Department of Biosystems and Agricultural Engineering, Oklahoma State University, Stillwater, OK; G. A. Fox*, Department of Biological and Agricultural Engineering, North Carolina State University, Raleigh, NC; Robert Mollenhauer, Ph.D., and Shannon Brewer, Ph.D., U.S. Geological Survey, Oklahoma Cooperative Fish and Wildlife Research Unit, Stillwater, OK.

*Corresponding author: Dr. Garey A. Fox, Professor and Department Head, Department of Biological and Agricultural Engineering, North Carolina State University, Campus Box 7625, Raleigh, NC 27695, Phone: 919-515-6700; E-mail: garey_fox@ncsu.edu.

Abbreviations: NSE, Nash–Sutcliffe Efficiency coefficient; RMSE, root mean square error; TEM, thermal equilibrium method; WASP, Water Quality Analysis Simulation Program

Groundwater Flux Estimation in Streams: A Thermal Equilibrium Approach

Abstract. Stream and groundwater interactions play an essential role in regulating flow, temperature, and water quality for stream ecosystems. Temperature gradients have been used to quantify vertical water movement in the streambed since the 1960s, but advancements in thermal methods are still possible. Seepage runs are a method commonly used to quantify exchange rates through a series of streamflow measurements but can be labor and time intensive. The objective of this study was to develop and evaluate a thermal equilibrium method as a technique for quantifying groundwater flux using monitored stream water temperature at a single point and readily available hydrological and atmospheric data. Our primary assumption was that stream water temperature at the monitored point was at thermal equilibrium with the combination of all heat transfer processes, including mixing with groundwater. By expanding the monitored stream point into a hypothetical, horizontal one-dimensional thermal modeling domain, we were able to simulate the thermal equilibrium achieved with known atmospheric variables at the point and quantify unknown groundwater flux by calibrating the model to the resulting temperature signature. Stream water temperatures were monitored at single points at nine streams in the Ozark Highland ecoregion and five reaches of the Kiamichi River to estimate groundwater fluxes using the thermal equilibrium method. When validated by comparison with seepage runs performed at the same time and reach, estimates from the two methods agreed with each other with an R^2 of 0.94, a root mean squared error (RMSE) of 0.08 (m/d) and a Nash–Sutcliffe efficiency (NSE) of 0.93. In conclusion, the thermal equilibrium method was a suitable technique for quantifying groundwater flux with minimal cost and simple field installation given that suitable atmospheric and hydrological data were readily available.

Keywords. Groundwater flux; Thermal equilibrium; Seepage run; Stream water temperature; Temperature signature

Introduction

The interaction of stream water with groundwater influences water quality and quantity and plays an essential role in aquatic ecosystems. Streams with high groundwater interactions are often characterized by high biological and microbial diversity and activity due to elevated solute transport and nutrient exchange across the streambed interface (Laursen and Seitzinger, 2005; Schmidt et al., 2007). Groundwater flux can also limit benthic invertebrate exposure to low oxygen and contaminants (Malard and Hervant, 1999), and provide thermal refugia and microbial food supply for fishes (e.g., salmon) (Kurylyk et al., 2013). The importance of groundwater to stream biota has led to increased efforts to quantify the effects of groundwater on both stream temperatures (Constantz, 1998) and energy sources (Barlocher and Murdoch, 1989). However, the complex nature of stream-groundwater hydrological connectivity can make quantifying those interactions difficult and labor intensive.

Over the past few decades, many approaches have been developed to quantify stream and groundwater interactions that can be generally categorized into Darcian, streamflow, water budget and tracer methods (Table 1). Extensive reviews of these approaches have been provided by Kalbus et al. (2006), Brodie et al. (2007), and Turner (2009), but are briefly overviewed below. Darcian methods calculate point stream-groundwater exchange flux as the product of measured hydraulic gradient and conductivity based on Darcy's Law in a manner similar to that used to investigate water movement in porous media (Freeze and Cherry, 1979). Water budget methods use groundwater and watershed models, separately or in combination, to estimate groundwater and stream interactions as the unknown residual of the water budget by calibrating the model against streamflow records and estimated physical parameters of the aquifer.

Streamflow methods include a variety of approaches such as hydrograph separation, direct measurement using seepage meters and seepage runs (Harvey and Wagner, 2000). The hydrograph separation methods, such as recession-curve displacement and stream base-flow analysis, use various assumptions to separate a stream hydrograph into the different runoff, interflow, and baseflow components (Scanlon et al., 2002). The seepage meter method allows direct point measurement of stream and groundwater flux by calculating the rate of volume change of a collection bag over the area of the collecting bucket pushed into the streambed (Zamora, 2008). The incremental streamflow method for estimating groundwater flux (hereafter 'seepage run') involves measuring streamflow at multiple transects along the river (Donato, 1998; Harvey and Wagner, 2000). After eliminating contributions from tributaries, the surface-groundwater flux is assumed to be the flow rate difference between transects (Rosenberry and LaBaugh, 2008). Tracer methods estimate groundwater flux based on the mass balance of tracers. Introduced tracers, commonly chloride or dyes, are usually used in either dilution gauging or transient storage approaches (Zhou et al., 2016) while environmental tracers such as tritium and chlorofluorocarbons are used in hydrograph separation to provide information on groundwater flux. The limitations of these conventional methods are the high time and material cost for proper installation and maintenance (e.g., Darcian method with piezometer and seepage meter) (Berry et al., 2011) and the difficulty in parameter estimation (e.g., water budget methods) (Scanlon et al., 2002). Due to the ease of monitoring stream temperatures, thermal methods overcome some of these limitations and have gained increasing popularity in recent decades (Webb et al., 2008).

Thermal methods use heat as an environmental tracer with the analysis based on heat transfer and energy balance analogous to the mass balance of common chemical tracers. Thermal methods emerged as a versatile class of geophysical tools for monitoring focused recharge in arid and semiarid settings, but did not come into common use until the 1960s (Blasch et al., 2007) after analytical solutions to the coupled heat and water transport equations were established by Suzuki (1960), Stallman (1965), and Bredehoeft and Papaopulos (1965). The vertical thermal gradient method exploits the coupled relationship between heat and water advection and conduction processes to model vertical heat and water movement across the streambed (Anderson, 2005). By monitoring the temperature of stream water and saturated bed sediment at multiple depths over time, this method estimates the vertical movement of groundwater. This method has been used to investigate infiltration and percolation on the land surface (Suzuki, 1960), indicate gaining and losing reaches of stream channels (Lapham, 1989; Silliman and Booth, 1993; Constantz, 1998), and locate areas of inflow to lakes (Lee, 1985).

The stream thermal modeling approach typically uses a process-based model (Becker et al., 2004; Loheide and Gorelick, 2006) to simulate the heat budget of the stream using known hydrological and atmospheric variables and quantify heat introduced by groundwater flux as the residual of the known stream water heat budget. For example, Sinokrot and Stefan (1993) developed a numerical, finite-difference model for stream temperatures. In shallow streams, they noted the primary importance of incoming solar radiation but that other components of the heat balance (long-wave back radiation, evaporation, convection to the atmosphere, and conductive heat exchange between the streambed and water) are also significant. In an attempt to develop a tool for ecohydrological assessment in a watershed, Loinaz et al. (2013) applied a surface water-groundwater flow and heat transport model to predict stream temperatures. They noted the importance of spatially distributed flow dynamics for calibrating the model to match stream temperatures. A benefit of stream temperature modeling is that it can be performed at small spatial and temporal resolutions. For example, Westhoff et al. (2007) used data from a distributed temperature sensing system with 1.0 m spatial and two-minute temporal resolution to model stream temperature. Their results suggested that lateral groundwater inflow was a significant parameter for numerically predicting stream temperatures.

Despite these advances, there are still new potential applications for thermal methods. The vertical thermal gradient method provides a convenient alternative for quantifying groundwater flux at point scales, but the material and time costs are significant if the scale is to be expanded using multiple measurements. Stream thermal modeling methods estimate groundwater flux at a larger scale with relatively lower cost, but it loses the sensitivity of point estimations. Thus, there is still a need for accurate, convenient, and economical means of quantifying point groundwater flux that can be expanded to cover a predetermined area, e.g., stream reach.

