

UNDERSTANDING THE EFFECTS OF
FLOODPLAIN SHADE ON HYPORHEIC AND
STREAM CHANNEL TEMPERATURE CYCLES

by

Sarah Kathleen Fogg

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ABSTRACT

River reaches with coarse-grained alluvial floodplains have a breadth of lateral interaction between the channel and surrounding landscape, yielding extensive riparian zones and high rates of gross water exchange between the channel and substrate (i.e., hyporheic exchange). The lateral hyporheic zone on floodplain rivers is often near the ground surface, allowing for heat exchange between the atmosphere, unsaturated sediments, and hyporheic zone. We hypothesized that floodplain shade overlying lateral hyporheic water influences the conductive heat flux through unsaturated sediments, thus influencing hyporheic temperatures and temperatures in associated stream channels. We conducted simulation modeling experiments to test the potential effects of floodplain shade on hyporheic and stream channel temperatures. We found that scenarios with floodplain shade led to cooler hyporheic and stream temperatures than scenarios lacking floodplain shade under a variety of realistic floodplain conditions. We conclude that floodplain forest shade is a novel consideration for riparian management on floodplain river reaches and may be crucial in managing and maintaining cold-water habitat into the future.

INTRODUCTION

Temperature is a central driver of water quality and ecosystem processes in rivers and streams, with broad importance across multiple ecosystem scales [Caissie, 2006, Ouellet et al., 2020]. River temperatures vary across time and space, where persistent temporal cycles emerge daily, annually, and inter-annually [Dugdale et al., 2017]. Spatial patterns emerge across elevational gradients from headwaters to mouth, and across latitude, from equator to the poles [Pepin et al., 2022]. Human alterations to fluvial landscapes and global climate patterns have altered the spatial and temporal patterns of river temperature, thus affecting stream ecology and channel water quality [Caissie, 2006].

The term “thermal regime” refers to the temporal pattern of stream channel temperature at a particular location [Caissie, 2006]. On the daily scale, in the absence of precipitation or substantial cloud cover, minimum temperature occurs around sunrise, and maximum temperature occurs in the afternoon, creating a cyclical daily pattern reminiscent of a sine wave with a mean, amplitude, and phase. This cyclical, sine-like pattern also occurs seasonally. In the Northern Hemisphere, daily temperature means and amplitudes peak in late summer and are at a minimum in winter. Inter-annual trends in river temperature are associated with broader-scale climatic patterns such as the Pacific Decadal Oscillation (PDO) and the El-Nino Southern Oscillation (ENSO). These climatic patterns directly influence snow-pack and stream-flow [Pepin et al., 2022], which impacts water availability and river temperatures [e.g., Mote et al., 2003].

Broad spatial patterns in river temperature occur across latitudinal and elevational gradients [Isaak et al., 2017]. The warmest rivers occur near the equator and rivers become progressively cooler as latitudes increase toward the poles. Generally, rivers are cool at their

headwaters and temperatures follow a warming trend with decreasing elevation until they reach the ocean [Pepin et al., 2022]. However, these larger-scale patterns are more nuanced at smaller spatial scales. Reaches of cooler water can occur downstream of warmer reaches, given variations in local-scale landscapes [Ward and Stanford, 1983]. Such variation can be caused by a multitude of factors, such as localized groundwater inflows, a change in vegetative shade, or the release of cool water from a reservoir, among other causes. Temperature variation within a river reach occurs most often in highly heterogeneous fluvial landscapes, such as gravel-bedded floodplain rivers where pools, side channels, and upwelling hyporheic water may provide an array of temperatures different from the main channel [Hauer et al., 2016, Pander et al., 2024].

Over the past century, warming climates and changes to fluvial landscapes have been associated with a warming trend in river temperatures [Isaak et al., 2012, Kaushal et al., 2010]. Warming of river temperature regimes are predicted to continue into the next century [Liu et al., 2020]. Warmer river temperatures reduce oxygen saturation of water [Mackay and Fleming, 1969] and alter rates of nutrient cycling [Cross et al., 2015], decomposition [Ferreira et al., 2020], and pollutant processing [Noyes et al., 2009]. Temperature determines the metabolic rates of poikilothermic organisms [Sullivan et al., 2021]. Poikilothermic species—those which behaviorally regulate body temperature—are ubiquitous in freshwater ecosystems, thus temperature influences broad patterns in species distributions [Barbarossa et al., 2021]. Poikilotherms have optimal temperature ranges in which they live, adapted to either warm- or cold-water thermal regimes. Excessive or prolonged exposure to temperatures outside an organism’s optimal thermal range can cause stress and mortality [Bear et al., 2007, Selong et al., 2001]. Fish-kill events have been linked to elevated stream temperatures and concomitant lack of dissolved oxygen [Levy, 2003, Till et al., 2019]. As river temperatures warm, species distributions are expected to shift, producing habitats that may no longer be able to support populations of cold-water species

[Barbarossa et al., 2021, Isaak et al., 2012].

Documented declines in native cold-water salmonid populations have been associated with rising river temperatures across the intermountain western U.S. [Mantua et al., 2010, Mote et al., 2003]. In this region—spanning from the Pacific coastal ranges, in the west, to the Rocky Mountain front, in the east—salmonids are important economically, ecologically, and culturally. The average economic value of recreational fishing across Colorado, Idaho, Montana, Nevada, Utah, Oregon, Washington, and Wyoming was approximately \$458 million in 2016 [Miller et al., 2022]. Salmonids influence the stream and terrestrial ecosystem by transporting nutrients upstream (derived from the ocean in anadromous populations), engineering the ecosystem via spawning redds, and are an integral part of the stream food web [Bellmore et al., 2013, Walsh et al., 2020].

Populations of salmon in the western U.S. have been important both nutritionally and culturally for Native American peoples for at least 7,500 years [Campbell and Butler, 2010, Lynn et al., 2013]. For example, the salmon was the first animal to promise to take care of the peoples of the Confederated Tribes of the Umatilla Indian Reservation (CTUIR) in their creation belief [Jones et al., 2008a, Quaempts et al., 2018]. Ritualistic serving of “First Foods” in the CTUIR longhouses (community meeting places) begins with water and salmon which serves as a reminder to the people of their responsibility to respect and reciprocate care of these resources [Quaempts et al., 2018]. The First Foods mission of the CTUIR includes restoring native salmonid habitat and funding research that informs restoration strategies. The river temperature research presented within this dissertation has been funded by the CTUIR.

Historical fishing grounds of the CTUIR include the upper Columbia River basin, across the current U.S. states of Washington, Oregon, and Idaho, and is now concentrated near current reservation lands in the Umatilla River subbasin in Oregon [Quaempts et al., 2018]. Multiple dams and habitat degradations in the Columbia River Basin have led to salmonid

population declines and extirpation from their historic ranges [Kareiva et al., 2000, Lynn et al., 2013, Rieman and Isaak, 2010]. The population declines of Middle Columbia River steelhead and Columbia River bull trout have led to their listing as threatened or endangered under the Endangered Species Act [CTUIR, 2012]. The CTUIR aims to restore habitats and improve stream ecosystem processes to promote the expansion of steelhead and bull trout populations across the Umatilla River basin.

Coarse-bedded river floodplains are crucial habitats for the persistence of salmonid populations and may be critical in their future preservation as river temperatures are predicted to rise [Hauer et al., 2016]. River floodplains offer pools, riffles, side channels, and alcoves—features that provide heterogeneous temperature habitats [Amoros and Bornette, 2002, Wondzell, 2011]. Large-grained porous bed material allows for high rates of bidirectional water exchange between the channel and its alluvial aquifer, i.e., “hyporheic exchange” [Poole and Berman, 2001]. Hyporheic exchange buffers and lags stream temperature cycles on the daily and seasonal time scales, imparting heat storage and “thermal capacitance” to the channel-aquifer system [Arrigoni et al., 2008, Fogg et al., 2023]. During times of peak temperatures (summer or daytime), areas of upwelling hyporheic water are often cooler than the average temperature of the channel, whereas at times of minimum temperatures (winter or nighttime), upwelling water is warmer. This provides areas of potential thermal refuge for aquatic biota and leads to overall cooler (in summer) or warmer (in winter) channel temperatures in systems with high rates of hyporheic exchange.

Hyporheic exchange occurs across multiple spatial and temporal scales, where small spatiotemporal hyporheic exchange occurs across bed-form features such as dunes and large-scale hyporheic exchange occurs through larger features such as meander bends [Cardenas, 2015, Stonedahl et al., 2010]. Hyporheic flow is often described using flow path lengths or residence times. Across bedform-driven hyporheic exchange, flow path lengths and residence times in the hyporheic zone are small [Cardenas et al., 2008], contributing to temperature

changes of channel water at the daily scale. When hyporheic water flows through meander bends, flow path lengths, and subsurface residence times are much longer [Gomez et al., 2012]. The substantial lag of water flowing at this longer spatiotemporal scale contributes to the hyporheic influence on seasonal temperature cycles [Fogg et al., 2023]. The combination of small and large scale hyporheic exchange results in expansive hyporheic zones of coarse-grained river floodplains, where hyporheic waters can flow up to hundreds of meters away from the wetted channel in the most dynamic systems (e.g., the Umatilla River, Oregon [Poole et al., 2008], or the Nyack floodplain of the Middle Fork Flathead River, Montana [Hauer et al., 2016]).

Montane river floodplains have a unique fluvial landscape where conventional management strategies for reducing stream temperature are largely ineffective. For example, the most common strategy for reducing heat input to a stream is planting stream-bank vegetation to increase channel shade, thus reducing solar radiation [Beschta, 1997, Sweeney and Newbold, 2014]. Montane floodplain streams have snow-melt-dominated hydrology, causing high-energy spring floods which create an annual scour zone much wider than the summertime low-flow channel [Whited et al., 2007]. This wide annual scouring prevents young vegetation from reaching maturity and thus prevents the establishment of a substantial amount of vegetative shade for the low-flow stream channel.

The vegetation on river floodplains is rarely adjacent to the wetted channel at low flow conditions, yet the distant floodplain vegetation may indirectly affect stream channel temperatures by protecting shallow hyporheic water from solar radiation. Research on the effects of tree harvesting practices on streams has provided indirect evidence of hyporheic warming via solar radiation. Where forests are clear-cut outside of riparian buffer strips, subsequent stream channel warming sometimes occurred despite shade from the intact riparian canopy [i.e., Cole and Newton, 2013, Hewlett and Fortson, 1982, Moore et al., 2005]. These studies suggested that heating of shallow subsurface water from the increased solar

radiation in the clear-cut beyond the riparian buffer caused stream temperature warming, but did not directly test this hypothesis [Broszofski et al., 1997, Moore et al., 2005]. Other studies found temperature dynamics of hyporheic water were substantially impacted by heat conduction among the hyporheic zone, the bedrock, and the unsaturated sediments of the vadose zone [Molina-Giraldo et al., 2011, Vogt et al., 2012, 2010]. However, no studies have directly addressed how variation in floodplain shade might affect hyporheic water temperature and associated stream channel temperature response.

Simulation modeling experiments can test manipulations of the landscape in ways that would be either impossible or unethical to apply to a real stream. Physically based, hydrologic numerical models allow for investigators to analyze results from model manipulations are a reasonable approximation of what would happen if those manipulations were to occur in the real world. Hence, I use numerical simulation modeling to understand the effects of floodplain shade on hyporheic and stream channel temperatures. To investigate the interplay between floodplain shade, hyporheic heat exchange, and channel temperature, we used the numerical model HydroGeoSphere, which couples subsurface and surface flow and simulates physically-based heat transport [Brunner and Simmons, 2012].

In this dissertation, I investigate the effects of floodplain shade on temperatures in the hyporheic zone and stream channel. I hypothesize that (1) floodplain shade influences hyporheic temperatures by reducing solar radiation on the ground surface, thereby reducing the conduction of heat through the floodplain sediments overlying the hyporheic zone and (2) the pattern of floodplain shade and hyporheic flow paths govern the exchange of heat between the atmosphere, hyporheic zone, and channel, ultimately determining the magnitude of channel temperature response.

I present an *in silico* modeling experiment in Ch.2 aimed to theoretically constrain the influence of floodplain shade on hyporheic temperatures. Using a numerical model of a hyporheic flow path, I tested the effects on hyporheic temperatures of applying shaded,

unshaded, or no temperature signals to the ground surface of the model. Additionally, I varied the thickness of the modeled vadose zone and applied different hydraulic conductivity values to understand the potential influence of shade on hyporheic temperatures across a broad range of realistic floodplain conditions. In Ch. 3, I expand upon the theoretical modeling experiment and simulate shade distributions across a 3-dimensional model of a representative floodplain located on CTUIR tribal lands. I analyzed how the variable distribution of shade across the floodplain can influence temperatures of the ground surface, hyporheic zone, and stream channel. Finally, I investigate the impacts of two different channel re-meandering restoration strategies in Ch. 4. One strategy prioritized retaining maximum floodplain shade and the other did not and therefore had patches of bare floodplain. These shade scenarios were based on real distributions of floodplain shade before and after a channel re-meandering restoration. I test these simulations using either steady-state or transient hydrology and analyze how different hydrological boundary conditions impact simulation results.

EFFECTS OF FLOODPLAIN SHADE ON THE TEMPERATURE DYNAMICS OF A
SIMULATED LATERAL HYPORHEIC FLOW PATH

Contribution of Authors and Co-Authors

Manuscript in following chapter

Author: S. Kathleen Fogg

Contributions: Conducted research experiment, wrote HGS model, analyzed results and wrote manuscript

Author: Geoffrey C. Poole

Contributions: Advised on research analysis and model set-up, helped interpret results, edited manuscript

Author: Scott J. O'Daniel

Contributions: Deployed and retrieved temperature data loggers on Meacham Creek, funding contact, edited manuscript

Author: Byron A. Amerson

Contributions: Provided model code, advise and help on simulating water and heat flow in HGS

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S. Kathleen Fogg, Geoffrey C. Poole, Scott J. O'Daniel, and Byron A. Amerson

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Abstract

River reaches with coarse-grained alluvial floodplains have a breadth of lateral interaction between the wetted channel and surrounding landscape, yielding extensive riparian zones and laterally expansive hyporheic zones with high gross hyporheic exchange rates. The water table in hyporheic zones of floodplain rivers is often near the ground surface, allowing for heat exchange between unsaturated sediments warmed by the atmosphere and the hyporheic zone. We hypothesized that vegetative shade, or lack thereof, on a floodplain influences hyporheic zone temperatures, with the potential to impact biotic processing within the hyporheic zone and temperature in associated stream channels. To understand the potential effects of floodplain forest shade on hyporheic zone temperatures, we conducted a simple, *in-silico* modeling experiment on a simulated floodplain where we compared the effects of applying shaded versus unshaded vadose zone temperature signals on hyporheic temperatures. Additionally, we tested the sensitivity of simulated hyporheic temperatures to variable vadose zone thicknesses (0.5, 1.0, 2.0, and 3.0 m) and two hydraulic conductivity values of alluvial sediments (100 and 400 m day⁻¹). Results showed that conduction of differential temperature signals through the vadose zone substantially influenced the temperature patterns along our modeled hyporheic flow path, regardless of the vadose zone thicknesses tested. Our modeling results showed that floodplain shade has the potential to influence hyporheic zone temperatures of coarse-grained alluvial floodplains. We present floodplain forest shade as an important influence on hyporheic temperatures and a novel consideration for riparian management on floodplain river reaches.

Introduction

The thermal heterogeneity in hyporheic zones of gravel-bedded floodplain rivers impacts hyporheic water quality, subsurface aquatic ecosystems, and channel temperature regimes

[Evans et al., 1995, Hauer et al., 2016, Poole et al., 2008]. Interstitial water temperatures greatly affect biogeochemical reaction rates in the hyporheic zone, thus affecting the degradation of contaminants and the cycling of dissolved organic compounds [Hancock et al., 2005]. The influence of temperature on oxygen content, pH, and nutrient availability contributes to the habitat suitability of subsurface-dwelling macroinvertebrates, impacting their distributions across space and time [Stanford and Ward, 1988]. Additionally, the temperature of hyporheic water returning to the channel affects the channel environment. Areas of hyporheic upwelling are likely to have temperatures differing from the rest of the channel, providing thermal refuge to ectothermic species [Arscott et al., 2001]. In systems with sufficiently high gross hyporheic exchange rates, the cumulative effects of upwelling hyporheic water mixing with river water has the potential to change a river’s thermal regime at the reach scale [Arrigoni et al., 2008], causing cooler temperatures in the summer and warmer temperatures in the winter [Fogg et al., 2023].

Hyporheic water can be defined in multiple ways [White, 1993]; here we define hyporheic water as water stored in the alluvial aquifer which originated from the stream channel and will return to the channel [Poole et al., 2006]. Given this definition, the primary driver of hyporheic zone temperature dynamics is the temperature signal of downwelling river water. Upon downwelling into the hyporheic zone, the amplitude of the approximately sinusoidal daily and annual river temperature cycles are damped and the phase is lagged as flow path length increases [Luce et al., 2013, Molina-Giraldo et al., 2011, Poole et al., 2008]. At a certain distance from the point of downwelling and assuming one-dimensional (along the flow path) heat transport, the daily and annual sinusoidal patterns is damped completely—exhibiting a consistent temperature at increasing flow path distances.

The characteristic damping and lagging patterns of temperature cycles in hyporheic zones is caused by the dispersion of heat via advection and conduction along subsurface flow paths and can be modeled with solutions of the advection-diffusion equation (also known as

the “conduction-advection-dispersion” equation) [Hatch et al., 2006, Stallman, 1965]. Heat can be used as a tracer for inverse modeling of observed river and hyporheic temperature signals using the advection-dispersion equation [Anderson, 2005, Constantz, 2008, Fanelli and Lautz, 2008]. Heat tracer methods can characterize aquifer properties such as seepage rates, areas of upwelling versus downwelling and travel times hyporheic water [e.g., Vogt et al., 2012, 2010].

Many heat tracer techniques assume one-dimensional (1-D) transport behavior, but heat exchange in three dimensions has been shown to confound these approaches. Heat tracer methods utilize an analytical solution of the one-dimensional (1-D) advection-diffusion equation where a sinusoidal temperature signal is input at the beginning of the flow path and a constant groundwater temperature is assumed at the end [Stallman, 1965, Vogt et al., 2012]. This solution implies a predictable dissipation of heat in the subsurface and therefore predictable temperature signal attenuation with increasing distance from the beginning of the simulated flow path. However, many studies have shown that 1-D assumptions/simplifications of heat dissipation are insufficient in areas where river-water infiltrates, such as hyporheic zones [Hoehn and Cirpka, 2006, Molina-Giraldo et al., 2011, Vogt et al., 2012, 2010]. Heat exchanges occur in 3-dimensions in hyporheic zones and temperature signals other than the infiltrating river water interfere with the simplifying assumptions of the 1-D model.

Molina-Giraldo et al. [2011] compared results of the 1-D analytical model of temperature to a numerical model that included conductive heat exchange between the vadose zone, alluvial aquifer and aquitard. They found that the numerical model resulted in substantially different temperature damping and lagging compared to the 1-D solution, causing differences between travel time calculations based on temperature damping and lagging. Considering the daily temperature signal of river water infiltrating into a stream-bank, [Vogt et al., 2012] used temperature observations of the River Thur and numerical modeling techniques to show

that vertical heat exchange with the vadose zone also influences diel temperature oscillations.

Coarse-grained floodplain reaches with expansive hyporheic zones have characteristically high river water infiltration rates and shallow water tables, thus allowing for the conduction of the floodplain surface temperature signal into the hyporheic zone. The primary driver of the temperature signal of the ground surface is the daily and annual variation in solar radiation. Vegetation attenuates the solar radiative flux, producing a cooler ground surface temperature in shaded rather than unshaded areas. We hypothesize that the vertical heat flux between the atmosphere, unsaturated sediments and hyporheic zone is an important driver of the patterns of longitudinal hyporheic zone temperatures. We therefore expect that shade, vadose zone thickness and hydraulic conductivity are important mediators of hyporheic temperature patterns. If other components of the thermal energy budget are held constant, we predict that hyporheic zone temperatures will show substantial increases in temperature when shade is absent, especially when vadose zones are thin.

The aim of this study was to theoretically constrain the potential influence of floodplain shade on the temperature dynamics of hyporheic zones under a variety of reasonable approximations of floodplain conditions along hyporheic flow paths. We present an *in-silico* modeling experiment in which we compared the heat transport behaviors among hyporheic flow paths through shaded, unshaded, and control floodplain scenarios, where the control scenario was completely isolated from atmospheric heat exchange. Our modeling experiment included systematic variation in the degree of floodplain shade, thickness of the vadose zone overlying the hyporheic zone, and the hydraulic conductivity of the hyporheic sediments. We describe the ways in which each of these parameters affected hyporheic zone temperatures in our model, thus providing a basis of understanding the potential of thermal influence of floodplain forest shade on hyporheic temperatures in expansive, alluvial floodplain rivers.

Methods

We implemented our model scenarios in the groundwater flow and transport model HydroGeoSphere (HGS) [Brunner and Simmons, 2012]. HGS is a finite element solution of Richard’s equation and heat advection-diffusion equations for the simulation of water flow and heat transfer in 3-dimensions. HGS is therefore capable of simulating subsurface water and heat movement in both saturated and unsaturated conditions.

Experimental Design

We modeled a two-dimensional cross-section of an idealized floodplain in the vertical and longitudinal dimensions, approximating a configuration typical of expansive, coarse-grained alluvial aquifers in montane regions [Helton et al., 2012b, Jones et al., 2008b, Stanford and Ward, 1993]. River water infiltrated the hyporheic zone of the simulated cross section from a fully penetrating river at the “upstream” end of the cross section. The hyporheic zone was bounded by a 25 m thick bedrock layer below and a vadose zone above (Figure 2.1). Using a full factorial experimental design, twenty-four model scenarios were simulated to understand the potential influences of shade (three levels of no heat exchange, shaded, and unshaded), vadose zone thickness (four levels from 0.5 to 3 m) and hydraulic conductivity (two levels of 100 and 400 m day⁻¹ on hyporheic zone temperatures (Table 2.1). The range in vadose zone thickness encompassed observed values in expansive alluvial floodplains during base-flow [Helton et al., 2012b, Munz et al., 2017].

Values of hydraulic conductivity are representative of those observed in expansive, coarse grained alluvial floodplains of the intermountain western U.S., such as Meacham creek and the Umatilla river (Pendleton, OR), and the Nyack floodplain of the Middle Fork Flathead river (West Glacier, MT).

Since shading by plants directly effects ground surface temperature, we alternated the

specified temperature cycles of the top boundary condition of the vadose zone in our model to represent either shaded or unshaded floodplains. A “control” scenario was also simulated, which included the downwelling temperature signal from the river, but lacked a temperature cycle at the top of the vadose zone.

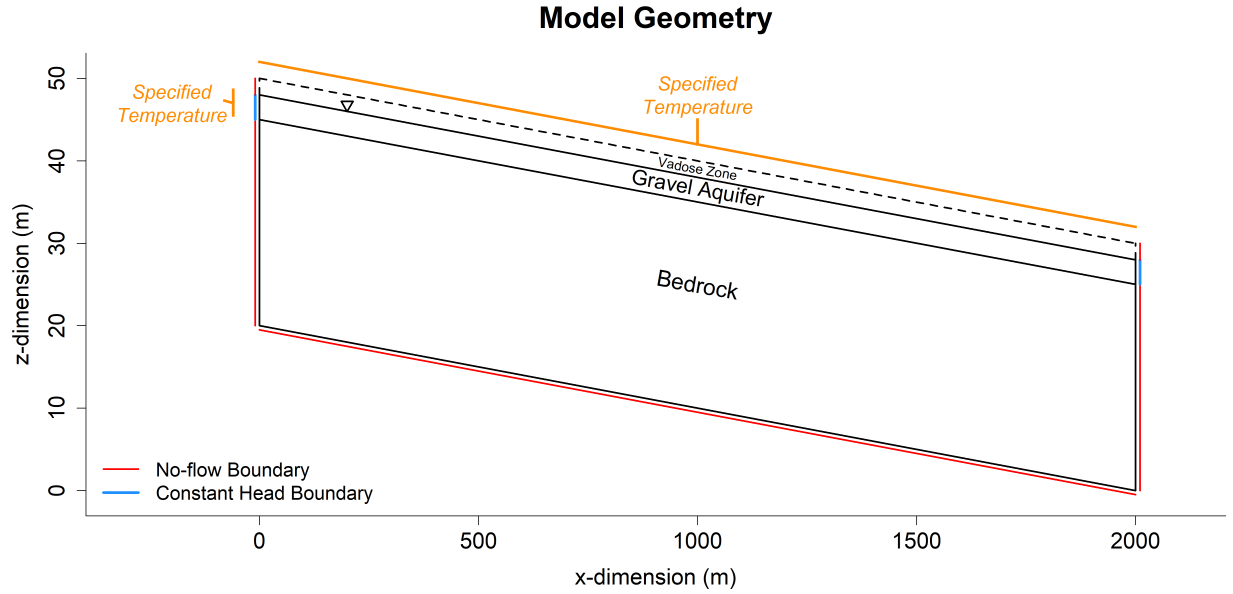


Figure 2.1: Model Geometry. The floodplain models were composed of three vertical layers: 25-m bedrock layer, 3-m thick gravel aquifer layer, and a vadose zone which varied with model scenario. Constant head boundary conditions were applied to the upstream and downstream boundaries of the gravel aquifer and an annually repeating compound sine wave signals of river and soil temperature were specified at the upstream end of the gravel aquifer and at the upper boundary of the vadose zone. Hydraulic gradient is exaggerated in this image due to the difference in x- and z-axes resolution.

