THERMAL CONTROLS ON HYPORHEIC FLUX AND STREAMBED FLOW DYNAMICS

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34 Pages

Heat is a naturally occurring and cost-effective tracer to study groundwater flow to, from, and throughout the subsurface. Often used for quantification of groundwater discharge, heat has been used to identify gaining and losing portions of streams and in determining flow parameters such as hydraulic conductivity (K) or velocity. Connecting ground and surface reservoirs is an area known as the hyporheic zone (HZ) where waters from both reservoirs interact. The flux of water throughout the HZ is controlled by stream bedforms, sinuosity, surface water velocity, local water table, seasonality, and sediment K. K is dependent on both the kinematic viscosity and density of water, and it is well established that temperature influences both variables. In most studies, these changes have been neglected because of the limited effect either has on K. However, these variations are important to understand because an increase in K will result in an increase in groundwater velocity, having implications relating to residence time and subsurface nutrient processing. To better understand how water temperature affects flow dynamics in the HZ, multiple two-dimensional models were created using the USGS software VS2DHI to map flow under both warm and cool thermal conditions. Data were collected from a series of varying temperature hydrologic flume trials where the effects of hyporheic flow altering variables like sinuosity, surface water velocity and volume, and bed-forms were controlled.

Results verify that K in the HZ will be greater under warm conditions and lower under cool conditions. Additionally, models indicate a faster speed of frontal movement under warm conditions than cool. Finally, the mapping of resultant Péclet numbers indicate a shallower advective extinction depth under cool conditions as opposed to warm. These variable thermal regimes provide much different conditions for flow amongst each other, and applying this, the significant differences in average seasonal water temperatures will introduce a spread of widely varying annual flow dynamics. Understanding these changes could help prepare us for future urban expansion, climate change, and other possibilities that could modify surface and ground water temperatures.

KEYWORDS: Hyporheic flux; Thermal tracing; Thermal impacts; Streambed flow dynamics

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J.W.R

CONTENTS

TABLES

FIGURES

CHAPTER I: INTRODUCTION

Heat is a useful tracer to study groundwater flow to, from, and throughout the subsurface. On top of already being a naturally occurring and cost-effective tracer, heat can be interpreted for tracking groundwater fluxes (Silliman et. al, 1995), delineating portions of gaining and losing streams (Silliman and Booth, 1993) and studying other parameters like hydraulic conductivity and flow (Lapham, 1989). To recall the development, Stallman brought attention to the use of subsurface temperature gradients as an indirect measurement of groundwater flow velocity and porosity from the use of partial differential equations (Stallman, 1963). Due to the low price and ease of temperature measurement, heat tracing to examine groundwater-surface water interactions resurfaced in the 1980's (Silliman and Booth, 1993).

Heat as a tracer has been employed primarily to quantify groundwater discharge and to identify areas of surface and groundwater interaction (Silliman and Booth, 1993; Harris and Peterson, 2020). Temperature data can be utilized for complex modeling of subsurface groundwater movement including three-dimensional velocity flow fields and alterations of flow redirected by in-stream structures (Menichino and Hester, 2014; Zlotnik and Tartakovsky, 2018). Two- and one-dimensional models of heat flow have been made increasingly user-friendly and accessible over recent years because of the release of free modeling software such as VS2DHI, which has the power to describe subsurface energy transport with a user-friendly and efficient GUI (Healy and Ronan, 1996).

The subsurface area directly beneath the stream hosting the mixture of upwelling groundwater and downwelling surface water is known as the hyporheic zone (HZ) (Conant, 2004; Harris and Peterson, 2020). The temperature gradient of the HZ is influenced by the individual temperatures of either end member (Bastola and Peterson, 2015). These differences in signatures form a gradient that is necessary for tracing the movement of either component (Stallman, 1963). This biologically active region is home to many processes that can be affected by the temperature gradient and it's contributing members (Lee and Rinne, 1980; Hester and Doyle, 2011). It is known that temperature effects the efficiency of denitrifying bacteria in the subsurface, where an increase in temperature results in greater efficiency (Hanaki et. al, 1990).

Direction and magnitude of hyporheic fluxes are controlled primarily by the hydrologic conditions of the streambed, and, regional water table forcing exchange via modifications to the local pressure field. Some streambed controls include hydraulic conductivity, sediment composition, channel morphology (bed-form, sinuosity), seasonality, and ultimately the local pressure field (Cardenas, 2009; Fox et. al, 2014; Singh et. al, 2018). For example, water table and channel morphology are significant drivers on the size of the hyporheic zone. Under losing conditions with little sinuosity and gaining conditions with high sinuosity, both the hyporheic zone and residence times are large (Cardenas, 2009). Under losing conditions with high sinuosity and gaining conditions with low sinuosity, both the hyporheic zone and residence times are small (Cardenas, 2009).

Bed-forms, such as pool and riffle sequences, help control the areas that water downwells and upwells. The sequences operate where the upflow face of the pool is dominant in surface water downwelling, and the downflow face of the pool is dominant in groundwater upwelling (Cardenas, 2009). Hydraulic conductivity, a function of the streambed medium and the density and viscosity of the fluid, impacts velocity and direction of hyporheic flux and can be defined as the ease in which fluid moves through the medium. Hyporheic exchange is unlikely or minimal in streambed mediums with low hydraulic conductivity.

Previous work studying the dynamism of substrate thermal responses to storm events concludes heat propagation within steam substrate differs seasonally, suggesting that temperature influences flow dynamics in the HZ (Bastola and Peterson, 2015; Oware and Peterson, 2020). While monitoring a low gradient, gaining stream, Oware and Peterson (2020) reported that during warm periods the thermal front in the HZ extends to 90 cm depth, while during cool periods in the same setting the front only extends to 60 cm. Often, the physical changes in viscosity and density of water from temperature changes are neglected because of the small effect viscosity and density has on hydraulic conductivity (Singh et. al, 2018); however, these changes are important to understand because an increase in hydraulic conductivity will result in an increase in groundwater velocity, which has implications relating to residence time and water volume, with implications on subsurface nutrient processing. With these observations in mind, the logical next step is to recreate a HZ environment where controls can be put in place to observe changes when the only variable is temperature.

Recognizing the impacts of temperature on the physical parameters of the HZ is significant because of its effects on biota and nutrient processing. Additionally, failing to integrate temperature variability may result in significant error in estimating hyporheic flux rates and residence times (Wu et. al, 2020). Using a similar method as described in Bastola and Peterson (2015), multiple two-dimensional inverse models were created using V2SDHI to determine hydraulic conductivity under both warm and cool thermal conditions. Experimental data were collected from a series of flume trials where variables that effect hyporheic flux (regional gradient, bedforms, etc.) were held constant except for water temperature. The implementation of consistent conditions using a flume allows for a detailed examination of temperature on flux, whereas high levels of scrutiny are required to analyze those specific controls on a highly variable in-situ stream (Cardenas, 2009). The experimental design provided an easy transition to a modeling software that uses a modified advection-dispersion equation to model fluid flow based on changes in temperature (Healy and Ronan, 1996). We hypothesize that flow in stream substrate will be more efficient (faster rate and greater depth) under warmer conditions and lower under cool conditions, warranting more stewardship in including the effects of temperature in modeling small scale environments. More specifically, how do HZ flow mechanisms differ under varied temperatures? Understanding these changes could help prepare us to mitigate the effects of future urban expansion, climate change, and other changes which could modify surface and ground water temperatures.

