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AN ASSESSMENT OF THE APPLICABILITY OF THE HEAT PULSE METHOD TOWARD THE DETERMINATION OF INFILTRATION RATES IN KARST LOSING-STREAM REACHES

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Abstract: Quantifying the rate at which water infiltrates through sediment-choked losing stream reaches into underlying karstic systems is problematic, yet critically important. Using the one-dimensional heat pulse method, we determined the rate at which water infiltrated vertically downward through an estimated 600 m by 2 m sediment-choked losing-stream reach in the Devil's Icebox Karst System of Central Missouri. The infiltration rate ranged from 4.9×10^{-5} to 1.9×10^{-6} m s^{-1}, and the calculated discharge through the reach ranged from 5.8×10^{-2} to 2.3×10^{-3} m^3 s^{-1}. The heat pulse-derived discharges for the losing reach bracketed the median discharge measured at the outlet to the karst system. Our accuracy was in part affected by significant precipitation in the karst basin during the study period that contributed flow to the outlet from recharge areas other than the investigated losing reach. Additionally, the results could be improved by future studies that deal with identifying areas of infiltration in losing reaches and how that area varies in relation to changing flow conditions. However, the heat pulse method appears useful in providing reasonable estimates of the rate of infiltration and discharge of water through sediment-choked losing reaches.

Introduction

Stream reaches that lose water diffusely through their substrate into underlying aquifers are common in karstic basins (White, 1988). In Missouri, 389 of 502 watersheds (~77%) have losing stream reaches, many of which are choked with sediment (Jacobson and Gran, 1999). Losing stream reaches are often hydrologically complex and typically lose water through diffuse infiltration over a length of streambed, direct flow into swallets, or a combination of both. Losing stream reaches are common in both karst and non-karst terrains and in all cases play an important function in aquifer recharge and ground-water – surface-water interactions (Babcock and Cushing, 1942; Burkham, 1970; Silliman and Booth, 1993). A number of studies have been performed to understand the flow dynamics of losing- and gaining-stream reaches (e.g., Constantz et al., 1994; Silliman et al., 1995; Constantz and Thomas, 1996; Ronan et al., 1998). Gaging flow through losing-stream reaches is difficult for a number of reasons. In many sediment-choked reaches, loss occurs via diffuse seepage into the streambed rather than at a discrete point. Furthermore, zones of infiltration can shift longitudinally due to seasonal changes in the adjacent aquifer levels. In karst losing reaches, discharge regimes are typically flashy, and stream sediments are mobile and subject to scour and fill sequences during high flow events (Dogwiler and Wicks, 2004). These characteristics limit the effectiveness of permanently installed continuous monitoring stage-based gaging stations (Constantz and Thomas, 1996).

As such, there is a critical need for a method of understanding and monitoring the flow dynamics of losing reaches. Numerical modeling of ground-water flow and contaminant transport in these hydrologic environments is dependent upon the ability to accurately determine infiltration rates in losing reaches (Wang and Anderson, 1982; Silliman et al., 1995). Furthermore, bigger-picture hydrologic characterizations of karst systems require a detailed understanding of the fluxes and mass balances of water, solutes, and sediments through these systems (e.g., Silliman and Booth, 1993; Wicks and Engeln, 1997; Halihan et al., 1998; Dogwiler and Wicks, 2004; Lerch et al., 2005). These needs are even more pressing in rapidly urbanizing areas like the Devil's Icebox Karst System (DIB) of Central Missouri.

Methods

In this study, we have applied the heat pulse technique to a karst losing reach. Our study site, the Bonne Femme Creek Losing Reach, is a sediment-choked losing stream reach that recharges, in part, the DIB in Central Missouri (Figs. 1 and 2). Wicks (1997a, 1997b) and Lerch et al.

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Figure 1. (A) The Bonne Femme Creek losing reach during a low flow period and (B) approximately 24 hours after a significant storm event.