Boundary Conditions

Across all scenarios, we modeled steady-state flow conditions and used a constant head boundary at the upstream and downstream ends of the simulated hyporheic zone. Although variation in flow conditions and hydraulic gradients is an important driver of hyporheic temperature, we chose to model steady-state conditions to remove confounding variation

Ground Surface Temperature Signal	Vadose Zone Thickness	Hydraulic Conductivity
None (control)	0.5 m	100 m day ⁻¹
Shaded	1.0 m	400 m day ⁻¹
Unshaded	2.0 m	
	3.0 m	

Table 2.1: Factors of experimental design. All permutative combinations were simulated, yielding 24 model scenarios.

that may be caused by variable aquifer recharge rates. The hydrologic gradient between the upstream and downstream constant head boundary conditions represented a 1% slope – a typical longitudinal gradient on alluvial floodplain reaches [Rosgen, 1994, Selby, 1985]. No-flow boundaries constituted the upstream, downstream, and bottom edges of the bedrock zone, and the upstream and downstream boundaries of the vadose zone (Figure 2.1).

Model Node Spacing

HGS calculates aquifer conditions at “nodes” distributed throughout the model domain. Our model used irregular node spacing; higher node densities were used in areas where temperatures were expected to be more dynamic and lower node densities in areas where temperatures were expected to be more spatially consistent. Specifically, diel temperature signals associated with the specified temperature boundaries cause steep thermal gradients over short distances, therefore the node spacing was smallest proximal to specified temperature boundaries. The horizontal node spacing at any location along the flow path (specified as the x dimension) was set using the parameters in Table 2.2. In each of the four “Zones”, the first node was set at the beginning of the zone. The second node was set at the distance indicated by “Initial Node Spacing”. Subsequent nodes were spaced at a distance of “Multiplier” times the prior node spacing, not to exceed “Max Node Spacing.”

In the vertical dimension, node spacing in the vadose zone and gravel aquifer was 0.1 m, and varied with depth in the bedrock (Table 2.3).

zone	Initial Node Spacing (m)	Multiplier	Max Node Spacing (m)
0-4 m	0.1	1	0.1
4-300 m	0.1	1.5	2
300-600 m	2	1.5	4
600-2000 m	4	1.5	10

Table 2.2: Parameters determining node spacing in the x-direction.

Horizontal Layer	Node Spacing (m)
vadose zone	0.1
gravel aquifer	0.1
bedrock:	
0-1.5 m	~0.5
1.5-5.5 m	~1
5.5-19 m	~2
19-25 m	~3

Table 2.3: Node spacing in the z-dimension.

Aquifer Properties

Aquifer properties other than geometry and hydraulic conductivity (Table 2.4) were based on field observations of Meacham Creek (B. Amerson, unpublished data), an expansive coarse grained alluvial aquifer-floodplain system located in the inter-mountain western USA. Meacham Creek is located in the Blue mountains east of Pendleton, Oregon, and flows

through a valley bottom of coarse-grained basaltic alluvium constrained below and laterally by bedrock. Wells and field observations show that the hyporheic zone of Meacham creek is approximately 3 meters thick and has high gross hyporheic exchange rates (B. Amerson, unpublished data).

Model parameters describing thermal properties of the aquifer were derived from published literature values for water and basalt (Table 2.5).

Driving Variables

Temperature time series applied at model boundary conditions were derived from models fit to observations of river and vadose zone temperatures on Meacham Creek. These thermal boundary conditions included input temperature signals for the downwelling river water, the top of the shaded vadose zone, and the top of the unshaded vadose zone.

We collected hourly temperature data in the thalweg of the Meacham Creek channel for approximately one year using an Onset HOBO Water Temperature Pro v2. We monitored vadose zone temperatures at approximately 30 cm depth at two different locations on the floodplain using Onset HOBO TidbiT MX temperature loggers. One vadose temperature site was located in an area of the floodplain that was shaded throughout the day whereas the other site was located in an unshaded area. Temperature data were collected at an hourly interval from July 2020 to October 2021.

We fit compound sine-waves that captured typical daily and seasonal variation to the observed temperature values collected in the stream channel, a shaded vadose zone site and the unshaded vadose zone site (Figure 2.2). Parameter values for the compound sine waves were inferred using non-linear least squares (nls function) in R to fit observed daily temperature means and ranges for each day of the year. We repeated the fitted compound sine waves across every year of simulation to remove the potential for confounding effects of interannual variation in temperature regime.

Parameter	Value	Unit
Alluvial Porous Media		
Porosity, θ_s	0.25	-
Dispersivity		
longitudinal, α_l	1.0	m
transverse, α_t	0.1	m
vertical transverse	0.1	m
Saturated Hydraulic Conductivity, K_s		
x-direction	<i>variable</i>	m s ⁻¹
y-direction	<i>variable</i> $\times$ 0.1	m s ⁻¹
z-direction	<i>variable</i> $\times$ 0.1	m s ⁻¹
Van Genuchten parameter, α	9.0	m ⁻¹
Van Genuchten parameter, β	3.1768	
Residual Saturation, S_r	0.1	
Pore Connectivity, l_p	0.5	
Bedrock		
Porosity, θ_s	1.0e ⁻²⁰	-
Dispersivity	1.0e ⁻²⁰	m
Saturated Hydraulic Conductivity, K_s	1.0e ⁻¹⁰	m s ⁻¹

Table 2.4: Hydrology model parameters for porous media and bedrock materials. Parameters from Freeze and Cherry [1979], Molina-Giraldo et al. [2011], Thoma et al. [2014].

Model Execution

We performed a spin-up procedure in order to determine the steady-state temperature distribution in the model domain that was not influenced by the initial conditions. The spin-up procedure began by applying the annual mean temperature of the downwelling river

Parameter	Value	Unit
Water		
Thermal Conductivity	0.5	$\text{W m}^{-2} \text{K}^{-1}$
Specific Heat Capacity	4185	$\text{J kg}^{-1} \text{K}^{-1}$
Density	1000	kg m^{-3}
Basalt¹		
Thermal Conductivity	1.6	$\text{W m}^{-2} \text{K}^{-1}$
Specific Heat Capacity	870	$\text{J kg}^{-1} \text{K}^{-1}$
Density	3000	kg m^{-3}

¹Waples and Waples [2004]

Table 2.5: Thermal parameters of Basalt and Water used in model

water (10.3°C) across the entire model. Each model was run in “spin-up” until annual cyclic steady state, where temperatures across all nodes stabilized to a maximum temperature difference of 0.1°C on 1-January between successive years.

The spin-up temperatures were then used as the initial conditions for each scenario simulation. Each scenario variant in our factorial experimental design was simulated for one year and one day of output was recorded every 15 days (Jan 1, Jan 16, ..., Dec 12), with a 1-hour output time step. These output days allowed us to capture the annual temperature cycle and diel cycles vary across seasons.

Analysis

We assessed model results by the patterns of amplitude damping of the temperature signal in the subsurface. We examined diel temperature signals in the first 10 meters from the downwelling boundary across time for five different days of the year and five different times within each of those days. We analyzed diel temperature signals in only the first 10

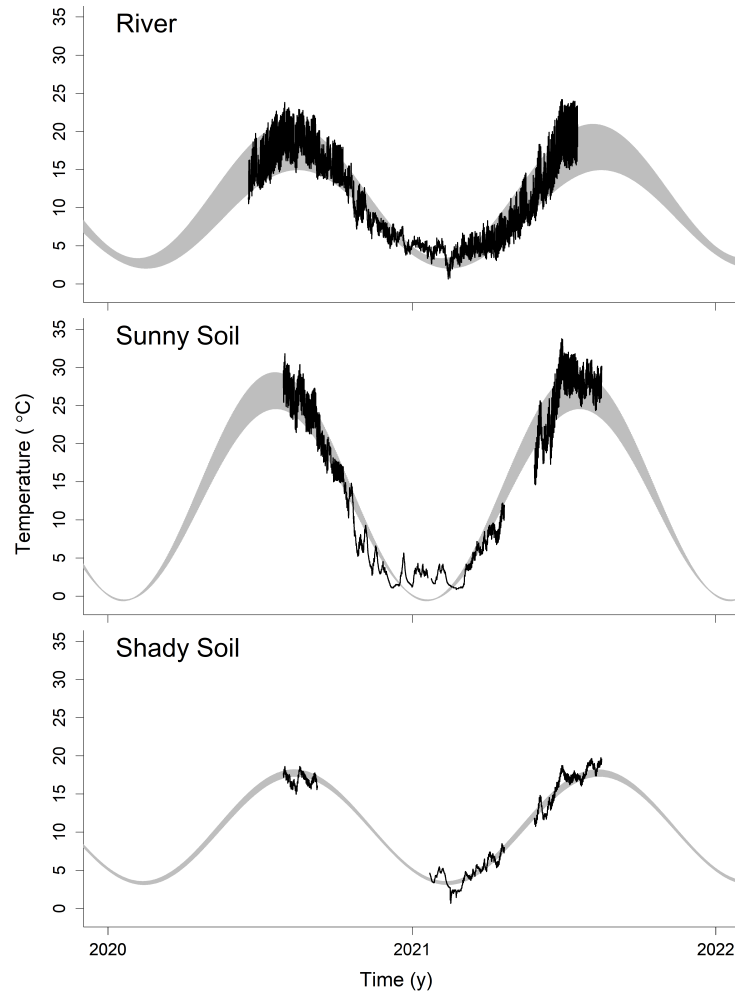


Figure 2.2: Compound sine wave models (gray) of observed temperature (black) of Meacham creek, top, and a shaded and unshaded site on the Meacham creek floodplain where we measured subsurface soil temperature.

meters of the flow path because these higher frequency signals typically damp out within a few meters from their source. We assessed the seasonal signals by examining the amplitudes of the annual signal of daily mean temperatures at all model nodes within the vadose zone, hyporheic zone, and the top 15 meters of the bedrock zone.

In shaded and unshaded model scenarios, annual temperature patterns in the subsurface reflect a convolution of two input temperature signals: one imparted by downwelling river water and the other imparted by the vadose zone. Temperature amplitude attenuation in soils

is commonly summarized using the damping depth—the distance in the subsurface where the temperature signal is reduced to e^{-1} , or ca. 36.8%, of the surface temperature signal [Holmes et al., 2008, Robson, 1989]. Here, we refer to the damping depth as the damping *distance* to reflect the fact that one input temperature signal moves primarily horizontally (river) and the other vertically (vadoze zone). Accordingly, we considered damping distances of the attenuation of each input signal: a_r , the annual amplitude of the river temperature signal, and a_v , the annual amplitude at the top of the modeled vadose zone (which also varied with the shaded and unshaded model scenarios). We compared patterns in contour lines in the model domain where the annual temperature amplitude equaled damping distances at $a_x e^n$ where n in $\{-1, -2, -3, -4, -5\}$ and the subscript refers to the input signal (r for river and v for vadose zone). For the daily signal we plot contours of damping distances at $a_r e^n$ where n in $\{-1, -2, -3\}$. These damping distances are more easily thought of as percentages of the original input amplitude, where e^{-1} , e^{-2} , e^{-3} , e^{-4} , and e^{-5} correspond with amplitude reductions of 36.8%, 13.5%, 4.9%, 1.8%, and 0.67%, respectively.

To compare hyporheic temperatures of shaded and unshaded floodplains, we first summarized the vertical variation in hyporheic zone temperatures. We averaged daily mean hyporheic zone temperatures across the 3 meter depth of the hyporheic zone for all flow path lengths (x-dimension locations). To allow examination of the spatiotemporal variation in the influence of floodplain shade on hyporheic temperatures across seasons, we subtracted the summarized longitudinal hyporheic temperatures of the shaded scenario from the unshaded scenario for each output day and displayed resulting values as a heat map of temperature differences across time and longitudinal x-distance in the aquifer.

Results

Patterns in Diel Hyporheic Temperatures

The cone-like pattern of hourly temperatures in the first 10 meters of the hyporheic flow path characterizes the decay in diel amplitude of the input diel temperature signal at the river-downwelling boundary (2.3). The height of the cone at 0 meters is the diel amplitude in surface waters. Results from all model scenarios show the diel amplitudes of downwelling river water (a_r) damped to less than 4.9% (e^{-3}) within the first 10 meters of the hyporheic flow path (Figure 2.3).

Model results showed minimal differences in longitudinal hyporheic zone temperatures (<0.01 °C) across scenarios with the same hydraulic conductivity values (K), regardless of differing vadose zone thickness and floodplain boundary temperature (Figure 2.3). Results from scenarios with $K = 100 \text{ m day}^{-1}$ show the diurnal amplitude damped to 4.9% of a_r at ~ 4 meters from the of downwelling model boundary. In scenarios with $K = 400 \text{ m day}^{-1}$, the same fractional reduction in amplitude occurred at ~ 9 meters.

Amplitude Attenuation of the Seasonal Signal

Control Scenarios The annual temperature amplitude of downwelling river water is attenuated with increasing distance along the hyporheic flow path (Figures 2.4 and 2.5). This pattern is most defined in the control scenarios because the surface water inflow is the only source of temperature variation in these simulations. The annual amplitude damps out to $a_r e^{-5}$ at shorter flow path distances as the vadose zone thickness increases. When $K = 400 \text{ m day}^{-1}$, amplitude contours were much further apart than results from scenarios with $K = 100 \text{ m day}^{-1}$.

The downwelling river temperature signal travels through the modeled hyporheic zone horizontally, thus the amplitude contours are vertically oriented. Contours become closer

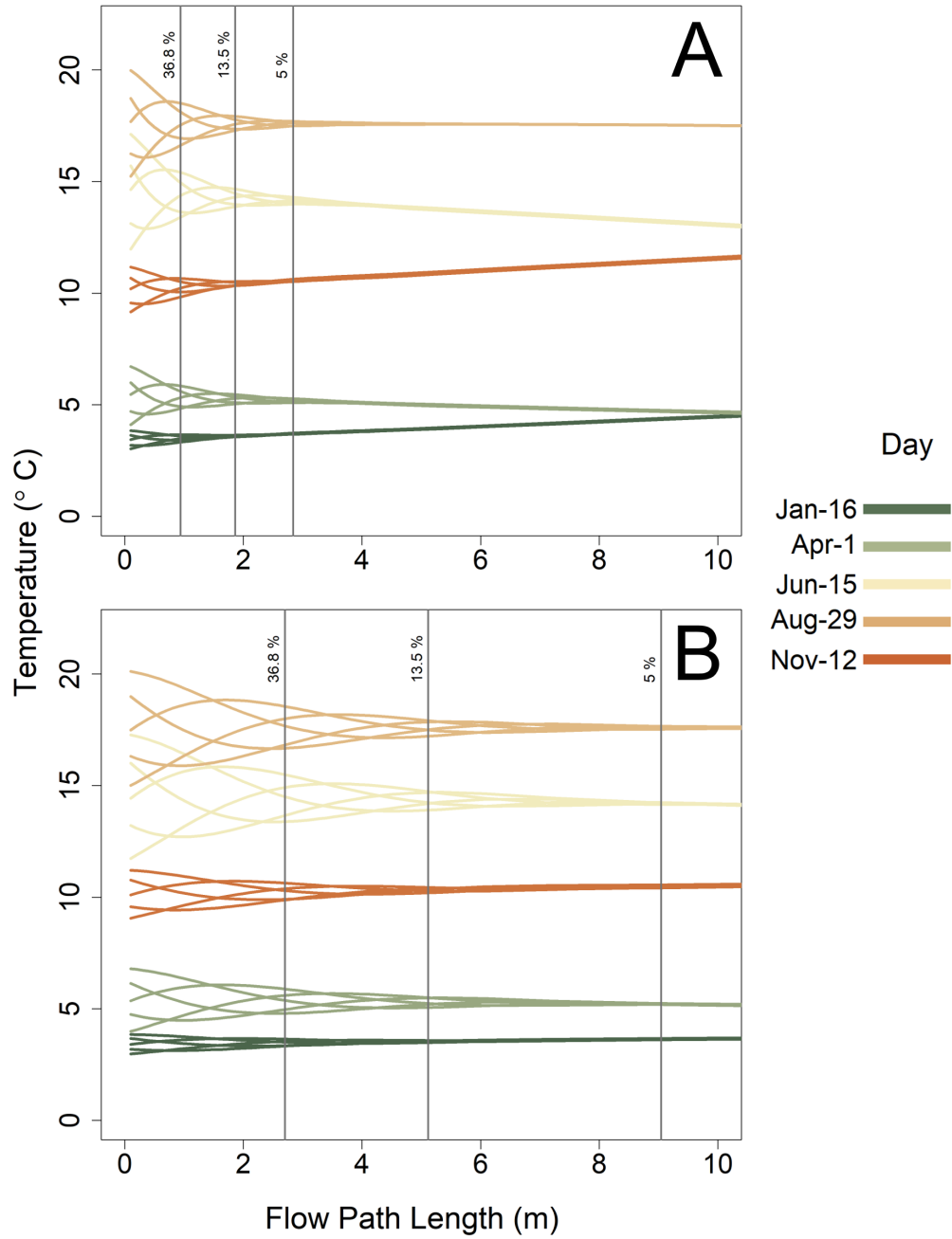


Figure 2.3: Temperatures in the hyporheic zone for scenarios where $K = 100 \text{ m day}^{-1}$ (A) and $K = 400 \text{ m day}^{-1}$ (B). Variation in diel temperature patterns were less than 0.01°C for scenarios with same K values and different vadose zone thicknesses and temperature boundaries. Five days across the year are differentiated by color, and each of the five lines for a given day represent 0100, 0600, 1100, 1600, and 2100 time of that day.

together as the model's vadose zone thickness increases, indicating that more dispersive and conductive spreading of heat is occurring in scenarios under thicker vadose zones, causing the annual amplitude to attenuate below $a_r e^{-5}$ closer to the downwelling boundary (Figures 2.4 and 2.5).

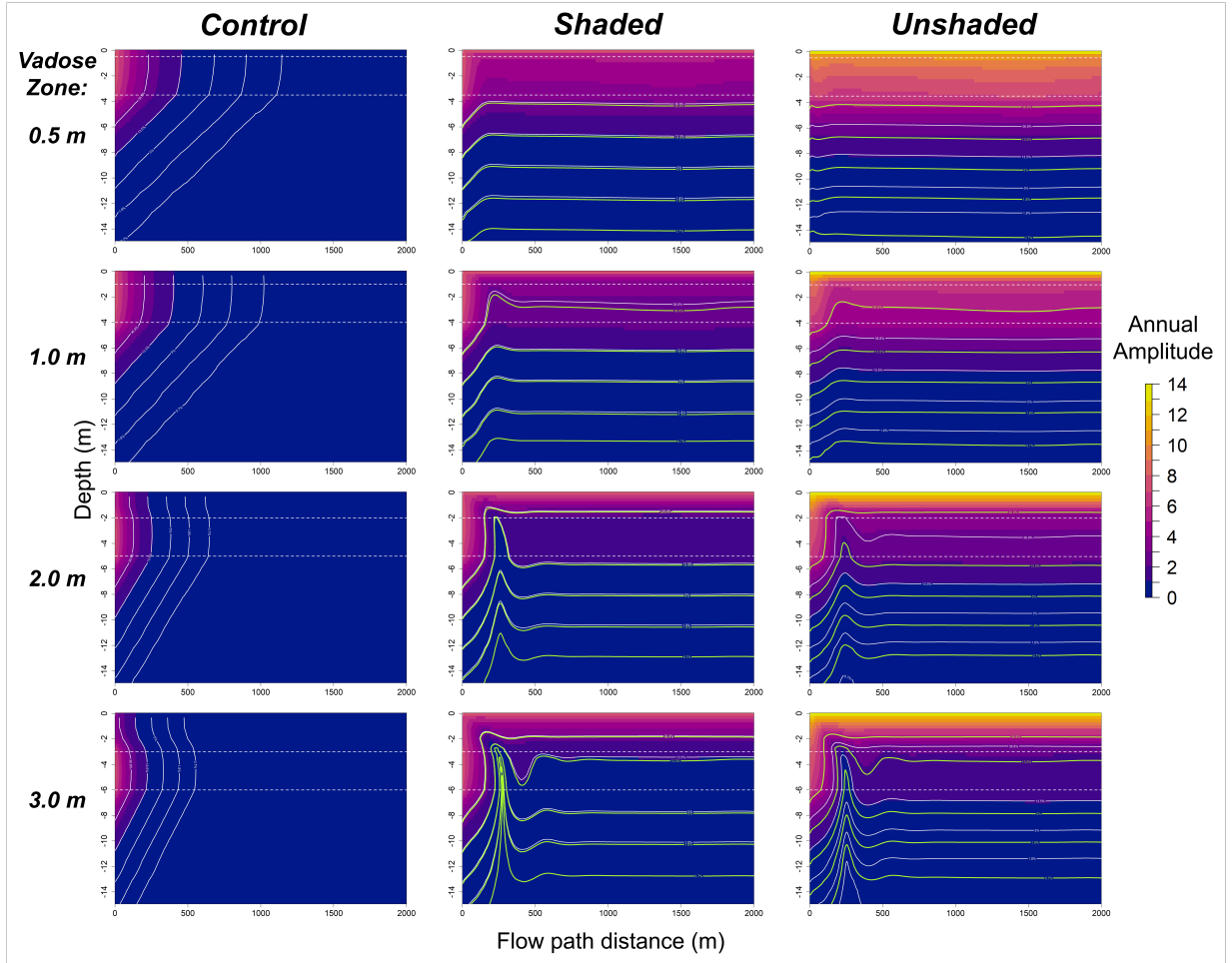


Figure 2.4: Annual temperature amplitude of scenarios with $K = 100 \text{ m day}^{-1}$. Dotted white lines delineate the vadose zone, hyporheic zone, and bedrock (shown up to 15-m depth). Solid white lines display the contours of $a_r e^n$ and solid green lines display the contours of $a_f e^n$ for n in $\{-1, -2, -3, -4, -5\}$. These contour lines show where in space the annual amplitudes are a constant value. Amplitudes attenuate in a direction perpendicular to the contour lines and the distance between contours is the rate at which amplitudes attenuate across space. Labels of contours lines display e^n as a percentage reduction.

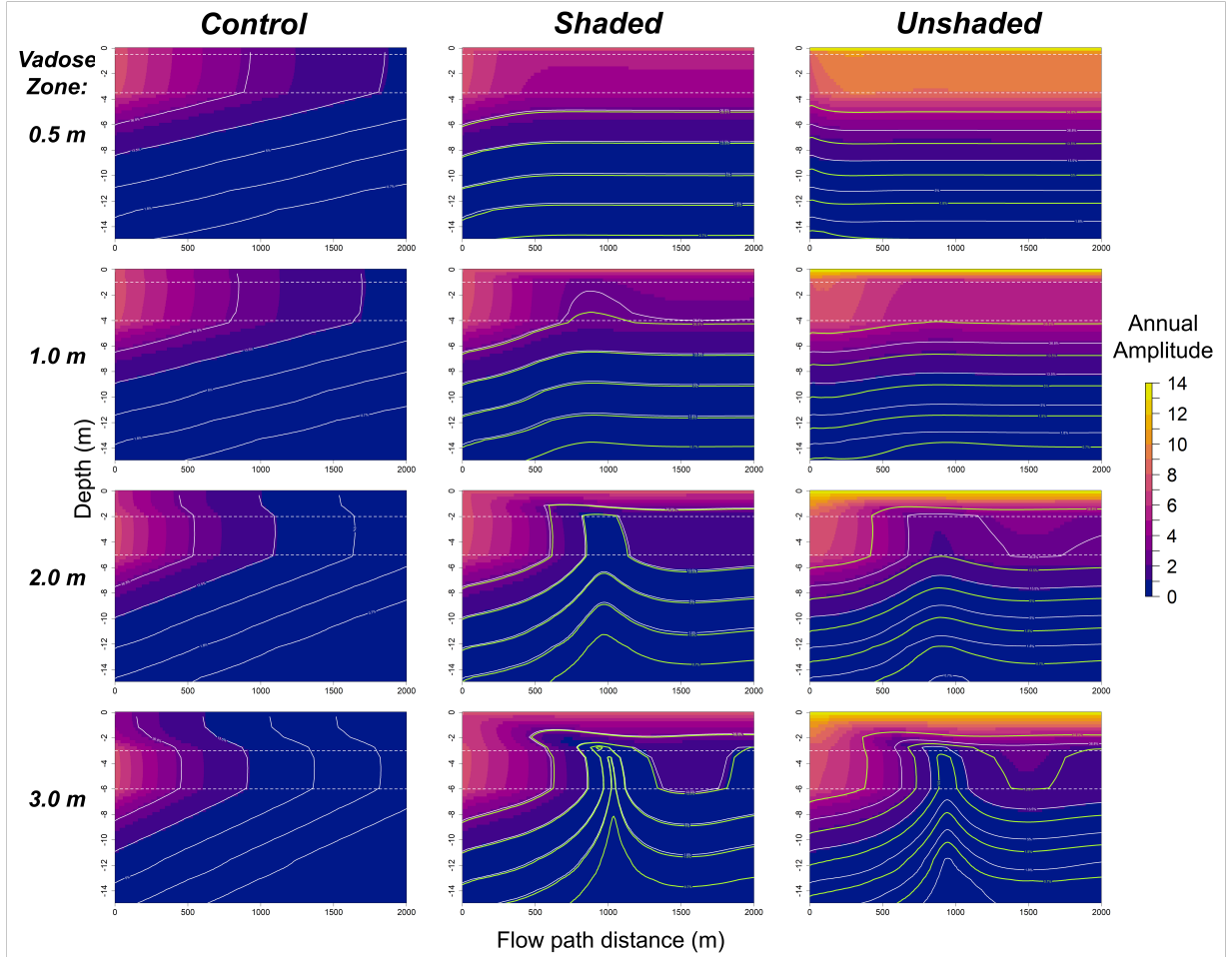


Figure 2.5: Simulated annual temperature amplitude for scenarios with $K = 400 \text{ m day}^{-1}$. Dotted white lines delineate the vadose zone, hyporheic zone, and bedrock (shown up to 15-m depth). Solid white lines display the contours of $a_r e^n$ and solid green lines display the contours of $a_f e^n$ for $n \in \{-1, -2, -3, -4, -5\}$. These contour lines show where in space the annual amplitudes are a constant value. Amplitudes attenuate in a direction perpendicular to the contour lines and the distance between contours is the rate at which amplitudes attenuate across space. Labels of contours lines display e^n as a percentage reduction.