CHAPTER II: METHODS

2.1 Flume

The development of our understanding of heat flow as a proxy to groundwater flow, coupled with improvements in modeling techniques utilizing heat as a tracer, enables us to accurately analyze ground-surface water interactions (Constantz, 2008; Bastola and Peterson, 2015). In 2009, a series of flume trials were conducted within a hydraulic channel (flume) that measured 4.5m long x 0.32m wide x 0.4m deep (Fig. 1). To account for the controls on hyporheic flux and isolate temperature, the flume (Fig. 2) was filled with a homogeneous, well-sorted, very coarse-grained sand (roughly %80), with the remainder being coarse and medium grained sand. For each trial, slope, air temperature, and water volume were all kept constant. Slope was measured consistently at %0.5. Additionally, water input rate slightly varied, where the pumping rate for 4 of the trials examined was set to 8.5 L/s, and the pumping rate for 2 of the trials examined was set to 4.9 L/s. Water temperature within the system was the only significant changing variable. While natural streams experience temporal rearrangement of the streambed (Peterson et al. 2008), sediment redistribution of the flume was not incorporated into the experimental design. Along the flume, seven arrays, with a longitudinal spacing of 0.5 m starting at 1.0 m from the upstream end of the channel of four temperature loggers were installed. The loggers had a vertical spacing of 0.07 m starting from the bottom of the channel (Fig. 1). Temperature was recorded every 5 to 15 seconds (some trials varied) for the duration of the trial. The streamflow through the channel was maintained at constant velocity until the surface water temperature reached an equilibrium with room temperature, approximately 22 °C. Trials typically lasted about 24 hours. Subsurface temperature was not required to reach equilibrium to complete the trials; however, a uniform temperature throughout the system was required before a subsequent trial was initiated. A total of 34 trials were conducted alternating between warm and cool surface water inputs. To simulate streams under common mid-latitude conditions, for half of the trials, cool water and ice were added to the 950-liter reservoir to lower the water temperature to 17 °C (simulating spring or fall conditions). For the other half of the trials, warm water was added to raise the water temperature to 27 °C (simulating summer conditions). Once the reservoir water reached the required temperature, the trial was initiated by starting flow through the channel.

Table 1. Physical model parameters.

Flume Trial	*Slope (%)*	*Pump Rate (L/s)*	*Air Temp (°C)*	*Input Temp (°C)*
Cool 1	0.5	8.5	21.5	16.1
Cool 2	0.5	8.5	22.3	13.6
Cool 3	0.5	4.9	22.3	15.6
Warm 1	0.5	8.5	22.0	29.9
Warm 2	0.5	8.5	22.4	30.1
Warm 3	0.5	4.9	21.9	30.5

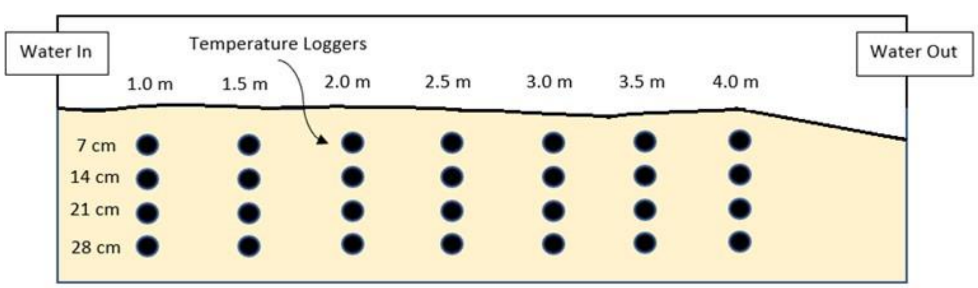

Figure 1. Conceptual model of hydraulic channel.

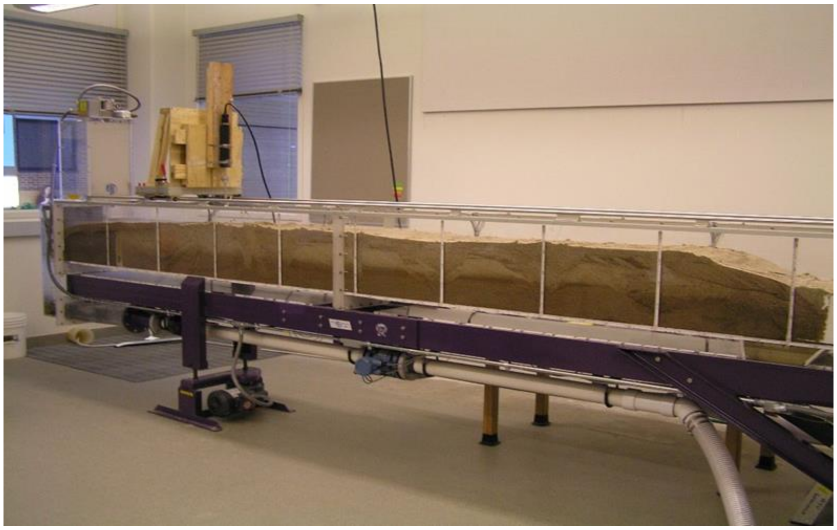

Figure 2. Photograph of the hydraulic flume used for data collection.

2.2 VS2DHI

A transient, 2-D, homogeneous model was setup to simulate the system in VS2DHI (Healy and Ronan, 1996), a public-domain thermal modeling software from the USGS. Model domain was ascertained from laser line measurements provided before each flume trial. Initial model conditions included temperature, initial equilibrium profile, head at each modeled boundary condition, and thermal and other physical transport properties of both the medium and fluid (Table 1). The time-step varied from 5 to 15 seconds depending on the resolution of the temperature loggers. Boundary conditions were set as follows: left, right, and bottom sides as no-flow no-flux boundaries, and the top of the domain was split into 9 equal length specified head and temperature boundaries which result in a head gradient that creates flow. Head was steady-state, but temperature varied with recharge period. In VS2DHI, recharge periods are considered timeframes of equal temperature. Excel was used to reduce surface water temperatures into those recharge periods. This was done by taking the average of the three present surface water probes at each time step, rounding each resulting average to the nearest 0.5 °C, and grouping like temperatures into recharge periods (depending on if temperature was on its rising or falling limb). Recharge periods were appended to each specified head boundary to specify input temperatures.

Observation points were added to the model at the same location temperature probes were installed in the flume to compare outputs during the calibration process. Model simulations were run to analyze differences in flow with different thermal conditions solved via the advection- dispersion equation (Eq. 1):

$$\frac{\partial}{\partial t}[\theta C_w + (1-n)C_s]T = \nabla \cdot K_t(\theta)\nabla T + \nabla \cdot \theta C_w D_h \nabla T - \nabla \cdot \theta C_w v T + q C_w T^{\circ} \quad (1)$$

where θ represents volumetric moisture content, C_w is heat capacity of water, C_s is heat capacity of dry solid, n is porosity, T is temperature, K_t is the thermal conductivity of the water and solid matrix, D_h is the coefficient of hydrodynamic dispersion, q is the rate of the source fluid, and T° is the temperature of the source fluid in °C (Healy and Ronan, 1996). Initial values for the parameters are provided in Table 2. VS2DH uses equation (1) to solve energy transport problems in variably saturated porous media, where the left side represents the energy stored within a volume over time, and the terms on the right side of the equation describe conduction, dispersion, advection, and heat sources or sinks, respectively (Healy and Ronan, 1996; Bastola and Peterson, 2015). In addition to the models based on flume data, an additional theoretical cold temperature model was created utilizing the same domain. This was to allow for a larger range of model temperatures to include in our interpretation.