(2005) provide thorough hydrologic descriptions of the DIB basin. Bonne Femme Creek is the source of allogenic water to the cave stream that flows through DIB (Lerch et al., 2005). This portion of the losing reach is bounded on the upstream side by a 1 m deep pool that represents the limit of perennial surface-water flow, and on the downstream side by a large swallet. Except during extreme high-flow events, surface water is lost diffusely through the sediment substrate and no surface flow enters the swallet at the downstream end of the reach.

The heat pulse method provides a means of monitoring infiltration rates through losing-stream reaches using thermal variations in the stream substrate as a proxy for infiltration rates. The technique was developed and has been used in ephemeral streams in the arid southwestern U.S. (Constantz et al., 1994; Constantz and Thomas, 1996) and perennial streams in the humid Midwestern U.S. (Silliman and Booth, 1993; Silliman et al., 1995). The heat pulse method employs tracing thermal pulses, which occur in the surface water due to diurnal forcing or the rapid influx of thermally unequilibrated waters during storm events, through the recharge zone. Because suitable high-resolution temperature loggers are economical and readily available, the heat pulse method is an attractive approach for monitoring infiltration rates through losing-stream reaches. This study utilizes a portion of a dataset generated by Dogwiler (2002) and analyzed, for a purpose separate from this study, in Dogwiler and Wicks (2005). Here we have re-examined these data and applied the heat pulse method to them in order to determine infiltration rates through the Bonne Femme Losing Reach.

The temperature series data[1] were collected using Stowaway Tidbit temperature dataloggers positioned at different depths in the stream substrate at three stations in a portion of the ~600 m Bonne Femme Losing Reach from May 27, 2000 to July 10, 2000 (Fig. 2). Station 1 was located approximately 8 m downstream of the end of the perennial stream reach. Stations 2 and 3 (most downstream station) were spaced longitudinally downstream in 15–20 m intervals. The depths at which the dataloggers were placed are in the general range recommended by Silliman et al. (1995). In placing the temperature dataloggers, there is a trade-off between detecting the magnitude of the pulse, which decreases with depth, and the phase shift of the signal, which increases with depth (Silliman et al., 1995). Additionally, as depth increases, the assumption that flow is one-dimensional becomes more unreasonable. Many karst streams, including Bonne Femme Creek, have thin veneers of bed sediment. Therefore, relatively shallow placement of temperature sensors is both ideal and necessary. We positioned the temperature dataloggers at each station at depths of 1–3, 7–10, and 15–20 cm. The range for each depth represents both the difficulty of precisely burying the dataloggers in a streambed, as well as the actual thickness of the dataloggers. For the purpose of our calculations (described below), we have focused on the data from the upper and lower dataloggers at each station and quantified the depths as 2.0 cm and 17.5 cm, respectively. Each datalogger was programmed to record temperature at 15 minute intervals.

At each of the three stations, we identified heat pulses and used them to track the thermal fluxes through the recharge zone. The rate of thermal flux, which is assumed to be controlled by downward advection of the surface water, becomes a direct proxy for infiltration rate toward the underlying aquifer (Silliman et al., 1995). The heat pulse method involves measuring the lag time between a maximum (or minimum) surface temperature and a corresponding maximum (or minimum) subsurface temperature and the distance between the surface and subsurface monitoring locations (Taniguchi and Sharam, 1990). The velocity υ_t at which a heat pulse migrates downward through sediment is proportional to the rate q_i at which water infiltrates through that sediment according

[1] Available as a comma-delimited ASCII file on the *JCKS* website or through the NSS archives and library.