This research proposed a thermal equilibrium method (hereafter TEM) to estimate the time-averaged groundwater flux to a stream using monitored stream water temperature data at a single point and atmospheric and hydrological data. The proposed approach significantly reduces the need and cost of data collection while maintaining the sensitivity and independence of a point measurement. The research validated the performance of the TEM by comparison with estimates from seepage runs (i.e., streamflow measurements).

Materials and Methods

Thermal Equilibrium Method

The TEM was developed based on the assumed thermal equilibrium of all heat transfer processes in the stream including both atmospheric heat transfer and groundwater interactions. Equilibrium stream water temperature, calculated based on atmospheric conditions (atmospheric equilibrium water temperature, hereafter T_{AE}), has traditionally been used as an approximation to stream water temperature (hereafter T_S) (Smith, 1981). Recent research showed that the T_{AE} calculated on a weekly or coarser temporal scale was linearly related, but not equal to T_S (Bogan et al., 2003). The differences between T_{AE} and T_S were attributed to external water inputs, primarily groundwater interactions for 80% of 596 sites in the eastern and central USA (Bogan et al., 2003; Bogan et al., 2004; Webb et al., 2008). In the current study, we assumed that by including groundwater interactions, a more comprehensive equilibrium water temperature (hereafter T_E) could be calculated to appropriately represent T_S on a weekly or coarser temporal scale. In another words, we assumed streams were at thermal equilibrium with the combination of atmospheric conditions and groundwater interactions.

A stream temperature model was applied to simulate the atmospheric heat transfer processes (i.e., heat conduction, shortwave solar radiation, longwave atmospheric radiation, etc.) based on the upstream boundary of monitored T_S and atmospheric and hydrological conditions of the monitored point. The monitored stream point was represented by an expanded continuous model domain (Figure 1), allowing the model to stabilize and predict T_{AE} at the downstream boundary. Based on the thermal equilibrium assumption, the difference between T_S (upstream boundary) and predicted T_{AE} (downstream boundary) was attributed to groundwater flux. Therefore, if the predicted T_{AE} of the downstream boundary differed from the upstream T_S, a groundwater flux could be applied to the domain and calibrated until the difference between the two boundaries was minimized ($T_E = T_S$). The magnitude of the flux required for thermal equilibrium would provide an estimate of the unknown groundwater flux at the monitoring point.

In this study, the Water Quality Analysis Simulation Program (WASP) was used to simulate stream heat transfer with an output temporal resolution of 1 hr. WASP, developed by the U.S. Environmental Protection Agency (EPA) (Wool et al., 2006), is a dynamic compartment-modeling program for pollutant transport in aquatic systems. The time-varying processes of advection, dispersion, point and diffuse mass loading and boundary exchange are represented in the basic program. In the WASP temperature module, heat transfer is computed based on the following one-dimensional advection-diffusion equation:

$$\frac{\partial T_s}{\partial t} = -\frac{\partial}{\partial x}(V_x T_s) + \frac{\partial}{\partial x}\left(D_x \frac{\partial T_s}{\partial t}\right) + \frac{H_n A_s}{\rho_w C_p V} + S \qquad (1)$$

where T_S is the stream water temperature (℃), V_x is the advective velocity (m/s), D_x is the dispersion coefficient (m^2/s), V is the segment volume (m^3), A_s is the segment surface area (m^2), ρ_w is the density of water (997 kg/m^3), C_p is the specific heat of water (4179 J/kg ℃), H_n is the net surface heat flux (W/m^2), and S is the loading rate include boundary, direct and diffuse loading (℃/s). The net surface heat flux includes the effects of a number of processes computed as (Cole and Buchak, 1995):

$$H_n = H_s + H_a + H_e + H_c - (H_{sr} + H_{ar} + H_{br}) \qquad (2)$$

where H_n is the net heat flux across the water surface (W/m^2), H_s is the incident short wave solar radiation (W/m^2), H_a is the incident long wave atmospheric radiation (W/m^2), H_{sr} is the reflected short wave solar radiation (W/m^2), H_{ar} is the reflected long wave radiation (W/m^2), H_{br} is the back radiation from the water surface (W/m^2), H_e is the evaporative heat loss (W/m^2), and H_c is the heat conduction (W/m^2).

In this study, a one-dimensional conceptual domain with a length of 2 km was constructed in WASP and divided into twenty 100 m-long segments (Figure 2). A monitored T_S time series was input as the upstream boundary and the initial temperature for each segment was set to the T_S at the first time step. The geometry and flow rate in the main channel of the model were assumed to be uniform and described by parameters acquired from transect measurements at the monitored point (see seepage runs below). An atmospheric time series was obtained from the nearest Oklahoma Mesonet station and input into the WASP model to compute heat transfer at each time step. The Oklahoma Mesonet includes 121 automated weather-monitoring stations distributed throughout Oklahoma with observations every 5 minutes (http://mesonet.org, Brock et al., 1995). The effect of canopy cover was neglected because the studied reaches were located on unshaded areas of higher order streams. Thermal interaction of groundwater flux was represented by a uniform flow input across the twenty segments and incorporated in the model via hydrological connections (Figure 2).

Using measured or estimated groundwater temperature, atmospheric and hydrologic variables, estimates for thermal variables such as the dispersion coefficient, and constants for thermal properties, the approach then calibrates the magnitude of the groundwater flux until the sum of squared error (SSE) was minimized between the predicted T_E at the downstream boundary and T_S at the upstream boundary. When the temperature at the two boundaries matched, the net heat transfer across the conceptual domain was zero and all the heat transfer processes were equilibrated. The estimated flow represented the optimal groundwater flux required for the T_S to equilibrate as indicated in the thermal equilibrium assumption.

In this study, the groundwater temperature time series was estimated from air temperature with a 1.5-month time lag as recommended by Pluhowski (1970) (Figure 3). The air temperature offset assumed in this research allows easy application of the TEM. Alternatively, practitioners can measure or estimate GW temperatures using a method of their choice and utilize those data in the TEM methodology. Measuring local groundwater temperatures can allow the consideration of more local conditions. Also, note that the length and number of segments constructed in the model did not physically represent the monitored point, but served only as a model domain that allowed the model to stabilize.

Study Areas

To validate the TEM by comparison with seepage runs (i.e., streamflow measurements), five sampling reaches were chosen on the Kiamichi River (Figure 4). The Kiamichi River watershed in southeast Oklahoma has an area of about 4800 km^2, with elevation ranging from 270 to 810 m (Pyron et al., 1998). The sedimentary rocks of the area have been deformed into tightly folding anticlines and synclines forming steep east-west trending ridges separated by a broad and flat-bottomed stream valley. The area was expected to have substantial groundwater storage potential as well as permeability to allow stream and groundwater interactions.

Nine additional sampling reaches were located on different streams in the Springfield Plateau in the Ozark Highland ecoregion of Missouri, Arkansas and Oklahoma (Figure 4). The Springfield Plateau comprises the southwest portion of the Ozark Plateau with an area of approximately 26,700 km^2 including parts of west-central and southwest Missouri, northeast Oklahoma, southeast Kansas and northern Arkansas (Adamski et al., 1995). Elevations range from 300 to 520 m with mostly gentle topographic relief except for Eureka Springs Escarpment that separates the Springfield and Salem Plateaus. Most streams in Springfield Plateau drain radially from the plateau center (Adamski et al., 1995; Nigh and Schroeder, 2002). The limestone bedrock in the region is intermittently soluble, producing regionally abundant sinkholes, springs, and caves (Nigh and Schroeder, 2002). The Springfield Plateau overlies the Ozark Plateau aquifer system, which extends throughout southern Missouri, eastern Oklahoma, southeast Kansas and a large area of northwest Arkansas (Miller and Appel, 1997). Extending sites to the Ozark Highlands allowed us to test the TEM on streams with higher expected groundwater contributions due to the predominant karst topography. All of the study sites were chosen for near-natural flow characteristics. Examination of the study reaches in a GIS showed that most of the reaches were near small farm ponds (≤ 1 km), some of the reaches were relatively near household water wells (~200-300 m), and one reach contained a permitted surface water irrigation diversion, although it was unlikely to be active during the winter when the seepage run and temperature monitoring were conducted. None had instream impoundments.