Table 2: Initial model parameter values.

Parameter	*Value Range*
Hydraulic Conductivity (m s^{-1})	$9 \cdot 10^{-7} – 6 \cdot 10^{-3}$[a]
Heat Capacity (J m^{-3} K^{-1})	
Solid	$1.1 \cdot 10^{6} – 1.3 \cdot 10^{6}$[b]
Liquid	$4.2 \cdot 10^{6}$[b]
Saturated Solid	$2.5 \cdot 10^{6} – 3.2 \cdot 10^{6}$[b]
Porosity	0.30 – 0.50[c]
Thermal Conductivity (W m^{-1} K^{-1})	1.4 – 2.2[b]
Dispersivity	0.0005[b]

a Values sourced from Domenico and Schwartz (1990)
b Values sourced from Bastola and Peterson (2015)
c Values sourced from Earle (2015)

2.3 Calibration

To manually calibrate a model, it is best to start with data sets with layouts that can instantiate each other. Upon initial comparison, the layout of the flume data was completely different from the model output, where each row in the flume data set represented all temperature data at a specific time step, while each row in the model output represented temperature data for a specified temperature logger at a specific time step. This, along with other discrepancies between data sets were addressed by manipulating the dataset using R (more details can be found in appendix A).

Model calibration is a fine-tuning process that involves making small modifications to your model until it best represents your original data set. Since VS2DHI does not have a calibration function, R was employed to streamline the process. Calibration was based on the error between the measured temperature to the simulated temperature. The goal was to reach an RMSE of 0.7 ℃. Hydraulic conductivity, heat capacity of the solid and liquid, dispersivity, and saturated thermal conductivity were all used in tuning the model. In the end, hydraulic conductivity was the most sensitive parameter and calibration was based primarily off that value.

The approach was to utilize a ratio of modeled over experimental temperature under a variety of specified conditions to give insight on the model's shortcomings and to calculate a RMSE to place an overall statistical value on the model. The further the ratio was from zero, the worse the model's prediction. Above zero, the model over predicted temperature, and below zero, the model under predicted. Some of the specified conditions were node clusters at individual x locations throughout a model run as well as node clusters at individual y locations throughout a model run to locate more precise spatio-temporal shortcomings to better determine what in the model needed to be changed for better predictive power. This process was repeated until a sufficient RMSE (0.7 ℃) was reached for each model.

2.4 Data Processing

Expected hydraulic conductivity (K) was calculated based on water temperature throughout each trial using the following equation (Eq. 2):

$$K = \frac{k\rho g}{\mu} \tag{2}$$

where k is the permeability of the medium, ρ is fluid density, g is gravitational acceleration, and μ is fluid viscosity. Fluid density and viscosity are temperature dependent variables and are expected to change throughout time. The expected K value was then compared to the calibrated K value and other trials to analyze in what direction it deviates under warm or cool conditions.

Using the flume data as input, VS2DHI interpolates heat flux using a finite-difference model and returns the resultant Darcy flux throughout the model. Darcy flux was used along with the model parameters mentioned above to solve for Péclet numbers (Pe) at each node in the model. The Péclet number is a ratio of advective to conductive transport and provides insight about the movement of fluid in a system. Advection can be described as the transference of heat via the physical movement of water, and conductive can be described as the transference of heat via molecular spreading. The ratio is calculated as follows (Eq. 3):

$$Pe = \frac{q\Delta z\rho C}{2\lambda} \tag{3}$$

where q is the specific discharge at a cell in the model, Δz is the representative depth, ρ is fluid density, C is fluid specific heat, and λ is thermal conductivity of the medium (Deming, 2002). Average Darcy flux at each depth in the model domain was also compared between all trials to see its deviation with temperature and depth.

CHAPTER III: RESULTS

Results of the six computed models are reported in Table 3. Only 3 models of each temperature were created as they provided consistent results, and more models would not provide much more utility in answering our research questions. Results were recorded using data outputs and figures generated from VS2DH output files in both Microsoft Excel and R. We believe the RMSEs (0.33-0.65°C) for these models are reasonable as surface temperature changes are rapid and sheer upon slug input. Each model has an associated heat-map overlain with horizontal Darcy flux (Eq. 3) and average Péclet numbers. An additional theoretical cold trial was simulated to represent a temperature of 5°C to analyze changes at average local cold season stream temperatures, as experimental runs only highlighted slight changes in temperature.

The Darcy flux (q) of these models help describe how the speed of water varies under different thermal conditions (Fig. 3). The range of Darcy flux for cold trials was $1.3*10^{-5}$ - $4.3*10^{-5}$ m/s, mostly less than the range of flux for warm trials, $2.2*10^{-5}$ - $7.6*10^{-5}$ m/s, with some overlap. This overlap is expected as the temperature ranges are similar, and some similarities between the maximum q of cold trials and the minimum q of warm trials is likely. Our theoretical cold trial highlights how dramatic this change can potentially be over a wider range of temperatures with a Darcy flux of $9.3*10^{-6}$ m/s, nearly 4 times less than the average warm flux (Table 3).

Péclet numbers can be interpreted to determine whether advective or conductive transport dominates at a given location. The design of the flume experiment initiated advective transport, which would become more prominent with increasing velocity and conduction being more significant with decreasing velocity (Eq. 3). The advective thermal extinction depth is representative of the maximum depth of physical movement of the input water within the flume. Maps of calculated Péclet numbers over the model's domain indicate a distinct depth of advective extinction for each trial (Fig. 4). For this analysis, the depth of extinction was treated as the depth where the average Péclet number reaches 2. Despite a Péclet number of 1 being the logical extinction depth, a Péclet number of 2 was used as some of the models did not have a depth in which 1 was reached, and 2 represents a definitive advective signal. In cool runs, the average extinction depth was 21 cm, and the system lost advective signals between hyporheic flux sites where there was little interfacing between reservoirs (Fig. 4, a-c). The average extinction depth in warm trials was 25.6 cm (Fig. 4, d-e). The cold theoretical trial's extinction depth was interpreted as 13 cm, establishing a trend of decreasing extinction depths as input water temperature decreases (Fig. 5, f).

Comparing Darcy flux and Péclet ratios with their associated thermal regimes for each trial confirms the expected relationship between temperature and velocity (an increase under warm conditions, a decrease under cool conditions), and ties in predictable changes in extinction depth associated with temperature (an increase under warm conditions, a decrease under cool conditions). Flow can be interpreted as being parallel to Darcy Flux contours and concentrated at areas of higher Péclet number. While warm and cool trials follow similar flow paths, the depth and speed at which bulk flow propagates is greater under warm conditions than cool. It is also interesting to note the significant range of Péclet numbers found near-surface in warm trials, indicating more flux between surface and ground reservoirs under these conditions. Comparing the maximum warm and minimum theoretical cold q(s), we observe an increase in just over an order magnitude, with a reduction in viscosity of ~50% from cold to warm. Combined results of flux and Péclet ratios imply that warmer temperatures allow for enhanced subsurface flow due to a greater ease of movement resulting from a reduced viscosity.

Table 3: Model results of cool runs 1-3, warm runs 1-3, and the theoretical cold run (a).

