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[1] We monitored summertime base flow water temperatures of hyporheic discharge to surface water in main, side, and spring channels located within the bank-full scour zone of the gravel- and cobble-bedded Umatilla River, Oregon, USA. Diel temperature cycles in hyporheic discharge were common, but spatially variable. Relative to the main channel's diel cycle, hyporheic discharge locations typically had similar daily mean temperatures, but smaller diel ranges (compressed by 2 to 6°C) and desynchronized phases (offset by 0 to 6 h). In spring channels (which received only hyporheic discharge), surface water diel cycles were also compressed (by 2 to 6°C) and desynchronized (by −4 to 6 h) relative to the main channel, creating diverse daytime and nighttime mosaics of surface water temperatures across main, side, and spring channels, despite only minor differences (<1°C) in daily mean temperatures among the channels. The river's hyporheic zone received and stored heat from the channel, yet hyporheic return flows carried heat back to the channel minutes to months after removal. Associated surface water temperature dynamics were therefore complex. Hyporheic discharge was not simply "cooler" or "warmer" than main channel water. Instead, instantaneous temperature differences between channel water and hyporheic discharge typically arose from diel temperature cycles in hyporheic discharge that were buffered and lagged relative to diel cycles in the main channel.

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1. Introduction

[2] Water temperature is an overarching habitat characteristic that influences the physiology, distribution and abundance of aquatic biota [*Allen*, 1995]. In streams and rivers, water temperatures are influenced by a diverse set of hydrological and ecological processes [*Webb and Zhang*, 1997; *Poole and Berman*, 2001]. Two drivers have been well described in the scientific literature; shading of channels by riparian vegetation and the thermal influence of groundwater influx into rivers (Table 1 defines groundwater and other terms for the purposes of this paper). Riparian shade is an important driver of stream temperature in small forested streams where canopy closure (or near closure) can occur overtop streams [e.g., *Beschta and Taylor*, 1988; *Johnson and Jones*, 2000; *Malcolm et al.*, 2004; *Jones et al.*, 2006]. However, as stream width increases, less of the channel's surface area can be shaded by riparian vegetation. Thus in larger rivers, the influence of riparian shading on stream temperature wanes [*Poole and Berman*, 2001].

[3] The effects of groundwater on stream temperature can be explained by considering the unidirectional discharge of groundwater to river channels [e.g., *Hewlett and Fortson*, 1982; *Mellina et al.*, 2002; *Gaffield et al.*, 2005; *Brown et al.*, 2007], which is an appropriate conceptual model in some geomorphic contexts. However, in alluvial rivers and streams, the prevalence of bidirectional water flux between the river channel and alluvial aquifer ("hyporheic exchange") is well established [*Stanford and Ward*, 1988; *Findlay*, 1995; *Brunke and Gonser*, 1997; *Malard et al.*, 1999], and researchers have recognized tight coupling between the thermal dynamics of river channels and their underlying hyporheic zones.

[4] Because the channel and hyporheic zone are two components of an integrated hydrologic system [*Huggenberger et al.*, 1998; *Cardenas et al.*, 2004; *Poole et al.*, 2008], their associated temperature dynamics are interdependent [*Petts*,

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Table 1. Hydrologic Terminology

Term	Definition
Hyporheic zone	Portion of the alluvial aquifer that contains some proportion of water derived from the stream channel [*White*, 1993].
Hyporheic exchange	The bidirectional flux of water across the streambed, between the channel and hyporheic zone.
Hyporheic flow	Water movement within the hyporheic zone.
Hyporheic water	Water within the hyporheic zone; always includes water derived from the channel but may also include water derived from groundwater.
Hyporheic discharge	Hyporheic water emerging onto the Earth's surface; also known as "upwelling" or "outwelling."
Hyporheic recharge	Channel water entering the hyporheic zone; also known as "downwelling" or "inwelling."
Hyporheic flow path	Route of water flow through the hyporheic zone, from point of hyporheic recharge to point of hyporheic discharge.
Groundwater	Water beneath the water table that has not mixed with water from a stream channel [*White*, 1993].
Surface water	Any water atop the Earth's surface.
Channel water	Surface water contained in a stream channel.

2000; *Anderson*, 2005]. The magnitude of hyporheic exchange in a river system depends on characteristics of the river channel and alluvial aquifer, such as the range, frequency, and spatial variation in hydraulic conductivity and hydraulic gradients, as well as variations in floodplain and streambed topography and geomorphology [*Wondzell and Swanson*, 1999; *Dent et al.*, 2001; *Gooseff et al.*, 2006; *Wondzell*, 2006]. Thus the thermal regimes of the channel and hyporheic zone are apt to be most tightly integrated in rivers with highly conductive floodplain aquifers, and complex channel patterns or streambed topography, all of which enhance rates of hyporheic exchange [*Cardenas et al.*, 2004; *Poole et al.*, 2006]. Such conditions are most prevalent in anabranched rivers with alluvium dominated by gravel and cobble. In these rivers, hyporheic zones are especially expansive [e.g., *Stanford et al.*, 1994; *Jones et al.*, 2007], and hyporheic flow paths commonly range in length from meters to kilometers [*Poole et al.*, 2008].

[5] Where the magnitude of hyporheic exchange is sufficient, channel water temperatures are both influenced by and affect hyporheic water temperatures [e.g., *White et al.*, 1987; *Evans et al.*, 1995; *Constantz and Thomas*, 1997; *Evans and Petts*, 1997; *Arscott et al.*, 2001; *Malard et al.*, 2001; *Johnson*, 2004; *Fernald et al.*, 2006; *Loheide and Gorelick*, 2006]. In small streams, hyporheic discharge commonly affects channel temperature across the entire channel [*Storey et al.*, 2003; *Johnson*, 2004; *Loheide and Gorelick*, 2006], while in larger streams and rivers, hyporheic discharge has been shown to create thermal heterogeneity [*Arscott et al.*, 2001; *Fernald et al.*, 2006].