Incremental Streamflow Method: Seepage Runs

Seepage runs were performed at each site to validate the TEM (Figure 1, Table 2). Reaches were selected from candidate streams without flow contributions from tributary streams or major springs. Once identified, each reach was divided into three to five transects separated by 200 to 500 m (Figure 1, Table 2). Discharge at each transect was measured with a RiverSurveyor M9 Acoustic Doppler Current Profiler (SonTek, San Diego, CA; hereafter ADCP). The enhanced density of transects per reach was established to achieve a smaller spatial scale which more closely matched the model setup used in the TEM while maintaining accurate groundwater flux estimation in consideration of instrument accuracy (error ≤ ± 0.015 m^3/s). At each reach, the ADCP-measured discharge at each transect was normalized for any flow changes detected at nearby USGS gauges during the sampling period to remove any temporal variation.

The normalized transect discharges were then regressed against the separation distance (upstream to downstream) with the slope of the regression representing the flux between the stream and groundwater for the specific reach (Rosenberry and LaBaugh, 2008). Each seepage run included a flow and transect measurement at each logger site that were used to describe the channel geometry and hydrology in the model. The groundwater flux measurements were normalized by the streambed area (i.e., stream length and average ADCP transect width). Using T_S measured instantaneously by the ADCP, the T_S difference among transects within each seepage run was determined to be less than 2℃, with this temperature variation likely due to diurnal temperature variations.

Stream Temperature and Atmospheric Time Series

Temperature loggers (HOBO Water Temp Pro v2) were placed in the thalweg at each of the selected reaches. Hourly averaged T_S readings were recorded over a 15-d period in September 2016 on the Kiamichi River and June and December 2016 in the Ozark streams. Those times covered an extended low-flow period without any significant precipitation event. In the shallow Ozark streams, the loggers were placed at a depth between 0.3 and 1 m, and 1.0 to 1.5 m in the deeper Kiamichi River.

A time series of air temperature, wind speed, solar radiation, and relative humidity was obtained for each site from the nearest Oklahoma Mesonet site (OCS, 2016), with the largest distance from a stream site being approximately 35 km for the Kiamichi River and approximately 40 km for the Ozark streams. The Mesonet stations are automated and collect data at 5-min increments, and reported an hourly average corresponding to the T_S time series. Note that there was some regional variation from site to site (Table 3), but the variation was not substantial. For the TEM approach, if users have meteorological data closer to their sites, they can easily use that data. In other words, the TEM approach should utilize the best available meteorological and hydrological data.

Statistical Evaluation

To validate the TEM, the FITEVAL software was used to evaluate the fit between groundwater fluxes measured from seepage runs and predicted by the TEM. FITEVAL is a software tool that uses procedures presented by Ritter and Muñoz-Carpena (2013) to incorporate both data and model uncertainty into standardized model evaluation. FITEVAL conducts model evaluations using a combination of graphical illustrations, absolute value error statistics (root mean square error, RMSE), and normalized goodness-of-fit statistics (Nash–Sutcliffe Efficiency coefficient, NSE). Bias corrected confidence intervals are calculated based on approximated probability distributions derived from bootstrapping, followed by hypothesis test results of the indicators, helping to reduce subjectivity in the interpretation of the model performance (Ritter and Muñoz-Carpena, 2013).

Results and Discussion

TEM versus Seepage Runs

Model validation suggested that TEM was a suitable technique for estimating groundwater flux into streams. The groundwater flux into the streams measured via seepage runs ranged from 0.01 to 1.09 m/d and from 0.00 to 0.95 m/d with the TEM (Table 4). The estimated groundwater flux at the Ozark sites was generally higher than at the Kiamichi sites as expected. The resulting RMSE and NSE for the TEM fit to the seepage run data from FITEVAL were 0.08 (m/d) and 0.93, respectively, indicated a very good fit. Linear regression analysis showed a uniform variance across the range of estimates with an R^2 of 0.94 (Figure 5). However, the TEM tended to under predict the seepage run flux estimates by -5.7% (Figure 5). An example continuous application of the TEM at a weekly time scale is shown in Figure 6 for one of the sites. The corresponding groundwater flux estimated by the seepage run is also shown in the figure.

Deviations are to be expected between the two measurements. The seepage run represents a spatially integrated flux estimate over a small temporal scale (~2 hr), whereas the TEM represented a temporally integrated flux estimate of a small spatial scale. More specifically, during a seepage run, locally alternating gaining and losing sections of the stream are integrated into this spatially integrated measurement. These two estimates were similar suggesting that the groundwater flux into these streams may not vary widely over the approximately 1.5 km of stream length or the 15-d time period used in this study. Also, the seepage runs measured the net groundwater exchange. If the groundwater discharge along a reach was exactly balanced with the groundwater recharge, then the net change of streamflow would not be detected. The TEM method, however, will quantify the groundwater inflow because it generates a temperature signal. This difference may explain situations when the TEM method overestimated the groundwater flux compared to that measured during the seepage runs, e.g. Kiamichi River. Future research should examine the prediction from TEM further by comparison against estimates derived from other methods at different times of the year and with temporal scales that align better with TEM.

Other than the groundwater flux, the expanded domain length was the only model parameter not represented by measurements; therefore, it was important to examine that parameter and its influence on the groundwater flux results. The model domain in the TEM was a 2-km conceptual stream reach composed of twenty segments of 100 m each. To test the effect of model domain length, groundwater flux for Spavinaw Creek in northwest Arkansas was estimated with TEM using alternate total domain lengths of 0.2 and 20 km, each with twenty equal length segments. The estimated groundwater fluxes (indicated by the minimum of SSE between the upstream and downstream boundaries) were identical for the 0.2- and 2-km domains, but larger for the 20-km domain. This is likely due to the accumulation of groundwater flux over an extensive simulation distance that significantly changed the heat capacity of the stream. For example, at Spavinaw Creek the average flow rate was 2 m^3/s, and the total estimated groundwater flux accumulation over a 2-km model domain was 0.1 m^3/s; a difference that is unlikely to significantly change the energy balance of the stream. In contrast, the total groundwater flux accumulation over a 20-km model domain was 1.0 m^3/s with the same rate of groundwater flux, an increase that greatly affected the stream heat capacity.

Since the design of the model domain also affects runtime, some test runs with different domain dimensions may be helpful to balance accuracy and processing time. The temperature module of WASP applied the given thermal and stream parameters sequentially to each segment using a variable internal time-step to reach satisfactory convergence. For the test simulations mentioned above, the run time of the 0.2-km model domain extended to over an hour, whereas the 2-km domain took only 7 to 10 minutes. This was likely due to the extra iterations required for time-dependent thermal processes to converge in the reduced length of the smaller domain.

Where is TEM applicable?

Due to the one-dimensional nature of the temperature model used in TEM, the method was most appropriate for shallow, well-mixed streams that were unlikely to exhibit stratified zones of temperature or flow. A groundwater temperature signature, defined as the temperature difference between T_{AE} and T_E caused by groundwater flux, was required for the TEM to predict effectively. Streams with low flow and no groundwater flux tend to equilibrate at a high temperature during warm weather conditions ($T_E = T_{AE}$). In contrast, streams with groundwater flux cooler than stream water equilibrate at a lower temperature during warm weather conditions ($T_E < T_{AE}$), causing a temperature signature that could be used to quantify groundwater flux through TEM. However, groundwater recharge (i.e., losing streams) would not result in a similar temperature signature, and thus could not be quantified by the TEM. Similarly, when T_S approximate groundwater temperatures during certain times of the year (Figure 3) (Briggs et al., 2016; Kurylyk et al., 2016), the temperature signature of the groundwater flux would be difficult to detect. Therefore, the TEM is most effective where the temperature signature of groundwater flux is strong, i.e., gaining reaches and seasons when groundwater temperatures deviate from T_S. Nevertheless, the change in heat capacity caused by the loss or addition of stream water volume will lead to an altered T_S temporal variance. Future research with higher data precision may be able to identify the altered T_S variance and use it to quantify groundwater interactions similarly to TEM.