Trial	*Ex. Depth (cm)*	*Darcy Flux (m/s)*	*Avg. T (HZ) (°C)*	*RMSE (°C)*	*Modeled K (m^3/s)*	*Calculated K (m^3/s)*
Cool 1	21	2.1*10-5	20.5	0.55	0.0009	0.0010
Cool 2	21	2.8*10-5	20.9	0.62	0.0018	0.0009
Cool 3	21	2.1*10-5	20.8	0.57	0.0018	0.0010
Warm 1	26	4.3*10-5	23.8	0.65	0.0022	0.0011
Warm 2	26	3.5*10-5	23.3	0.33	0.0020	0.0011
Warm 3	25	3.1*10-5	22.9	0.59	0.0020	0.0011
Cold a	13	9.3*10-6	8	-	0.0001	0.0007

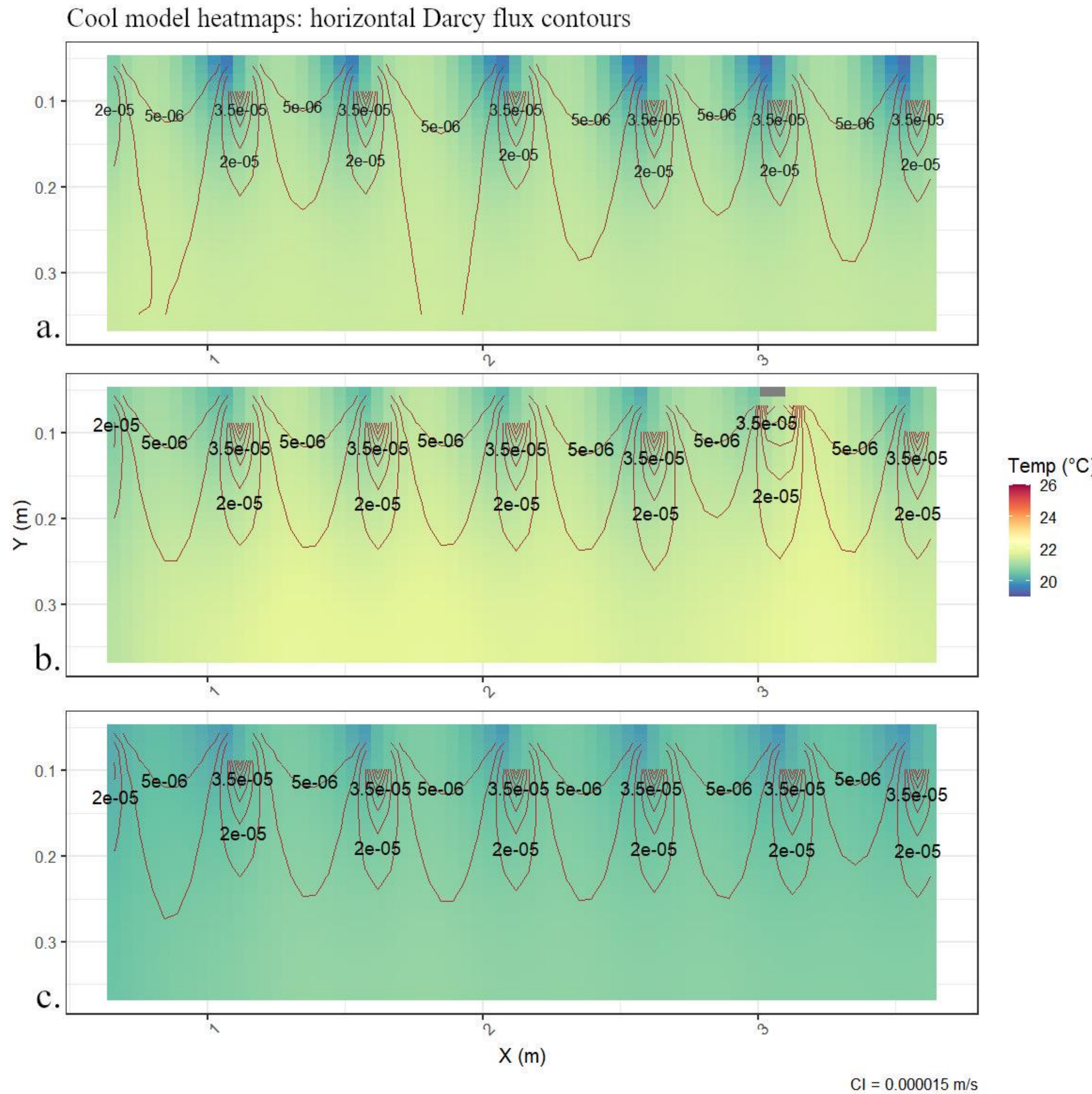

Figure 3. Temperature heat maps of cold runs (a-c) and hot runs (d-f) overlain with contours of horizontal Darcy flux with a contour interval of 0.000015 m/s. Notice areas of highest horizontal flux are associated with temperature signatures of downwelling zones. Additionally, the shape of the contours is driven by the modeling software's 'step-down' type gradient, the actual contours would be more laterally continuous (see: 4.1).