[6] The specific influence of hyporheic discharge on channel water temperature varies. Researchers have reported that hyporheic discharge both cools and/or warms stream channel water [e.g., *Ward et al.*, 1999; *Cozzetto et al.*, 2006; *Fernald et al.*, 2006]. Similarly, hyporheic water temperature may decrease with depth [e.g., *Evans and Petts*, 1997; *Schmidt et al.*, 2006] or increase with depth [e.g., *Hanrahan*, 2007]. Such conflicting findings might be rectified by considering water temperature cycles (diel or annual) rather than instantaneous readings. For instance, in temperate climates the annual range in hyporheic water temperature is typically less than that of channel water. Thus hyporheic discharge is commonly cooler than channel water in spring and summer, and warmer than channel water in the winter and fall [*Evans et al.*, 1995; *Poole et al.*, 2008]. Similarly, hyporheic discharge commonly reduces the diel range in channel water temperature [*Johnson*, 2004; *Loheide and Gorelick*, 2006]. Along short hyporheic flow paths (e.g., ones to tens of meters in coarse-grained alluvial aquifers) the channel's diel temperature cycle can penetrate into the hyporheic zone via advection and/or conduction, though the timing of maximum and minimum temperatures in hyporheic discharge if often lagged relative to channel temperature cycles [*Malard et al.*, 2001]. Thus understanding the effects of hyporheic exchange on channel water temperature requires a detailed assessment of temperature cycles, both in channel water and hyporheic discharge.

[7] Any regular cyclical pattern (e.g., sine wave) can be fully described by its mean, amplitude, phase, and period. Temperature cycles with in streams have two distinct periods: diel and annual. We use the terms cooled, buffered, and lagged as three basic terms to describe differences in the mean, amplitude, and phase between two temperature cycles (Figures 1a–1c). Cooled (or warmed) denotes a difference in means, buffered denotes a difference in range,

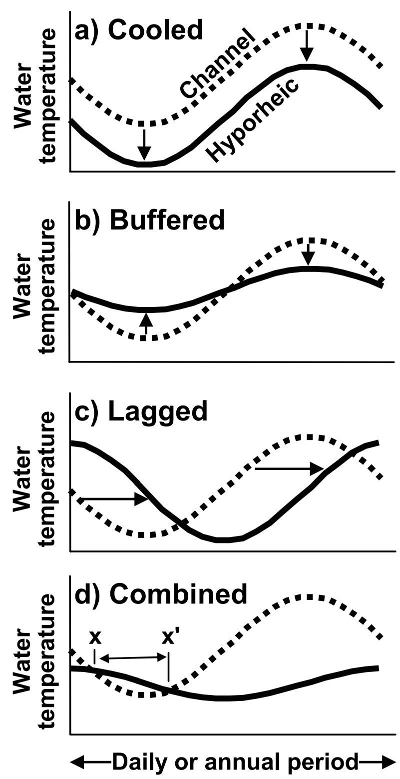

Figure 1. Metrics for describing differences between temperature cycles; (a–c) archetypical differences between two temperature cycles (d) can be combined to describe any observed difference.

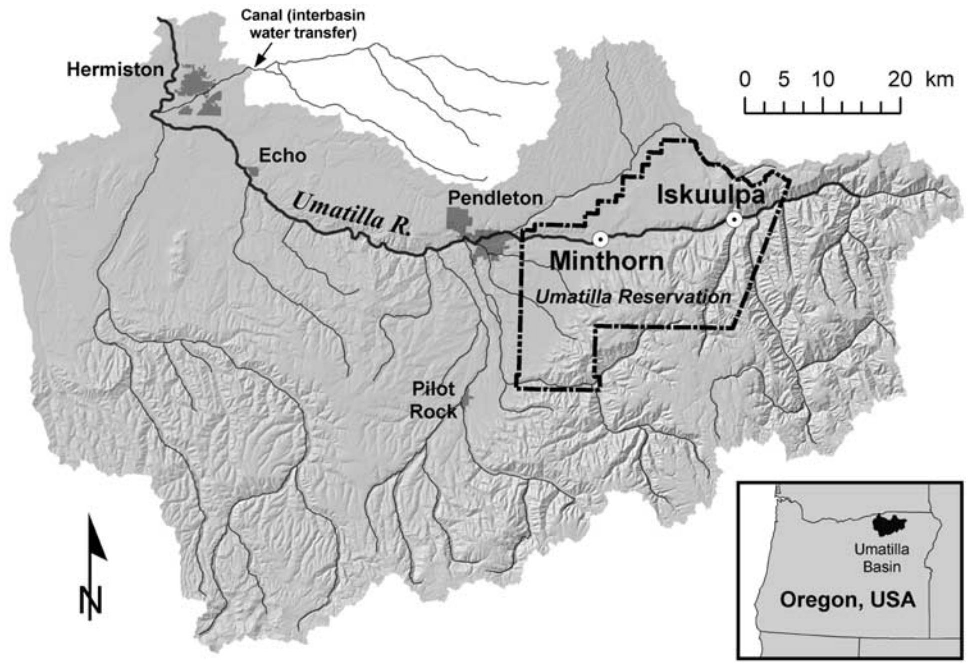

Figure 2. Location of the Minthorn and Iskuulpa study sites in the Umatilla River basin.

and lagged denotes a difference in phase. Differences in temperature cycles between channel water and hyporheic discharge arise as water moves along hyporheic flow paths [*Poole et al.*, 2008]. We present the patterns in Figures 1a–1c as simple archetypical representations of differences between two temperature cycles. In practice, these affects are apt to be combined in order to describe any observed difference between two cycles. (e.g., Figure 1d).

[8] Whenever hyporheic discharge and channel water have different temperature cycles, hyporheic discharge has the potential to influence channel water temperature. At any point in time, temperature readings of both will suggest that hyporheic discharge has either a warming or cooling effect on channel water. Notably, however, isolated readings can be misleading. For instance, if the temperature cycle of hyporheic discharge is lagged, an isolated reading can suggest that hyporheic discharge warms channel water, even if the diel cycle in hyporheic discharge is cooler than that in channel water (e.g., a reading taken between x and x′ in Figure 1d). Thus a holistic understanding hyporheic influence on channel water temperature requires analysis of channel and hyporheic temperature cycles. Such an approach can disentangle patterns of temperature change (sensu Figure 1) and help explain the variety of reported effects of hyporheic discharge on channel water temperatures.