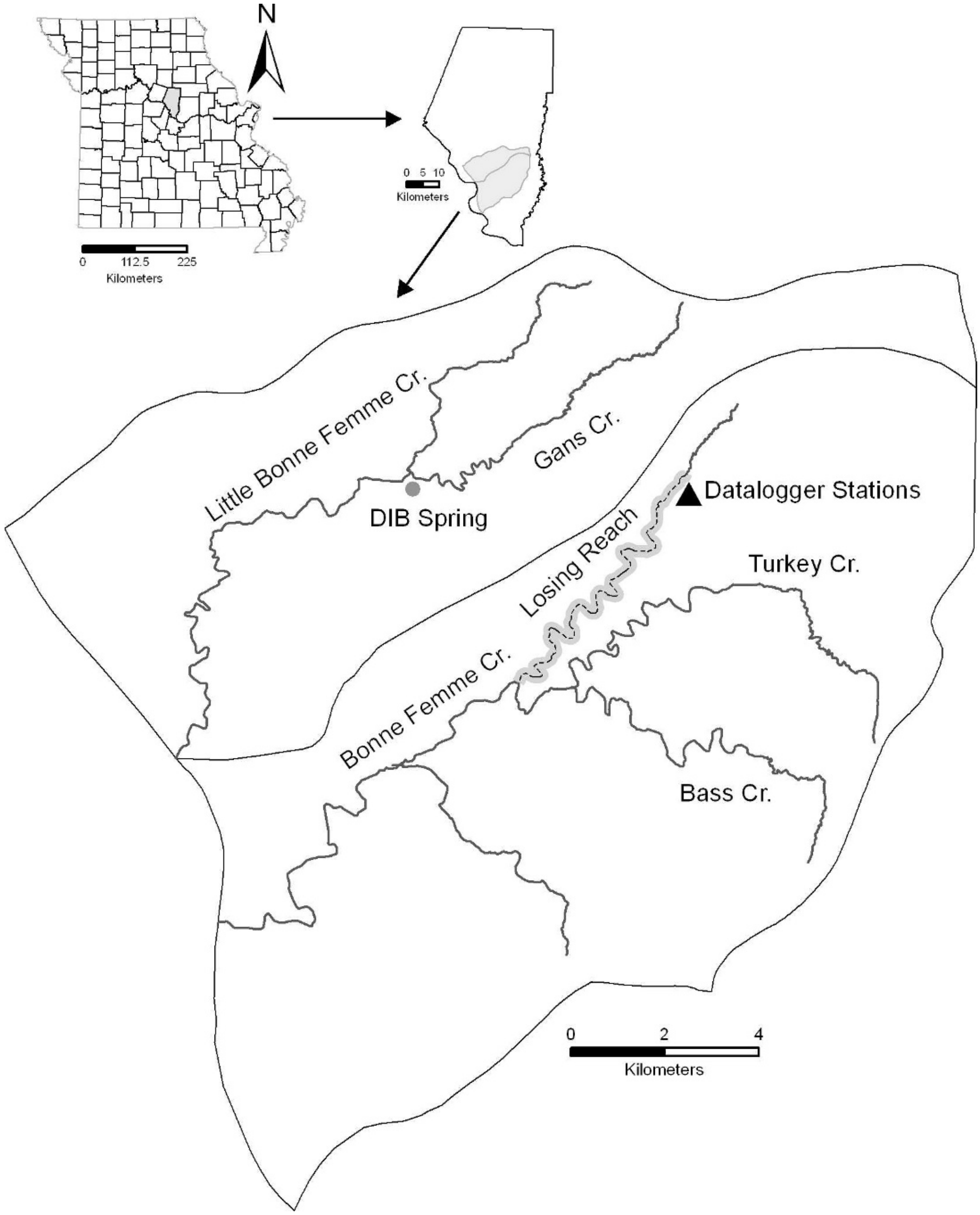

Figure 2. Location and hydrologic setting of the Devil's Icebox Karst System (DIB), including the Bonne Femme Creek Losing Reach (indicated by gray highlighting and dashed line). The total recharge area for the DIB is approximately 34.0 km^2 (Lerch et al., 2005). Water is pirated via the losing reach under the topographic drainage divide that separates the Bonne Femme Creek and Little Bonne Femme Creek. Ultimately, the water drained by the DIB re-emerges as surface flow at Connor's Cave Spring (indicated as DIB Spring on the map).

to

$$q_i = v_t \left(\frac{c_s \rho_s}{c_w \rho_w} \right) \tag{1}$$

where c_w and c_s are the specific heats of the water and wet sediment, respectively and ρ_w and ρ_s are the density of water and bulk density of the wet sediment, respectively (Constantz and Thomas, 1996). Thus, $c_s\ \rho_s$ and $c_w\ \rho_w$ are the volumetric heat capacities of the water and wet sediment, respectively (see Table 1 for values used for these constants).