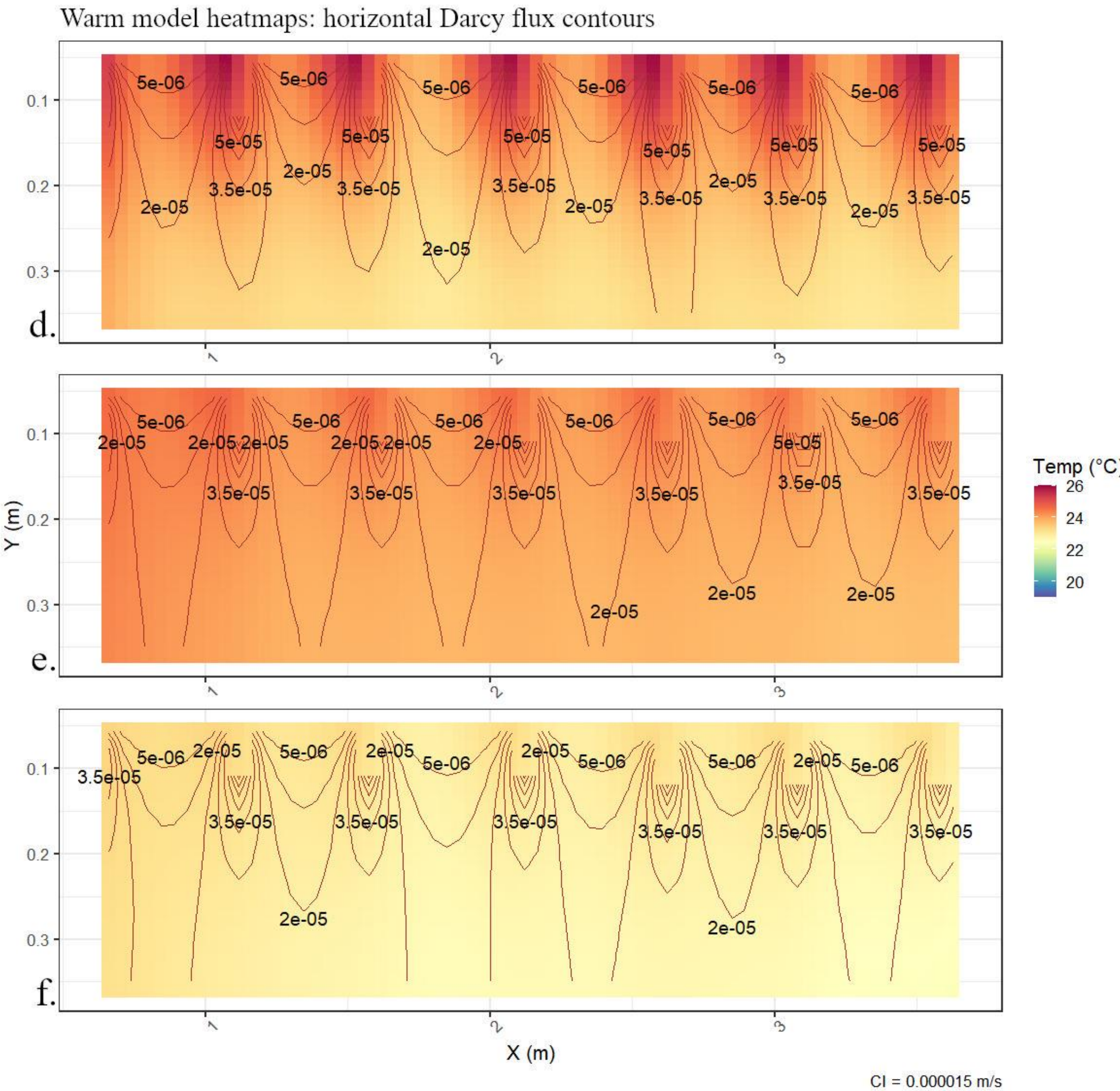

Figure 3. continued.

Cool model heatmaps: Péclet number contours

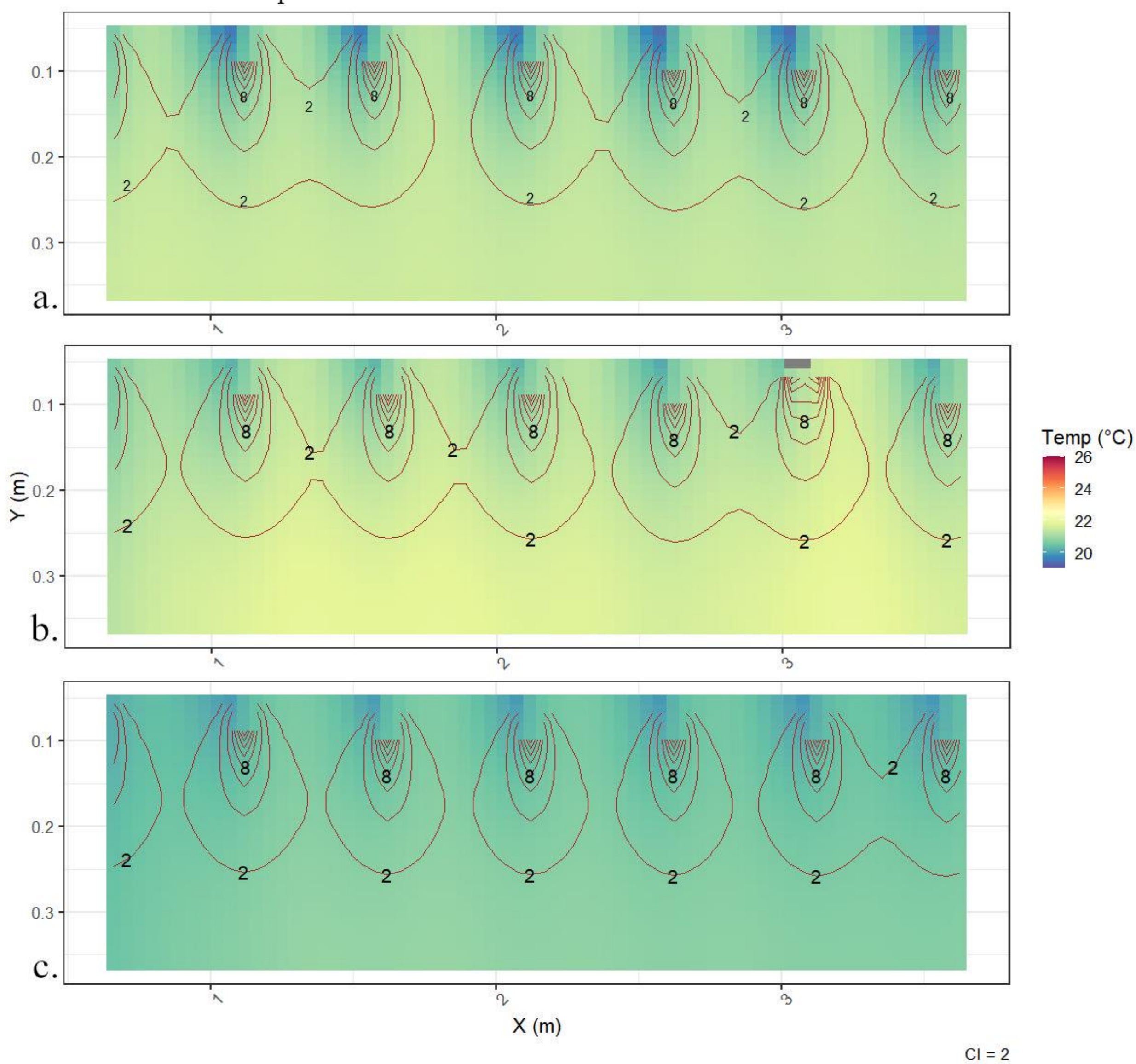

Figure 4. Temperature heat maps of cold runs (a-c) and hot runs (d-f) overlain with contours of Péclet numbers with a contour interval of 2. Notice areas of highest Péclet number are associated with temperature signatures of downwelling zones. Additionally, the shape of the contours is driven by the modeling software's 'step-down' type gradient, the actual contours would be more laterally continuous (see: 4.1).