[9] In the late 1990's, we collected summertime thermal remote sensing data [sensu *Torgersen et al.*, 2001; *Loheide and Gorelick*, 2006] on the Umatilla River, Oregon. Those data (S. O'Daniel, unpublished data, 1999) revealed unexpected variation in channel water temperature among the main channel, side channels, and spring channels (channel types defined in section 3.1). Because these temperature patterns were spatially uncorrelated with riparian shading, and stream gauging data revealed that this high-desert river receives negligible inputs of groundwater during the summer, we hypothesized that the observed channel temperature patterns resulted from hyporheic exchange within the river. This paper describes a field-based study designed to test our hypothesis and, if correct, document whether observed spatial variation in summertime channel water temperature is associated with diel temperature cycles in hyporheic discharge that are cooled, buffered, and/or lagged.

2. Study Site

2.1. Site Location and Characteristics

[10] The Umatilla River Basin (5930 km^2) is located in the high desert region of northeastern Oregon, USA (Figure 2). Precipitation in the Umatilla basin occurs mostly from October to May as rainfall on the floodplain (0.3 m/a), and as rain or snow in the Blue Mountains (0.8 to 1.8 m/a) (U.S. National Oceanographic and Atmospheric Administration). On average, daily air temperatures in January range from −2.8°C to 4.4°C and in July range from 13.9°C to 30.6°C (U.S. National Oceanographic and Atmospheric Association). The Umatilla River flows west out of the Blue Mountains for 185 km to its confluence with the Columbia River. The river's discharge is flashy; the USGS gauge at Pendleton (gauge ID 14021000) shows that base flow is ~1 m^3/s, while spring snowmelt and rain-on-snow events in the Blue Mountains induce typically brief (24–36 h) winter and spring freshets and flood events with flows ranging approximately from 50 m^3/s to >300 m^3/s.

[11] Our study took place on reservation lands of the Confederated Umatilla Tribes, east of the town of Pendleton, Oregon, and upstream of the significant agricultural water withdrawals that occur lower in the basin. On the reservation, the river forms a dynamic, anabranched channel (Figure 3) occupying a well-developed (0.3–1 km wide) floodplain that is underlain by a ~3 to 4 m thick unconfined alluvial floodplain aquifer bounded by basalt. As flood waters expand to fill the bankfull channel, the river scours vegetation from a broad swath of the floodplain. Low flows then retreat into a much narrower channel, leaving the base

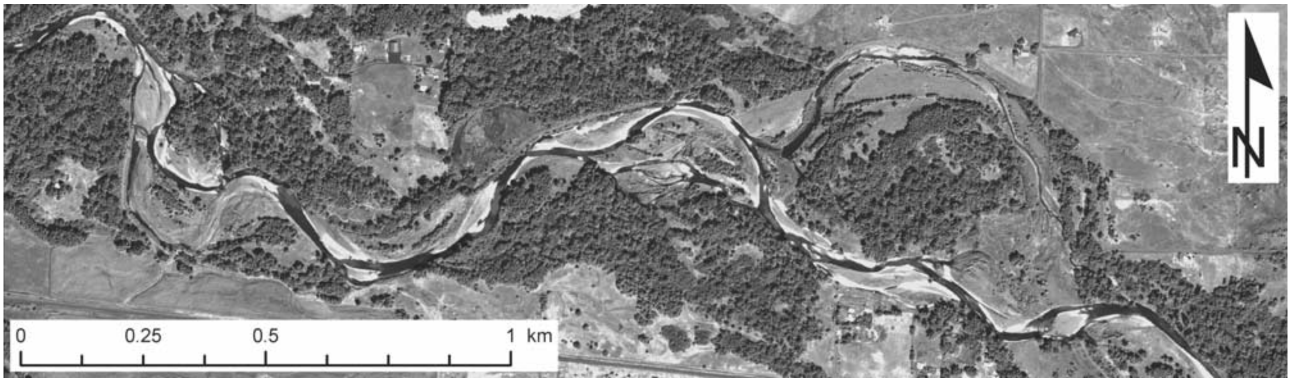

Figure 3. Quickbird satellite image of the Umatilla River showing the anabranched channel pattern and unshaded low-flow main channel.

flow channel generally unshaded by riparian vegetation (Figure 3). During the summer, USGS gauges at the east (upstream) and west (downstream) boundaries of the Umatilla Reservation document a slight (~10%) reduction in river discharge from the upstream to downstream ends of this 40 km section of floodplain, suggesting summertime losses to evapotranspiration are somewhat greater than any additional flow provided by the river's small tributaries.

[12] We conducted our research on two study reaches, "Iskuulpa" (pronounced e-SKUL-puh) and "Minthorn" (Figure 4). Each was located on a larger study site with the same name, encompassing the entire width of the floodplain. The Iskuulpa reach was located ~32 km upstream of the Minthorn reach, and the Minthorn reach was located ~9 km upstream of the western reservation boundary (Figure 2). Both reaches had complex channel patterns. The main channel at the Iskuulpa reach was sinuous and the reach included several in-channel bars that created smaller side and spring channels. The Minthorn reach was located in a section of the river where channel adjustments are common during freshets and floods. The main channel was less sinuous, but the reach contained abundant side and spring channels separated by multiple in-channel bars.

2.2. Hyporheic Hydrology of the Umatilla River

[13] At both study reaches the saturated portion of the alluvial aquifer averaged ~2–3 m in thickness; on the floodplain adjacent to the channel, the overlying unsaturated zone ranged from 0 to ~2 m. The aquifer consisted of basalt gravel, cobbles, and boulders, with some sand and silt intermixed, and was underlain by basalt. The hydraulic conductivity of the aquifer was high (300–700 m d^{-1}) (B. Boer, unpublished data, 2003). Geochemical analyses of channel water and water collected from monitoring wells installed in the alluvial aquifer showed that the hyporheic zone extended across the entire floodplain, and that hyporheic water throughout the aquifer was dominated by water derived from the channel [*Jones et al.*, 2007].