Results and Discussion

For all depths and at all stations, diurnal variations were noted in the temperature data, although during some

Table 1. Constants used in heat pulse calculations (see Equation 1 in the text).

Parameter	Value	Units	Description
c_s	0.840	kJ kg^{-1} °K^{-1}	Specific heat of wet sediment
ρ_s	2,820.0	kg m^{-3}	Bulk density of wet sediment
$c_s\ \rho_s$	2,368.0	kJ m^{-3} °K^{-1}	Volumetric heat capacity of wet sediment
c_w	4.184	kJ kg^{-1} °K^{-1}	Specific heat of water
ρ_w	1,000.0	kg m^{-3}	Density of water
$c_w\ \rho_w$	4,184.0	kJ m^{-3} °K^{-1}	Volumetric heat capacity of water

time periods precipitation-induced flow temporarily reduced the variations (Fig. 3). Variations of up to ~7°C were common at a depth of 2.0 cm at all stations, whereas the maximum variation observed at 17.5 cm depth was ≤ 4°C. At the 2.0 cm depth, the maximum and minimum daily temperatures typically occurred between 3:00 and 6:00 p.m. CDT and between 6:00 and 9:00 a.m. CDT, respectively. The maximum and minimum daily temperatures at the 17.5 cm depth generally lagged approximately five to six hours behind those recorded at the 2.0 cm depth. In addition to the observed diurnal variation, there were weekly cooling (or warming) trends.

Using data from the 2.0 cm and 17.5 cm depths, 190 heat pulses were noted in the record. Equation (1) was applied to all 190 recognizable heat pulses, and Table 2 contains a summary of the descriptive statistics for the calculated υ_t and Table 3 contains the same analysis for q_i. The overall median heat pulse velocity υ_t was 1.1×10^{-5} m s^{-1} and the derived overall median infiltration rate q_i was 6.5×10^{-6} m s^{-1}. The rates of infiltration we found are comparable to rates determined in investigations utilizing the heat pulse method in other geologic settings (i.e., Constantz and Thomas, 1996; Constantz et al., 1994; Ronan et al., 1998; Silliman, 1993; Silliman, 1995), and therefore, appear reasonable.

An important assumption in the heat pulse method is that the rate of thermal flux υ_t is controlled by the downward advection of the surface water and is therefore a proxy for the infiltration rate q_i. The relative importance of conduction and convection in the substrate of the losing reach can be assessed based on the calculated Peclet number. Following the approach of Silliman et al. (1995), we calculated the range of Peclet numbers for the losing reach as 0.3 to 4.0 with a mean of 1.2. For infiltration rates greater than 3.7×10^{-6} m s^{-1}, the Peclet number is greater than 1.0 and advection dominates. At the lower end of our range of calculated infiltration rates, we push the assumption of dominantly advective transport. Of the 185 heat pulses we tracked, only 20 (11%) of the resultant q_i values are less than the 3.7×10^{-6} m s^{-1} threshold for advectively dominated transport. All of our median and mean q_i values are well above the threshold for advective transport (Table 4).

Based on the range of our derived q_i values, the volumetric discharge of water through the ~600 m by 2 m losing reach was between 2.3×10^{-3} and 5.8×10^{-2} m^3 s^{-1} with a median of 7.8×10^{-3} m^3 s^{-1} (Table 4). Over the same period, the median discharge from DIB (Fig. 3 and Table 4) was 4.2×10^{-2} m^3 s^{-1} (Lerch et al., 2005). Thus, the heat pulse-derived discharge range reasonably brackets the measured discharge at DIB. However, the median heat pulse-derived discharge through the losing reach is much