Shaded Scenarios In contrast to results from the control scenarios, the shaded model scenarios show annual temperature amplitudes are sustained at a further distance into the hyporheic zone from the downwelling model boundary. When the vadose zone is thin, an annual amplitude greater than the first damping depth is sustained in the hyporheic zone at all flow path lengths. The horizontal orientation of amplitude contours in the 0.5-m vadose zone scenarios indicates that the surface temperature boundary is driving the pattern in amplitude attenuation.

When the thickness of the vadose zone increases, the pattern in amplitude attenuation becomes more similar to the corresponding control scenarios. At shorter flow paths, the direction of annual amplitude attenuation occurs in the direction of downwelling water, whereas at longer flow paths, the direction of amplitude attenuation occurs at increasing depth into the subsurface. Thus, at the beginning of the hyporheic flow path, the temperature patterns are dominated by downwelling river water whereas at longer flow paths the spatial pattern in temperature is dominated by the temperature pattern on the soil surface.

Coincidentally, the amplitude contour lines of the vadose zone input amplitude, a_v (green contours), are approximately equal to contours of the downwelling river amplitude, a_r (white contours) in shaded model scenarios. Similarities in the patterns of river water and shaded atmosphere temperature signal attenuations are due to the fact that the annual amplitude of the shaded temperature signal (7.2°C) and the river temperature (7.6°C) are similar.

Similar to the control scenarios, the difference between the $K = 100 \text{ m day}^{-1}$ and 400 m day^{-1} is the longitudinal spread of amplitude attenuation when hydraulic conductivity is greater.

Unshaded Scenarios Results of unshaded model scenarios show similar patterns in amplitude attenuation as the shaded scenarios, however, the annual amplitudes are greater

in the vadose and hyporheic zones of these scenarios. In contrast to results from the shaded model scenarios, the unshaded scenarios show a marked difference in amplitude contour lines calculated from a_v (green) than a_r (white) in the unshaded model scenarios. The annual amplitude of the unshaded soil temperature signal is approximately 6°C greater ($a_v = 13.7$) than the river temperature signal ($a_r = 7.6$). Contours based on attenuation of a_v occur at greater temperature amplitudes than contours based on attenuation of a_r because the same fractional attenuation is being applied to different source amplitudes.

Because the amplitude contours are calculated as a percentage of the *input* temperature amplitude, the contours based off of a_v occur at greater temperature amplitudes than contours based off of a_r .

Longitudinal Hyporheic Temperature Differences

The surface temperature of the unshaded floodplain was warmer in the summer and cooler in the winter than the shaded floodplain—an annual pattern which transferred into the differences in hyporheic zone temperatures between shaded and unshaded floodplains. The transfer of heat from the floodplain surface to the hyporheic zone was most pronounced in scenarios with thin vadose zones (0.5 or 1.0 m), and was subdued in scenarios with thicker vadose zones (2.0 or 3.0 m).

Across most months of the year, and most pronounced in the summer months, the unshaded floodplain scenarios had warmer simulated hyporheic zone temperatures than the shaded floodplain scenarios (Figure 2.6). Thin vadose zones demonstrated the greatest temperature differences between the shaded and unshaded floodplains. The temperature signal of the unshaded floodplain surface was cooler than the shaded floodplain in the winter, resulting in cooler wintertime hyporheic zone temperatures in the unshaded scenarios.

Close to the downwelling boundary, temperature differences were minimal between shaded and unshaded scenarios ($0 \pm 0.5^\circ\text{C}$; white areas of Figure 2.6) because all scenarios

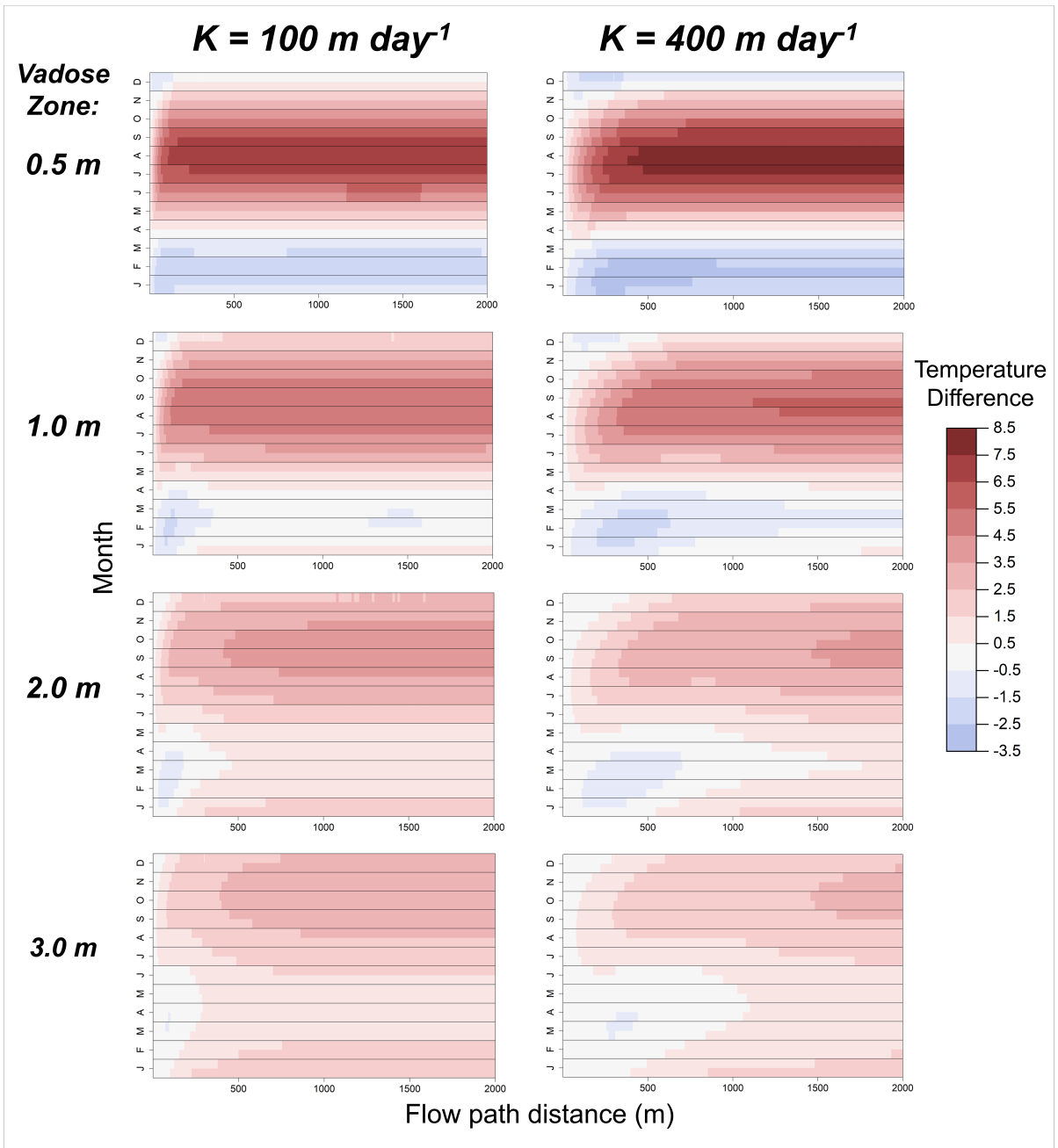


Figure 2.6: Differences in simulated longitudinal hyporheic temperature of the unshaded floodplain compared to the shaded floodplain. Colors denote the vertically averaged temperature difference in the hyporheic zone. Horizontal lines differentiate the month of the year.

had the same input downwelling river temperature signal. Thicker vadose zones attenuated the atmospheric input signal before reaching the hyporheic zone, thus the length of hyporheic flow-path with minimal temperature difference between shaded and unshaded scenarios increased correspondingly with thicker vadose zones.

Discussion

Influence on Heat as a Tracer Methods

Our modeling experiment shows how the vertical heat flux between the atmosphere, vadose zone, and hyporheic zone is likely to be an important driver of the patterns in lateral-longitudinal hyporheic flow path temperatures. Results demonstrate how attenuation patterns of annual temperature amplitude in our simulated hyporheic zone were determined by an interaction of two temperature signals: (1) downwelling river water, traveling horizontally, and (2) the ground surface temperature traveling vertically through the vadose zone (Figures 2.4, 2.5). The balance between propagation of the river signal and atmospheric signal was strongly influenced by the thickness of the vadose zone above the hyporheic flow path and the hydraulic conductivity of the hyporheic sediments.

The interaction of the horizontal and vertical input temperature signals was not apparent in simulated hyporheic temperatures at the daily time-scale, as results showed little temperature sensitivity to surface boundary temperature or vadose zone thickness (Figure 2.3). This result implies that horizontal heat advection and dispersion of the river signal, rather than vertical conduction of the temperature signal of the vadose zone, are the primary heat transfer mechanisms at the daily spatiotemporal scale. We expect diel temperature signals may be affected by vertical heat conduction if we simulated lower hydraulic conductivity values, thus representing a system with less gross hyporheic exchange than the highly connected river-floodplain system we aimed to represent. Lower hydraulic conductivities would slow interstitial water velocities and would allow for conductive heat

exchange at this scale, thus altering the diel temperature signal across scenarios with differing surface temperatures and vadose zone thicknesses.

Our simulation results show how multiple temperature signals may interact in expansive, coarse-grained hyporheic zones. Interaction of vertical and horizontal signals caused differential amplitude attenuations of the seasonal temperature cycle of downwelling river water compared to the amplitude attenuation displayed in scenarios with no vadose zone temperature input (i.e., the control scenarios). The interaction of multiple temperature signals is not accounted for in 1-D analytical solutions of the advection-dispersion equation commonly used in heat tracer methods of vertical subsurface temperature profiles [Hatch et al., 2006, Stallman, 1965, Vogt et al., 2010]. Our study supports research done by Molina-Giraldo et al. [2011] and Vogt et al. [2012] which conclude that heat conduction between the vadose zone, hyporheic zone, and bedrock influences lateral hyporheic zone temperatures. Our study builds on this work by showing how a shaded ground surface temperature cycle leads to different patterns of amplitude attenuation and generally cooler hyporheic flow path temperatures than an unshaded temperature cycle. Heat conducted from the vadose zone interacts with the horizontally moving downwelling river temperature signal to impede the full damping out of the annual temperature amplitude (Figures 2.4 and 2.5). Thus, our results corroborate Molina-Giraldo et al. [2011] conclusions that the effect of conduction between the vadose zone and hyporheic zone can badly confound the use of temperature signals to age hyporheic waters.

Relevance to Real River-Floodplain Systems

The assumptions of our *in-silico* modeling experiment are that parameters and driving variables, although simplified, represent a range of real-world floodplain conditions. All model scenarios showed that floodplain shade influenced hyporheic thermal regimes, thus hyporheic temperatures in real river-floodplain systems are likely influenced by the shade

regime on the floodplain surface. These results provide an initial basis for understanding the potential impact of floodplain forest shade on hyporheic zone temperatures.

Real-world shade distributions overlying hyporheic zones are more complex than our simplified representation of the entire length of a hyporheic flow path being either fully shaded or unshaded. However, patterns of shade along hyporheic flow paths likely yield generalizable patterns [Huggenberger et al., 1998]. At short hyporheic flow-path lengths, near the river channel, the overlying floodplain surface is more likely to be within the annual scour zone and therefore bare and exposed to conditions similar to our unshaded floodplain scenarios. As distance along a flow-path gets further away from the channel and away from the annual scour zone, the overlying floodplain is likely to be more vegetated. More realistic distributions of shade patterns along a lateral hyporheic flow path in future modeling experiments would yield a greater understanding of how shade can alter temperature patterns in hyporheic zones. Additionally, a gradient of vadose zone thickness also likely exists along hyporheic flow paths—another gradient that could be simulated in future numerical modeling simulations.

Our simulation results represent the potential effects that floodplain shade may have on hyporheic zone temperatures. In real river-floodplain systems, hyporheic flow paths encounter much more heterogeneous conditions than the hyporheic zones represented in our simplified modeling experiment. Not only did we use idealized driving variables (Figure 2.2), but floodplain sediments have variation in hydrologic and thermal properties which we hold constant in our modeling experiment, thus temperature differences between shaded and unshaded floodplains should be interpreted as general patterns of potential effects.

Implications for River-Floodplain Management

The hyporheic zone acts as a physical, chemical, and biological filter, therefore performing an important ecosystem service which improves water quality in the river-

floodplain system [Haack and Bekins, 2000, Hancock et al., 2005, Peralta-Maraver et al., 2018]. Our results suggest that shade on the floodplain surface can mediate hyporheic temperatures, thus influencing the rates of biogeochemical reactions and water quality in the hyporheic zone.

We demonstrate the potential warming effects of unshaded floodplain sediments on hyporheic zone temperatures. The loss of floodplain forest shade can be a result of many causes, including encroaching land development, wildfire [Tranmer et al., 2020], and forest harvesting in the riparian buffer which may not encompass the entire width of the floodplain where lateral hyporheic flow paths occur [Hewlett and Fortson, 1982]. Concomitant with reduced floodplain shade and greater hyporheic zone temperatures is altered biogeochemical processing rates, redox potential, oxygen content, and microbial and macroinvertebrate community structure in the underlying hyporheic zone [Acuña et al., 2008, Mori et al., 2003, Nogueira et al., 2021]. Additionally, warming of hyporheic temperatures can result in warmer in-stream areas of upwelling hyporheic water, thus reducing the impact of thermal refuges provided by these areas of upwelling hyporheic exchange in the summertime [Baxter and Hauer, 2000, Faulkner et al., 2020].

Compared to scenarios representing shaded floodplains, unshaded model scenarios show additional warming of hyporheic zone waters (Figure 2.6). Heating of hyporheic water via conduction through unshaded floodplain sediments is thus an additional influence to consider when modeling, managing, or restoration of hyporheic zone temperatures.

Conclusion

Our *in-silico* modeling experiment shows the potential of floodplain shade to affect the thermal regimes of hyporheic zones under a variety of realistic floodplain conditions. Modeled floodplains with unshaded vadose zones resulted in warmer simulated hyporheic temperatures than under shaded vadose zones, with greatest temperature difference occurring when vadose

zones were thin and during the summer months. The simulation results presented here show that floodplain shade has the potential to impact hyporheic zone temperatures, which could alter biogeochemical reaction rates and the temperature of upwelling hyporheic water if shade regimes on floodplains are altered.

A SIMULATION MODELING EXPERIMENT TO UNDERSTAND SHADE-MEDIATED
HEAT EXCHANGE BETWEEN THE HYPORHEIC ZONE AND ATMOSPHERE AND
ITS EFFECTS ON STREAM TEMPERATURES

Contribution of Authors and Co-Authors

Manuscript in following chapter

Author: S. Kathleen Fogg

Contributions: Conducted research experiment, wrote HGS model, analyzed results and wrote manuscript

Author: Geoffrey C. Poole

Contributions: Advised on research analysis and model set-up, helped interpret results, edited manuscript

Author: Scott J. O'Daniel

Contributions: Provided essential data on Meacham Creek, provided funding through the CTUIR, helped interpret results, edited manuscript

Author: Byron A. Amerson

Contributions: Provided model code, advise and help on simulating water and heat flow in HGS

Manuscript Information

S. Kathleen Fogg, Geoffrey C. Poole, Scott J. O'Daniel, and Byron A. Amerson

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Abstract

Promoting hyporheic exchange in gravel-bedded floodplain rivers can lead to buffered and cooled channel temperature regimes. Hyporheic zones of these floodplain streams often have a wide lateral extent and occur at a shallow depth beneath the ground surface, allowing for shade on the broader floodplain to influence ground surface temperatures, thus affecting hyporheic and channel temperatures. We hypothesized that shade on floodplain soils that overlie lateral hyporheic water influences the conductive heat flux through the vadose zone, altering hyporheic zone temperatures and therefore altering stream channel temperatures. To test our hypothesis, we conducted a simulation modeling experiment in which we varied the areal percentage of floodplain shade (0%, 50% and 100%) and channel shade (0% and 100%), while holding all other parameters and driving variables constant. Results from our experiment showed that increasing levels of floodplain shade cooled hyporheic and stream channel temperatures, regardless of whether the channel was fully shaded (0%) or unshaded (100%). We found floodplain shade had a similar potential to effect stream temperatures as channel shade—increasing channel shade from 0% to 100% cooled summertime channel temperatures an average of 6.3°C, whereas increasing floodplain shade 0% to 100% caused an average of 6.0 °C of channel cooling. Our results show how floodplain shade can be an important component of temperature dynamics in streams with expansive, coarse-grained hyporheic zones and how the management of floodplain forests may be crucial in managing and maintaining cold-water habitat.

Introduction

Changes to the climate and landscape have caused warming of average stream channel temperatures [Hester and Doyle, 2011, Poff et al., 2002]. The warming trend of stream temperatures is especially detrimental to the ecological functioning of cold-water river

ecosystems where warm water temperatures may cause hypoxic conditions [Helton et al., 2012b, Mackay and Fleming, 1969], exacerbate harmful algae blooms [Elliott, 2012], increase the prevalence of disease [Harvell et al., 2002, Marcos-López et al., 2010] and the toxicity of pollutants [Peralta-Maraver et al., 2018]. These stressors, among others, can threaten the survival of native cold-water species—a concern widely recognized by managers and researchers of rivers in the intermountain western U.S. [Al-Chokhachy et al., 2016, DeBano et al., 2016, Eliason et al., 2011, Fuller et al., 2022].

The primary restoration tool river managers use to reduce channel temperatures is to increase channel shade by planting vegetation near the stream bank [Katz et al., 2007, Wondzell et al., 2018]. Shade restoration works well in reaches where the flow regime and channel geometry allow for mature stream-bank vegetation but can be insufficient in montane streams with dynamic floodplains (hereafter referred to as “floodplain streams”) where snow-melt hydrology causes springtime floods and a wide annual scour zone that prevents the establishment of mature bank-side vegetation [Huggenberger et al., 1998]. Wide annual scouring limits the viability and sustainability of planting stream-bank vegetation to increase shade and cool stream channel temperatures. In floodplain stream reaches, most vegetative shade occurs away from the wetted channel at low flows and is assumed to have little to no effect on stream channel temperatures.

Floodplain streams of the intermountain western U.S. are often gravel-bedded, with large hyporheic zones that extend laterally far from the channel, providing a connection between the channel and broader landscape, including floodplain forests [Hauer et al., 2016]. Lateral hyporheic flows can extend up to 100s of meters away from the low flow channel, e.g., in the well-characterized hyporheic zones of the Middle Fork Flathead River in western Montana [Stanford and Ward, 1993] and the Umatilla River in eastern Oregon [Poole et al., 2008]. In addition to the broad lateral extent, the water tables of these hyporheic zones tend to occur at a shallow depth (often less than 4 m from the ground surface) [Helton et al.,

2012a, Poole et al., 2008]. Hyporheic flows in these expansive alluvial aquifers traverse areas of the floodplain outside of the parafluvial zone and thus flow beneath mature floodplain forest vegetation, where shade may be a strong control on the thermal dynamics of shallow hyporheic water.

High rates of hyporheic exchange occurring in gravel-bedded floodplain reaches promote cooler daily mean temperatures in the summer and reduce temperature ranges of both daily and seasonal temperature cycles [Arrigoni et al., 2008, Fogg et al., 2023]. The amount of shade occurring on the broader floodplain has the potential to affect hyporheic temperatures, and therefore, the amount of floodplain shade could hinder or help the ability of hyporheic exchange to cool and buffer channel temperature cycles. An investigation by Vogt et al. [2012] of the River Thur in Switzerland showed that substantial heat exchange occurs between the unsaturated vadose zone and river water infiltrating into the riparian zone, thus altering diel temperature cycles of hyporheic water. However, the relationship between vegetative shade overlying the lateral hyporheic zone and its potential influence on stream channel temperatures via altering dynamics of the hyporheic thermal energy budget has yet to be explored.

We hypothesize that the amount of shade on the floodplain surface alters the hyporheic heat zone heat budget via conductive heat flux through the vadose zone, altering hyporheic water temperatures and thus altering stream channel temperatures in floodplain reaches with high hyporheic exchange rates. We conducted a modeling experiment using a 3-dimensional numerical simulation model of Meacham Creek—a montane floodplain reach located in eastern Oregon, USA, where exceedingly warm temperatures are the primary water quality concern [Powell, 2001]. Holding channel morphological structure and floodplain flow conditions constant, we tested how temperatures in shaded and unshaded stream channels responded to various levels of shading of the channel and the floodplain surface overlying hyporheic flow paths.

Methods

We used HydroGeoSphere (HGS) [Brunner and Simmons, 2012] to simulate the hydrologic behavior and thermal dynamics of the Meacham Creek floodplain. HGS is a 3-dimensional multiphysical finite element solver that couples surface and subsurface transport of water and heat, simulates unsaturated conditions, advective, dispersive, and conductive heat fluxes in the subsurface, and atmospheric heat exchanges with soils and water surfaces. We varied shade distribution on the model surface to investigate floodplain shade influences on hyporheic and channel temperatures. We simulated steady-state, low flow hydrologic conditions, thus eliminating any stream channel thermal responses due to variation in channel of subsurface flow patterns.

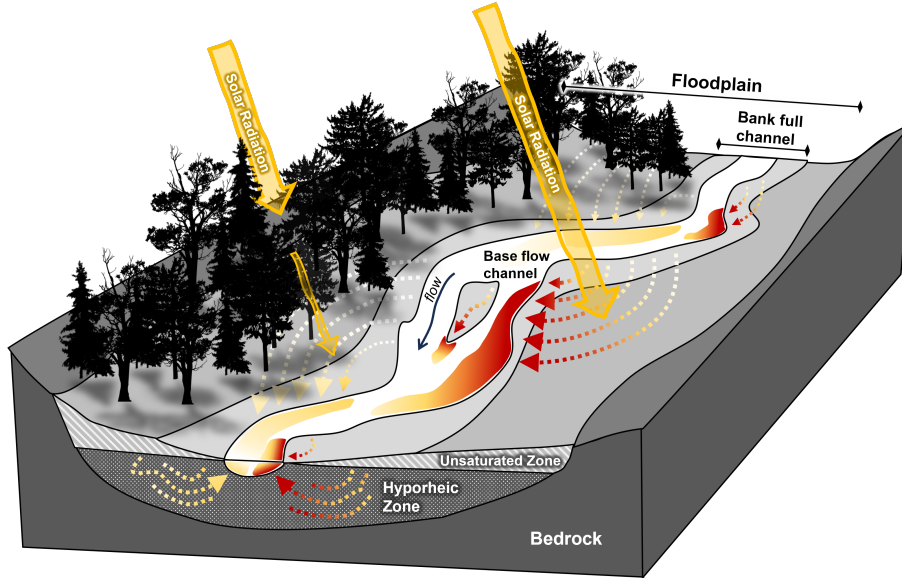


Figure 3.1: Conceptual model of how shade affects heating of lateral hyporheic flow paths and upwelling water to the stream channel. The left side of the image shows an intact riparian forest, which provides shade, thus attenuating solar radiation and leading to less warming of lateral hyporheic flow paths (dotted lines) than the right side of the floodplain image which is unvegetated, thus allowing for un-attenuated solar radiation to heat lateral hyporheic flow paths, thus causing areas of hotter upwelling water in the stream channel than the flow paths flowing beneath the shaded side of the floodplain.

Experimental System

Coarse-grained alluvial floodplains of mountainous regions often occur in previously glaciated valleys constrained by bedrock. The base-flow stream channel in these valleys is inset within a wider annual scour zone (i.e., bank-full channel), that is set within a broader floodplain that floods inter-annually (Figure 3.1) [Stanford et al., 2005]. Each year, young vegetation establishes within the scour zone, only to be removed by high-energy spring flooding before it can mature [Bywater-Reyes et al., 2015, Hauer et al., 2022, Wilcox and Shafroth, 2013]. Accordingly, mature vegetation is often only found beyond the annual scour zone. In these systems, hyporheic exchange is extensive, and lateral hyporheic flow paths traverse meander bends, point bars, and gravel bars and occur at a shallow depth beneath the ground surface. We propose that this geometry allows for differential heating of hyporheic water beneath floodplains that are shaded (left bank of the conceptual model in Figure 3.1) versus floodplains that are not shaded (right bank, Figure 3.1), leading to warmer upwelling water temperatures in floodplains with less vegetative shade, and therefore warmer channel temperatures.

Meacham Creek and its associated floodplain epitomize a coarse-grained alluvial floodplain system akin to the conceptual model in Figure 3.1 and is thus used as our experimental system. Meacham Creek is an alluvial floodplain stream with high gross hyporheic exchange rates. It is located in the Blue Mountains east of Pendleton, Oregon (OR), U.S.A., and flows south to north over an alluvial floodplain composed of coarse-grained basaltic substrate [Hogenson, 1964]. The bedrock of the mountain valley constrains both the sides and bottom of the alluvial aquifer of the Meacham Creek floodplain, and the entire alluvial aquifer is considered the hyporheic zone. Springtime rain-on-snow events create annual floods orders of magnitude greater than low summer flows, creating a wide scour zone that prevents the establishment of mature vegetation.

We used a 3.5-km section of Meacham Creek for our modeling experiment (Figure 3.2).

The section labeled ‘Analysis Reach’ has been extensively studied because it underwent a restoration that returned the channel to the center of its historic floodplain. A spring channel remains against the western valley wall, where the pre-restoration stream once flowed. Data from a Light Detection and Ranging (LiDAR) flight conducted in 2013 was used to define an elevation model of the ground surface and to replicate actual shade patterns on the floodplain in the verification of our model (Section 3). Groundwater monitoring wells were installed at different locations in the Meacham Creek floodplain in 2011 (Figure 3.2), and hyporheic temperature data were collected in each well, recording at an hourly interval somewhat continuously between 2011 and 2015. These hyporheic temperatures were used to verify our model’s hydrologic and thermal parameters. Solar radiation data (W m^2) were collected at sunny and shaded sites using ONSET pyranometers installed on a weather station in the summer of 2015. We used these data to estimate shade attenuation of typical vegetation found on Meacham Creek and to reduce a satellite-derived clear-sky radiation signal to a ‘shaded’ solar radiation signal (Section 3). Air and water temperatures were collected between June 2020 and September 2021 at hourly intervals using Onset HOBO TidbiT temperature data loggers. Water temperatures were recorded in the thalweg of Meacham Creek, and air temperatures were recorded at a site adjacent to the channel. We used these data to create archetypal temperature time series for model driving variables (see Section 3).