Warm model heatmaps: Péclet number contours

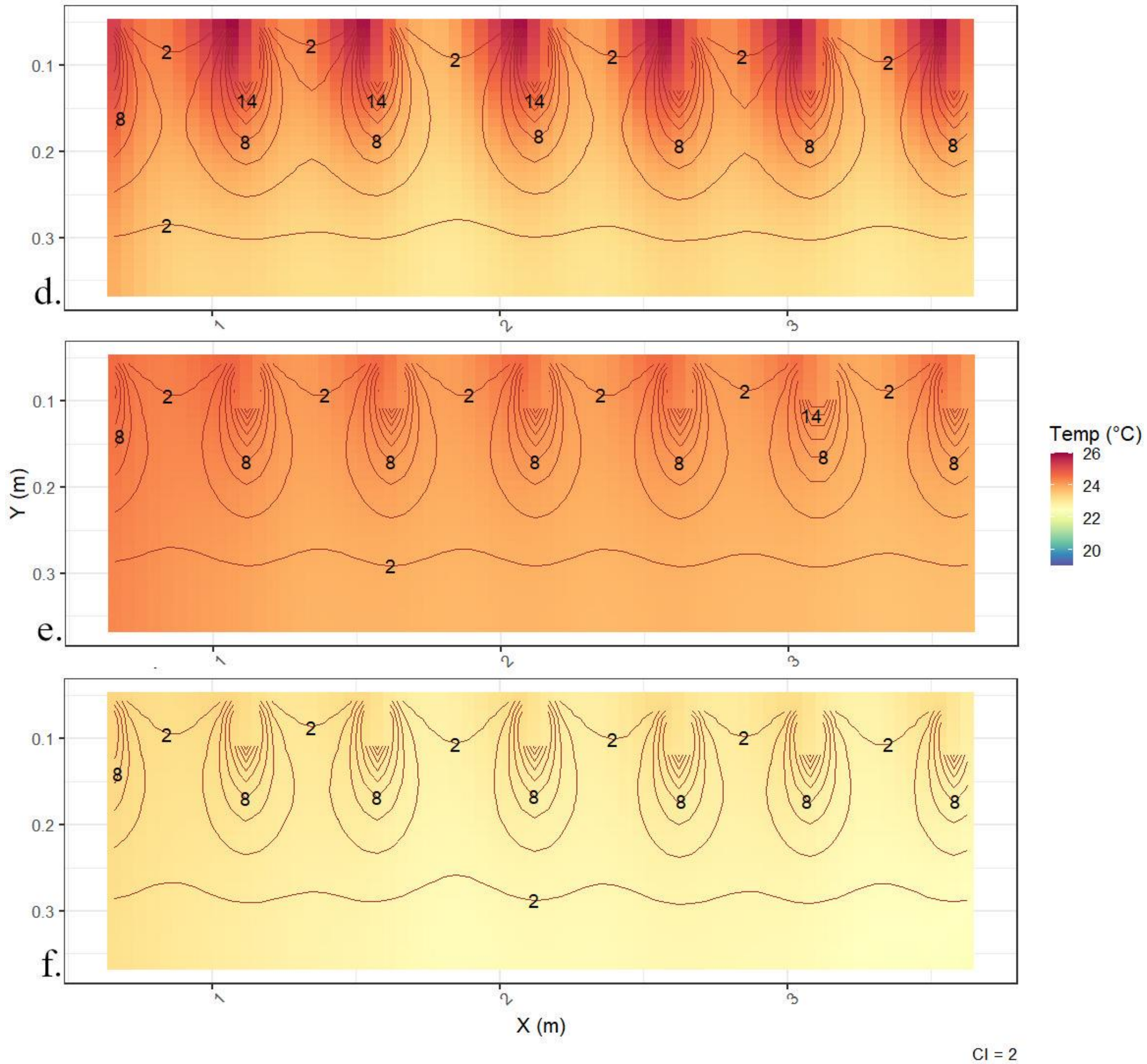

Figure 4. continued.

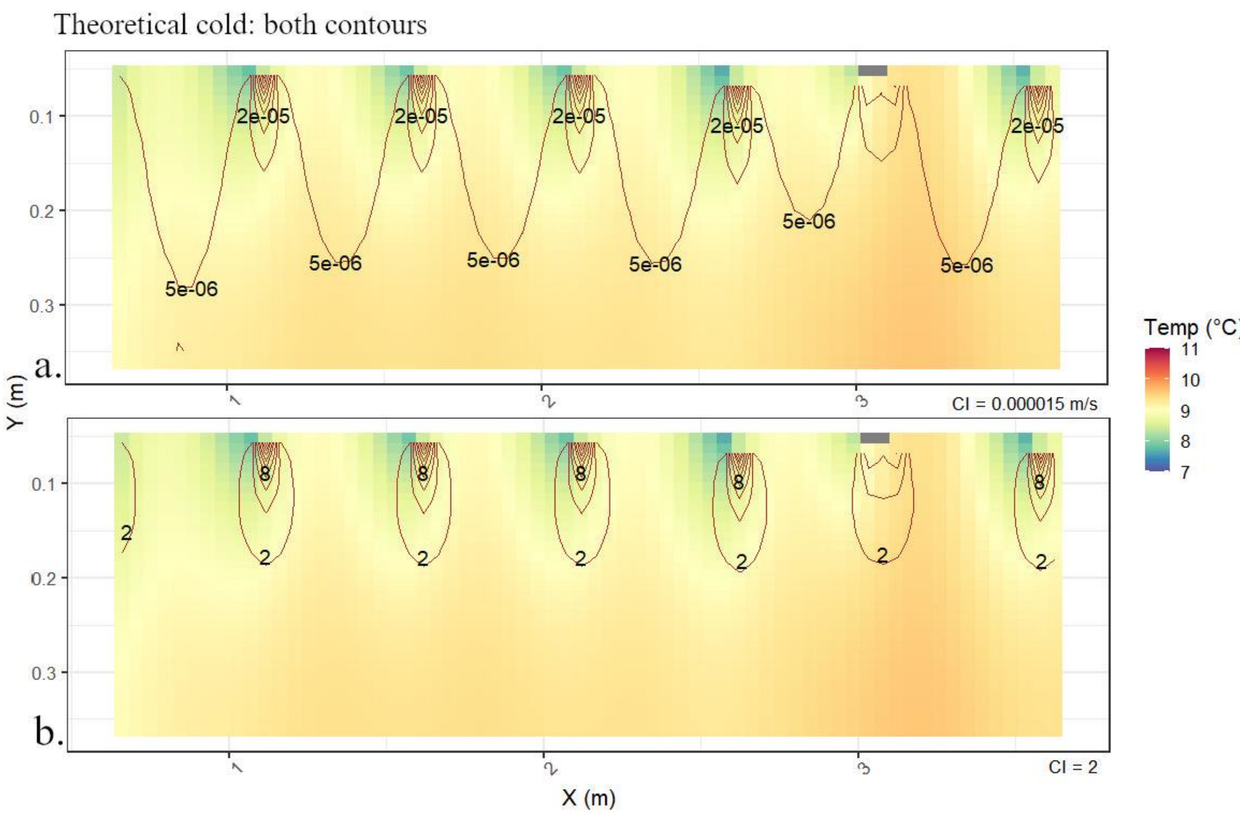

Figure 5. Temperature heatmaps of the theoretical cold trial with a starting ambient temperature of 10 degrees Celsius and input temperature of 5 degrees Celsius ran for 60,000 seconds. Figure 5a is overlain with contours of horizontal Darcy flux (CI = 0.000015 m/s). Figure 5b is overlain with contours of Péclet number (CI = 2). Additionally, the shape of the contours is driven by the modeling software's 'step-down' type gradient, the actual contours would be more laterally continuous (see: 4.1).

CHAPTER IV: DISCUSSION

4.1 On the Nature of our Model

When interpreting these results, it is critical to keep in mind the dynamics introduced by VS2DHI's limitations. VS2DHI does not allow for a traditional steady head gradient, instead, each boundary condition of the model is assigned a head, and the abrupt (albeit small) change in pressure head between boundaries drives downwelling in the system. In the actual flume, downwelling is likely forced along the upward sloping bedform on the far left and upwelling along the downward sloping bedform on the far right (Fig. 1). It is important to note the shape of the contours overlaying our heatmaps represents this aspect, and the contours realistically should be more laterally continuous. Despite these discontinuities, understanding what is controlling flux zones and the differences between our flume and the model gives a better understanding of the hydrologic modifications of temperature in this system. Since the primary driver of SWI (surface water interface) exchange is differences in pressure, we understand why this discrepancy arises, and our gap in understanding may be reduced (Singh et. al, 2018). Additionally, the model assumes its results hold consistent with a more regular groundwater temperature.

4.2 Comparisons and Applications

Comparing our results to Cardenas and Wilson (2007) confirms the hydrologic changes associated with thermal setting. Cardenas and Wilson studied the influence of bedforms using temperature as a tracer, those of which behave similarly to flux between our step-down type gradient. The study however fails to assess alone the impacts of temperature. Both results agree that fluid flux is proportional to water temperature (Cardenas and Wilson, 2007). Both also observe strong temperature variation in downwelling zones, and a return to ambient temperature in upwelling zones. Cardenas and Wilson conclude with the importance of sediment permeability, where temperature has little effect in low permeability mediums. This makes sense as in these settings SWI exchange will be limited, and the water that does exchange will primarily transfer heat via conduction and the slow rate of flux.