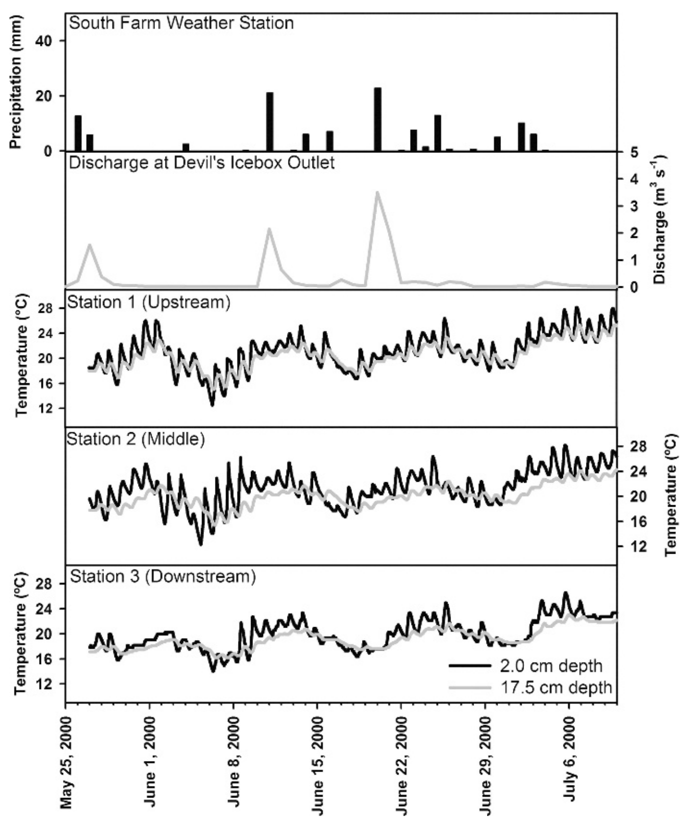

Figure 3. Graph of the time-series data utilized in this study. The uppermost plot shows 24-hour precipitation totals from the University of Missouri South Farm Weather Station located ~1 km north of the Bonne Femme Creek losing reach. The second plot from the top shows the discharge measured at the DIB spring (see Figure 2). The lower three plots depict water temperature at the 2.0 cm and 17.5 cm depths for the three datalogger stations. Note that Station 1 is the upstream-most station and Station 3 is furthest downstream.

Table 2. Calculated heat pulse velocities v_t in the Bonne Femme Creek Losing Reach.

Station	Median (m s^{-1})	Mean (m s^{-1})	Range Low (m s^{-1}) High (m s^{-1})	Stan. Dev. (m s^{-1})	No. of Pulses
1	1.6×10^{-5}	2.2×10^{-5}	$6.9 \times 10^{-6} - 8.6 \times 10^{-5}$	1.7×10^{-5}	73
2	1.0×10^{-5}	1.4×10^{-5}	$3.4 \times 10^{-6} - 8.6 \times 10^{-5}$	1.3×10^{-5}	62
3	8.4×10^{-6}	1.1×10^{-5}	$3.5 \times 10^{-6} - 3.4 \times 10^{-5}$	6.9×10^{-6}	55
Overall	1.1×10^{-5}	1.6×10^{-5}	$3.4 \times 10^{-6} - 8.6 \times 10^{-5}$	1.4×10^{-5}	190

smaller than the median measured discharge from DIB. The mismatch between the median calculated and measured discharges is likely due to uncertainty relating to the area through which infiltration occurs, rather than due to poor assumptions or parameterization of our infiltration calculations. As Figure 1 suggests, estimating the infiltration area of the losing reach is difficult. Future work is needed to define the area of streambed that serves as an infiltration boundary and to determine how that area varies as a function of flow conditions.