Simulation Model

Model Grid We constructed a model of the Meacham Creek floodplain in HGS using 5 m by 5 m grid cells. The model domain encompassed the entire width of the Meacham Creek floodplain and extended upstream 1 km and downstream 900m of our area of interest to minimize the influence of upstream and downstream boundary conditions (‘Analysis Reach’, Figure 3.2). We applied bare-earth elevations collected during the 2013 LiDAR flight for the

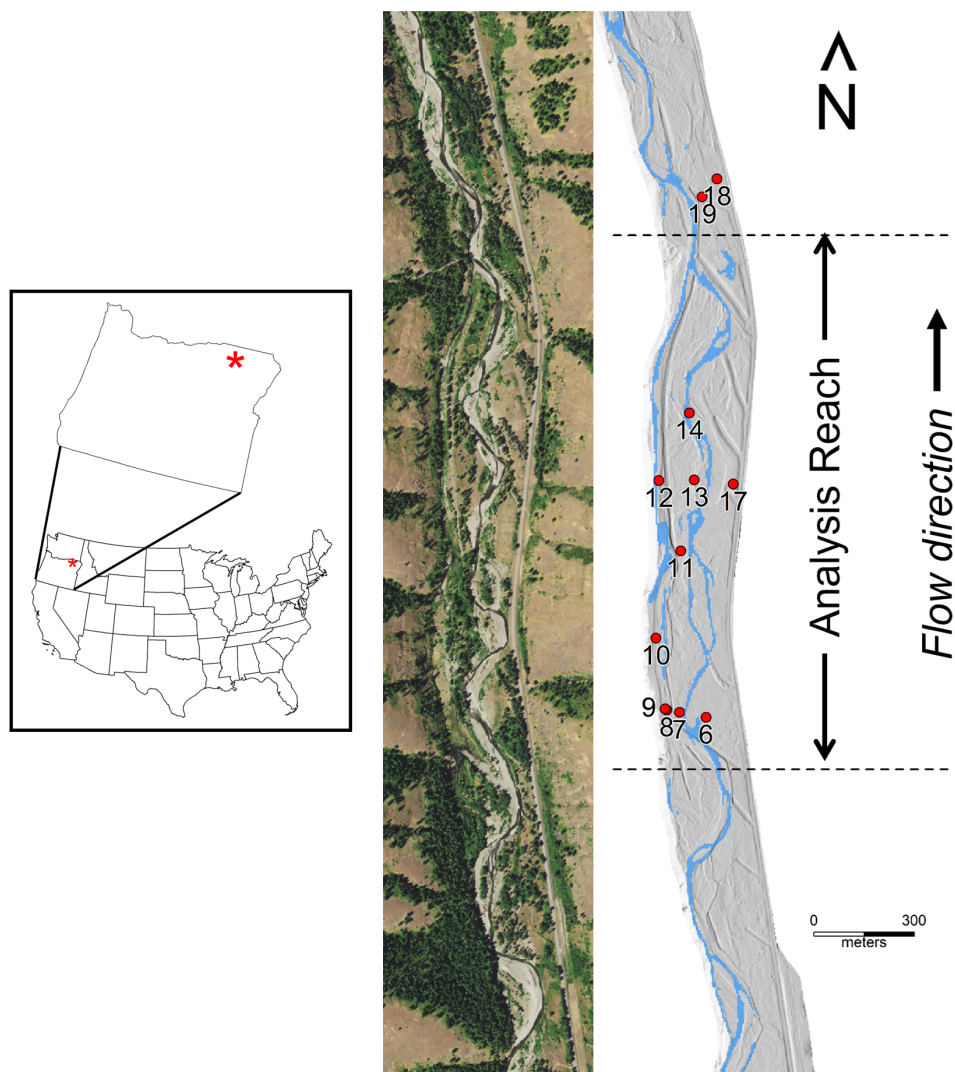


Figure 3.2: Location of Meacham Creek in eastern Oregon, USA (left), orthoimage of modeled area (2020) (center), and hillshade model of Meacham Creek floodplain based on LiDAR elevation data collected in 2013. Base-flow wetted channel of model output is shown in blue on the hillshade depiction. Red circles show the locations of floodplain wells.

surface elevations of our model. Elevation data were collected at a 1-m by 1-m resolution. The elevations of HGS model nodes were defined by bilinear interpolation from the raster file.

For the model subsurface, we created a porous media zone, bounded by a bedrock zone underneath to represent the Meacham Creek hyporheic zone. The layer differentiating the porous media zone from the bedrock was a plane with the same average slope as the stream channel. Floodplain surface elevations were variable; therefore, the thickness of the porous media zone varied between 2.5 to 5.7 meters thick. Our modeled bedrock zone was 25 meters thick to ensure the complete vertical attenuation of annual temperature amplitude to an annual mean temperature.

Materials Properties Thermal and hydrologic properties of the bedrock and alluvium materials were representative of coarse-grained basaltic alluvium and are displayed in Table 3.1. Table 3.2 outlines overland flow properties. We used Manning’s roughness coefficient (n) of 0.043 for the stream channel—a value estimated on the Grand Ronde River at La Grande, Oregon [Barnes, 1969]. We used the published value of n of the Grand Ronde River because it is located in the same region of eastern Oregon as Meacham Creek and has similar coarse-grained bed materials.

Boundary Conditions We used steady-state hydrology representing the summertime base-flow channel of Meacham Creek (Figure 4.1). The specified flux at the upstream model boundary was 0.274 m s^{-3} for the stream channel and $4.62 \times 10^{-5} \text{ m s}^{-3}$ for the alluvium (this value represents a darcy flux of 400 m day^{-1} Poole et al. [2008], a value that resulted in approximate equilibrium of exchange between ground and surface water in the model). We applied a specified flux to the upstream boundary of the bedrock layer at $3.162 \times 10^{-10} \text{ m sec}^{-1}$. We used a critical depth boundary condition for the surface nodes of the downstream model boundary and a specified head boundary for the downstream subsurface nodes.

Alluvium

Parameter	Value	Unit	Cite
Longitudinal hydraulic conductivity	0.0046	m s ⁻¹	Poole et al. [2008]
Transverse hydraulic conductivity	0.00046	m s ⁻¹	1/10 Longitudinal; Freeze and Cherry [1979]
Longitudinal dispersivity	1.4	m	Rau et al. [2012]
Transverse dispersivity	0.4	m	Rau et al. [2012]
Thermal conductivity	1.6	W m ⁻¹ K ⁻¹	Waples and Waples [2004]
Specific heat capacity	840	J kg ⁻¹ K ⁻¹	Waples and Waples [2004]
Density	3000	kg m ⁻³	Waples and Waples [2004]
Porosity	0.25	-	Freeze and Cherry [1979]
Van Genuchten α	22.0	m ⁻¹	Thoma et al. [2014]
Van Genuchten n	1.79		Thoma et al. [2014]
Residual saturation	0.04	-	Thoma et al. [2014]
Minimum relative permeability	1.995e-06		Thoma et al. [2014]

Bedrock

Parameter	Value	Unit	Cite
Hydraulic conductivity (isotropic)	1.0e-10	m s ⁻¹	Freeze and Cherry [1979]
Dispersivity (isotropic)	1.0e-20	m s ⁻¹	(value suitably close to zero)
Thermal conductivity	1.6	W m ⁻¹ K ⁻¹	Waples and Waples [2004]
Specific heat capacity	840	J kg ⁻¹ K ⁻¹	Waples and Waples [2004]
Density	3000	kg m ⁻³	Waples and Waples [2004]
Porosity	0.04		Zakharova et al. [2012]

Table 3.1: Material Properties

Driving Variables We used idealized air and water temperature time series as model driving variables to eliminate the possibility that differences in model scenario thermal response were caused by atmospheric and upstream conditions. We created annually-repeating best-fit compound sine-waves using the observed air and water temperatures collected on Meacham Creek (Figure 3.3). Parameter values for the compound sine wave models were estimated using the non-linear least-squares, ‘nls()’, function in R [R Core Team, 2021].

The air temperature signal was applied to the atmospheric heat transfer component of all model surface nodes. The water temperature signal was applied to the in-flowing water at the upstream boundary of the stream channel, hyporheic zone, and bedrock nodes.

Experimental Design

We used a 2×3 factorial experimental design for our simulation experiment with two levels of shade on the stream channel, 0% and 100%, and three levels of shade on the floodplain, 0%, 50%, and 100% (Figure 3.4). Experimental shade percentages refer to the percent of the surface area covered by shade, not the percent reduction of a clear-sky solar radiation signal (described below). For the 50% floodplain shade scenarios, we assigned shaded cells following a random binomial distribution.

The six scenario shade maps (Figure 3.4) were used to apply different solar radiation time series to HGS model nodes. A clear-sky solar radiation signal was applied to model nodes that fell within an “unshaded” map cell (Figure 3.5). A vegetation-attenuated solar radiation signal was applied to nodes within map cells classified as “shaded”.

The clear-sky solar radiation data used in our model was a land surface, direct shortwave radiation flux data product (SYN1degEd41) from the National Aeronautics and Space Agency’s (NASA) Clouds and the Earth’s Radiant Energy (CERES) project [Rutan et al., 2015]). This data product is a calculation of the clear-sky shortwave radiation flux on the

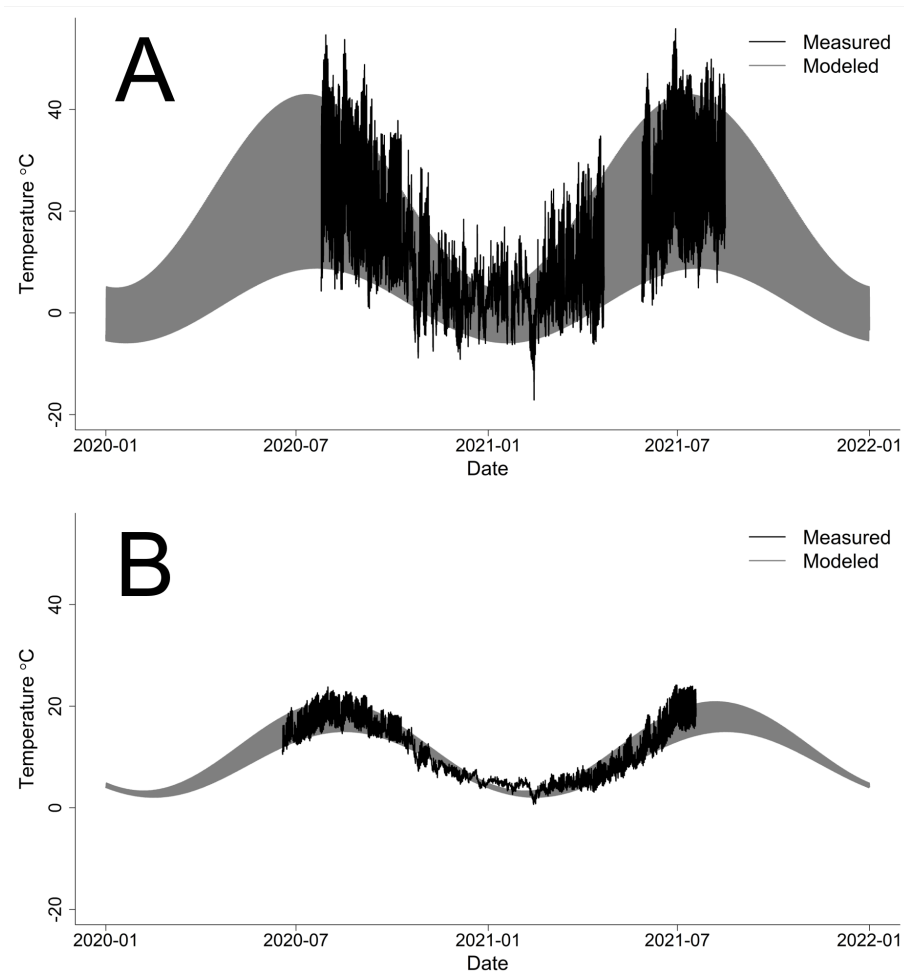


Figure 3.3: Air temperature (A) and water temperature (B) used as input to model simulations. Input temperatures were compound sine-wave models (labeled ‘Modeled’) of measured temperatures taken from a weather station on the Meacham Creek floodplain (A, air temperature) and in the thalweg of Meacham Creek (B, water temperature) between June 2020 and September 2021.

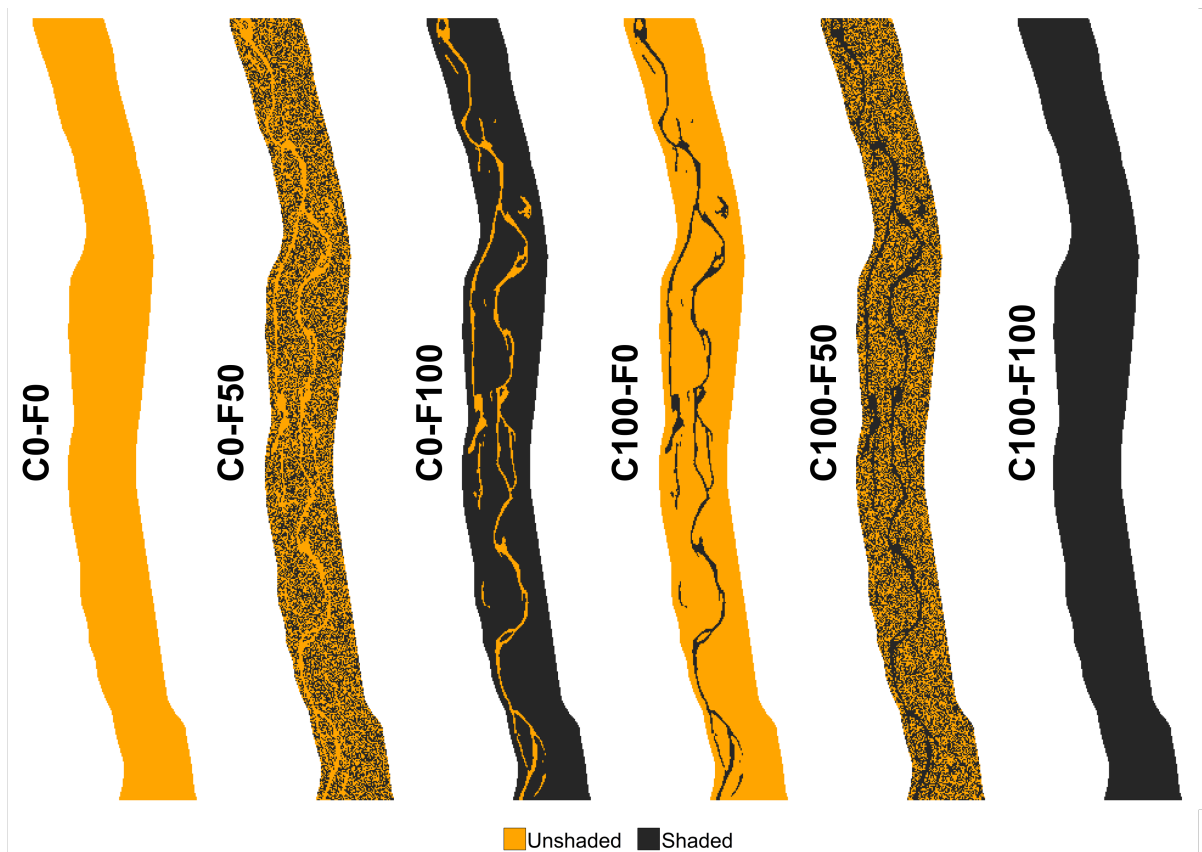


Figure 3.4: Model scenarios depicting the pattern and percentages of shade on the floodplain and channel. From left to right, the first 3 model scenarios have 0% channel shade (C0) and increasing floodplain shade (F0, F50, F100). The last 3 model scenarios have 100% channel shade (C100) and increasing floodplain shade from left to right (F0, F50, F100).

ground surface at a $1^\circ\text{latitude} \times 1^\circ\text{longitude}$ resolution, calculated at an hourly interval. We used radiation data from the year 2020 at a data grid that included the analysis reach of the Meacham Creek site, $(45.379, -118.611) - (45.879, -118.111)$.

We estimated a vegetation-attenuated solar radiation signal from observed solar radiation values at a shaded and unshaded site on Meacham Creek. The data were collected using silicon pyranometers (HOBO brand, model S-LIB-M003) installed on weather stations, which concurrently logged solar radiation values every 15 minutes at the unshaded and shaded site from 31 July 2015 through 2 August 2015. The shaded site was located in a cottonwood-conifer stand and was shaded in the afternoon hours (12:00-17:00); therefore, only data taken between 12:00 and 17:00 were used in the linear regression shown in Figure 3.5A. We used this linear model to reduce the daytime solar radiation values of the SYN1degEd41 signal used for the “unshaded” cells (displayed in yellow; Figure 3.5B) to create the shade-reduced signal used for “shaded” cells (black; Figure 3.5B).

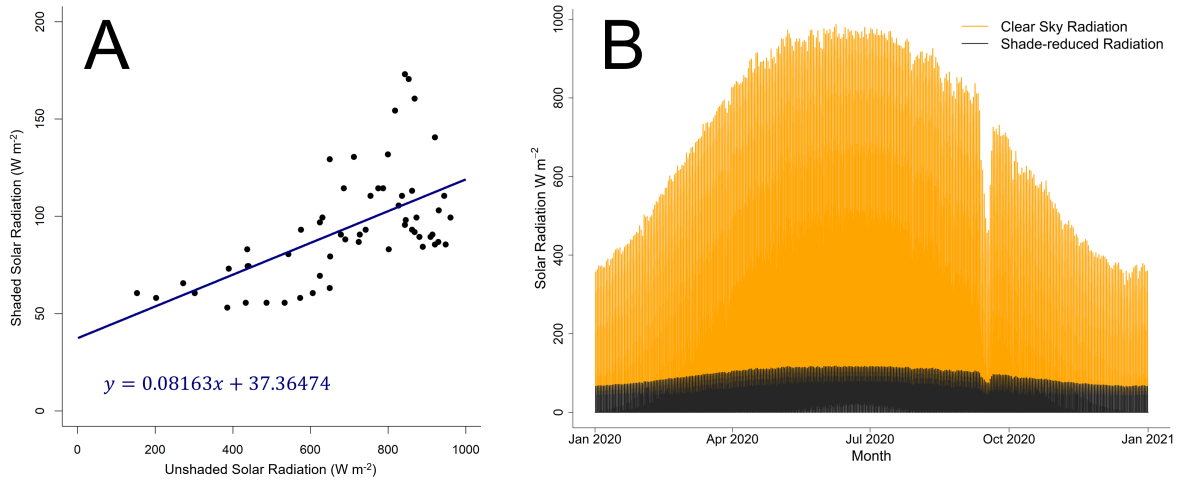


Figure 3.5: Calibration curve of shaded solar radiation (A) and solar radiation model input (B). Yellow is the signal used for clear-sky (0% shade) solar radiation at Meacham Creek and the black line shows the shade-reduced signal, created from the linear relationship on the left. The linear model predicts shaded solar radiation from sunny solar radiation, both measured on Meacham Creek.

Verification Scenario

Confidence in the ability of our simulation model to approximate the potential effects of floodplain shade on hyporheic and channel temperatures relies on having model parameters that represent the physics of study systems similar to Meacham Creek. To assess model parameters, we verified that the model would produce reasonable hyporheic zone temperature dynamics under real-world conditions. To make this assessment, we compared field observations from wells on the Meacham Creek floodplain (Figure 4.1) to model results from a simulation using the observed distribution of shade on the floodplain (Figure 3.6).

We used LiDAR data from Meacham Creek (2013) to estimate the pattern of shade on the floodplain (Figure 3.6). We created a binary raster of shaded and unshaded cells by first calculating tree heights by subtracting LiDAR-derived bare-earth elevations from the highest hits elevations. Then, we filtered out all tree heights less than 3 meters to simulate only mature vegetation that contributed substantial shade. The remaining cells in the raster were considered “shaded”. We aggregated the 1-m resolution binary raster of shade cells from LiDAR to a 5-m resolution raster for the HGS model by calculating the percent shade coverage of each 5-m grid cell. If the 5-m cell had 50% or greater shaded cells, then that cell was assumed to be “shaded” in the 5-m grid. As in the experimental scenarios, we left the base-flow channel of Meacham Creek unshaded in the verification shade raster because Meacham Creek has minimal vegetation on the base-flow channel bank. Apart from the shade pattern applied to the model surface, all other model parameters and driving variables were the same as the experimental modeling scenarios.

Initialization and Spin-up

We initialized the model with heads equal to the surface elevation at all node depths. HGS solved for the correct head values for each node given the flow conditions, and we used these head values to initialize the simulation of each experimental scenario and the

Floodplain

Parameter	Value	Units	Cite
x-, y- friction	0.3	$\text{m}^{-1/3} \text{ s}$	ABDUL (?)
Rill storage height	0.005	m	

Wetted Channel

Parameter	Value	Units	Cite
x-, y- friction	0.043	$\text{m}^{-1/3} \text{ s}$	Barnes [1969]
Rill storage height	0.015	m	

Table 3.2: Overland Flow Properties



Figure 3.6: Pattern of floodplain shade used for the verification of model parameters. Shade pattern is estimated from 2013 LiDAR data.

verification run. The hydrologic component of the model ran in steady-state; therefore, no spin-up was required.

All model node temperatures were initialized at 10.3 °C, the mean annual input channel water temperature. Each model was simulated for eight years to allow for spin-up of temperatures. Temperatures of the models were output every 30 days beginning on January 1 until 12 output days were reached. Within each day, output was recorded at each hour, thus yielding 288 output times.

Model Analysis

Verification We compared measured well temperatures to modeled well temperatures as the verification metric. Well locations on Meacham Creek are shown in Figure 3.2, and modeled “wells” (points in the simulated aquifer for which output was recorded) were created in the same locations. Modeled wells recorded temperatures approximately 0.5 m above the bedrock model layer in the saturated porous media zone of the model.

Experimental Scenarios We calculated daily mean temperatures for all model nodes for all 12 output days. We summarized and analyzed the modeled temperatures for three components of the floodplain: surface water, hyporheic water, and the ground surface. We delineated surface water from the ground surface using water depth. If the model node had a water depth greater than 1 cm, it was considered part of the channel. Otherwise, the node was considered part of the floodplain surface. We extracted hyporheic zone temperatures from the porous media zone where the simulated pressure head was equal to 1. This iso-surface was located between the bottom of the hyporheic zone and the hyporheic water table.

We examined temperature maps of the floodplain ground surface temperature and hyporheic zone iso-surface for two days of the year, one day representative of summer, 29 July, and the other day representative of winter, 31 January. We observed the simulated

change in stream channel temperature from upstream to downstream across the analysis reach according to the locations labeled in Figure 3.2.

Results

Verification of Model Parameters

Our Meacham Creek model showed similar patterns in simulated versus observed daily mean hyporheic well temperatures measured in the hyporheic zone of Meacham Creek (Figure 3.7). Simulated temperatures were closest to the observed temperatures for the annual minimum of daily mean temperatures (averaged across sampling wells). However, simulated annual maximum of daily mean temperature ($X^{\circ}\text{C}$) was cooler than the observed maximum temperature (Figure 3.7). Simulated diel variation in temperatures were similar to those of observations except for late in the year (ca. 200–250 day of year) when simulated temperature ranges were less than observed diel ranges.

Simulated temperatures in wells mimicked the annual patterns of observed well temperatures (Figure 3.8), although observed well temperatures generally show a modestly greater annual lag in the sinusoidal temperature signal than the simulated signal. The largest temperature differences between the observed and simulated well temperatures occur at the maximum and minimum of the annual sinusoidal signal, where the simulated amplitude was typically lower than the observed amplitude (Figure 3.8).

Effects of Shade on Temperatures

Regardless of location (on the channel or the floodplain), hyporheic and channel temperatures were inversely related to percent shade cover in our simulations. Channel shade and floodplain shade caused the temperature patterns observed on the ground surface, in the hyporheic zone and stream channel (Figures 3.9–3.12). Surfaces that were assigned a ‘shaded’ cell and therefore received a reduced solar radiation signal were cooler than surfaces that

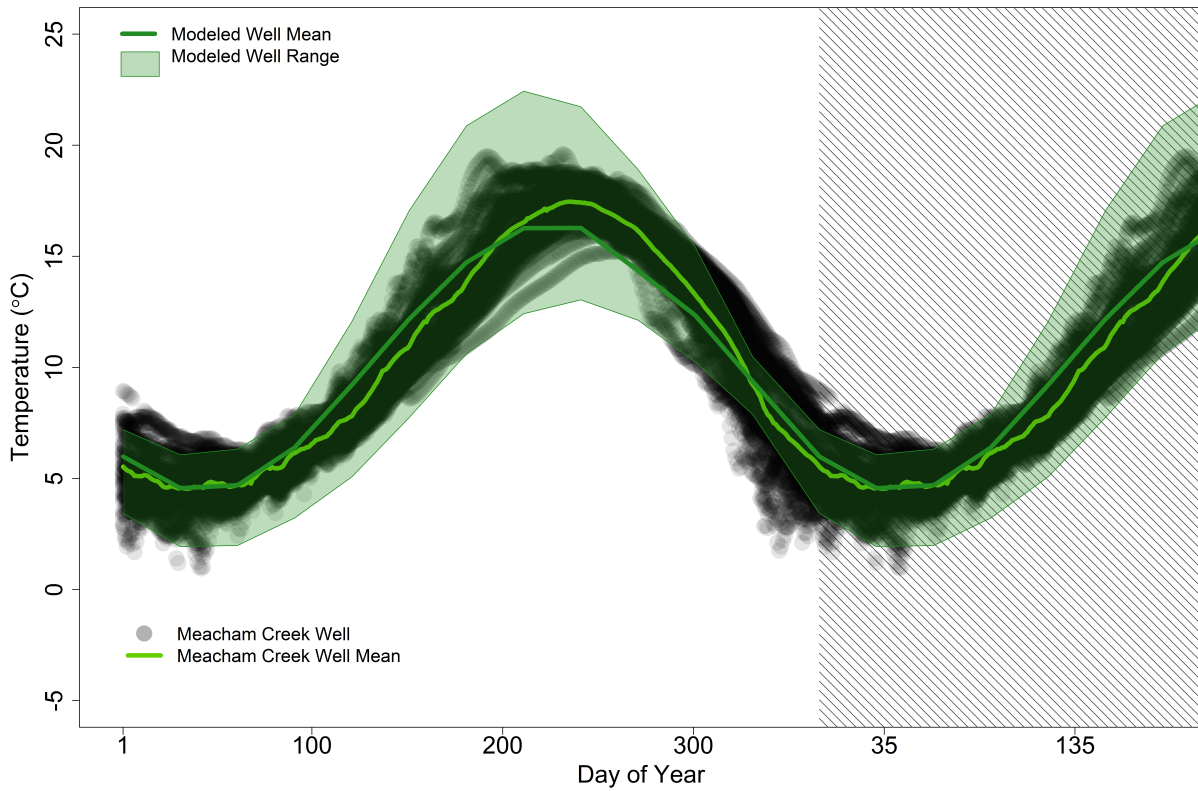


Figure 3.7: Observed hyporheic well temperatures on the Meacham Creek floodplain compared to simulated well temperatures using verification shade scenario. Measured temperatures and day-of-year mean are shown as grey points and solid light green line. The simulated day-of-year mean well temperatures are displayed as solid lines, day-of-year range in well temperatures is shown as translucent polygon around the mean line. One-hundred and eighty days of data was repeated in the shaded area of the plot to better display the annual pattern in temperatures.