[14] In such a porous and conductive aquifer, hyporheic flow path lengths are among the longest reported. *Poole et al.* [2008] used simulation modeling and data from 48 monitoring wells to show that hyporheic flow distances commonly range from tens to thousands of meters. *Poole et al.* [2008] also demonstrated that hyporheic water traversing short flow paths (less than ~100 m) had a daily mean temperature nearly identical to that of the river, but diel variation that was buffered or lagged somewhat (Figure 5a). Water flowing through longer hyporheic flow paths (greater than ~100m) had no diel variation and a daily mean temperature that diverged from that of the main channel (i.e., cooler in the summer and warmer in the winter) due to buffered and lagged annual temperature cycles (Figure 5b; see also *Bundschuh* [1993]).

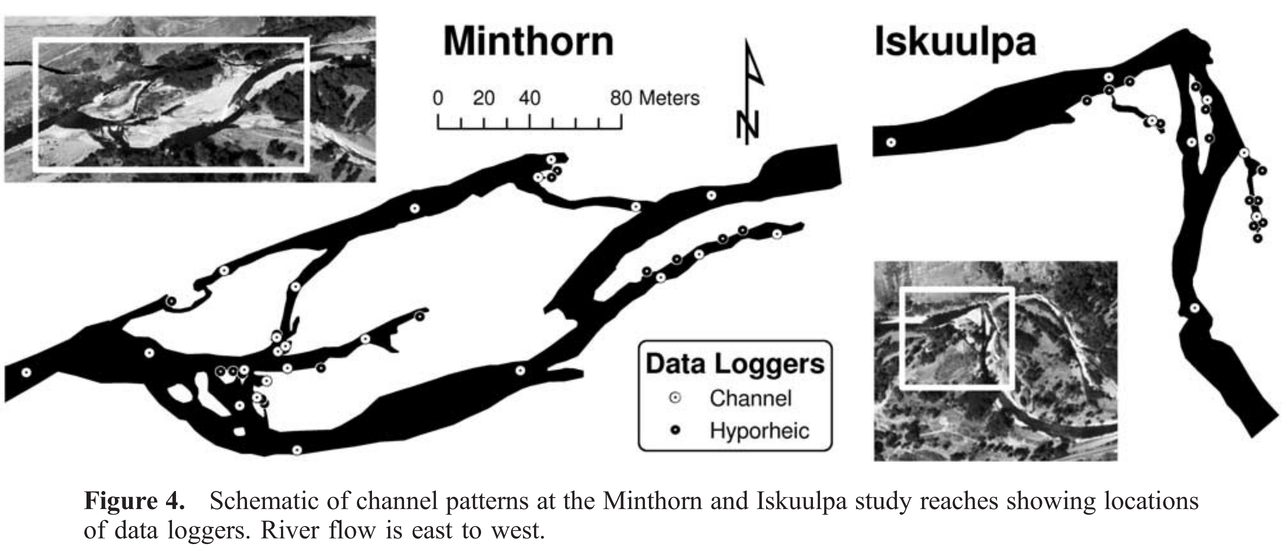

Figure 4. Schematic of channel patterns at the Minthorn and Iskuulpa study reaches showing locations of data loggers. River flow is east to west.

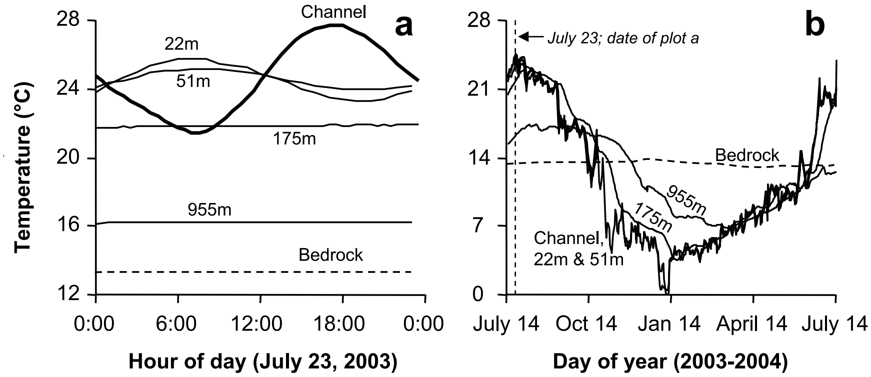

Figure 5. (a) Diel and (b) annual temperature cycles for hyporheic flow paths of different lengths within the Umatilla River alluvial aquifer. Data from *Poole et al.* [2008] (Copyright © 2008 by John Wiley & Sons, Inc. Reprinted by permission of John Wiley & Sons, Inc.).

[15] Thus previously published results from our work provide background data on Umatilla River hyporheic hydrology that support the water temperature analyses described in this paper. First, on the Umatilla River floodplain, any water emerging from the subsurface to the surface was hyporheic water and most of this hyporheic discharge was originally derived from the channel [*Jones et al.*, 2007]. Second, temperature cycles in hyporheic discharge can provide a rough indication of the associated hyporheic flow path length (Figure 5) [*Poole et al.*, 2008]. Third, such observations of hyporheic temperature cycles across the alluvial aquifer revealed the range of variation in diel and annual cycles across the hyporheic zone, and therefore defined the breadth of potential influences that hyporheic discharge may have on channel water temperature cycles.

3. Methods

[16] In early July, 2003 (just prior to our study), preliminary field reconnaissance surveys revealed a diversity of summertime water temperatures among channel habitats, consistent with remote sensing data collected in the late 1990s. The patterns appeared closely related to channel and floodplain geomorphic features (e.g., the arrangement of bars, secondary channels, etc.), further suggesting that hyporheic exchange was a primary driver of observed thermal variation in channel water. To document the differences in channel water temperature and determine whether those differences were consistent with temperature patterns in hyporheic discharge, we monitored diel temperature cycles in channel water and hyporheic discharge across channel habitats in both study reaches. We compared temperature cycles of channel water and hyporheic discharge within and among spring channels, side channels, and the main channel.