To improve the robustness of the thermal equilibrium assumption, it is important to consider the location of the T_S monitoring point and the sampling duration. When groundwater flux changes gradually, stream water remains at thermal equilibrium and therefore $T_S = T_E$ (Figure 7). In contrast, upwelling zones, where there would be an abrupt change in groundwater flux, would cause a loss of thermal equilibrium that recovers over some downstream distance ($T_S \neq T_E$). Groundwater flux estimates made at any point at thermal equilibrium represented the true magnitude of groundwater flux into the stream at that point. Estimates made at points where thermal equilibrium is recovering would yield an inaccurate groundwater flux because the T_S does not meet the primary assumption of the TEM. Although an investigator is unlikely to have prior knowledge of the spatial distribution of groundwater interactions in a particular stream, it would be advantageous to avoid placing temperature loggers at locations with drastically varying temperatures. Moreover, based on previous research, we suggest that at least one week of T_S time series should be collected for the thermal equilibrium assumption to be met (Bogan et al., 2003).

Conclusions

The TEM proposed in this research has several advantages to researchers interested in characterizing stream and groundwater interactions as long as the primary assumptions of the approach are met. With this approach, only T_S is needed at a single point to monitor groundwater flux. This can also potentially add significant value to T_S data typically collected in stream biology studies (Hawkins et al., 1997), and T_S data are readily available at a number of USGS gage locations. Although a minimum of one week of T_S data is recommended to satisfy the thermal equilibrium assumption (Bogan et al., 2003), the TEM can be used to estimate groundwater flux at any temporal scale coarser than one week (i.e., monthly, seasonally or yearly). Similarly, the proposed method has the potential to economically quantify spatial differences in groundwater fluxes at multiple stream points or to create a flux estimate for a large area if applied in an array. The main limitation of the TEM is that it requires a detectable and equilibrated temperature signature of groundwater flux. Another weakness of the method, and one that it shares with other model approaches, is that the precision of groundwater flux is heavily dependent on the availability and quality of the input data. Our study utilized atmospheric and solar radiation data from the Oklahoma Mesonet; however similar systems exist or are being installed in other many other states. Finally, the approach performs best in well-mixed shallow streams because those conditions most closely match the one-dimensional model structure.

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Table 1. Comparison of common methods for estimating groundwater discharge to/from streams (adapted from USGS, 2017).

Category	Method	Spatial Scale	Temporal Scale	Typical Quantity Estimated	Ease of Use	Data Needs	Relative Cost	Reference
Water Budget	Groundwater Modeling	Local / Regional	Month to Years	Recharge	Moderate	High	High	Sophocleous and Perkins (2000)
	Watershed Models	Watershed /Regional	Days to Years	Recharge	Moderate	High	High	Sophocleous and Perkins (2000)
Darcian	Piezometers	Point	Instantaneous	Potential Recharge	Moderate	Low	High	Stofleth et al. (2008)
Streamflow	Seepage Meters	Point	Event to Months	Potential Recharge	Moderate	Low	Low	Taniguchi and Fukuo (1993)
	Stream Base-Flow Analysis	Watershed	Years	Net Recharge	Easy	Low	Low	Arnold et al. (1995)
	Incremental Streamflow Method (Seepage Run)	Local	Instantaneous	Potential Recharge	Easy	Low	Low	Rosenberry and LaBaugh (2008)
	Recession-Curve Displacement Method	Watershed	Event to Years	Net Recharge	Moderate	Low	Low	Rutledge (1998)
Tracer	Chloride	Point	Years	Recharge	Easy	Moderate	Moderate	Eriksson and Khunakasem (1969)
	Chlorofluorocarbons	Local	Month to Years	Recharge	Difficult	Moderate	High	Cook and Solomon (1997)
	Temperature	Point	Days to Years	Recharge	Moderate	Moderate	High	Constantz (2008)
	Tritium	Point	Month to Years	Recharge	Moderate	Moderate	High	Allison and Hughes (1975)

Table 2. Characteristics of the seepage runs performed to quantify groundwater flux into streams.

Site Name	Upstream Boundary		Downstream Boundary		Date Performed	Measurement Period	Number of Transects
	Latitude	Longitude	Latitude	Longitude	Year/Month/Day	(hr)	
NDN	34.6597	-95.0307	34.6578	-95.0415	2016/7/6	2.4	4
Robins	34.6361	-95.125	34.6270	-95.1267	2016/7/7	1.8	4
JFC up	34.5986	-95.3281	34.5976	-95.336	2016/7/7	1.4	4
JFC down	34.5959	-95.3368	34.5895	-95.3395	2016/7/7	1.2	3
Payne	34.4255	-95.5765	34.4190	-95.5727	2016/7/8	1.8	3
Spavinaw	36.3245	-94.7063	36.3214	-94.7142	2017/1/11	1.1	3
Honey	36.5401	-94.7036	36.5428	-94.7111	2017/1/10	1.3	3
Caney	35.7927	-94.8475	35.7886	-94.8499	2016/6/7	2.2	4
Buffalo	36.6396	-94.6273	36.6356	-94.6303	2017/1/11	1.3	3
Saline	36.2896	-95.0847	36.2850	-95.0917	2017/1/11	1.0	3
Caney 2	35.7927	-94.8475	35.7886	-94.8499	2017/1/12	1.6	4
Greenleaf	35.7523	-95.0472	35.7410	-95.0591	2017/1/13	2.0	5
Spavinaw 2	36.3495	-94.5666	36.3335	-94.6386	2017/6/22	2.3	4
Spavinaw 3	36.3297	-94.6468	36.3271	-94.6685	2017/6/22	2.2	3

Table 3. Regional variations of air temperature (^oC), dew point temperature (^oC), wind speed (m/s), and solar radiation (W/m^2) monitored at Mesonet stations located near the Kiamichi River in September 2016 and in the Ozark Highland Ecoregion in June and December 2016, showing their averages, errors and p-value of paired t-tests. Atmospheric conditions monitored at Clayton Station were compared to those at two nearby Mesonet Stations: Talihina (36 km away) and Antlers (67 km away). Atmospheric conditions monitored at Jay were compared to those at the nearby Mesonet station at Westville (67 km away).

		Air Temperature (^oC)	Dew Point (^oC)	Wind Speed (m/s)	Solar Radiation (W/m^2)
Kiamichi River – September 2016					
Clayton	Average	23.63	18.58	1.56	232.43
Talihina	Average	23.45	18.45	1.83	233.16
	Error	-0.18	-0.13	0.27	0.73
	p-Value	0.00	0.00	0.00	0.74
Antlers	Average	23.57	18.99	1.55	226.03
	Error	-0.05	0.41	-0.01	-6.41
	p-Value	0.26	0.00	0.76	0.01
Ozark Highland Ecoregion – June 2016					
Jay	Average	25.16	20.31	1.77	305.34
Westville	Average	25.28	19.83	1.91	301.08
	Error	0.12	-0.48	0.14	-4.26
	p-Value	0.00	0.00	0.00	0.07
Ozark Highland Ecoregion – December 2016					
Jay	Average	4.58	-1.90	3.02	84.93
Westville	Average	5.33	-0.85	3.36	81.76
	Error	0.75	1.05	0.34	-3.18
	p-Value	0.00	0.00	0.00	0.43

Table 4. Comparison of groundwater flux (m/d) estimated by seepage run and thermal equilibrium methods for each sample site. Stream water and air temperature (℃) during the simulation period were averaged and reported as T_w and T_a respectively.

Watershed	Site Name	Groundwater Flux Estimate: Seepage Run, SR (m/d)	Groundwater Flux Estimate: Thermal Equilibrium Method, TEM (m/d)	Relative Difference* (%)	Stream Temperature, T_w (°C)	Air Temperature, T_a (°C)
Kiamichi River	NDN	0.11	0.08	-27	24.5	19.2
	Robins	0.14	0.10	-29	22.2	19.2
	JFC up	0.08	0.11	38	22.4	19.2
	JFC down	0.01	0.12	1100**	22.4	19.2
	Payne	0.01	0.08	700	22.8	19.2
Ozark Highland Ecoregion	Spavinaw	0.38	0.46	21	9.8	3.3
	Honey	0.74	0.65	-12	10.4	8.1
	Caney	0.35	0.32	-9	9.3	5.7
	Buffalo	0.61	0.63	3	10.7	2.9
	Saline	0.66	0.54	-18	10.6	2.9
	Caney 2	0.15	0.05	-67	9.5	3.4
	Greenleaf	0.09	0.00	-100	7.1	3.4
	Spavinaw 2	0.56	0.61	9	21.0	24.4
	Spavinaw 3	1.09	0.95	-13	21.2	24.4

* Calculated as (TEM-SR)/SR x 100%

** Large relative percent differences were due to low groundwater fluxes measured during the seepage run

Figure Captions

Figure 1. Diagram of the application of the thermal equilibrium method (TEM) and seepage run segment at one of the sites. Diamonds in the seepage run segment represent measurement transects. The TEM assumes stream water temperatures are at thermal equilibrium with the combination of atmospheric conditions and groundwater interactions at the monitoring point. The monitoring point was expanded to a hypothetical model domain to investigate the thermal equilibrium reached at the monitoring point and consequently solve for the unknown groundwater flux.