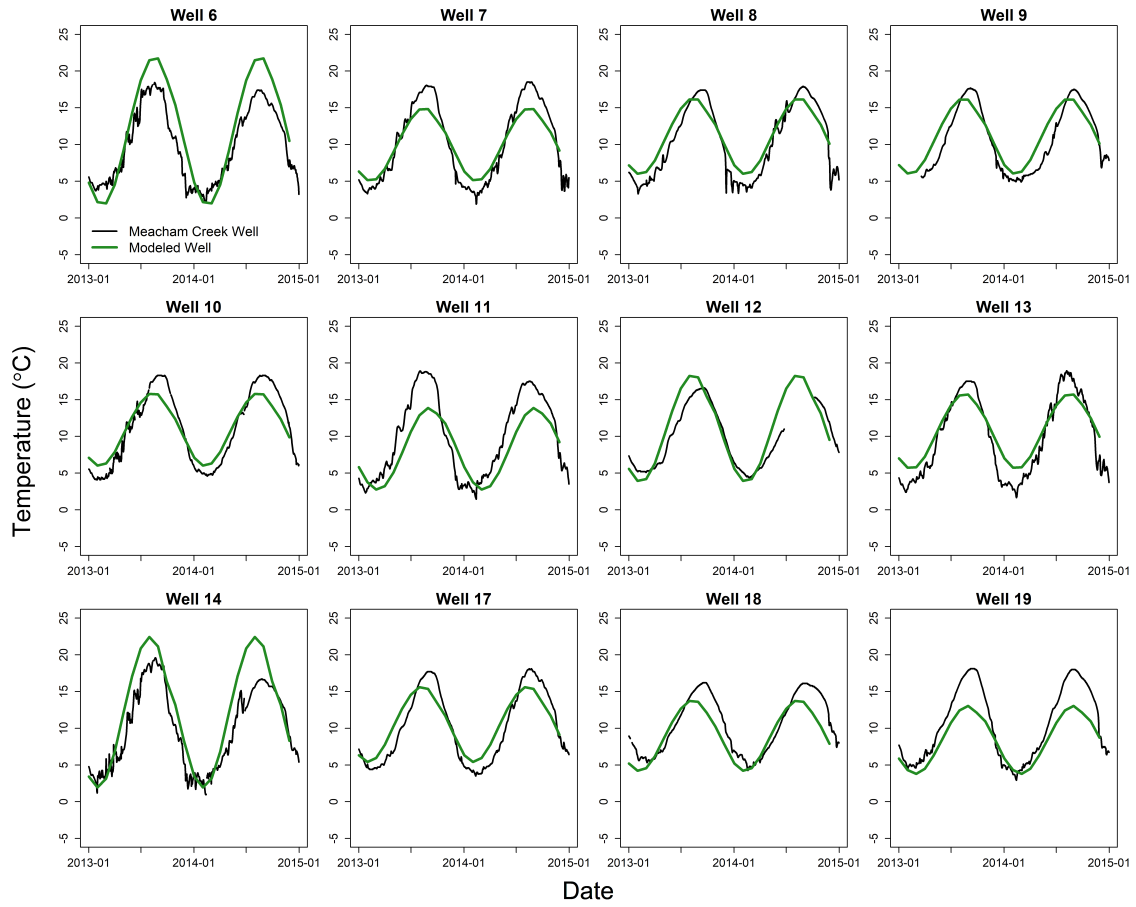


Figure 3.8: Comparison of measured Meacham Creek well temperatures to well temperatures simulated using verification shade pattern on the floodplain. Measured well temperatures are displayed in black and modeled well temperatures are displayed in green.

were assigned ‘unshaded’ and received the clear-sky solar radiation signal (Figures 3.9, 3.11). The cooler surface temperatures of shaded areas caused cooler subsurface temperatures in the vadose and hyporheic zones of our simulation models (Figures 3.10, 3.12).

Varying channel shade influenced simulated channel temperatures but had little influence on ground temperatures. Increasing channel shade from 0% to 100% caused an average reduction of 4.8°C of daily mean channel temperatures on the summer day (29-Jul) and 1.6°C on the winter day (31-Jan) across scenarios with the same floodplain shade (Figure 3.13). The amount of channel shade had minimal influence on ground temperatures of the floodplain, where increasing channel shade from 0% to 100% caused a maximum reduction of daily mean ground temperature of 1.3°C (Figure 3.13A).

In contrast to channel shade, floodplain shade affected both the ground temperature and channel temperatures in our simulations (Figures 3.9, 3.11, and 3.13). Increasing floodplain shade by 50% caused an average decrease in daily mean ground temperature of 5.7°C in summer (29 July) and 2.7 °C in winter (31 January). Channel temperatures decreased by an average of 3.0°C for each 50% increment increase in floodplain shade in summer and 1.6 °C in winter.

Similar to the response of ground temperatures to floodplain shade, hyporheic temperatures decreased when floodplain shade increased (Figures 3.9–3.12). A 50% increase in floodplain shade yielded a 4.3°C reduction in summertime mean hyporheic temperature and a 3.0°C reduction in mean winter hyporheic temperature when channel shade was held constant (Figure 3.13).

In summer, warmer ground surface temperatures created stronger vertical thermal gradients in the vadose zone, causing heat to conduct into the cooler hyporheic zone (Figure 3.10). As floodplain shade increased, cooler ground surface temperatures led to less steep vertical thermal gradients in the vadose zone, leading to less warming of hyporheic zone temperatures (Figure 3.9, 3.10). In contrast to the summertime pattern, winter ground

temperatures were cooler than the hyporheic zone, driving heat conduction out of the hyporheic zone. Increased floodplain shade caused greater cooling of hyporheic temperatures in the winter (Figures 3.11, 3.12).

The thickness of the overlying vadose zone mediated rates of summertime warming and wintertime cooling of hyporheic temperatures. In summer, hyporheic temperatures beneath thick areas of the vadose zone were cooler than areas where the vadose zone was thinner (Figure 3.10). The opposite effect was observed in wintertime model results, where warmer hyporheic water was insulated from conductive cooling where the vadose zone was thickest (Figure 3.12).

Just as channel temperatures were influenced by upwelling hyporheic water temperature, hyporheic temperatures were influenced by downwelling channel temperatures (Figures 3.9–3.12; Marzadri et al. [2013], White et al. [1987]). Conduction between the channel and streambed and advection of heat via downwelling water created the temperature gradient occurring from the channel-hyporheic interface into the hyporheic zone. This gradient is most visually apparent in the C0-F100 scenario in the summer (Figures 3.9, 3.10) and in the C100-F0 scenario in the winter (Figures 3.11, 3.12). The gradient of warm to cool in near-channel summer hyporheic temperatures in the C0-F100 scenario was caused by the lack of channel shade, which warmed channel temperatures that then down-welled into the cooler hyporheic zone—which was protected from solar heating via 100% floodplain shade (Figure 3.10). A cool-to-warm temperature gradient is seen in the wintertime model results of the C100-F0 scenario. The fully shaded channel led to cooler water temperatures, thus cooler near-channel hyporheic temperatures, gradually becoming warmer further away from the channel, where the floodplain was completely unshaded (Figure 3.12).

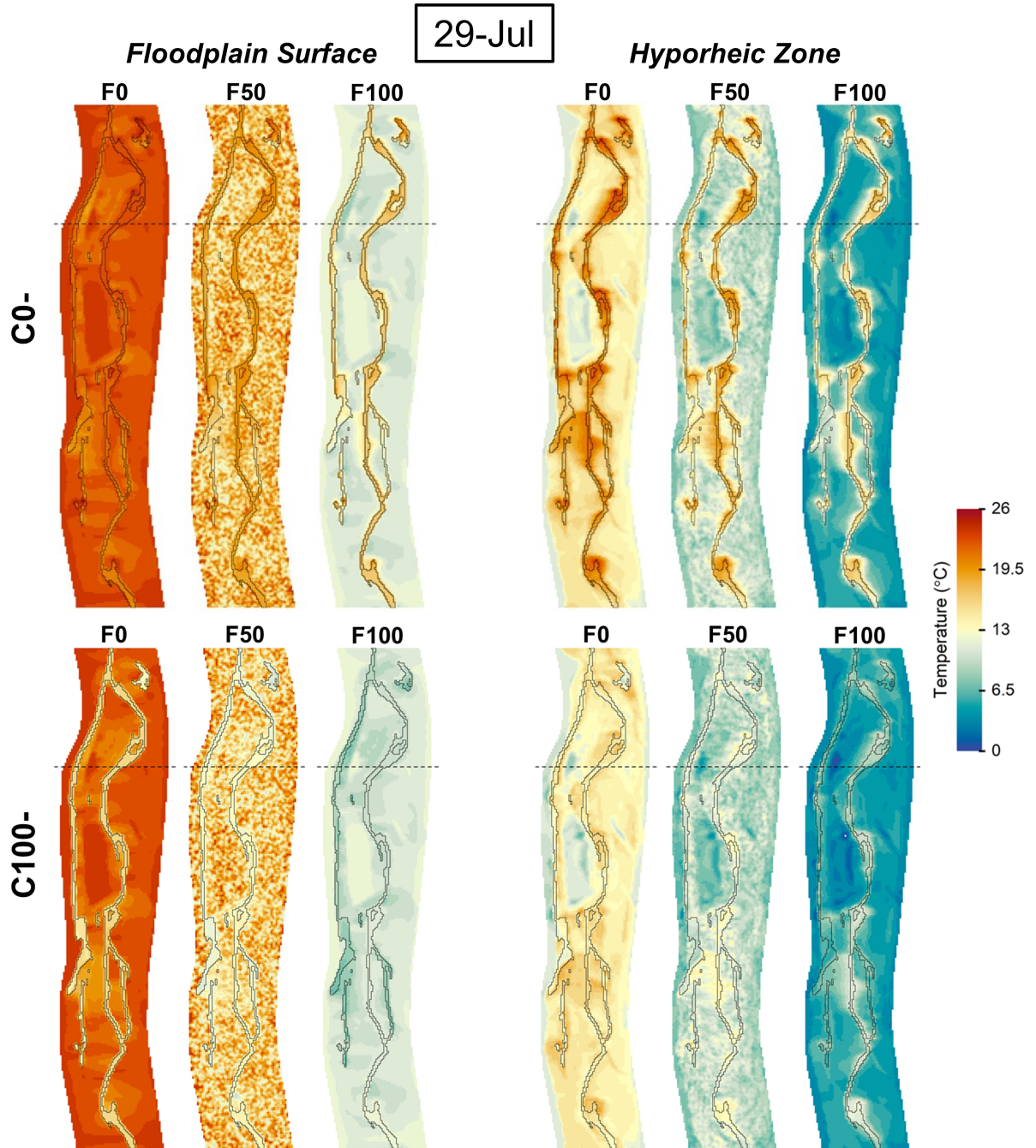


Figure 3.9: Maps of daily mean temperature output for 29-July, across the floodplain surface (left), hyporheic zone (right), and the stream channel (outlined in all). Scenarios are organized by increasing floodplain shade from left to right and increasing channel shade from top to bottom. Dotted line corresponds to the vertical slice shown in Figure 3.10.

Discussion

In general, the spatial location of shade profoundly impacted temperatures of the hyporheic zone and stream channel in our simulation experiment. Specifically, higher percentages of shade on the broader floodplain led to cooler ground surface temperatures, cooler hyporheic zone temperatures, and cooler stream channel temperatures in our Meacham Creek model (Figures 3.9–3.13). The cooling effect of floodplain shade on channel temperatures was observed across all model scenarios, whether or not the channel was 0% or 100% shaded.

Since floodplain shade did not directly influence any channel-atmospheric heat exchanges within our simulations, the only pathway that floodplain shade could have affected stream temperatures was via its effects on hyporheic temperatures and hyporheic heat exchange with the channel (Figure 3.13B). Our modeling experiment showed that increasing levels of floodplain shade cooled lateral hyporheic flows (Figures 3.10, 3.12) and the high rate of hyporheic exchange in our model of Meacham Creek (400 m day^{-1}) lead to cooler channel temperatures (Figure 3.13C). The relatively thin vadose zone (average of 1.4 m across analysis reach) allowed for heat conduction between the ground surface and the hyporheic zone. Thicker areas of the vadose zone acted as an insulating layer, reducing the influence of ground surface temperatures—and therefore floodplain shade—on hyporheic zone temperatures in the localized area (Figures 3.10 and 3.12).

In floodplain streams of the intermountain western U.S., lateral hyporheic flows play an important role in the transport of solutes, nutrients, and heat [Buffington and Tonina, 2009, Gomez-Velez et al., 2017, Wu et al., 2020]. Our representative study system typified the common geometry of lateral hyporheic zones—wide and shallow—thus allowing for hyporheic-ground surface heat exchange to occur and be mediated by floodplain shade. The temperature buffering potential of hyporheic exchange has been well studied in reached

with coarse-grained alluvium [Arrigoni et al., 2008, Arscott et al., 2001, Johnson, 2004, White et al., 1987]. More recent investigations have shown that vertical heat exchange between the vadose zone, hyporheic zone, and confining layer (often bedrock) significantly influences temperature dynamics in lateral hyporheic zones [Molina-Giraldo et al., 2011, Vogt et al., 2012]. However, these investigations lacked consideration of vegetative shade on hyporheic temperatures and the subsequent alteration of stream temperatures as our simulation experiment, presented here.

Water resource managers often focus on the restoration of stream-side shade to promote desirable thermal regimes in rivers and streams [Cormack, 1949, Fuller et al., 2022, Katz et al., 2007]. Our results showed that shading the channel did indeed have a marked effect on the accumulation of heat in surface water as it traversed the simulated channel reach (Figures 3.9, 3.11, 3.13C). As expected, if substantial channel shade were possible in streams with wide scour zones, our model predicts that channel water temperatures would be consistently cooler under complete channel shade (C100 scenarios) than when the channel is unshaded (C0 scenarios) (Figure 3.13C).

Investigations on the effects of forest harvesting on stream temperatures have shown mixed results of leaving intact riparian buffers to mitigate channel warming [Cole and Newton, 2013, Hewlett and Fortson, 1982, Janisch et al., 2012, Macdonald et al., 2003, Miralha et al., 2024, Story et al., 2003]. As an extreme example, Cole and Newton [2013] observed greater warming (up to 5.3°C) at sites *with* riparian buffers than those without. Some of these studies suggested that coarse streambed substrate is correlated with greater warming following forest harvest [e.g., Janisch et al., 2012, Miralha et al., 2024], and others proposed that heating via shallow groundwater inflows or hyporheic exchange is the cause of unexpected channel warming [e.g., Hewlett and Fortson, 1982, Story et al., 2003]. However, these studies did not directly investigate the influence that shade may have on groundwater or hyporheic temperatures or its potential effects on channel temperature.

A comparison of our experiment’s C0-F100 and C100-F0 scenarios exposes the importance of floodplain shade (Figure 3.13). Simulated stream temperatures under these two scenarios were nearly identical, showing that 100% shade on the floodplain had a similar effect on channel temperatures as did 100% shade on the channel itself. Our results suggest that the loss of floodplain shade is potentially an unrealized degradation in important salmonid-spawning streams [see, e.g., Bisson et al., 2009, Gregory and Bisson, 1997]. Accordingly, the restoration or retention of floodplain shade is an unexplored management practice for promoting cooler channel temperatures. For restorations, concomitant with an accurate diagnosis of the stream degradation [sensu Bilby et al., 2023] is understanding the mechanisms that caused the degradation. Channel-atmosphere heat exchanges are often the only mechanisms considered important to channel temperatures. However, we show that subsurface heat exchanges—even those occurring far from the channel—can also be an important driver of channel temperature.

Conclusion

We examined the effects of floodplain shade on hyporheic and stream channel temperatures using a simulation modeling experiment and found that shade on the floodplain surface led to cooler simulated temperatures of the hyporheic zone and stream channel in our model of Meacham Creek. Meacham Creek is a representative montane floodplain reach, with coarse-grained alluvium and high hyporheic exchange rates, common among reaches in the intermountain western U.S. and other mountainous areas worldwide. Our results highlight the potential influence of floodplain shade on stream channel temperatures. As rising water temperatures threaten native cold-water species in the intermountain U.S., our modeling results suggest that protecting floodplain shade in systems with expansive hyporheic zones is beneficial for maintaining cool-water thermal regimes in expansive, coarse-grained floodplain rivers.

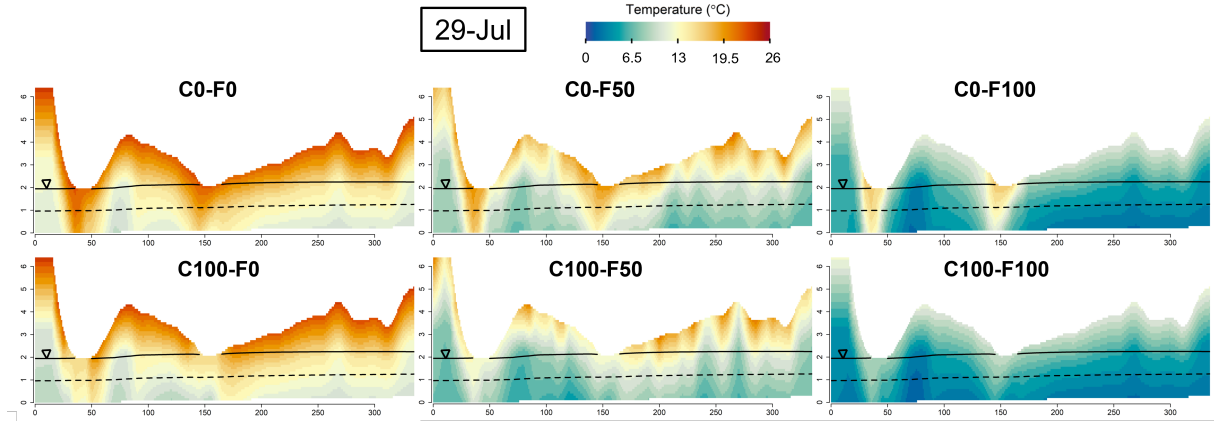


Figure 3.10: Temperature maps of a cross-sectional profile of the floodplain across all model scenarios on our representative summer day, 29-July. Images are vertically exaggerated, shown by the meter scales bars as the x- and y- axes. Solid line is the surface of the hyporheic water table, where pressure head equals 0, and the dotted line shows the location of the horizontal isosurface displayed as ‘hyporheic zone’ in Figure 3.9. Scenarios are organized by increasing floodplain shade from left to right and increasing channel shade from top to bottom.

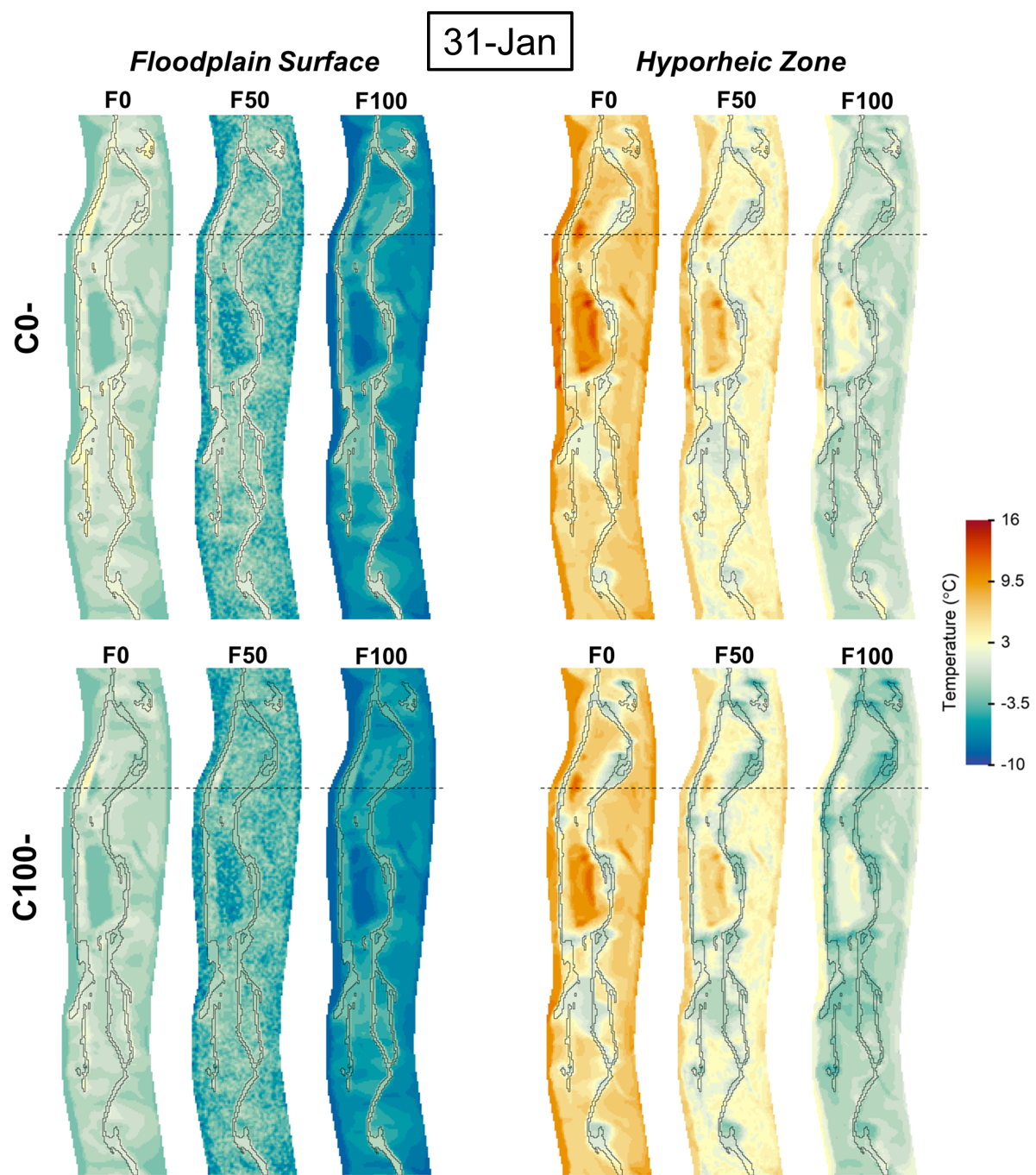


Figure 3.11: Maps of daily mean temperature output for 31-January, across the floodplain surface (left), hyporheic zone (right), and the stream channel (outlined in all). Floodplain shade increases from left to right. Results with no channel shade are presented in the top row and results with a fully shaded channel are presented in the bottom row. Dotted line shows the location of the vertical slice in Figure 3.12.

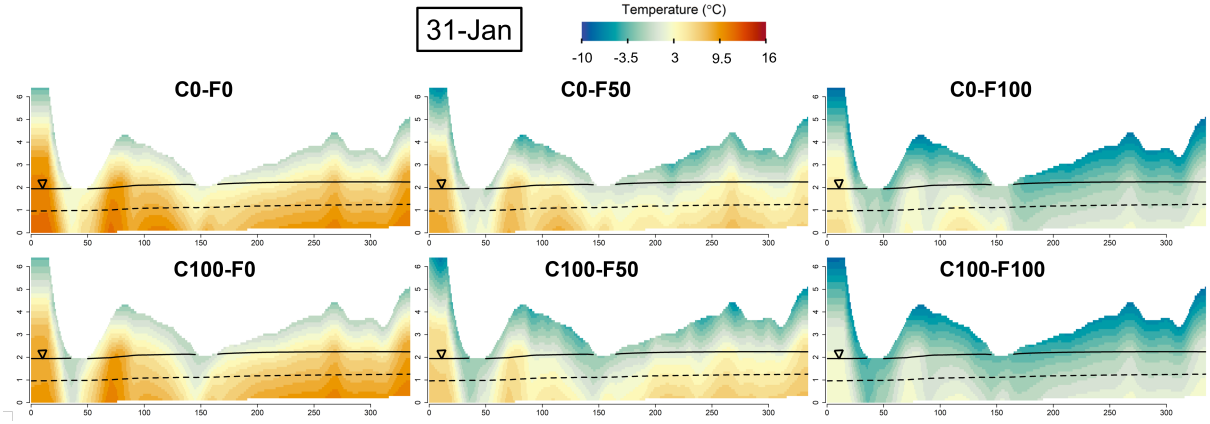


Figure 3.12: Winter temperature maps of a vertical slice of the floodplain on our representative winter day, 31-January. Images are vertically exaggerated, shown by the meter scales bars as the x- and y- axes. Solid line shows the surface of the hyporheic water table, where pressure head equals 0, and the dotted line shows the location of the horizontal isosurface displayed as ‘hyporheic zone’ in Figure 3.11. Scenarios are organized by increasing floodplain shade from left to right and increasing channel shade from top to bottom.