[17] Field observations revealed the locations of hyporheic discharge were primarily on the downstream edge of gravel bars or in side channels and spring channels – suggested that hyporheic discharge was derived from the adjacent main channel and therefore, traversed relatively short (<50 m) hyporheic flow paths [see also *Fernald et al.*, 2006]. To verify this assumption, we employed two lines of empirical evidence. First, we created maps of the potentiometric surface across gravel bars to estimate the direction of hyporheic water movement. Second, we compared the temperature cycles of hyporheic discharge to cycles associated with hyporheic flow paths of different lengths, as determined by *Poole et al.* [2008] and shown in Figure 5.

3.1. Channel Typology

[18] Within our study reaches, we classified channels into three categories (main channel, side channel, or spring channel) based on their source water and rate of discharge. The main channel was the contiguous channel flowing through the site that maintained the highest discharge during the course of the study. Side channels were all other channels originating from and returning to either the main channel or another side channel. Spring channels were flowing channels that had no upstream surface water connection to main or side channels, but instead derived flow only from hyporheic discharge.

3.2. Creating Gravel-Bar Potentiometric Surface Maps

[19] To estimate hyporheic flow direction through bars, we surveyed the water surface elevation using a Topcon electronic theodolite along channel margins and where piezometers were installed in bars and islands. We started and ended each survey at known benchmarks established on the floodplain by a professional surveyor. Ending elevations for our survey loops agreed with the professional survey elevations within ±2 cm. A potentiometric surface map was created from resulting survey data by interpolating the water table between measured survey points using a spline algorithm in ArcGIS 9.1 (ESRI).

3.3. Monitoring Water Temperature Cycles

[20] We monitored temperature cycles in channel water and hyporheic discharge at multiple locations across the three channel types using two types of data loggers: Onset Tidbit data loggers (temperature range: −4°C to +37°C), and smaller iButton data loggers (temperature range: −5°C to +26°C). All data loggers were calibrated in a 22°C water bath (the approximate mean temperature of the river) for 30 min in a laboratory. From these data, a correction factor (either positive or negative) was established for each logger and added to the field results recorded by that logger. More extensive calibration tests by *Johnson et al.* [2005] showed that expected error in similarly calibrated temperature measurements among loggers was <0.25°C.

[21] The Tidbit data loggers were used to measure channel water temperatures in spring channels, side channels,

and the main channel. The loggers were deployed at mid depth throughout the study sites by attaching them to 1.24 cm diameter rebar installed in the streambed. The smaller iButton data loggers were used to monitor the diel temperature cycles in hyporheic discharge. At the channel margin of spring channels, side channels, and the main channel, seeps of hyporheic water emerged from the gravel at the channel margin, elevated ~1 to 5 cm above the river surface. At 30 such locations, a 30 cm × 3.8 cm diameter perforated (every 3 cm) PVC piezometer was installed in the stream bank, about ~10 cm laterally from the channel margin. iButton loggers were deployed inside the piezometers, ~10 cm below the water table, wrapped in a small nylon mesh bag with an attached nylon string (used for retrieval) extending out of the top of the piezometer.

[22] Eight channel water loggers and 14 hyporheic discharge loggers were deployed at Iskuulpa (Figure 4). Twenty-one channel water loggers and 16 hyporheic discharge loggers were deployed at Minthorn; however, one channel water logger and one hyporheic discharge logger were not recovered after deployment. The Tidbit and iButton data loggers recorded temperature every 15 min. Loggers were deployed between 10 and 17 July and retrieved between 13 and 14 August 2003.

3.4. Data Analysis

[23] The "typical" diel temperature cycle at each sampling location was determined from the logged temperature measurements by calculating the mean temperature for each hour of the day across the sampling period (defined as 4–12 August 2003, the period for which we had a complete data set for all loggers). Diel temperature cycles closely approximated a sine wave (see Results); therefore, we fit the mean daily temperature (M) in °C, diel temperature range (R) in °C, and phase, (P) in hours to the observed typical diel cycle for each temperature logger by minimizing RMSE for the following sinusoidal equation:

$$T_h = (0.5R)\cos((h - P)c) + M \tag{1}$$

where T_h is the mean hourly water temperature (°C) for a given hour of the day (h), and c is $2\pi/24$, a constant to convert radians to hour of the day. Since P is cyclical over a 24 h period, we constrained P to range between 0 and 24 when fitting parameters. Fitting the data to equation (1) enabled us to use each logger's full data set to objectively interpolate values of M, and R between the data recording precision of the data loggers (the increment of temperature readings; 0.16°C for Tidbits and 0.125°C for iButtons) and interpolate P between the time interval of the readings (15 min).

[24] Within each study reach, we averaged main channel values of M, R, and P to characterize a typical main channel temperature cycle for each study reach ($\bar{M}_m$, $\bar{R}_m$, and $\bar{P}_m$, where m denotes main channel). By comparing M, R, and P derived from each logger to values of $\bar{M}_m$, $\bar{R}_m$, and $\bar{P}_m$, we calculated three metrics that quantify the difference between diel temperature cycles at each data logger location and the typical diel temperature cycle in channel water. The deviation in diel cycle mean temperature between a logger location and the main channel (ΔM) is:

$$\Delta M = M - \overline{M}_m \tag{2}$$

Negative values of ΔM represent a logger where the mean temperature of the diel cycle is less than that of the main channel. Therefore ΔM is an empirical measure of whether an observed temperature cycle is cooler or warmer than the main channel temperature cycle. The deviation in diel range between a logger location and channel water (ΔR) is:

$$\Delta R = R - \overline{R}_m \tag{3}$$

Negative values of ΔR represent a logger where the diel range of the observed water temperature cycle is buffered relative to that of the main channel. The deviation in phase between a logger location and the main channel (ΔP) is:

$$\Delta P = P - \overline{P}_m \tag{4}$$

Positive values of ΔP represent a logger where the timing of peak water temperature is lagged relative to that in the main channel. Since P and $\bar{P}_m$ are cyclical over a 24 h period, we adjusted all values of ΔP to be within the range of −12 h to +12 h.