The effects of heat are also consistent with observations made in Oware and Peterson (2020), which studies variations in a stream's thermal response to storms during cold and warm periods. During warm periods, they observed a thermal response to storm events at a greater depth than cold periods and suggest greater advective control during warm periods (Oware and Peterson, 2020). The differences in dampening of thermal amplitudes with depth under either condition are consistent with our results and can help explain the rates of forcing observed within their study site. Similarly, Beach and Peterson (2013) examined diel and seasonal hyporheic thermal profiles in mid-latitude streams, where groundwater temperatures were warmer than surface waters in fall and cooler than surface waters in summer. They observed hyporheic temperatures more like groundwater during fall, and more like surface water during summer, indicating more influence of the groundwater component during fall. The greater thermal influence of groundwater in cold conditions suggest a shallower depth of flux than during warm conditions. This paper concludes, "The transmission of diel signals is limited by the efficiency of advection…" (Beach and Peterson, 2013, p. 65), coinciding with our results, and confirming the thermal impact.

Our results can be used to modify additional hyporheic controls, for example, changes in flux under gaining and losing conditions with the addition of temperature perturbations. This can be best explained by defining "hydrologic/thermal forcing" to make separations between stream-aquifer and stream-aquifer-HZ relationships. For this section, gaining and losing will refer to aquifer-to-stream relationships and forcing will refer to stream-aquifer-HZ relationships. Forcing can be considered as directional flow influenced by either upwelling groundwater or downwelling surface water. To clarify, gaining and losing can be considered an adjective, and forcing can be considered as a verb. Singh et al. (2018) relates forcing with storm events and observes an increase in forcing depth during peak flow conditions, like Oware and Peterson (2020). It can be interpreted that hot water resists upward forcing and supports downward forcing, and vice versa. With an already small HZ under gaining conditions and a larger HZ under losing conditions, temperature will modify this size with its relationship to forcing (Cardenas, 2009). Since hot water supports downward forcing, the surface component of a gaining stream under hot conditions may have a slightly larger HZ than expected, while a gaining stream under cold conditions may have a much smaller HZ than expected (Fig. 6). In a losing stream under cold conditions, the HZ may be slightly smaller, while under hot conditions the HZ may be much larger (Fig. 6). The differentiation in the degrees of change is a result of cool water reducing ease of flow and hot water increasing it.

Under cool conditions, it may be expected that the cool surface water would move at a slower rate than the warmer groundwater; however, the concept of forcing helps explain this. With downwelling, water is forced through pores from differences in the pressure field. This pressure difference plays a greater role in the movement of water than temperature, outweighing differences in flow rates solely caused by temperature. Remember, temperature is not a primary driver as other controls like regional water table, which determines if a stream is gaining or losing, greatly outweigh it, but still has an impact and may be an accessory to other controls associated with SWI exchanges by altering the fluid's density and viscosity.

The flume system was initialized at room temperature (~22°C) and had no groundwater flow prior to the start of the test, and there was none of the groundwater upwelling or downwelling that would be expected in natural streams. This is important to note, for different groundwater signatures affect the subsurface viscosity and pressure field, resulting in different flow dynamics and attenuating the effects of thermal forcing.

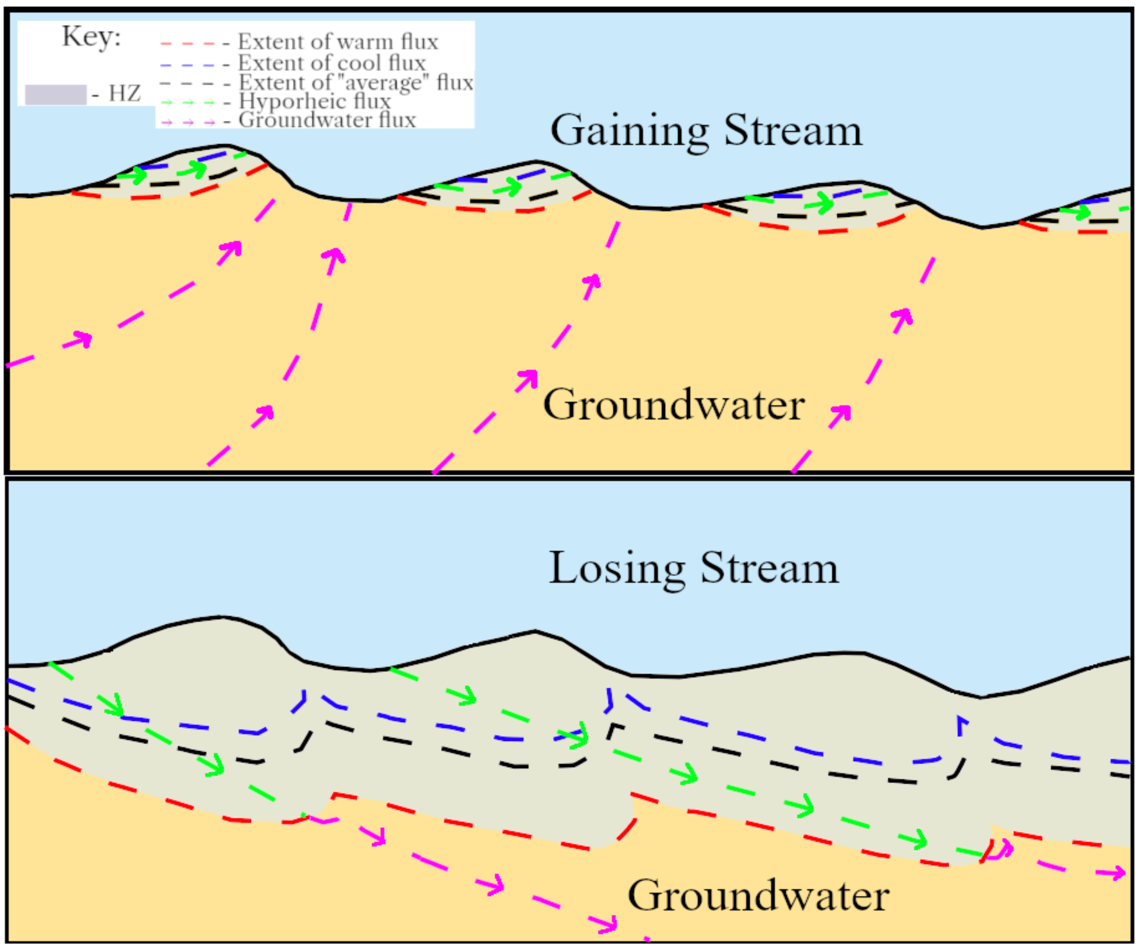

Figure 6. Conceptual model of alterations to HZ from temperature under gaining and losing conditions. The blue dashed line represents the theoretical extent of cool flux under either condition, and the red dashed line represents the theoretical extend of warm flux under either condition.

4.3 Implications for Seasonal and Diel fluctuations

Temporality is an essential aspect to consider while studying natural systems. While the surface water response is attenuated to atmospheric changes in temperature, the attenuation does not make subsurface temperature fluctuations by any means insignificant. Additionally, a continually decreasing diel amplitude of water temperature is observed with depth in the subsurface. However, this decrease is not constant and changes both seasonally and throughout the day in the top couple of meters of the subsurface. This variability may be due to changes in surface water temperature, which according to our results, has a reduced ability to infiltrate during cool conditions and an improved ability to infiltrate during warm conditions. The thermal gradient during cool periods (winter) may have a large slope, suggesting a rapid change in thermal amplitude with depth, and the thermal gradient during warm periods (summer) may have a more gradual change, suggesting a smaller change in thermal amplitude over the same depth. Observed changes in thermal amplitudes with depth are likely in-part due to changes in flow patterns associated with surface water temperatures. During periods when days are warming than nights, we may also see improved flow associated with an increase in temperature complimenting plant uptake and associated nutrient removal (Miller et. al 2018).