Figure 2. Temperature module of the thermal equilibrium method showing (a) the twenty segment model domain, (b) upstream boundary conditions derived from stream water temperature (T_S) and flow monitoring, (c) atmospheric heat transfer parameters applied to each model segment at each time step, and (d) the predicted equilibrium water temperature (T_E). The magnitude of the groundwater flux (e) at a given temperature is calibrated to minimize the sum of squared errors between the measured stream temperature and predicted equilibrium temperature at the downstream boundary. Most variables are defined in equations (1) and (2). Note that β is the fraction of short wave radiation adsorbed at the water surface, z is depth, C_c is Bowen's coefficient, ε is the emissivity of water (0.97), and σ is the Stephan-Boltzman constant.

Figure 3. Daily averaged stream water temperatures time series compared to groundwater for 2015. Stream water temperature time series were monitored on Big Cedar USGS gauge, and groundwater temperature was estimated using air temperature retrieved from Talihina Mesonet Station 15 miles away with 1.5-month time lag as recommended by Pluhowski (1970). Solid and dotted lines represent fitted sine curves for stream water and groundwater temperatures, respectively. The vertical lines indicate intersections of the fitted curves where there is no estimated difference between the measured stream water and the estimated groundwater temperatures, and the thermal equilibrium method cannot estimate groundwater flux.

Figure 4. Study sites on the Kiamichi River (bottom left) and Ozark Highland Ecoregion (top left). Oklahoma Mesonet stations and USGS gages are represented by triangle and diamond markers, respectively. Cross markers indicate monitoring sites where stream water temperature data were collected and seepage runs were performed.

Figure 5. FITEVAL evaluation and regression results comparing groundwater fluxes estimated by seepage run and thermal equilibrium method. Plots showing (a) regression of seepage run and thermal equilibrium method groundwater flux estimates, (b) FITEVAL plot of cumulative probability of Nash-Sutcliffe Efficiency (NSE), with the median value indicating the reported NSE, (c) FITEVAL model diagnostic report including hypothesis test results, outliers, and the

sensitivity of the indicators to model bias, and (d) scatter plot showing fit between seepage run and thermal equilibrium method groundwater flux estimates in order of the series. Actual values are shown in Table 2.

Figure 6. Example continuous application of the thermal equilibrium method (TEM) for Buffalo Creek for five weeks from 19 December 2016 to 22 January 2017. The TEM application shown was at a weekly time scale. The groundwater flux estimated from the seepage run performed on 11 January 2017 is also shown.

Figure 7. Temperature of a hypothetical stream in the presence of cooler groundwater flux. (a) Stream water temperatures (T_S) remain at equilibrium at the presence of gradual changing groundwater flux, and (b) loss of thermal equilibrium due to changing groundwater flux. The thermal equilibrium method provides an accurate estimate of the groundwater flux for any point at thermal equilibrium. Estimates made with the thermal equilibrium method where $T_S \neq T_E$ will not represent an accurate flux.

Groundwater Flux Estimation in Streams: A Thermal Equilibrium Approach

Research Highlights

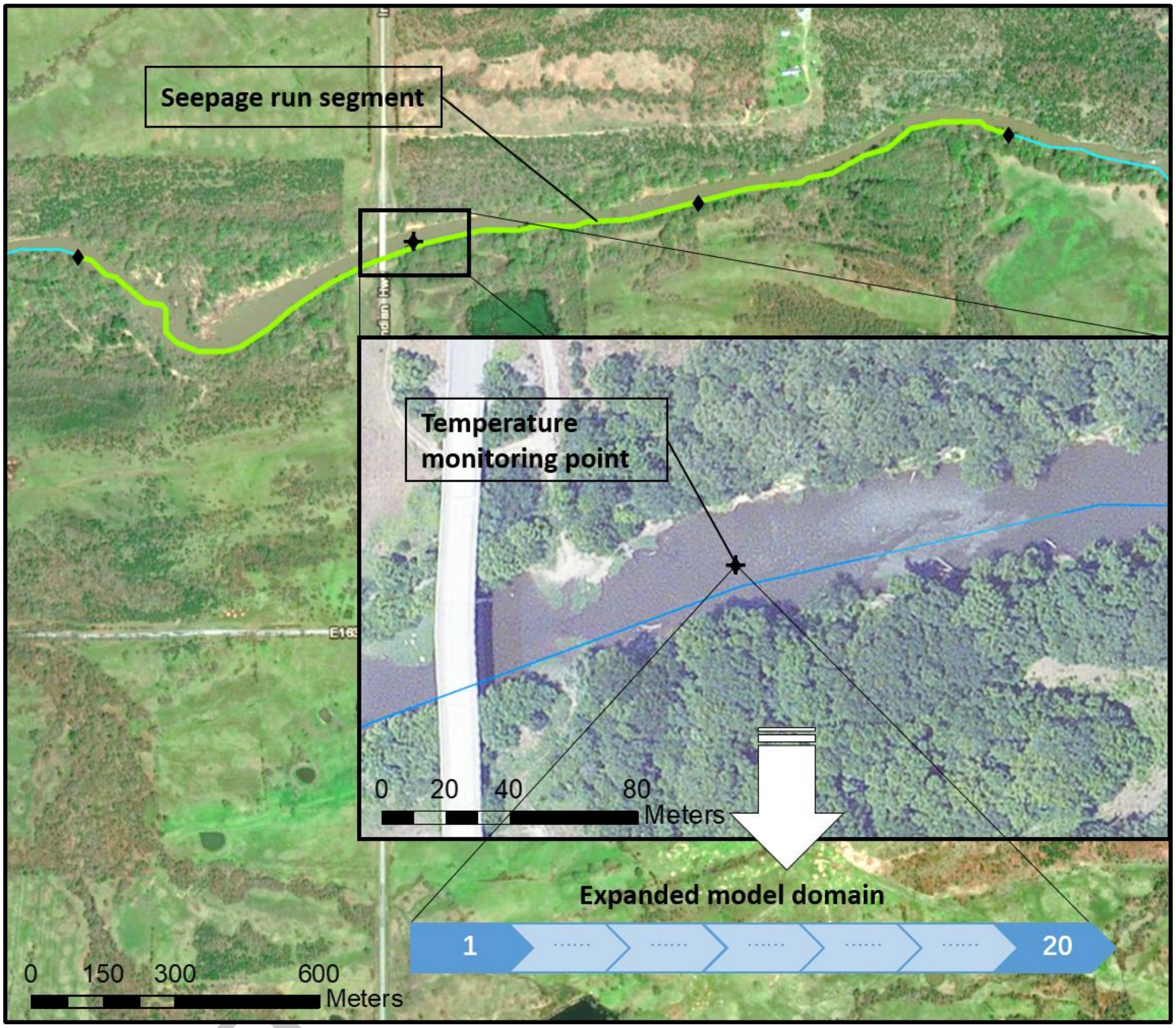
► A thermal equilibrium method was developed to quantify point groundwater flux.