[25] ΔM, ΔR, and ΔP are quantitative measures of the deviation of any observed temperature cycle from that of the main channel. As such, we used these values to assess whether the influences of hyporheic discharge on channel water temperature cycles are due to cooler (or warmer) mean temperatures, buffered ranges, and/or lagged phases in hyporheic discharge.

4. Results

4.1. Channel Classification

[26] Based on our categories of main, side, and spring channels, the Iskuulpa reach was dominated by the main channel habitat, with one short side channel and two spring channels. In contrast, secondary channels dominate the Minthorn reach, which contains four side channels, and two spring channels (Figure 6a). As expected, estimated hyporheic flow path directions through gravel bars (traced perpendicular to equipotential lines) suggested that hyporheic water beneath gravel bars flowed from main or side channels toward data loggers placed at observed locations of hyporheic discharge (Figure 7).

4.2. Diel Temperature Patterns

[27] Maps of water temperature variation through the diel cycle and among channel units (Figure 6b) reveal substantial temperature variation across both sites. Large temperature swings (~6°C) occur in the main and side channels over a 24 h cycle, while ranges in spring channels are buffered and out of phase with the main and side channels.

[28] Because equation (1) was fit to the diel temperature cycle derived from each logger, we have a RMSE and r^2 value for each logger (n = 57). The RMSE was <0.5°C for most loggers (Figure 8a), and the r^2 was >0.80 for all but one of the loggers (Figure 8b). Both RMSE and r^2 were strongly related to the magnitude of the diel temperature range (R). Not surprisingly, RMSE tended to be higher where R was high, while r^2 tended to be more variable as R decreased (because the signal-to-noise ratio in the data decreased with R). Regardless, equation (1) provided an excellent representation of logger data across the range of r^2

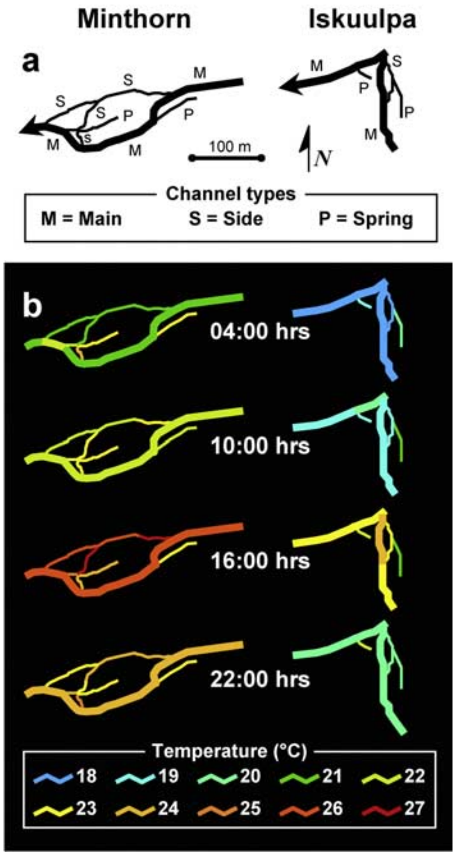

Figure 6. (a) Classification of channel types at the Minthorn and Iskuulpa reaches. (b) Mean hourly values from diel temperature cycles recorded by water temperature loggers deployed in channel water during study period (3–12 August 2003).

and RMSE values (Figures 8c–8e), and therefore provided reliable estimates of M, R, and P for each logger.

[29] Overall, the mean temperature of the diel cycle (M) in both channel water and hyporheic discharge was approximately 3°C cooler at the Iskuulpa site than at the Minthorn site (Table 2). Within both study sites, however, values of M for channel water and hyporheic discharge were generally similar to one another and similar across all channel units. The notable exception was hyporheic discharge within the Iskuulpa spring channels, where M ranged from 4.5°C below to 1.5°C above that of channel water. The mean and range among loggers for diel temperature range (R) and phase (P) are also reported in Table 2, but patterns in these data are most easily interpreted using plots of ΔM, ΔR, and ΔP, described next.

[30] ΔM, ΔR, and ΔP represent the diel cycles at sampling locations relative to the local average main channel diel temperature cycle. Therefore although main channel conditions differ between the Minthorn and Iskuulpa sites, the ΔM, ΔR, and ΔP data sets are relative measures and therefore Minthorn and Iskuulpa data can be combined and plotted together (Figure 9). With few exceptions (two high values and one low value in Figure 9f), ΔM is similar across the two sites and all channel units for channel water and hyporheic discharge (Figures 9a–9f), revealing that M in hyporheic discharge, side channels, and spring channels is typically similar to that in the main channel. The three extreme values in Figure 9f are all associated with hyporheic discharge to spring channels at Iskuulpa.

[31] Values of ΔR suggest that the diel range is typically similar among main and side channel water (Figures 9g–9h), although one side channel logger (which was situated near a strong hyporheic discharge location) exhibited a markedly depressed diel range relative to main channel water (Figure 9h). The diel range of spring channels, however, was consistently lower than that of main channel water (Figure 9i). For hyporheic discharge, the diel range is commonly reduced relative to main channel water, regardless of the channel unit type where the hyporheic discharge occurs (Figures 9j–9l).