Under low flow conditions, the volumetric flux of water through the losing reach should equal the discharge from DIB (Lerch et al., 2005). However, a total of 124 mm of precipitation fell at the University of Missouri South Farm Weather Station on 17 rain days during the 47 day study period (Fig. 3). At least some of the difference between the median calculated losing-reach discharge and the median measured outlet discharge is attributable to diffuse recharge from elsewhere in the karst basin during precipitation events. We tested this hypothesis by filtering the DIB spring discharge data to isolate flow rates for days representing baseflow conditions. Two different filters were applied to the data. Based on the discussion of the discharge rating curve for the DIB spring in Lerch et al. (2005), we used their criteria for low flow (based on their Equation 1) and re-calculated the flow statistics for days during our study period with flows of 0.43 m^3 s^{-1} or less (Table 4). We also filtered the data by identifying all days with, or preceded by, 2.54 mm or more of precipitation and defining those as days with above baseflow discharges. Both filtering methods improved the statistical overlap of the calculated and measured values of discharge at the losing reach and spring, respectively (Table 4). Although the filtered data still range higher than the heat-pulse derived estimations of discharge, the ability of the method to bracket the median discharge values measured at the DIB spring is encouraging with respect to the applicability of the method to karst systems.

Table 3. Calculated heat pulse-derived infiltration rates q_i in the Bonne Femme Creek Losing Reach.

Station	Median (m s^{-1})	Mean (m s^{-1})	Range Low (m s^{-1}) High (m s^{-1})	Stan. Dev. (m s^{-1})	No. of Pulses
1	9.3×10^{-6}	1.2×10^{-5}	$3.9 \times 10^{-6} - 4.9 \times 10^{-5}$	9.7×10^{-6}	73
2	5.7×10^{-6}	7.8×10^{-6}	$1.9 \times 10^{-6} - 4.9 \times 10^{-5}$	7.5×10^{-6}	62
3	4.8×10^{-6}	6.0×10^{-6}	$2.0 \times 10^{-6} - 1.9 \times 10^{-5}$	3.9×10^{-6}	55
Overall	6.5×10^{-6}	9.1×10^{-6}	$1.9 \times 10^{-6} - 4.9 \times 10^{-5}$	8.1×10^{-6}	190

Table 4. Comparison of total discharge through the Bonne Femme Creek Losing Reach as calculated with the heat pulse method and actual measured discharges at the DIB spring. The discharge data for the outlet is also presented using two filtering techniques to isolate baseflow discharges from storm-induced discharges. Because the losing reach has been shown to contribute nearly all of the flow to the DIB during baseflow conditions (Lerch et al., 2005), the filtered data provide the best comparison with the heat pulse-derived discharges.

Measurement Method	Median Discharge (m^3 s^{-1})	Mean Discharge (m^3 s^{-1})	Range Low (m s^{-1}) High (m s^{-1})
Losing Reach			
Heat Pulse Method	7.8×10^{-3}	1.1×10^{-2}	$2.3 \times 10^{-3} - 5.8 \times 10^{-2}$
DIB spring			
All Discharge Data	4.2×10^{-2}	2.8×10^{-1}	$1.3 \times 10^{-2} - 3.5 \times 10^{0}$
Baseflow Data (Rating Curve Filter)	3.8×10^{-2}	7.6×10^{-2}	$1.3 \times 10^{-2} - 3.7 \times 10^{-1}$
Baseflow Data (Precipitation Filter)	2.3×10^{-2}	4.9×10^{-2}	$1.6 \times 10^{-2} - 1.6 \times 10^{-1}$

Summary and Conclusions

Because the hydrology of the DIB is well-understood from previous investigations, the Bonne Femme Losing Reach provides an excellent opportunity for assessing the utility of the heat pulse method. This study suggests that the heat pulse method can be applied to sediment-choked losing stream reaches in karst areas to determine infiltration rates and approximate discharges. We found reasonable agreement in the mass balance of calculated discharge input to the system through the losing reach and the measured discharge at the outlet of the karst basin. Future work should focus on defining the infiltration area within losing reaches and assessing how that area varies with changing flow conditions.

The ability to estimate infiltration rates is critical to parameterizing the boundary conditions in numerical flow models (Wang and Anderson, 1982). Thus, the heat pulse method will provide better data that could lead to more robust modeling of karst systems. Additionally, the heat pulse method offers an alternative approach to quantifying discharge in karst stream reaches where traditional gaging strategies have proven inadequate due to the dynamics of flow, sediment transport, and the difficulty of maintaining rating curves in these highly variably fluvial systems.

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