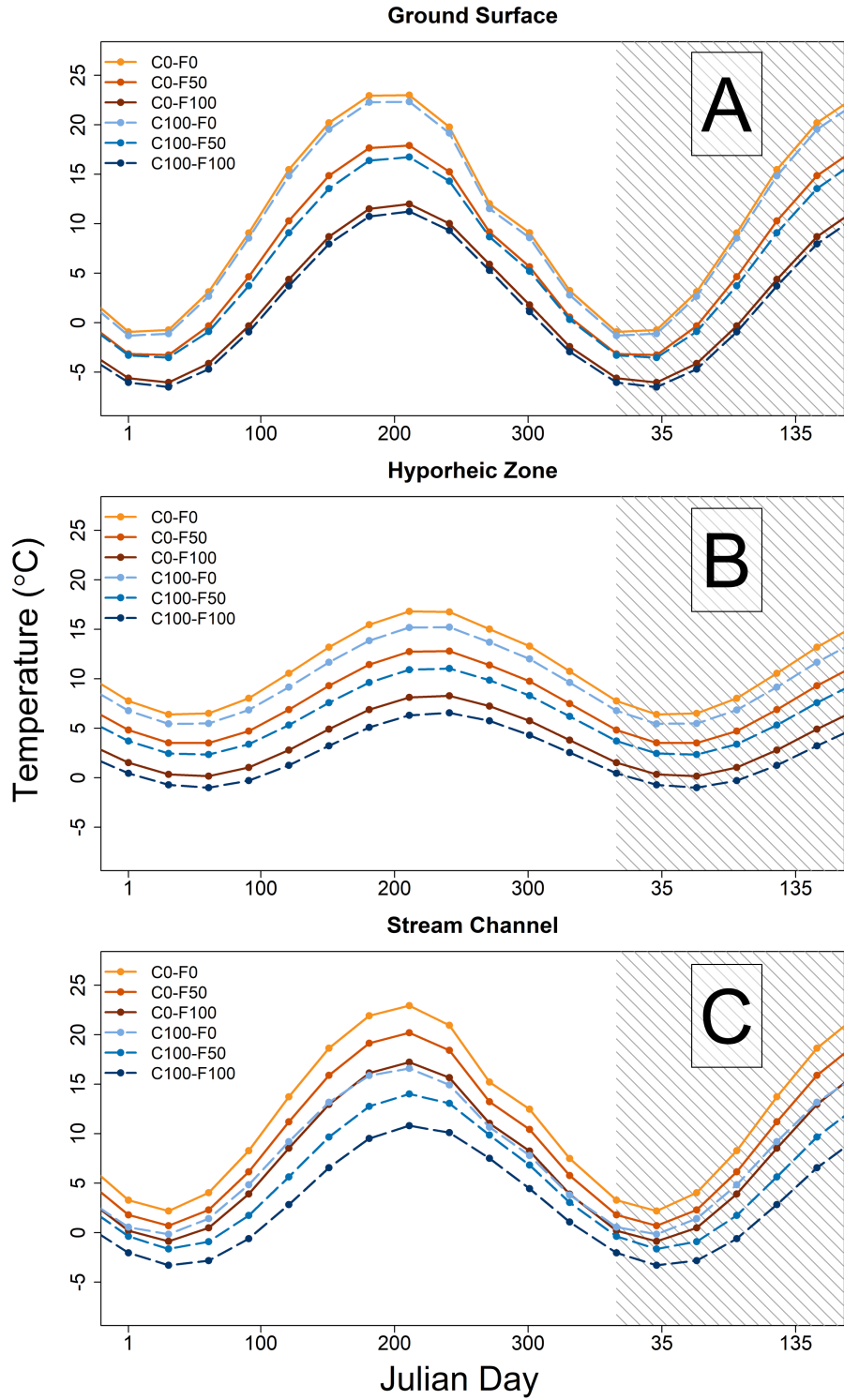


Figure 3.13: Seasonal pattern of simulated daily mean temperatures from each model scenario for the ground surface (A), hyporheic zone (B), and stream channel (C). Points are the calculated the daily mean temperature from simulation results at model output days; lines connecting points do not represent simulated temperatures. Grey shaded area indicates a repetition of the first 180 days of model output.

TRANSIENT HYDROLOGY AND FLOODPLAIN SHADE INFLUENCE CHANNEL
TEMPERATURES OF SIMULATED RESTORATION DESIGNS

Contribution of Authors and Co-Authors

Manuscript in following chapter

Author: S. Kathleen Fogg

Contributions: Conducted research experiment, wrote HGS model, analyzed results and wrote manuscript

Author: Geoffrey C. Poole

Contributions: Advised on research analysis and model set-up, helped interpret results, edited manuscript

Author: Scott J. O'Daniel

Contributions: Provided essential data on Meacham Creek, provided funding through the CTUIR, helped interpret results, edited manuscript

Author: Byron A. Amerson

Contributions: Provided model code, advise and help on simulating water and heat flow in HGS

Manuscript Information

S. Kathleen Fogg, Geoffrey C. Poole, Scott J. O'Daniel, and Byron A. Amerson
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Abstract

Floodplain forests have the potential to mediate atmosphere-aquifer heat exchange by shading the floodplain surface, thereby influencing stream channel temperatures via hyporheic exchange of water and heat. Channel re-meandering restorations cause large-scale disturbances to the landscape, including reducing floodplain shade, which could lead to the unintentional warming of hyporheic and channel temperatures. We propose a “shade retention strategy” (SRS) to restoration in which priority is given to retaining as much floodplain shade as possible. We tested the effects on channel temperature of this strategy using a simulation model of Meacham Creek (Oregon, USA). We compared temperatures results from the SRS scenario to a scenario representing a “business as usual” (BAU) floodplain shade pattern in which large patches of vegetation were cleared from the floodplain. Additionally, we simulated these shade scenarios using both steady-state base-flow and transient, snow-melt dominated annual hydrology input signals to understand the effects of different input hydrology on simulated channel temperatures. Model results showed that the greater amount of floodplain vegetation in the SRS scenario (5.7%) led to cooler hyporheic and stream channel temperatures than the BAU scenario. Simulated temperatures of scenarios using steady-state base-flow hydrology were warmer than temperatures simulated using transient hydrology—greatest temperature differences observed in the winter and spring when the differences in discharge between the two scenarios was greatest. Our results suggest that (1) adopting a “shade retention strategy” during channel restorations may help mitigate warming temperatures, thus lessening disturbance to the ecosystem and (2) simulating steady-state base-flow hydrology is not an acceptable simplification in numerically simulating the effects of hyporheic exchange on channel temperatures. In alluvial rivers where hyporheic zones extend far beyond the channel margins, consideration of floodplain riparian shade, both during and after restoration actions, may improve outcomes for stream

temperature management.

Introduction

The promotion and preservation of native cold-water fish populations is a widespread management priority in the inter-mountain western U.S. and river temperature is often a primary driver of water quality impairment [Beechie et al., 2013, Justice et al., 2017]. Fish experience metabolic stress when exposed to prolonged temperatures outside of their tolerance range [Gillooly et al., 2001, Motyka et al., 2017, Rodnick et al., 2004]. This stress exacerbates the toxicity of pollutants [Peralta-Maraver et al., 2018], virility of disease [Harvell et al., 2002, Marcos-López et al., 2010], and hypoxic conditions associated with higher temperatures [Helton et al., 2012b, Mackay and Fleming, 1969]—all which can eventually lead to fish deaths and declining populations [e.g., Levy, 2003]. Different species experience different thermal tolerances. For example, bull trout (*Salvelinus confluentus*) have one of the lowest upper thermal tolerance limits—approximately 20°C [Selong et al., 2001]. Steelhead or Rainbow Trout (*Oncorhynchus mykiss*) have a greater tolerance for warmer temperatures than bull trout, where thermal stress begins to occur at 21°C and incipient lethal temperature occurs at approximately 25°C [Boughton et al., 2015].

The persistence of cold-water salmonid populations depend on thermal refuges as critical habitat [Arscott et al., 2001, Brewitt and Danner, 2014]. Thermal refuges are areas of the riverscape that provide localized temperatures (typically cooler) that are used by poikilotherms avoiding less optimal (typically warmer) temperatures [Sullivan et al., 2021]. Gravel-bedded river floodplains provide a diverse array of habitats and temperatures via pools, side channels, alcoves and upwelling hyporheic water. Coarse-grained riverbeds provide the substrate for egg incubation in redds and are thus important landscapes to protect for species conservation [Hauer et al., 2016]. Hyporheic exchange is a large driver of the thermal landscape of gravel-bedded rivers, and these exchanges can buffer, lag and cool

stream temperature cycles across both daily and seasonal scales [Arrigoni et al., 2008, Fogg et al., 2023]. Promotion of hyporheic exchange across multiple scales can improve spawning sites for salmonids [Baxter and Hauer, 2000] and sites of upwelling hyporheic water provide potential thermal refuges [Kurylyk et al., 2015, Sullivan et al., 2021].

Restoration of hyporheic exchange processes include many benefits to the stream ecosystem, including improved temperature regulation and increased habitat for macroinvertebrates and salmonid spawning locations [Hester and Gooseff, 2010, Poole et al., 2008]. Small-scale improvements of hyporheic exchange include placing in-stream structures such as logs and boulders in the channel to create hydraulic gradients which induce downwelling of river-water into the hyporheic zone [Kasahara and Hill, 2008]. Alternatively, large scale hyporheic restorations involve re-meandering channels, which can induce lateral hyporheic flows across meander bends [Stonedahl et al., 2013].

Channel re-meandering can cause large-scale disturbance to the stream ecosystem. A goal of any restoration is to have the long-term benefit outweigh the temporary negative disturbance caused by the restoration [Bernhardt and Palmer, 2011]. The river ecosystem can take a decade or longer to show recovery from these types of large scale projects [Bilby et al., 2023, Roni et al., 2002]. Delayed onset of recovery makes it harder to determine the benefits of restoration via monitoring. Therefore, limiting disturbances or shortening the time-scale of recovery are important considerations for restoration. A consequence of channel re-meandering restorations is the broad removal of floodplain forest vegetation. Removal of vegetative shade overlying shallow hyporheic or groundwater flow-paths can result in warmer upwelling water temperatures in the channel (see Chapters 2, 3, *supra*). The warming of groundwater temperatures caused by deforestation far from the channel is often not considered in investigations of stream temperature dynamics. For example, multiple studies on the efficacy of riparian buffers within cutblocks have shown unexpected channel warming when timber harvesting occurred beyond the intact riparian buffers [e.g.,

Cole and Newton, 2013, Hewlett and Fortson, 1982, Story et al., 2003]. Many of these studies suggest that channel warming was caused by shallow groundwater inputs which were warmed in the logged area, outside of the riparian buffer zone, however, none directly tested that hypothesis.

In this study, we investigated the potential influence of floodplain shade on temperature outcomes of a channel re-meandering restoration. We define “floodplain shade” as vegetative shade which occurs on the broader floodplain, at a distance where vegetation would not shade the low flow wetted channel. This is in contrast to “channel shade” which casts shadows on the low-flow channel, thus influencing channel-atmosphere heat exchange. We used Meacham Creek (Oregon, U.S.A) to investigate the effects of floodplain shade on simulated temperatures because Meacham Creek underwent a channel re-meandering restoration and was well monitored before and after the restoration. The Confederated tribes of the Umatilla Indian Reservation (CTUIR) has an interest in restoring populations of the ESA-listed Middle Columbia River steelhead and Columbia River bull trout to Meacham Creek. The peoples of CTUIR, members and ancestors, have used Meacham Creek for fishing grounds since time-immemorial [CTUIR, 2012] but degradation of the stream in the last century (among other down-stream degradations) caused decreases in salmonid habitats. This catalyzed a multi-phased restoration effort on Meacham Creek beginning in 2006. Meacham Creek is a gravel-bedded floodplain reach that was re-meandered from a straight, incised channel flowing against a valley wall to a meandering channel in the center of the floodplain. The aim of this restoration was to increase hyporheic exchange, reduce longitudinal temperature warming across the reach, and increase salmonid habitat [CTUIR, 2012]. During the restoration, much of the vegetation on the broader floodplain was removed and used for heavy equipment staging areas.

Simulation modelling experiments testing alternative restoration designs are important for determining potential restoration outcomes [e.g., Bennett et al., 2008, Chu et al., 2013,

DeWalle, 2008, Johnson et al., 2015, Justice et al., 2017, Kasahara and Hill, 2008]. *A priori* modeling is especially important when the restoration is costly and may cause large-scale disturbance such as in channel re-meandering restorations. Simulation modeling of stream reaches using steady-state hydrology is simpler, thus cutting down on computational time and costs. However, the fluctuating flows of transient hydrologic model inputs have been shown to significantly influence hyporheic exchange flows [Gomez-Velez et al., 2017, Schmadel et al., 2015, Wroblicky et al., 1998]. Watson et al. [2018] found that flood flows influence the lateral movement of heat plumes in the hyporheic zone of the Lower Colorado River (Texas, U.S.A) and that large floods can alter hyporheic temperature dynamics long after flood recession. In addition to simulating alternative floodplain shade patterns, we investigated the influence of steady-state versus transient hydrology in our modeling experiment.

To minimize potential warming of shallow groundwater and stream channel, we propose a “shade retention strategy” for restoration practices which prioritizes keeping as much vegetation as possible on all parts of the floodplain. We predict that this strategy would prevent warming of shallow hyporheic water and thus promote cooler channel temperatures. We used a simulation modeling experiment to test the efficacy of the “shade retention strategy” (SRS) at producing cooler channel temperatures on the restored channel of Meacham Creek. We tested the SRS scenario against a “business as usual” (BAU) scenario in which large areas of floodplain vegetation were removed for channel re-meandering. The floodplain vegetation pattern that occurred due to restoration efforts was used as our “business as usual” scenario. The floodplain vegetation as it appeared before the restoration was used as our “shade retention strategy”, while simulating the newly re-meandered channel.

The goals of this study were to (1) examine the potential effects on channel temperature of retaining floodplain shade during channel restorations (“Shade retention strategy”) versus not prioritizing floodplain shade retention (“Business as usual”) via a simulation experiment and (2) understand the effects on simulated temperatures of using steady-state

base-flow hydrology versus a transient signal representative of annual snow-melt hydrology. We hypothesized that (1) the amount of floodplain shade, exemplified by our alternative restoration scenarios, influences the conduction of heat from the ground surface to the hyporheic zone, altering hyporheic temperatures, which in turn affects stream channel temperatures upon hyporheic upwelling. We also hypothesized that (2) the pattern of annual hydrology drive hyporheic exchange rates and channel-hyporheic heat exchange as well as channel-atmosphere heat exchange. We test these hypotheses by conducting a simulation experiment using a model of the re-aligned Meacham Creek channel. We altered the pattern of shade on the model surface where we retained the pre-restoration shade levels to represent the “Shade retention strategy” (SRS) versus the loss of floodplain shade associated with the channel re-alignment to represent the “Business as usual” (BAU) scenario. We tested the two shade scenarios with inflowing water either steady-state base-flow or an annual fluctuating hydrology. We predicted that (1) the scenario with greater floodplain shade (SRS) will lead to cooler simulated hyporheic and channel temperatures compared to the scenario with less floodplain shade (BAU) and (2) scenarios simulated using base-flow hydrology year-round will result in warmer channel temperatures during parts of the year which flood in the transient scenarios.

Study Site and Background

The Meacham Creek floodplain is located in a bedrock-confined valley of the the Blue Mountains in eastern Oregon (OR), USA (Figure 4.1). The stream flows south to north and is a major tributary of the the Umatilla River. The confluence of the Umatilla River and Meacham Creek is approximately 23 miles east of the city of Pendleton, OR. Meacham Creek drains a catchment area of approximately 461 km² of foothill terrain [CTUIR, 2012]. Springtime rain-on-snow events create annual floods orders of magnitude greater than base-flow. The floodplain of Meacham Creek extends from valley wall to valley wall. The alluvial

aquifer material is composed of coarse-grained, basaltic gravels [Powell, 2001] and saturated thickness is approximately 3 meters (well log data, B. Amerson). The alluvial aquifer is constrained vertically and laterally by bedrock and the entire volume of the alluvial aquifer is considered the hyporheic zone of Meacham Creek.

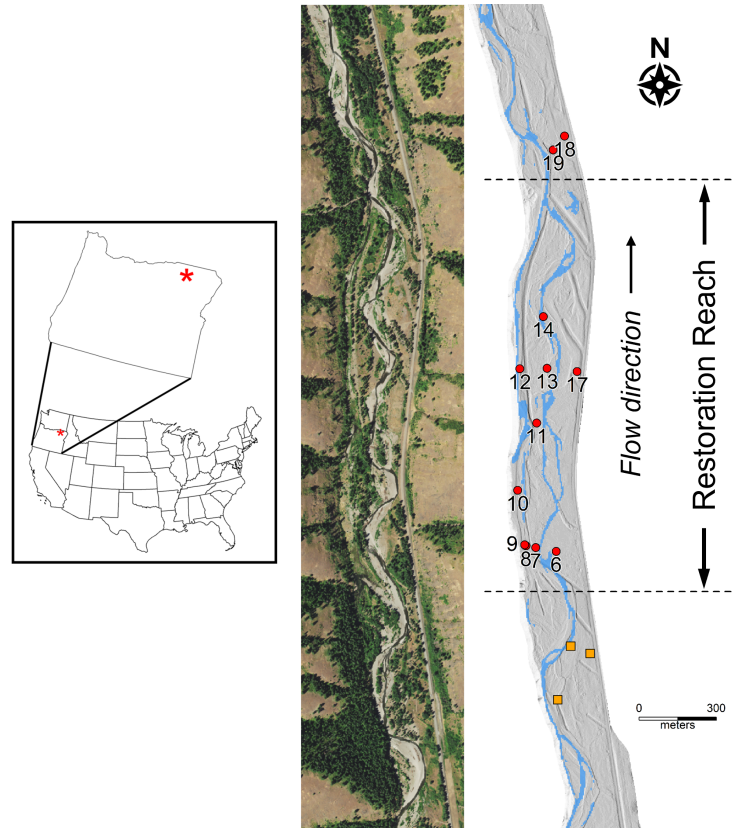


Figure 4.1: Location of the Meacham Creek floodplain in northeastern Oregon, USA (left); 2020 orthoimage of model extent (center); hillshade model using 2013 LiDAR data (right). The base-flow channel is shown in blue, wells used for verification of hyporheic temperatures are shown as red circles and labeled accordingly, and wells used to produce the input hyporheic temperature signal are shown as orange squares.

The primary ownership in the Meacham Creek watershed includes the United States Forest Service, the Confederated Tribes of the Umatilla Indian Reservation (CTUIR) and Boise Cascade (lumber production) [Powell, 2001]. Historical land use in the watershed includes timber harvesting and sheep and cattle grazing [CTUIR, 2012]. A Union Pacific

rail line runs through Meacham Canyon and was built against the eastern valley wall through much of the canyon length. 20th century construction of levees and spur dikes to protect the rail line against potential flood damage caused the greatest impact to the hydrology and geomorphology of Meacham Creek [CTUIR, 2012]. Historically free-flowing in the center of the valley floodplain, levees and dikes restricted the natural hydraulic processes of Meacham Creek, leading to channel straightening, incision, and disconnection from its historic alluvial aquifer and floodplain.

Our study site is the 1.6 km (1 mile) reach of Meacham Creek that underwent a channel re-meandering restoration in 2011 (Figures 4.1, 4.2). The restoration included 28.33 hectares of adjacent floodplain and occurred between river kilometer 9.6 (river mile 6) and 11.3 (river mile 7) [CTUIR, 2012]. Prior to the restoration, this reach flowed against the western valley wall through a straight and incised channel (Figure 4.2).

The 2011 restoration removed a 853-meter levee and removed or modified multiple spur dikes in the floodplain (Figure 4.2). The channel was re-aligned toward its historic channel and was meandered in the center of the floodplain, allowing the natural channel migration between the west valley wall and constraining floodplain features on the east (i.e., right-of-way road, modified spur dikes). The restored Meacham Creek channel has a coarse-grained alluvial aquifer, a broad floodplain and high hyporheic exchange rates.

The restored section of Meacham Creek has been extensively monitored both before and after restoration. Pre-restoration (2009) and post-restoration (2013, 2016, 2020) Light Detection and Ranging (LiDAR) flights recorded the topographic and vegetation changes of the restoration (Table 4.1). Groundwater monitoring wells were installed in the alluvial aquifer in 2011 (Figure 4.1) and sensors recorded hourly temperature measurements ($^{\circ}\text{C}$) for multiple years. Stream channel and air temperature ($^{\circ}\text{C}$) were recorded across multiple sites across multiple years. Weather stations (ONSET brand) were installed for a short period (1-week) during the summer of 2015; a silicon pyranometer recorded solar radiation (W m^2)

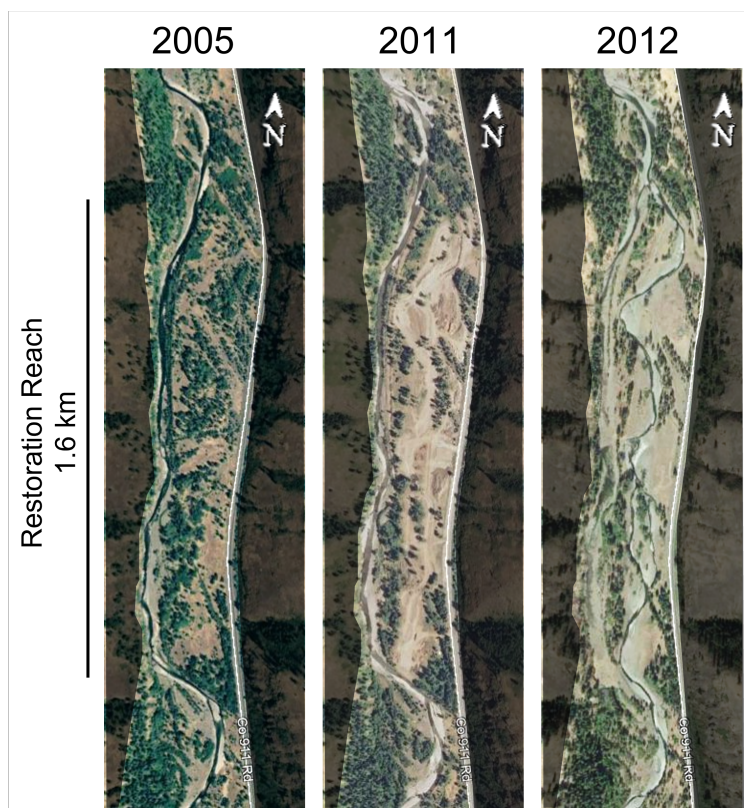


Figure 4.2: Floodplain restoration progression on Meacham Creek, Oregon, USA. Pre-restoration image (2005; left) shows the channel constrained to the western valley wall of Meacham Canyon; mid-construction of the new channel (2011; center); restored channel flows in center of floodplain (2012; right). Note the substantive loss of floodplain shade over the course of the restoration.

at a shaded site under a cottonwood-conifer stand and at a site in full sun. These datasets were used to parameterize our model scenarios. Further description in how these data were used can be found in Sections 4, 4, and 4.

LiDAR flight year	% vegetation coverage
2009	72.7%
2013	67.1%
2016	68.7%
2020	68.9%

Table 4.1: Percent vegetation (> 3 m) coverage of restoration reach across years.

Simulation Experiment

We tested the effects of floodplain shade patterns and variable flow conditions on the temperature of Meacham Creek using a simulation experiment. HydroGeoSphere (HGS)—a 3-dimensional control-volume finite element simulation model [Brunner and Simmons, 2012]—was used to simulate hydrology and heat exchanges in the Meacham Creek floodplain. HGS couples surface and subsurface transport of water and heat, simulates unsaturated conditions, advective, dispersive, and conductive heat fluxes in the subsurface, and atmospheric heat exchanges with soils and water surfaces.

We developed experimental scenarios by varying floodplain shade pattern and input hydrology to our model of Meacham Creek. We tested the response of the restored Meacham Creek channel to two alternative shade scenarios—one representative of a “Shade Retention Strategy” (SRS) which included all floodplain vegetation prior to the restoration, and the other representative of the floodplain shade pattern after the restoration, i.e. the “Business as usual” (BAU) scenario, reflecting the shade loss associated with operating

heavy construction equipment for channel realignment. Both SRS and BAU shade scenarios had no channel shade on the base-flow simulated channel. We tested the effects of steady-state (“-ss”) and transient (“-t”) hydrology on the temperature response of both shade models, totalling four experimental scenarios (SRS-ss, SRS-t, BAU-ss, BAU-t). The steady-state hydrology represented low flow conditions typical of late summer through winter, whereas scenarios with transient hydrology represented a now-melt driven hydrologic regime typical of the inter-mountain western U.S, with high flows in the spring and early summer and low flows in the later summer through early spring (Figure 4.5).

Our analysis of model results focused exclusively on the 1.6-km restoration reach. The simulation model, however, included reaches 1000-m upstream and 900-m downstream of the restoration reach (3.5-km total length) in order to minimize the potential influence of the simplified model boundary conditions on simulated temperature dynamics within the restoration reach (Figure 4.1).

Model Description

Model Grid

We used a 5-m by 5-m grid in the horizontal x- and y- directions for each model layer. Surface elevations of the grid intersections (i.e. “model nodes”) were set to bare earth elevations from the 2013 post-restoration LiDAR flight of Meacham Creek. The elevation data were collected at a 1-m by 1-m grid resolution, thus assignment of the model nodes was done by bilinear interpolation in HGS. Three layers were established below the surface elevation and represented 2.5 m of hyporheic alluvium. Two additional layers totaling 25 m (layer thickness = 1 m and 24 m) were created below the alluvium to represent bedrock (Figure 4.3).