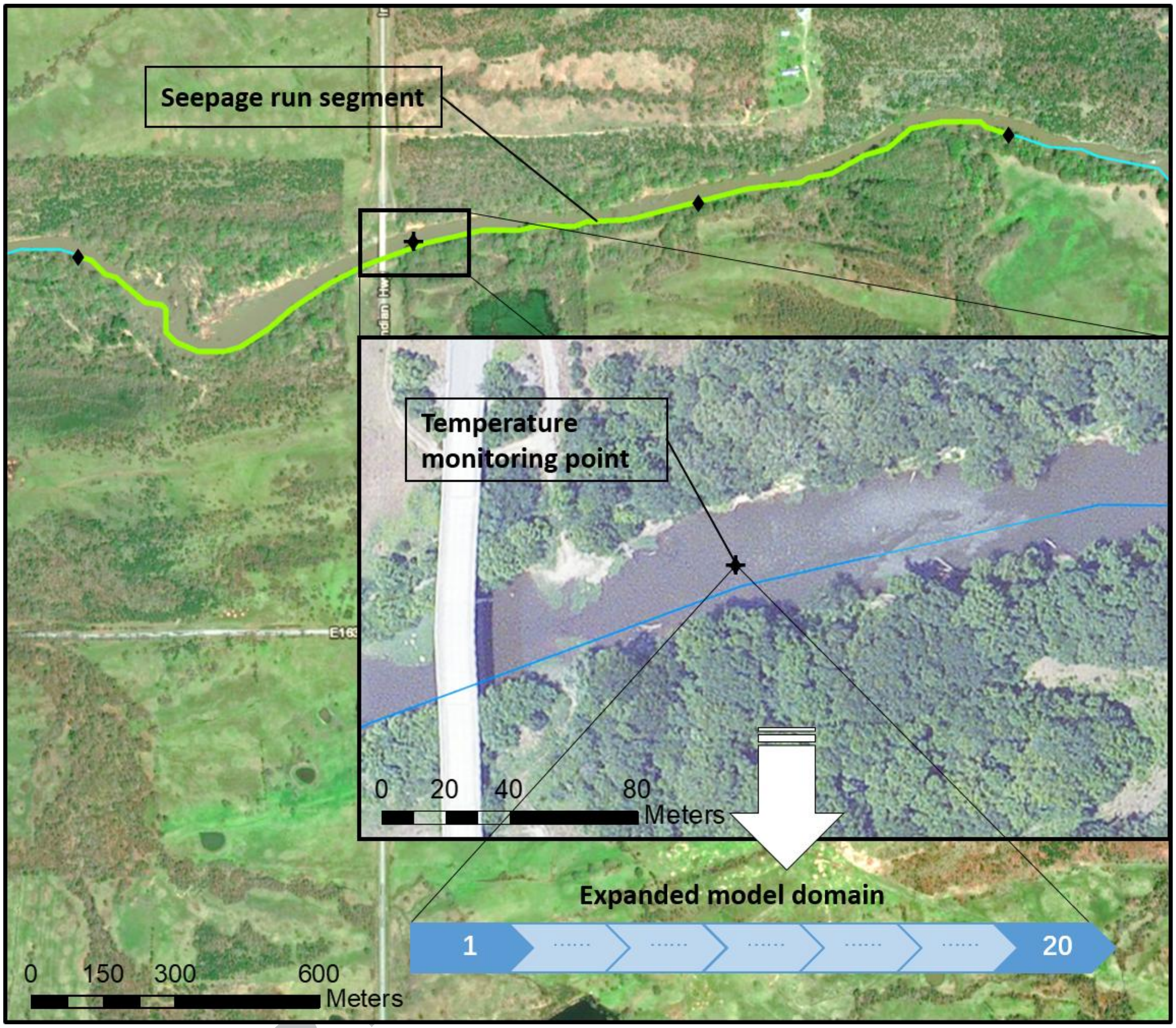
► The primary assumption was thermal equilibrium at the temperature monitoring point.

► The predictions matched well statistically with measurements from seepage runs.

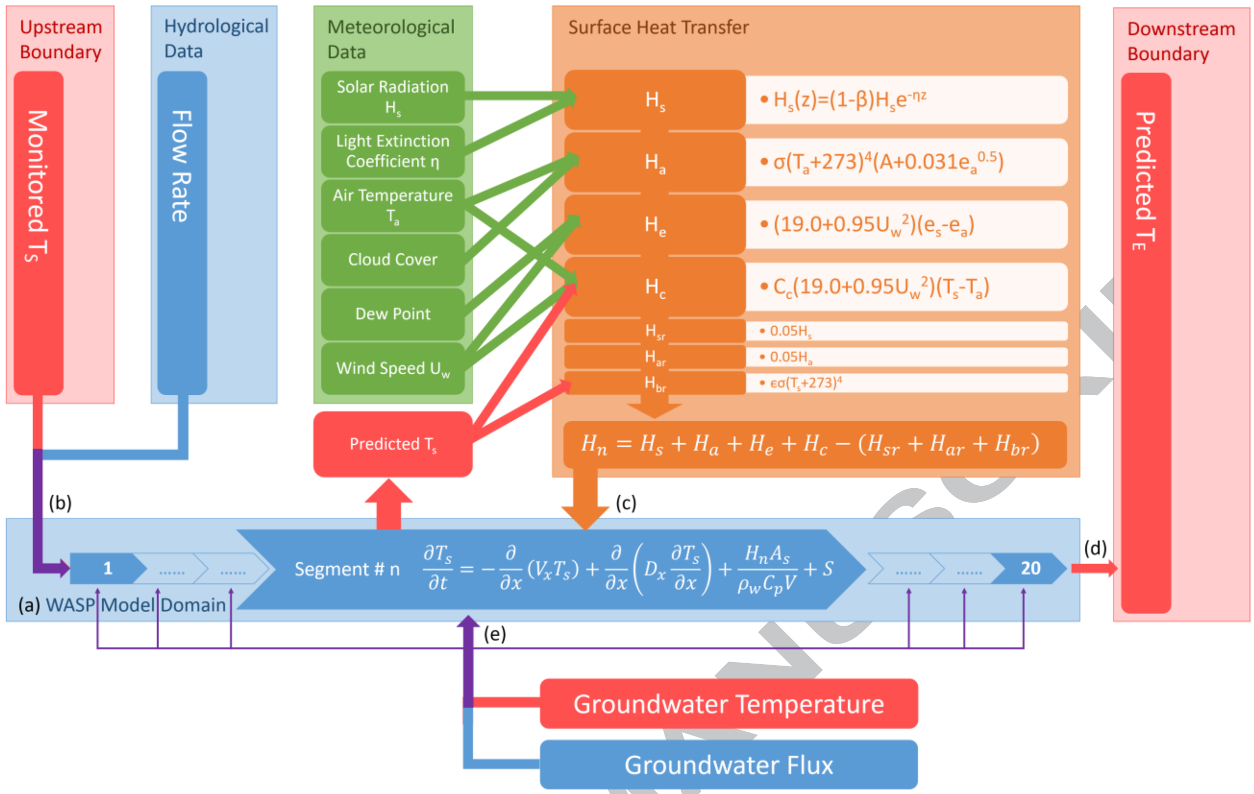
► The method requires a groundwater temperature signature to work effectively.