[32] Values of ΔP show that the timing of peak temperatures in main and side channel water is largely synchronized, but water temperature in spring channels peaks anywhere from 3 h before to 6 h after main channel water (Figures 9m–9o). ΔP for hyporheic discharge most commonly ranges from 0 to 6 h, although values were observed across the entire range of −12 to 12 h (Figures 9p–9r). Negative values of ΔP are responsible for much of the observed variation in ΔP across the study reaches. However, truly negative values of ΔP are unlikely. Instead, negative values of ΔP are most easily interpreted as diel cycles that were lagged >12 h, although we made no attempt to differentiate between negative values of ΔP and values >12 h. Additionally, some negative values of

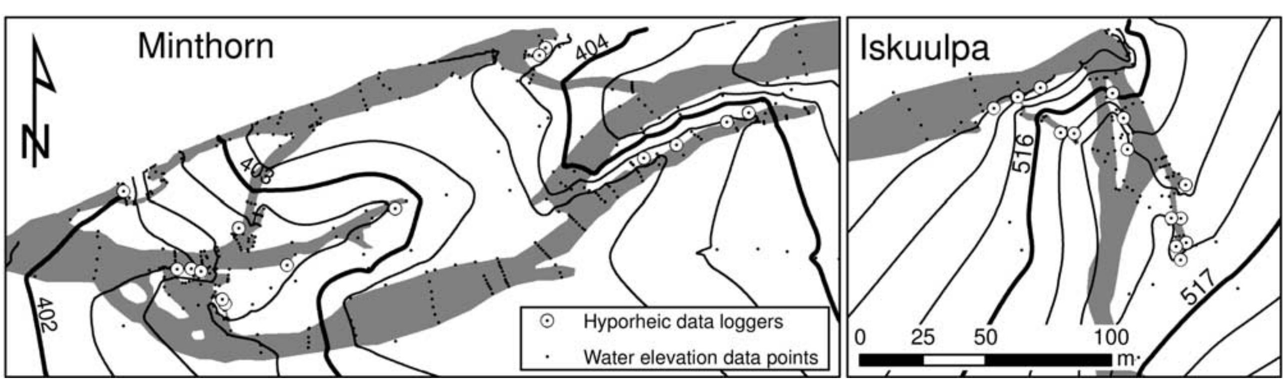

Figure 7. Interpolated potentiometric surfaces in the Minthorn and Iskuulpa study reaches derived from surveys of surface water stage and water table elevation (black points).

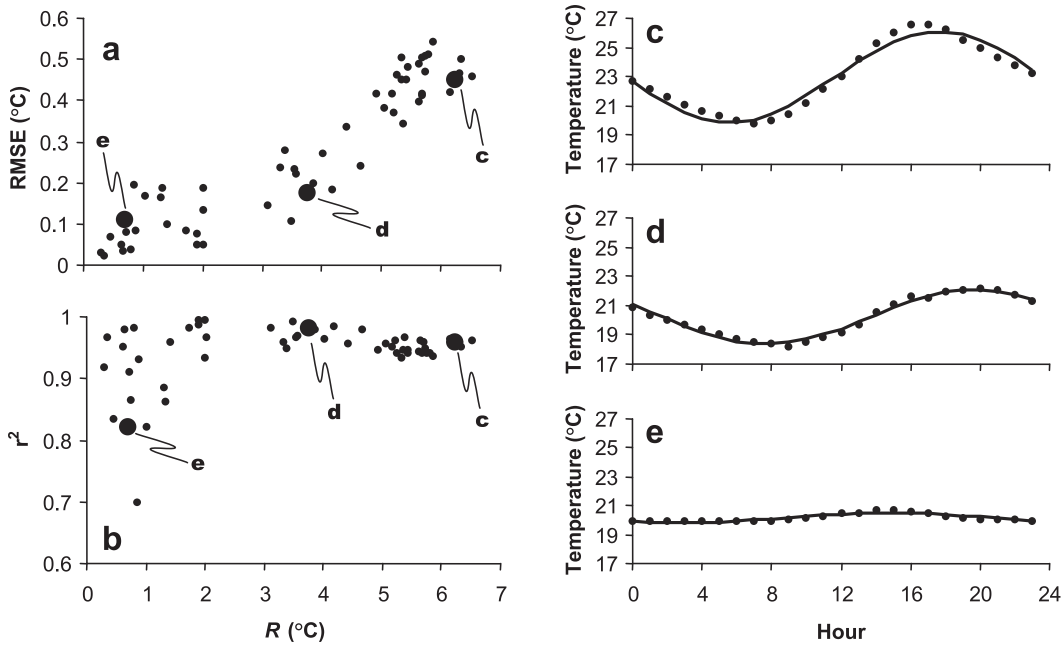

Figure 8. "Goodness of fit" for equation (1) to the mean diel temperature cycle recorded by each temperature logger. (a) Root mean squared error (RMSE) and (b) r^2 for each logger were related to *R* (the diel range in water temperature recorded by each logger). (c–e) Examples of fit by equation (1) (line) to observed mean diel temperature cycle (points) for three data loggers that span the approximate rage of RMSE and r^2.

ΔP were associated with sites where *R* was negligible. As *R* approaches zero, *P* becomes ambiguous and, therefore, ΔP becomes rather arbitrary (e.g., Figure 8e).

5. Discussion

5.1. Hyporheic Flow Path Direction and Length

[33] Previously published data show that the hyporheic zone of the Umatilla River extended throughout the underlying alluvial aquifer [*Jones et al.*, 2007] and, thus, that all water emerging to the floodplain surface from the alluvial aquifer was hyporheic (see section 2.1). Our current results contain two lines of evidence showing that most hyporheic discharge locations in our study reaches are fed by relative short (generally, less than ~100 m) hyporheic flow paths. First, potentiometric surfaces on our study sites were consistent with channel water entering the hyporheic zone at the upstream end of gravel bars, passing through the bars, and reemerging at the downstream end of the bars (Figure 7). Hyporheic flow directions were always directly toward our hyporheic discharge data loggers. Second, the diel temperature cycles at most hyporheic discharge locations

Table 2. Summary of 24-h Mean Temperature (*M*), Diel Temperature Range (*R*), and Diel Temperature Phase (*P*) Derived by Fitting the Typical Diel Cycle for Each Logger to Equation 1 (Main Text)[a]