Materials Properties

Hydrologic and thermal properties of alluvium and bedrock were assigned values from the literature which represented the thermal properties of basalt and the hydrologic properties of gravel-dominated alluvium (Table 4.2). Overland flow properties applied to the channel and floodplain are found in Table 4.3. The stream channel had a Manning's roughness coefficient, n , of 0.043, a value measured on the near-by Grand Ronde River at La Grande, Oregon, which has similar geomorphology [Barnes, 1969].

Boundary Conditions

Hydrological The amount of in-flowing water to the models were applied using specified head boundary conditions for the upstream and downstream flows. A critical depth boundary condition was applied to surface nodes of the downstream boundary to allow water to freely drain out of the model. Inflowing water at the upstream boundary was either steady-state, representing base-flow or transient, representing annually fluctuating hydrology. Steady-state scenarios had an upstream head value of 628.16 m and a downstream head value of 597.6 m. Transient hydrology represented an archetypal year, loosely approximating flows in 2013, which had spring-time floods and a long base-flow period in the summer and fall (Figure 4.6).

Thermal In HGS, we assigned temperature signals to the inflowing water at the upstream boundary and the atmospheric heat exchange component. The temperature signals of the inflowing water at the upstream boundary were based on temperature observations recorded in the channel and in floodplain wells on Meacham Creek.

Channel water temperature observations from July-2020 and October-2021 were used to create an annually-repeating compound sine wave for the channel temperature signal input (Figure 4.4) using the non-linear least-squares function 'nls()' in R [R Core Team,

Alluvium

Parameter	Value	Unit	Cite
Longitudinal hydraulic conductivity	0.0046	m s ⁻¹	Poole et al. [2008]
Transverse hydraulic conductivity	0.00046	m s ⁻¹	1/10 Longitudinal; Freeze and Cherry [1979]
Longitudinal dispersivity	1.4	m	Rau et al. [2012]
Transverse dispersivity	0.4	m	Rau et al. [2012]
Thermal conductivity	1.6	W m ⁻¹ K ⁻¹	Waples and Waples [2004]
Specific heat capacity	840	J kg ⁻¹ K ⁻¹	Waples and Waples [2004]
Density	3000	kg m ⁻³	Waples and Waples [2004]
Porosity	0.25	-	Freeze and Cherry [1979]
Van Genuchten α	22.0	m ⁻¹	Thoma et al. [2014]
Van Genuchten n	1.79		Thoma et al. [2014]
Residual saturation	0.04	-	Thoma et al. [2014]
Minimum relative permeability	1.995e-06		Thoma et al. [2014]

Bedrock

Parameter	Value	Unit	Cite
Hydraulic conductivity (isotropic)	1.0e-10	m s ⁻¹	Freeze and Cherry [1979]
Dispersivity (isotropic)	1.0e-20	m s ⁻¹	(value suitably close to zero)
Thermal conductivity	1.6	W m ⁻¹ K ⁻¹	Waples and Waples [2004]
Specific heat capacity	840	J kg ⁻¹ K ⁻¹	Waples and Waples [2004]
Density	3000	kg m ⁻³	Waples and Waples [2004]
Porosity	0.04		Zakharova et al. [2012]

Table 4.2: Material Properties

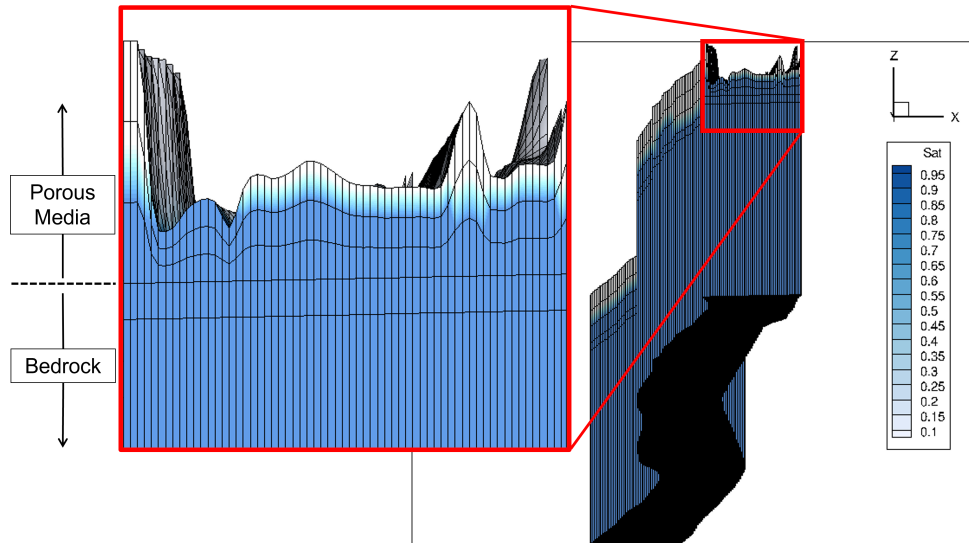


Figure 4.3: Model grid layers at upstream model boundary. Inset image shows the three porous media layers and the two bedrock layers. Coloration shows gradient of saturation values on 28-Oct. Vertical exaggeration is apparent in this grid image: the horizontal scale is indicated by the 5 m mesh spacing; in contrast, the vertical scale is indicated by the 1 m thick of upper bedrock layer.

Floodplain

Parameter	Value	Units	Cite
x-, y- friction	0.3	$\text{m}^{-1/3} \text{ s}$	
Rill storage height	0.005	m	

Wetted Channel

Parameter	Value	Units	Cite
x-, y- friction	0.043	$\text{m}^{-1/3} \text{ s}$	Barnes [1969]
Rill storage height	0.015	m	

Table 4.3: Overland Flow Properties

2021]. The temperature input signal applied to inflowing subsurface water was calculated from temperature data collected in groundwater wells installed on the Meacham Creek floodplain. The wells used to estimate this signal were located upstream of the restoration reach. Temperature observations were collected between 2011 and 2015. We calculated the daily mean groundwater temperature, averaged across all years and all wells and used the resulting curve as the input signal for water flowing into the subsurface nodes (Figure 4.5).

The air temperature applied to all surface nodes of the model was based on measured air temperature collected on the Meacham Creek floodplain between July-2020 and October-2021. We calculated an annually-repeating compound sine-wave using ‘nls()’ in R (Figure 4.4).

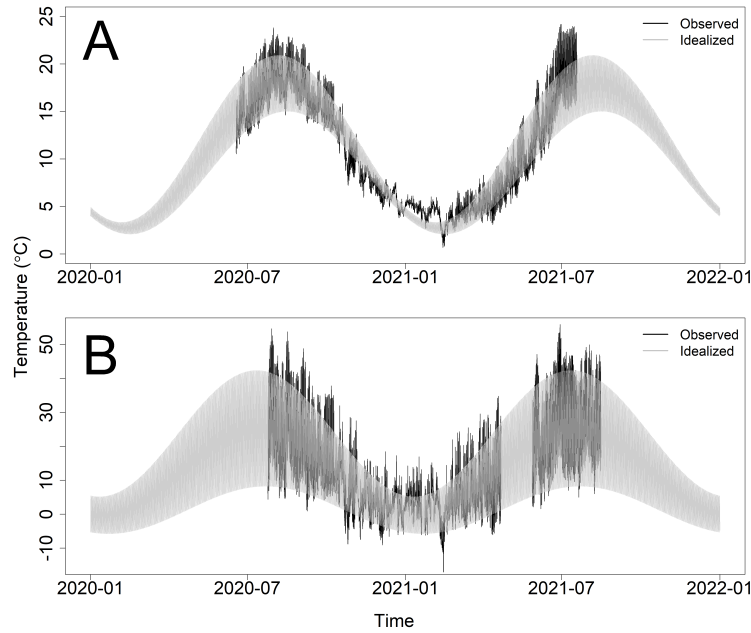


Figure 4.4: Observed vs. idealized air and surface water temperatures used as model boundaries. (A) Observed air-temperature data plotted over top of the compound sine wave used to represent idealized air temperature regime at floodplain surface boundary. (B) Observed channel water temperature plotted over top of the compound sine wave used to represent surface water temperature at the surface water input model boundary. Observed signals were collected on Meacham Creek on a weather station (A) or in the thalweg of the channel (B) between 06-2020 and 09-2021.

Representation of Floodplain Shade Scenarios

Alternative shade patterns were applied to the atmospheric heat exchange component of the Meacham Creek HGS model using binary rasters (Figure 4.7). Vegetative shade was represented by applying a reduced solar radiation signal in the shaded areas of the floodplain, compared to a clear-sky solar radiation signal applied to areas of the model that were unshaded (Figure 4.8). The pattern of shade and therefore solar radiation across model nodes was applied via binary rasters: nodes which were located in raster cells categorized as “unshaded” were given the clear-sky solar radiation signal, whereas nodes located in raster cells categorized as “shaded” were given the shaded solar radiation signal (Figure 4.8).

The binary shade rasters used for the SRS and BAU scenarios were created using LiDAR data. The SRS scenario used data collected prior to the restoration (2009) and the BAU scenario used data collected after the restoration (2013). First, tree heights were calculated by subtracting the bare earth elevations from the highest hits elevations. Any cells with bare ground or vegetation heights less than 3 m were considered “unshaded” and any cells with vegetation height greater than 3 m were considered “shaded”. We filtered tree heights at 3 meters to be conservative in our estimation of floodplain shade as this would represent near the minimum reasonable estimation of shade. Our binary application of either “shaded” or “unshaded” cells does not represent the dappled nature of shade nor the geometric casting of shadows which changes across the day as the Earth rotates. Our goal was to capture the general influence of shading with a reasonable computation time, though we acknowledge this strategy results in a more discontinuous spatial distribution of energy exchange at the floodplain surface than would be occurring in the real system.

The LiDAR data were collected at a 1-m grid resolution and our HGS model grid resolution was 5-m. We calculated the percent shade coverage within each 5-m grid from the 1-m grid and determined “shaded” cells where the percent shade was greater than or equal to 50%. Cells with less than 50% shade coverage were “unshaded” in our final rasters.

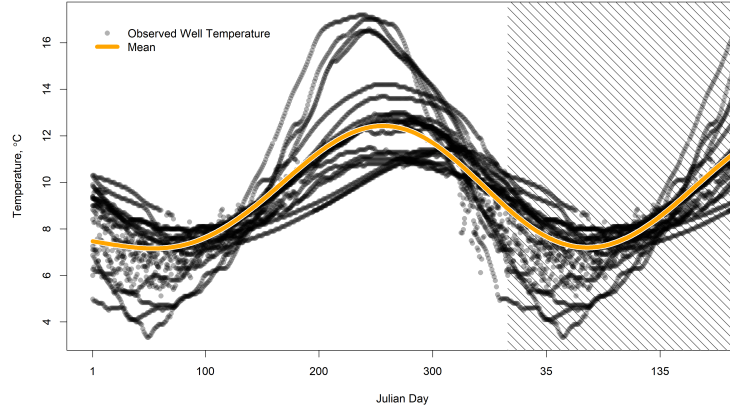


Figure 4.5: Groundwater temperatures observed across the years 2011–2015 in the three Meacham Creek floodplain wells upstream of the restoration reach (orange squares, Figure 4.1). Orange line is the daily mean groundwater temperature signal used as input hyporheic zone and bedrock temperatures at upstream model boundary. The shaded area on the right side of the graph is a repetition of the first 182 days of data.

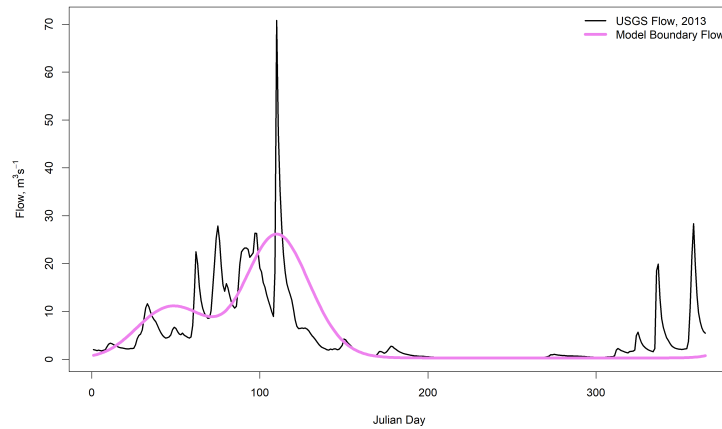


Figure 4.6: Example flow (black; $\text{m}^3 \text{s}^{-1}$) of Meacham Creek from USGS gauge at Gibbon, Oregon (site no. 14020300), located approximately 10km downstream of restoration reach before confluence with Umatilla River. Smoothed input signal used as the upstream boundary condition for model scenarios with transient hydrology (violet; $\text{m}^3 \text{s}^{-1}$).

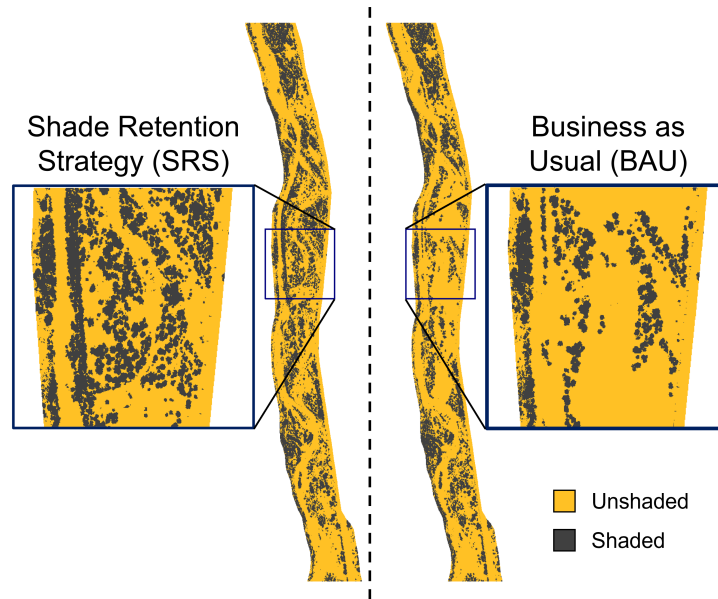


Figure 4.7: Binary rasters used for alternative shade scenarios. “Business as usual” (BAU) shade pattern (right) shows a decrease in vegetative shade compared to the “Shade retention strategy” (SRS) shade pattern (left). The base-flow channel was unshaded in both scenarios.

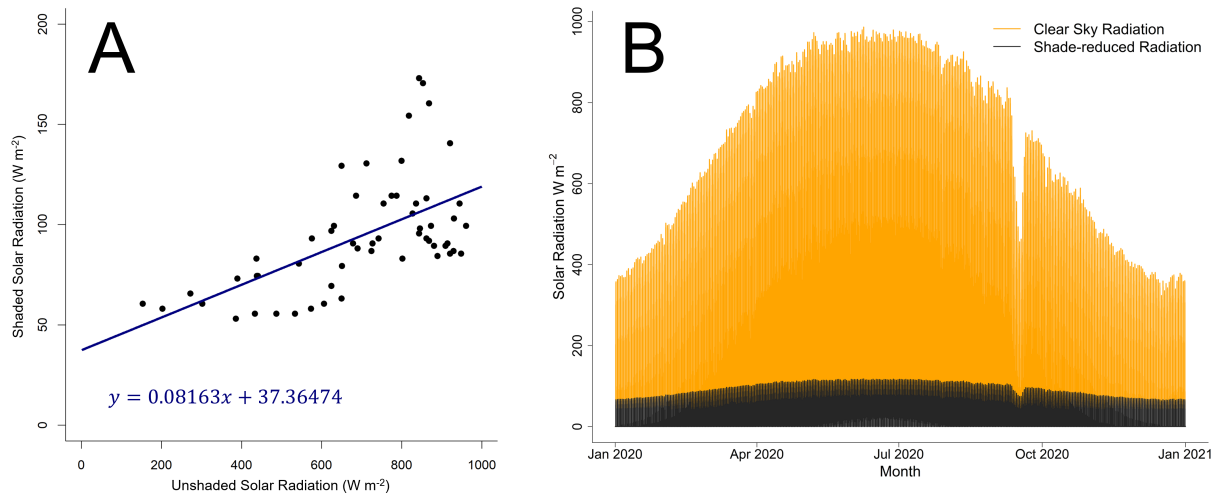


Figure 4.8: Linear model of vegetation-attenuated solar radiation (A) and annual signals of solar radiation model inputs (B). The clear sky radiation signal was applied to “unshaded” model cells and the shade-reduced signal was applied to all “shaded” model cells. The linear model predicts shaded solar radiation typical of Meacham Creek vegetation from unshaded solar radiation measured on Meacham Creek.

The base-flow channel of Meacham Creek was left “unshaded” in both models because the base-flow channel has little shade. This simplification not only is a realistic representation of shade conditions at the base-flow channel, but allows us to standardize the shade contribution to the base-flow channel of all model scenarios, regardless of shade pattern on the broader floodplain.

We used a clear-sky solar radiation signal from the National Aeronautics and Space Agency’s (NASA) Clouds and the Earth’s Radiant Energy (CERES) project for solar radiation signal applied to the “unshaded” model nodes (Figure 4.8). Specifically, we used the surface, clear-sky direct shortwave radiation flux data product (SYN1degEd41) [Rutan et al., 2015]), which is a calculation of the shortwave radiation flux on the ground surface at a 1°latitude x 1°longitude resolution, calculated at an hourly-interval. Radiation data were retrieved for the year 2020, at a data grid which included the restoration reach of Meacham Creek, (45.379,-118.611)-(45.879,-118.111).

We calculated a reduced solar radiation signal from the clear-sky signal using a linear model predicting shaded solar radiation from clear-sky solar radiation using observations collected on Meacham Creek (Figure 4.8A). Solar radiation data was collected using silicon pyranometers (HOBO brand, model S-LIB-M003) at two weather stations on Meacham Creek—one at a shaded site and the other at an unshaded site. The loggers recorded solar radiation values every 15 minutes between 31-July-2015 and 2-August-2015. The shaded site was located in a stand of mixed cottonwoods and conifers. The pyranometers at this site received full-sun solar radiation in the morning hours and vegetation-attenuated solar radiation in the afternoon hours (12:00-17:00), therefore only data taken between 12:00 and 17:00 were used to calculate the linear relationship of shaded solar radiation and unshaded solar radiation shown in Figure 4.8A.

Model Spin-up

We ran a simulation run of HGS to solve for head values at each model node. The model was initialized with head values equal to surface elevation at all depths and then simulated for one year with transient hydrology (Figure 4.6) but no variation in temperature. Head values at each node were output on 01-January and were used to initialize model scenarios with transient hydrology.

Temperatures of all model nodes were initialized at the mean annual water temperature of the input signal, 10.3°C. All scenarios were simulated for 8 years for temperature spin-up and the 9th year was output and used for analysis. We verified 8 years was a long enough run time for temperatures to reach a cyclic equilibrium.

Model Analysis

Verification

We evaluated the thermal and hydrologic parameterization of our model by comparing modeled groundwater temperatures to measured groundwater temperatures. We used simulated groundwater temperatures from the BAU-t model scenario because it had the most realistic parameterization for comparison to well temperatures collected between 2011 and 2015. Wells located downstream of the upstream restoration boundary were used for this analysis (red circles, Figure 4.1).

Experimental Scenarios

Output of the experimental model scenarios was recorded on 01-January and every 30 days thereafter until output totaled 12 model days. Output at each hour of the 12 days was recorded, totalling 288 output times. We calculated the daily mean temperature for all model nodes, resulting in 12 days of daily mean temperature output. We categorized model nodes according to their location. Ground surface nodes were differentiated from channel

nodes by surface water depth. All nodes with depths less than 0.1 cm were classified as ground surface, whereas nodes with water depth greater than 0.1 cm were considered in the stream channel. The hyporheic zone nodes were all subsurface nodes between the bedrock layer and the model surface nodes which were fully saturated.

Physiological Habitat Analysis

We analysed differences in simulated channel temperature distributions between scenarios to understand the potential impacts of floodplain shade on the amount of channel area that could be used for physiological refuge or areas that could cause thermal stress to important salmonids species. We focus our analysis of physiological refuges to the hottest months of the year, July, August and September, when physiological refuge may be most important. We plotted the distribution of channel temperatures as histograms with 0.5°C breaks overlaying a backdrop differentiating physiologically important temperatures. We delineate temperatures at 20°C, 17°C, and 13°C. Although variable by species, 20°C is the temperature which is generally delineated as the stressor for cold-water species in general [Armstrong et al., 2021, Magnuson et al., 1979]. 13°C was the the thermal guideline temperature for bull trout and 17°C for steelhead outlined in the impairment analysis of Meacham Creek [CTUIR, 2012].

Longitudinal Temperature Trend Analysis

We analyzed the change in daily average channel temperatures from upstream to downstream to understand longitudinal temperature trends across the restoration reach. We calculated the difference in temperature between stream water entering the reach and leaving the reach as a measure of the change in water temperature as water flowed through the reach. Negative values thus indicate longitudinal cooling and positive values indicate longitudinal warming.

Results

Verification

Simulated daily mean well temperatures were generally similar to the well temperatures observed on Meacham Creek across the years 2013–2014 (Figure 4.9). Temperatures in some simulated wells over- or under- estimated maximum or minimum annual temperatures, but the annual means are well approximated. Considering that the model was parameterized independently and not calibrated to the observed temperature data, the similarities of the simulation to the observations suggest the model was a reasonable analog to the hydrological and thermal behavior of Meacham Creek.

Daily Mean Temperatures

BAU scenarios, with reduced floodplain shade, resulted in warmer daily mean ground, hyporheic and stream temperatures in the restoration reach compared the the SRS scenarios, which had more floodplain shade (Figure 4.10). The greatest temperature differences between shade scenarios occurred in the summer months.

Comparison of transient vs. steady-state flow conditions had much less effect on ground and hyporheic temperatures than differences in shading, whereas the largest differences due to flow conditions occurred when flows were highest. (Figure 4.10). Channel temperatures showed the strongest differences with flow conditions, again during the highest stream flows. The greatest difference between the two hydrology scenarios occurred during springtime, when the transient scenarios had high-flows and the steady-state scenarios were still at the level of a base-flow channel. During spring, transient hydrology scenarios simulated cooler channel temperatures than steady-state scenarios. Between scenarios with the same shade pattern, but differing hydrology, the same heat flux was applied to different volumes of water, with different surface areas. During the springtime, when transient flows peaked, channel

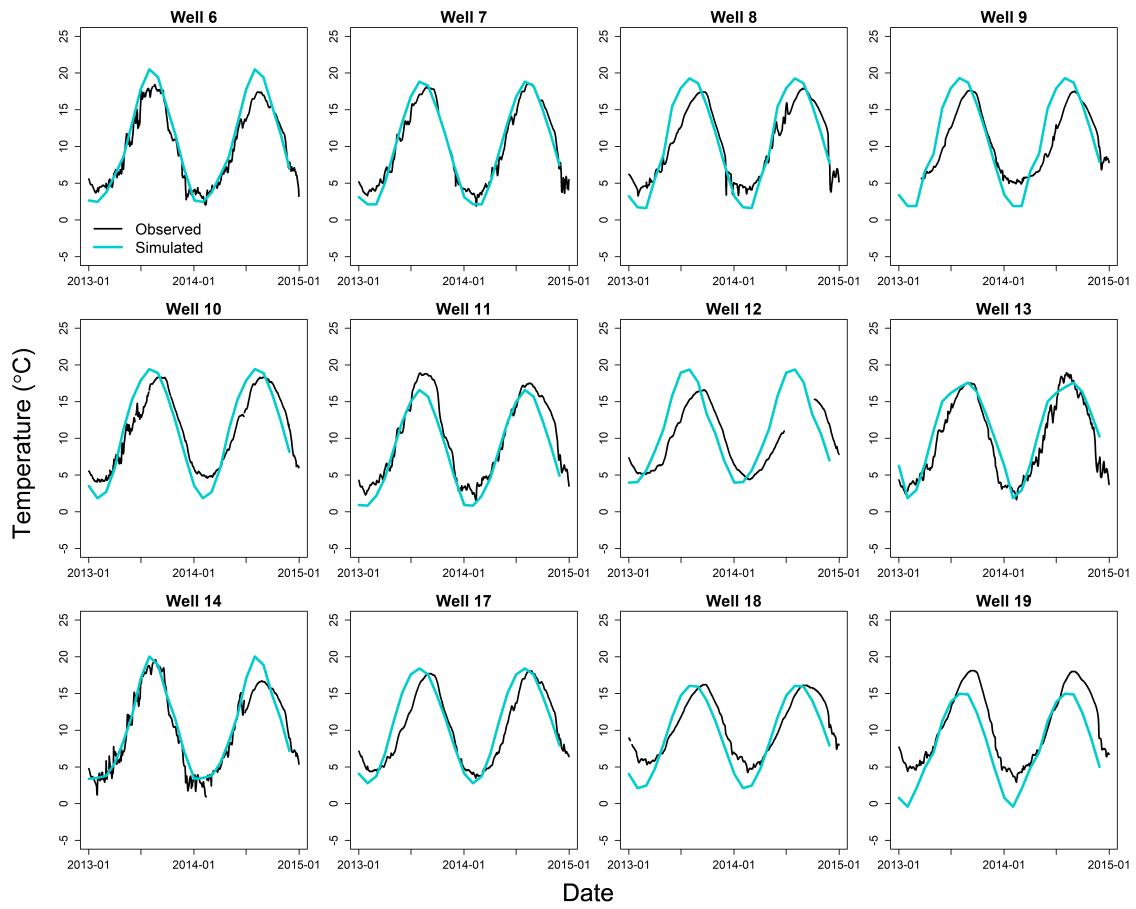


Figure 4.9: Comparison of observed and simulated well temperatures across the Meacham Creek floodplain. See Figure 4.1 for well locations.

temperature was more stable than under steady-state assumptions. When the transient models returned to base-flow discharge in late summer, the differences between simulated channel temperatures under the steady-state and transient scenarios was negligible (Figure 4.10).