Harris and Peterson (2020) assess stage as a potential control of vertical hyporheic exchange. Like Oware and Peterson (2020), their raw data exhibit a deeper thermal response to storm events during winter than summer. Despite Harris and Peterson’s conclusion that summer exhibits a shallower hyporheic zone than during winter, storm responses in their data relay the opposite, where the downwelling associated with peak flow events extends to a greater depth during warm periods than cool periods, highlighting the influence of surface water temperature.

The discrepancy between baseline summer and winter flux (comparing Harris and Peterson to this project) is due to changes in regional groundwater gradient, which outweighs the controls of temperature on hyporheic flow depth and is a clear limitation of utilizing a flume.

4.4 Implications for Contaminant Transport and Biogeochemistry

In surface water reservoirs like agricultural streams or ponds, there are high nutrient levels due to tile drainage systems that serve as a direct conduit to streams for removing excess water from irrigated soils. Nitrates and other contaminants like phosphates in these tile waters may promote algal blooms in reservoirs both immediately effected by contaminated runoff as well as downstream which have the potential to cause hypoxia and suffocate wildlife.

A change from 1-25°C equates to a 50% decrease in viscosity, thus twice the hydraulic conductivity, twice the velocity water, and twice the transport of solutes. Generally, each 10°C increase results in twice the chemical reaction rate and efficiency of nutrient processing biota. Therefore, from that 25°C increase not only is there twice the transport of solutes, but an additional 5 times the rate and efficiency of processing those solutes with sufficient labile DOC (Peterson and Hayden, 2018). Additionally, in warm conditions we have a greater depth of downwelling thus greater residence time or more time for a complete reaction, and combined with an increased efficiency, create nutrient processing "hotspots" (Zheng and Cardenas, 2018). Under cold conditions, not only is there less solute transport and processing efficiency, but also a shallower downwelling depth and shorter residence time increasing the potential for an incomplete reaction. In the case of denitrification, an incomplete reaction can result in the release of ozone gas instead of molecular nitrogen (Briggs et al., 2015, Singh et al., 2018).

4.5 Future Work and Limitations

It would be beneficial to retest this with a more robust modeling software. VS2DHI determines head at boundary conditions, and thus cannot be made into a smooth gradient, but more of a "step-down" gradient. This induced downwelling at these "step-down" areas, and despite calibrating well, was not a perfect representation of the flume. The flume itself initialized at room temperature and did not represent groundwater conditions, so a full hyporheic thermal gradient could not be established. It would be helpful to return to this using a deeper flume to completely capture the extent of downwelling or finding a representative in-situ location. The model also does not account for movement perpendicular to flow, but in a real stream this may only be significant nearest stream banks, and our model may serve as a good representative for conditions below the thalweg. Additionally, applying the temperature concept to other controls of SWI exchange like stream slope or regional groundwater table may help in creating a fuller understanding on the association of temperature with hyporheic flux.

Two additional observations were made within the results that could be further explored. First, the calculated and modeled hydraulic conductivities varied. Modeled hydraulic conductivity was consistently higher than calculated hydraulic conductivity in 5 of the 7 models. This is likely because the 'g' component in eq. 2 is not the only force acting on water. In the case of the hyporheic zone, water is being 'forced' beneath the subsurface, and the downward force of downwelling water adds an additional unconsidered member to the formula. The second observation is found within models cool 3 and warm 3. These differ from the other 4 models as the pump velocity of the flume was 4.9 L/s instead of 8.5 L/s. Within these runs, cooler thermal maxima (warm 3) and minima (cool 3) are observed than the remaining models of their respective temperature. This connection may be related back to the forces of downwelling water, but further work must be done to better elucidate these connections.

CHAPTER V: CONCLUSION

Analyzing data from an experimental flume trial, which mitigated the effect of common in-situ flow drivers, proved effective in identifying the influence of the thermal regime on hyporheic flow dynamics. While some papers addressed this question, none to our knowledge have utilized an experimental system or addressed solely the influence of temperature (Cardenas and Wilson, 2007a; Wu et. al, 2020). Based on our results, we were able to interpret the following:

1. Warmer waters have increased kinematic viscosity thus improved efficiency of flow in both horizontal and vertical directions. A deeper hyporheic zone along with greater and more spatially spread advective flux was observed when temperature was increased in the flume.
2. Colder waters have reduced kinematic viscosity thus reduced efficiency of flow in both horizontal and vertical directions. A shallower hyporheic zone along with reduced and less specially spread advective flux was observed when temperature was decreased in the flume.
3. The depth of advective thermal transport is greater in warm runs than in cold runs.
4. A significant difference in flow exists between our warmest trial and the theoretical cold run (the temperature difference of which represents a typical yearly max and min) implying a definitive impact of thermal conditions on hyporheic forcing.

These interpretations were then applied to other theoretical models of hyporheic controls like the regional groundwater table and second were able to be observed within seasonal in-situ datasets to confirm our results (Beach and Peterson, 2013; Oware and Peterson 2020; Harris and Peterson

2020). Additionally, thermal influence needs to be further analyzed while coupled with other flux drivers to determine how they interact. An exact quantification of residence time would also be beneficial to understanding more potential flow changes, to examine if the deeper flow is extending residence time, as flow velocity increases simultaneously. Finally, the hydraulic conductivity formula utilized should be reconsidered as gravity is not the only downward force in the system.

With an increasingly warming climate thermal influence on processes need to be thoroughly analyzed. In the streambed for example, it is doubly important to understand this influence. Changes in temperature not only have an impact on flow (which alone can modify the size of the biochemically active region), but also changes in the transmission of heat itself have the potential to alter biogeochemical processes important to scrubbing harmful nutrients from stream reservoirs (Zheng and Cardenas, 2018; Wu et. al, 2020).

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APPENDIX A: REARANGEMENT OF DATA WITH R

First, the model output data were processed. Time (in seconds), temperature node (simplification for the name, and coordinates in 2-D space of observation points in the model), temperature, and Darcy velocity in both horizontal (X) and vertical (Y) directions were selected for processing. Each node cluster (all nodes at the same X location) was divided into individuals and labeled 1 through 4 depending on their Y location. Each node cluster was then labeled as a nest starting at 1 and increasing by 1, until all clusters in the data set were numbered (same clusters at new time steps were considered new clusters). This numbering is essential as it allows us to ensure that the model output data set can be coupled with the flume data set. Some models also included a 50,000 second spin-up period to equilibrate, which were also removed from the data set. The flume data set was imported and processed next. The first step was to remove all unnecessary rows and columns, followed by re-categorizing our time steps from date-time to seconds as is in the model output. Each row in the data set was then pivoted so each row was representative of a specific temperature node and not all nodes at a single time step. The data were then numbered with a similar sequence and nesting scheme as the model output, and joined based on the time, the nest groups, and the sequences. Resulting was a single data set comprised of rows of individual temperature nodes at a given time step with their associated temperatures from both flume and model data sets, x and y locations of those nodes, and x and y Darcy flux at the node. This new data set simplified further data processing and model calibration.