Graphical abstract

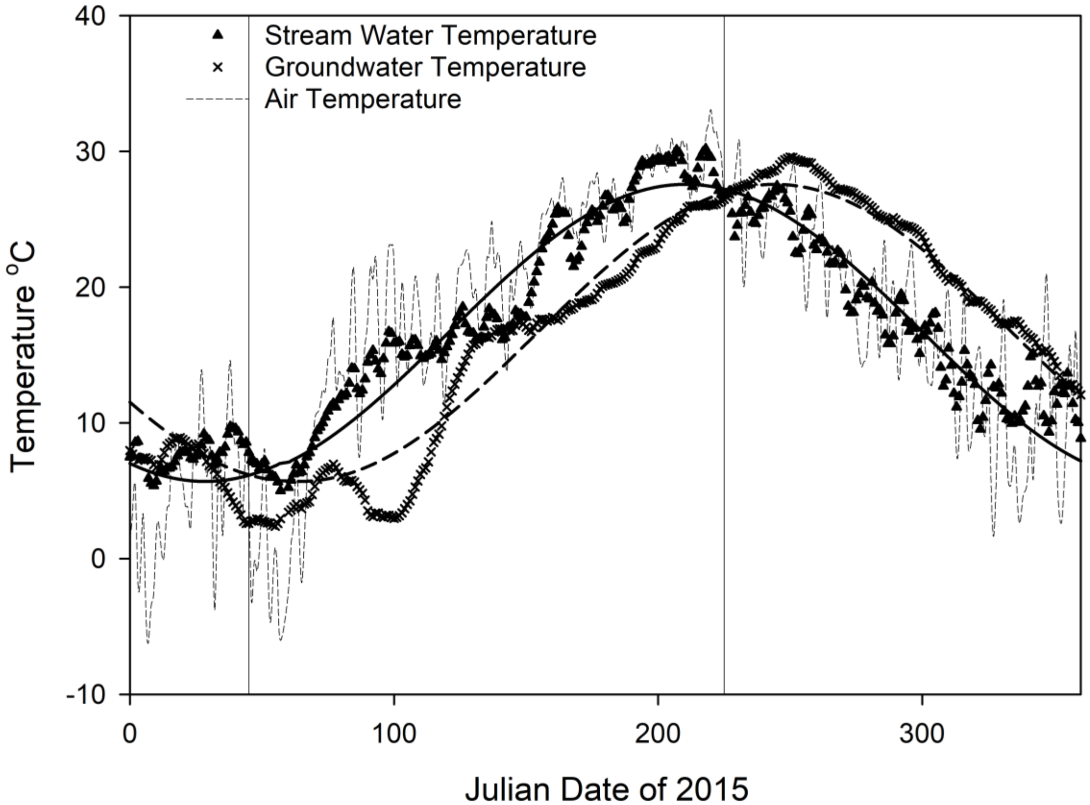




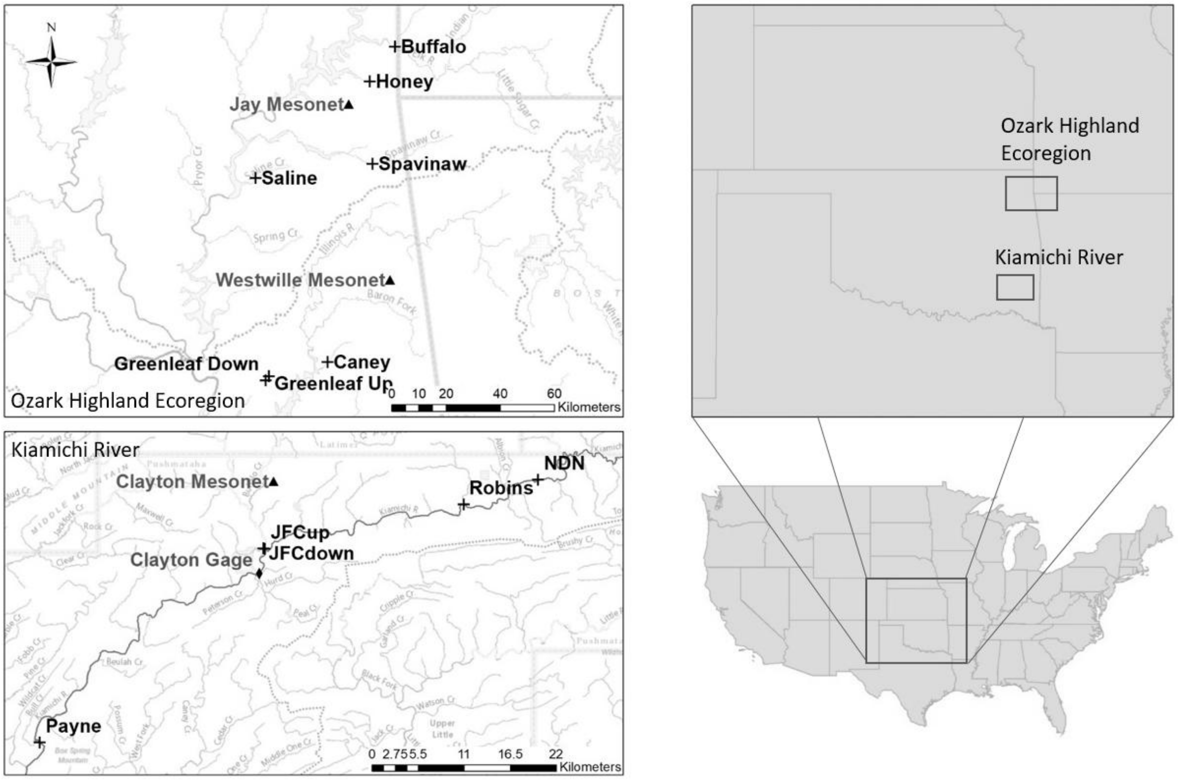
Seepage run segment
Temperature monitoring point
0 20 40 80
Meters
Expanded model domain
1
20
0 150 300 600
Meters