Summertime Channel Temperature Distributions

In the summer months, the additional shade from the SRS scenario provided the most additional potential thermal refuge compared to the BAU scenario. In August, the BAU-t scenario had 31.2% more channel temperatures $> 20^{\circ}\text{C}$ than the SRS-t scenario. There is little change in “optimal” physiological habitat (temperatures less than 17°C) across BAU and SRS shade scenarios. In July and August $< 2\%$ of channel temperatures were cooler than 17°C , and in September, $< 7.1\%$ (Figure 4.11). Still, greater areas of physiological refuge are expected in the SRS scenarios compared to the BAU scenarios.

Scenarios with steady-state flow conditions displayed more locations with higher temperatures than the transient scenarios, even though channel discharge of both hydrology scenarios are at base-flow in the late summer. The similar flow means the heat load on the channel surface is the same across scenarios with different flow dynamics, therefore the temperature differences are due to differences in hyporheic heat exchange associated with differing surface-subsurface water exchanges and not floodplain shade. Springtime high flows in the transient scenarios cause the hyporheic zone to expand and be “charged” with a greater amount of cool water than the steady-state scenarios. This generally cooler water is released from the hyporheic zone throughout base-flow, thus likely causing the late summer temperature differences between scenarios with the same floodplain shade but different flow conditions (i.e. BAU-t vs. BAU-ss; SRS-t vs. SRS-ss).

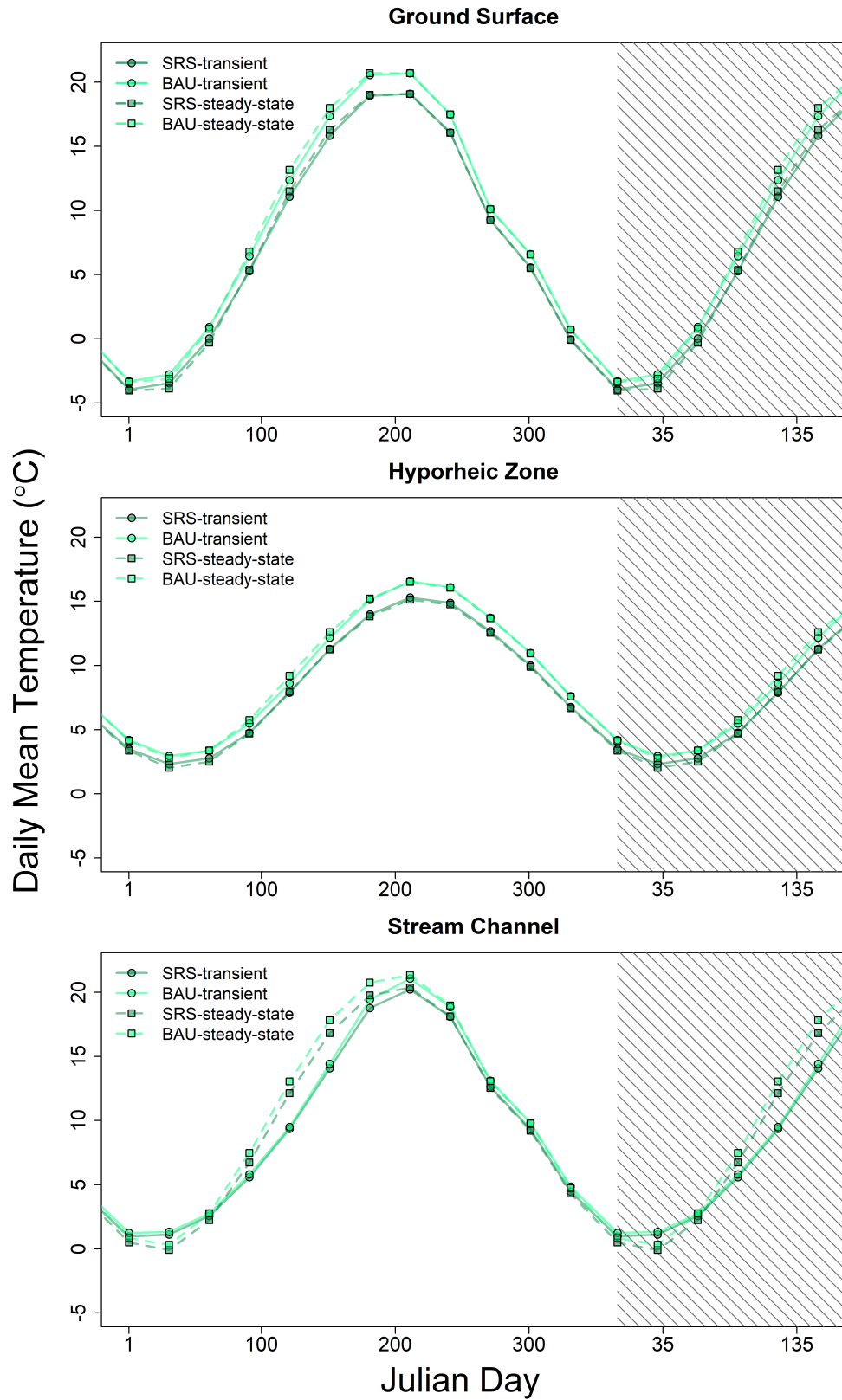


Figure 4.10: Daily mean temperatures of ground surface (top), hyporheic zone (middle), and stream channel (bottom) among model scenarios.

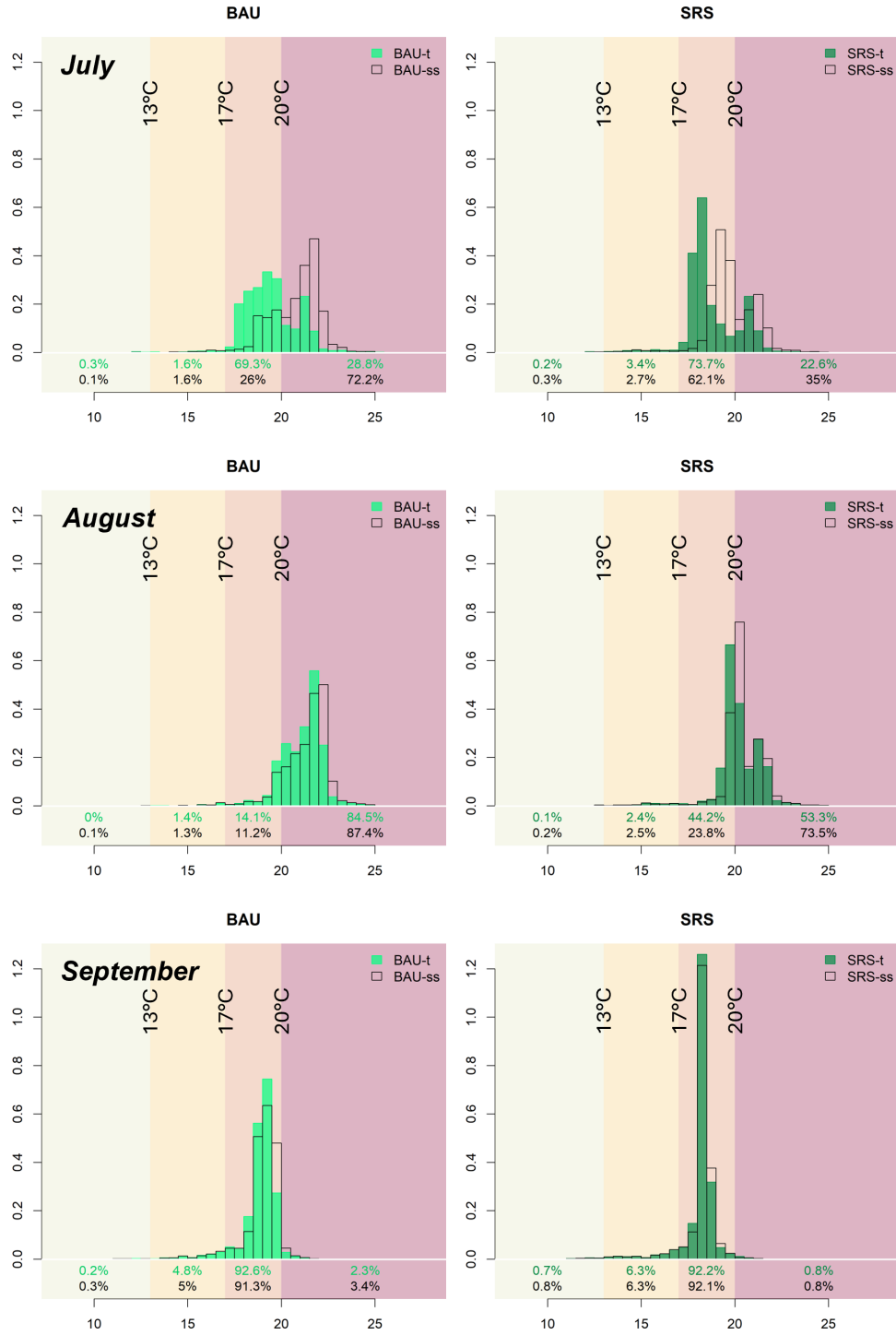


Figure 4.11: Distribution of channel temperatures for the BAU (left) and SRS scenarios (right). Transient scenarios are colored bars and steady-state scenarios are shown as bar outlines. Percentage of channel temperatures within each temperature range are displayed beneath distributions, colored text are percentages for the transient scenario and black text are percentages for the steady-state scenario.

Longitudinal Temperature Change

Among all model scenarios, longitudinal cooling of channel temperatures occurred in the winter and longitudinal warming occurred in the summer (Figure 4.12). In the steady-state scenarios (Figure 4.12A) BAU shade had greater longitudinal warming than SRS shade in the summer months. In winter, BAU shade caused less longitudinal cooling along the channel reach than the SRS shade scenario.

The higher flows in the transient flow condition scenarios caused substantially different longitudinal changes in channel temperature in the spring and early summer compared to the steady-state results (Figure 4.12B). Differences between daily average upstream and downstream temperatures were relatively small during higher flows from February–May in the BAU-t and SRS-t scenarios. Springtime flooding in transient scenarios resulted in channel transport dominating the translation of upstream temperatures to downstream, minimizing the potential for asding of hyporheic flow paths to have meaningful influence over the reach (Figure 4.12B). The amount of longitudinal warming began to diverge between BAU-t and SRS-t scenarios in June. In July and August, the reduced floodplain shade of the BAU-t scenario led to much greater longitudinal warming warming ($> 2^{\circ}\text{C}$) than the SRS-t scenario ($\sim 1^{\circ}\text{C}$).

Discussion

We found that a “shade retention strategy” for restorations (SRS model scenarios), which prioritizes non-streamside floodplain shade retention, led to cooler simulated channel temperatures than a “Business as usual” restoration, where retaining non-streamside floodplain vegetation was not prioritized. General conclusions of the effects of shade scenarios (SRS v. BAU) on simulated base-flow temperatures are similar between the steady-state and transient scenarios, i.e. reduced floodplain shade generally led to warmer channel

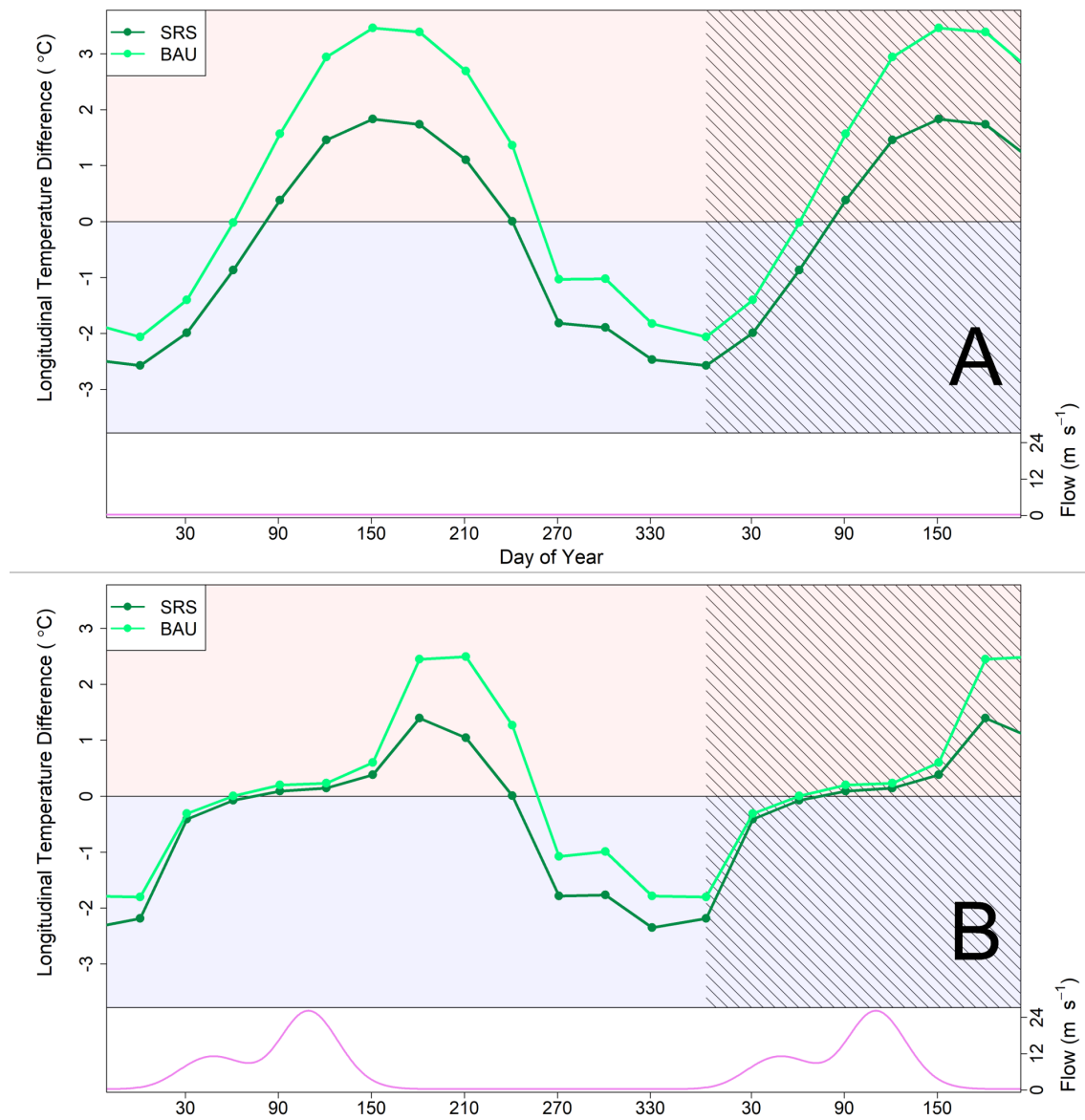


Figure 4.12: Difference in downstream and upstream daily mean stream temperatures across restoration reach comparing the effects of the “Business as usual” (BAU) shade scenario to the “Shade retention strategy” (SRS) for the steady-state hydrology models (A) and transient hydrology models (B). Positive values indicate downstream warming across the reach and negative values downstream cooling. Area of plots with diagonal shade lines indicates that these values are a repetition of the output values.

temperatures. However, the temperature magnitude of steady-state scenarios differed from transient scenarios and specific conclusions during times of the year which typically experience high flows are likely inaccurate. For instance, springtime floods diminished the effects of floodplain shade on channel temperatures in transient scenarios (similarity in spring channel temperatures of BAU and SRS, Figures 4.10 and 4.12A), whereas steady-state scenarios show that floodplain shade had a substantial effect on channel temperatures in the springtime (difference in spring channel temperatures of BAU and SRS, Figures 4.10 and 4.12B).

Comparing the channel temperatures for the steady-state and the transient flow scenarios during springtime high flows demonstrates how higher flows minimize the influence of the net atmospheric transfer of heat on mean temperatures (Figure 4.10) and on changes over the reach (Figure 4.12). High flows reduce the relative influence of atmospheric heat transfers compared to lower flows [Poole et al., 2008]. Temperature reflects the heat energy per unit volume of water, therefore when the volume increases and the heat flux does not, the temperature of the volume decreases. The surface area of the channel also increases with discharge, but the ratio of surface area to volume decreases, thus the greater channel surface area exposed to atmospheric heat exchange does not make up for the increase in volume of the channel, thus still lessening the effect of atmospheric heat loads. The interplay between heat load on the channel surface and channel volume is exemplified in scenarios run with transient flow conditions: mean channel temperatures are cooler in transient scenarios during high flows than in steady-state base-flow scenarios with the same atmospheric and floodplain shade conditions (Figure 4.10).

Fluctuating flows create head gradients between the channel and hyporheic zone, thus influencing hyporheic exchange. As river stage rises, the hyporheic zone expands, as more water is pushed into the alluvial aquifer [Watson et al., 2018, Wondzell and Swanson, 1999]. The opposite occurs when river stage falls—head gradients switch and more hyporheic water

upwells back to the channel. This expanding and contracting of the hyporheic zone influences hyporheic exchange, and therefore likely has a large impact on channel temperatures in coarse-grained alluvial floodplain rivers. Cool water downwells into the hyporheic zone as flood flows increase in the winter and spring, and upwell back to the channel in the falling limb and throughout the summer. We saw evidence of this phenomenon in our transient verses steady-state scenarios; transient scenarios had cooler simulated channel temperatures than the steady-state base-flow scenarios (Figure 4.10, 4.11).

Differences in the transient flow and steady-state flow scenarios demonstrate how an expanding and contracting hyporheic zone contributes to its thermal capacitance [sensu Fogg et al., 2023], beyond that created by steady-state hyporheic flow, alone. Although winter and spring river temperatures are often not a primary concern for river managers, accurate temperature simulation in winter/spring is important for accurate temperature simulation in summer/fall especially in streams with high gross hyporheic exchange rates. If results from transient scenarios were the “true” temperature of Meacham Creek and we simulated heat dynamics on Meacham Creek using steady-state hydrology instead of an annual fluctuating signal, simulation results would overestimate temperatures on the rising limb (winter/spring) of the annual cycle by a few degrees (Figure 4.10).

Previous research has shown that modeling transient flow conditions is import in understanding the exchange of water between the hyporheic zone and stream channel [Bobst et al., 2022, Gomez-Velez et al., 2017, Wroblicky et al., 1998, , among others], but few studies have shown the influence of transient annual hydrology on temperature dynamics in streams where a large fraction of channel flow is subject to hyporheic exchange. Our results show that spring floods significantly alter the temperature dynamics in floodplain streams with high hyporheic exchange, thus reinforcing conclusions made by Watson et al. [2018] who show that floods have a significant influence on temperatures in lateral, subsurface riparian flows.

Mitigating disturbance to the vegetation providing shade from restoration activity may aid in the recovery of the stream ecosystem. Our results show that adopting a “shade retention strategy” for channel re-meandering restorations can prevent warming of hyporheic and channel temperatures caused by a “business as usual” restoration. Patches of floodplain vegetation were cleared on Meacham Creek in the “business as usual” strategy (Figure 4.7). The somewhat minor reduction in vegetation > 3 m on the floodplain of the restoration reach (5.7%) caused warming of mean channel temperatures (Figure 4.10), higher percentage of channel temperatures $> 20^{\circ}\text{C}$ (Figure 4.11), and greater summertime longitudinal warming (Figure 4.12). Even minor temperature increases ($1\text{-}2^{\circ}\text{C}$) can have a large impact on thermally-sensitive cold-water species as metabolic rates are often non-linearly related to temperature [Brewitt and Danner, 2014]. Active retention of floodplain shade during restorations may decrease the prevalence of physiologically stressful thermal conditions (SRS scenario, Figure 4.11), which may otherwise be more prevalent (BAU scenario, Figure 4.11). Active floodplain shade retention may thus decrease the thermal disturbance to the ecosystem. Although grasses and shrubs grow back relatively quickly after a disturbance, Hewlett and Fortson [1982] shows that this type of vegetation does not provide a substantial enough attenuation of solar radiation to reduce shallow groundwater warming.

Conclusion

Large-scale restorations such as channel re-meanderings cause disturbance to the stream ecosystem and mitigating stressors to biota may aide in recovery of populations after the restoration. We show that a strategy which prioritizes retaining all vegetation on the floodplain (“Shade retention strategy”) may prevent warming of channel temperatures via heat exchange between the hyporheic zone and atmosphere. We urge an expanded consideration of floodplain shade, as the “riparian buffer” concept is naive in places where

hyporheic flows extend laterally into floodplain gravels. We also showed that simulating transient annual hydrology is necessary for accurate representation of heat and water exchange between the channel and hyporheic zone. As river temperatures are expected to rise with shifts in global climate patterns and human-caused alterations to fluvial landscapes, maintaining floodplain forests may be important in the management of channel temperatures of coarse-grained alluvial floodplain rivers.

CONCLUSION

The work presented here builds on the theoretical scaffolding for understanding how floodplain shade has the potential to affect the hyporheic heat dynamics that ultimately influences stream channel temperatures in coarse-grained alluvial floodplain rivers. I hypothesized that floodplain shade mediates the conduction of heat through the vadose zone, thus influencing hyporheic temperatures and stream channel temperatures upon hyporheic upwelling. I used simulation modeling experiments to progressively show how floodplain shade influences heat dynamics of a hyporheic flow path (Ch. 2), of the hyporheic zone and stream channel (Ch. 3), and how real distributions of floodplain shade may influence salmonid habitats and restoration outcomes (Ch. 4).

Using a numerical experiment in Chapter 2, I compared simulated hyporheic temperatures among scenarios with either a surface temperature boundary representative of shaded soil or unshaded soil, different vadose zone thicknesses (0.5–3.0 m), and different hydraulic conductivity values (100 vs. 400 m day⁻¹). I found that vertical heat exchange between the atmosphere, vadose zone, and hyporheic zone is an important driver of temperature along a hyporheic flow path. Interactions between vertical soil temperature signals and the horizontal downwelling river temperature signal caused differential amplitude attenuation in seasonal cycles of hyporheic temperature. Conclusions from our experiment in Chapter 2 expand on work done by Molina-Giraldo et al. [2011] and Vogt et al. [2010] which show that 1-dimensional simplifications of advective-dispersive heat exchange do not well represent amplitude attenuation in lateral hyporheic flow paths. Furthering understanding of the influence of vertical heat exchange on hyporheic temperatures, I show how differing soil temperatures caused by differences in vegetative shade cause differences in amplitude attenuation of the downwelling river temperature signal in the hyporheic zone.

In Chapter 3 I used a 3-dimensional simulation of a representative coarse-grained

alluvial floodplain river to test the effects of floodplain shade on simulated hyporheic and channel temperatures. I experimentally varied the distribution of shaded and unshaded solar radiation signals across the wetted channel and ground surface of the dry floodplain and found that the spatial location of shade profoundly impacted hyporheic and channel temperatures in our simulations. Results from this simulation experiment suggest that a fully shaded floodplain has a similar effect on channel temperatures as fully shading the channel itself. Although these experimental scenarios are idealized representations of naturally occurring shade distributions, our results show that loss of floodplain may be an unrealized degradation in coarse-grained alluvial floodplains and promotion of floodplain forests far from the channel has unrealized management potential.

In Chapter 4, I used real distributions of floodplain shade to compare two different shade retention strategies during channel re-meandering restoration: 1) a strategy which prioritizes the retention of floodplain shade and 2) a strategy which had no such priority, thus leading to area of the floodplain cleared of vegetation. I found that a floodplain shade retention strategy leads to cooler channel temperatures compared the restoration actions which reduced floodplain vegetation by $\sim 5\%$. The reduction of floodplain shade led to warmer hyporheic temperatures which led to greater longitudinal warming of channel temperatures and more physiologically stressful channel temperatures. These results suggest that mitigating disturbance of the floodplain vegetation may prevent excessive temperatures in the stream channel after a re-meandering restoration. In addition to testing the effects of floodplain shade retention strategies, I ran model simulations with steady-state base-flow or a transient signal representative of snow-melt hydrology. Our results suggest that spring floods alter the temperature dynamics in floodplain streams with high hyporheic exchange rates. This furthers work by Watson et al. [2018], who show that floods have a significant influence on temperatures in lateral hyporheic zones.

Stream-side vegetative shade is minimal on coarse-grained alluvial floodplain rivers of

mountainous regions [Whited et al., 2007] due to high flows from snow-melt that creates wide annual scour zones. This fluvial landscape does not allow for the establishment of substantial vegetative shade on the channel banks—a common restoration strategy for mitigating stream temperatures that are too warm to support cold-water species [Beschta, 1997, Johnson, 2004, Justice et al., 2017, Moore et al., 2005]. I show in this dissertation that floodplain shade, i.e., shade which does not block solar radiation from the base-flow channel but which occurs over lateral hyporheic flows, has the potential to mediate warming of shallow hyporheic water, leading to cooler upwelling hyporheic temperatures and cooler channel temperatures. Protection of floodplain forest shade in montane floodplain streams may be imperative for cooling stream temperatures and providing physiological refuges for cold-water species in a warming world.

Understanding of the interactions between floodplain shade, hyporheic heat exchange and channel temperatures may improve restoration and management outcomes for the CTUIR efforts in restoring “First Foods” (steelhead and bull trout) across the Umatilla River basin [Quaempts et al., 2018]. Insight on the drivers of channel temperature aides in the CTUIR river vision and its aim to restore stream ecosystem processes [Jones et al., 2008a]. The successful restoration of steelhead and bull trout to CTUIR lands, including on Meacham Creek, is important for sustaining cultural traditions and religious ceremony for the peoples of the CTUIR [Quaempts et al., 2018].

Anthropogenic manipulations of fluvial landscapes, along with shifts in global climate patterns have caused the warming of temperature patterns in many rivers [Isaak et al., 2012]. Warming stream temperature regimes can lead to degraded water quality, altered ecosystem functions, and stress on important cold-water biota [Sullivan et al., 2021]. Floodplain rivers with coarse-grained alluvial beds are critical habitat for persistence of salmonid populations [Hauer et al., 2016]. Protecting floodplain forest shade and hyporheic exchange processes may thus be important for the persistence of cold-water habitat in river-floodplain systems.

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