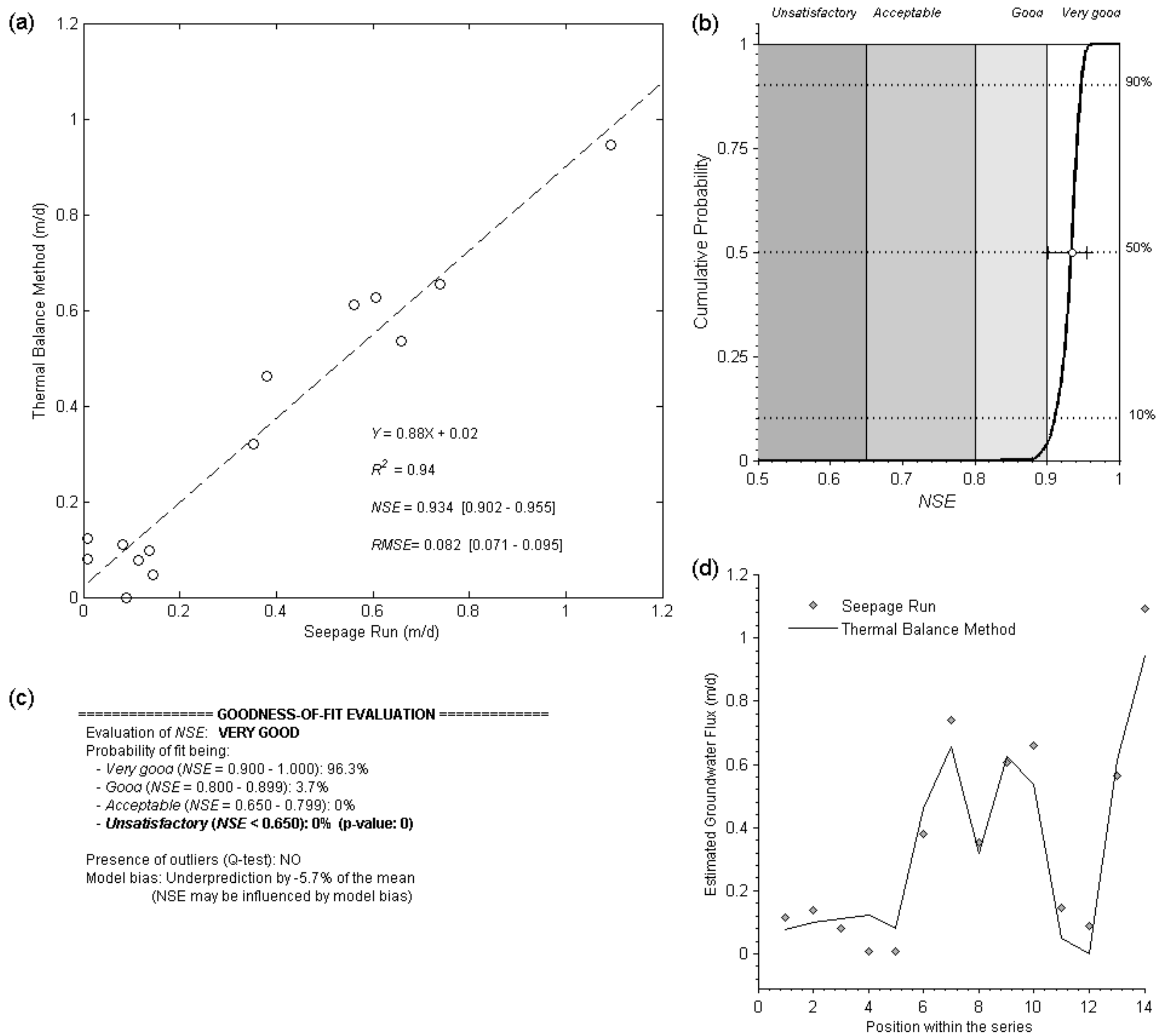


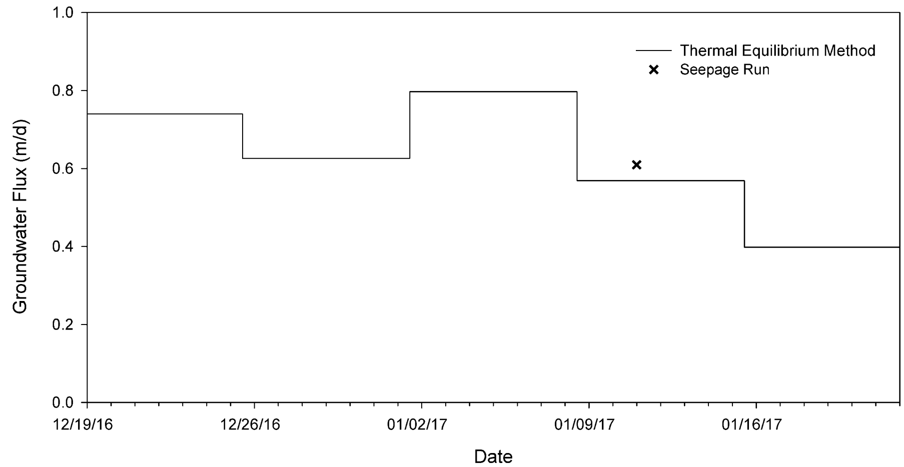
Upstream Boundary
Monitored Ts
Hydrological Data
Flow Rate
Meteorological Data
Solar Radiation Hs
Light Extinction Coefficient η
Air Temperature Ta
Cloud Cover
Dew Point
Wind Speed Uw
Predicted Ts
Surface Heat Transfer
Hs
• Hs(z)=(1-β)Hse-ηz
Ha
• σ(Ta+273)4(A+0.031ea0.5)
He
• (19.0+0.95Uw2)(es-ea)
Hc
• Cc(19.0+0.95Uw2)(Ts-Ta)
Hsr
• 0.05Hs
Har
• 0.05Ha
Hbr
• εσ(Ts+273)4
$H_n = H_s + H_a + H_e + H_c - (H_{sr} + H_{ar} + H_{br})$
Downstream Boundary
Predicted Ts
(b)
(c)
(d)
(e)
1
20
Segment # n
$\frac{\partial T_s}{\partial t} = -\frac{\partial}{\partial x}(V_x T_s) + \frac{\partial}{\partial x}\left(D_x \frac{\partial T_s}{\partial x}\right) + \frac{H_n A_s}{\rho_w C_p V} + S$
(a) WASP Model Domain
Groundwater Temperature
Groundwater Flux

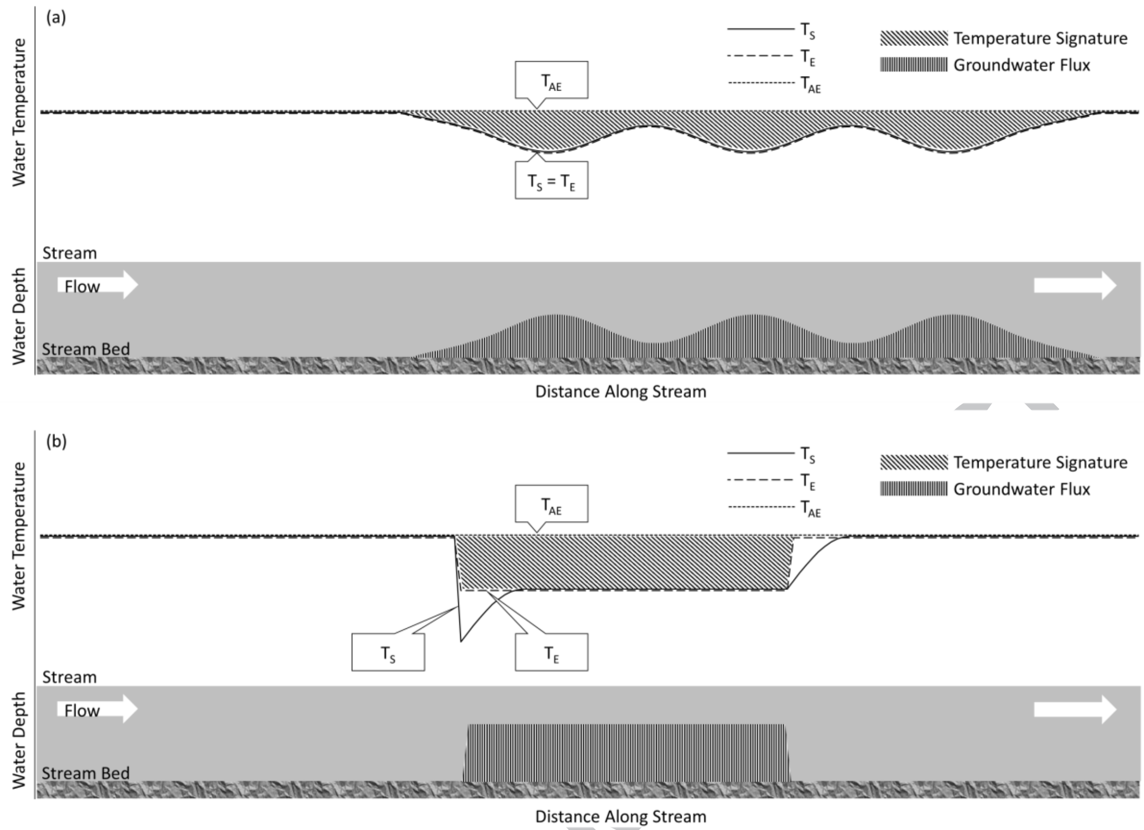


40
30
20
10
0
-10
Temperature °C
0
100
200
300
Julian Date of 2015
Stream Water Temperature
Groundwater Temperature
Air Temperature


Buffalo
Honey
Jay Mesonet
Saline
Spavinaw
Westwille Mesonet
Greenleaf Down
Caney
Greenleaf Up
Ozark Highland Ecoregion
0 10 20 40 60 Kilometers
Kiamichi River
Clayton Mesonet
NDN
Robins
JFCup
JFCdown
Clayton Gage
Payne
0 2.75 5.5 11 16.5 22 Kilometers
Ozark Highland Ecoregion
Kiamichi River


(a)
Thermal Balance Method (m/d)
Seepage Run (m/d)
Y = 0.88X + 0.02
R2 = 0.94
NSE = 0.934 [0.902 - 0.955]
RMSE= 0.082 [0.071 - 0.095]
(b)
Unsatisfactory
Acceptable
Good
Very good
Cumulative Probability
NSE
90%
50%
10%
(c)
================ GOODNESS-OF-FIT EVALUATION =============
Evaluation of NSE: VERY GOOD
Probability of fit being:
- Very good (NSE = 0.900 - 1.000): 96.3%
- Good (NSE = 0.800 - 0.899): 3.7%
- Acceptable (NSE = 0.650 - 0.799): 0%
- Unsatisfactory (NSE < 0.650): 0% (p-value: 0)
Presence of outliers (Q-test): NO
Model bias: Underprediction by -5.7% of the mean
(NSE may be influenced by model bias)
(d)
Seepage Run
Thermal Balance Method
Estimated Groundwater Flux (m/d)
Position within the series


Thermal Equilibrium Method
Seepage Run
Groundwater Flux (m/d)
1.0
0.8
0.6
0.4
0.2
0.0
12/19/16
12/26/16
01/02/17
01/09/17
01/16/17
Date


(a)
Water Temperature
T_S
T_E
T_{AE}
Temperature Signature
Groundwater Flux
T_{AE}
$T_S = T_E$
Stream
Water Depth
Flow
Stream Bed
Distance Along Stream
(b)
Water Temperature
T_S
T_E
T_{AE}
Temperature Signature
Groundwater Flux
T_{AE}
T_S
T_E
Stream
Water Depth
Flow
Stream Bed
Distance Along Stream