Channel Type	n	*M*	R	P
		Minthorn		
Channel water				
Main	6	23 (22.9–23)	5.8 (5.2–6.3)	17.8 (17.6–18.2)
Side	8	23.2 (23–23.6)	5.1 (1.7–6.5)	18.1 (17–22.7)
Spring	6	22.7 (22.2–23.1)	1.7 (0.6–3.4)	19.3 (17–22.5)
Hyporheic discharge				
Main	3	22.6 (22.4–22.7)	3.5 (0.9–5.4)	14.3 (2.6–21.5)
Side	4	22.8 (22.6–22.9)	3.8 (1.9–5.3)	15.7 (2.7–23.4)
Spring	8	22.6 (22.1–22.9)	2.7 (0.8–6.2)	14.4 (0.3–23.4)
		Iskuulpa		
Channel water				
Main	4	20 (19.9–20.2)	5.5 (4.9–5.8)	17.1 (16.4–17.3)
Side	1	19.8	5.6	17.4
Spring	3	20.1 (19.8–20.2)	1.9 (0.7–3.8)	16.8 (14.8–19.7)
Hyporheic discharge				
Main	3	20.1 (19.6–20.5)	2 (0.4–3.5)	12.8 (5.9–20.1)
Side	3	19.8 (19.6–19.9)	5.7 (5–6.3)	18.4 (17.8–19.3)
Spring	8	19.7 (15.5–21.5)	1.8 (0.3–4.7)	19.9 (17.7–22.1)

[a]Values listed are the mean and range of values among deployed loggers.

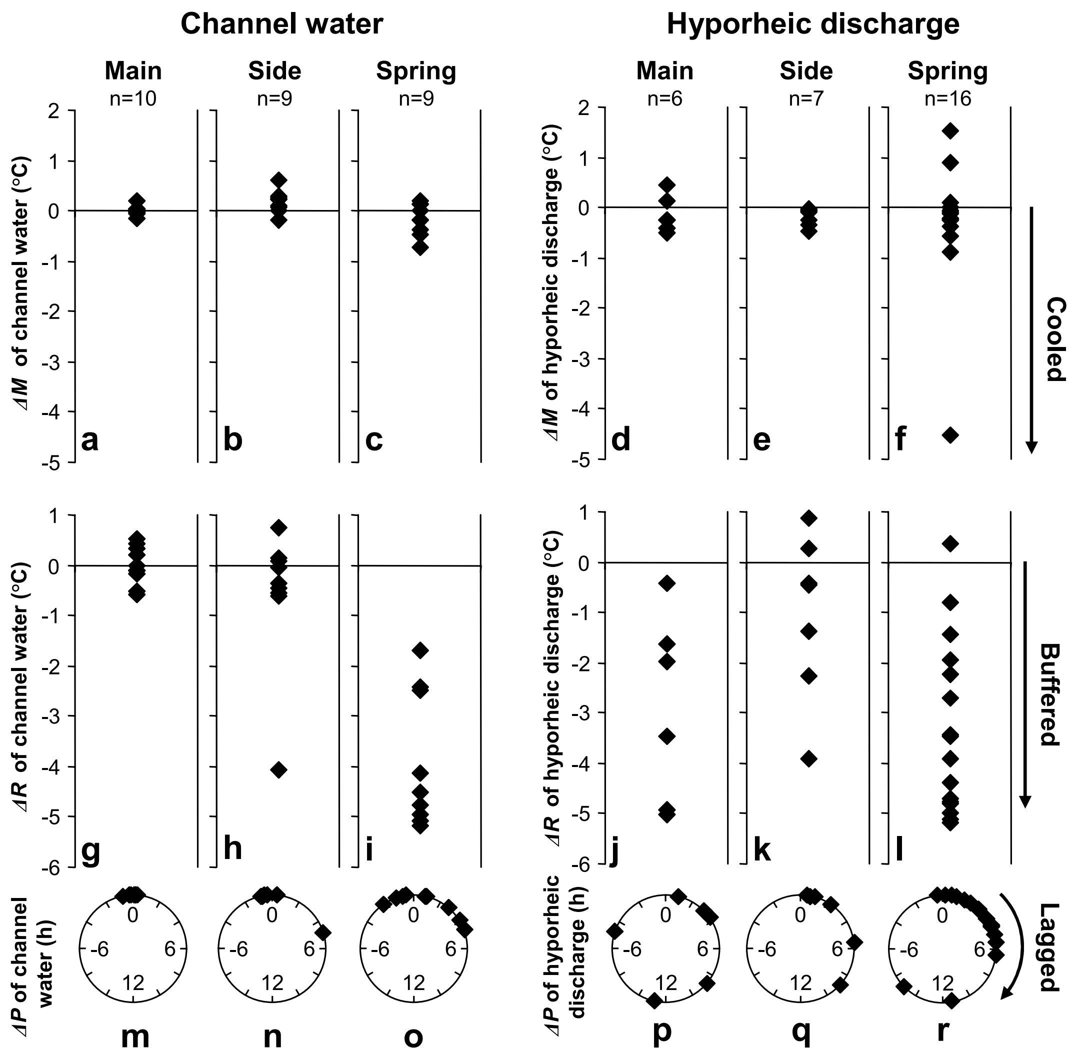

Figure 9. Components of the diel temperature cycle recorded by each data logger, plotted relative to components of the typical main channel diel cycle and organized by water type (channel water versus hyporheic discharge) and channel type (main, side, and spring). Negative values of ΔM represent a reduction in the mean of the diel temperature cycle relative to that typical of the main channel (cooler diel cycles). Negative values of ΔR represent a reduction in diel temperature range relative to that typical of the main channel (buffered diel cycles). Positive values of ΔP represent a delay in the timing of maximum diel water temperature relative to that typical of the main channel (lagged diel cycles).

were consistent with temperature cycles associated with flow paths of less than ~100 m in the Umatilla River aquifer (Figure 5). Specifically, the mean diel cycle temperature of all but three hyporheic discharge locations was within ~1°C of that in main channel water (Figures 9d–9f), and most hyporheic discharge locations maintained diel temperature variation, although some hyporheic discharge sites were more buffered (Figures 9j–9l) and lagged (Figures 9p–9r) than others. Only one hyporheic discharge location exhibited a diel temperature cycle consistent with a long (greater than ~100 m) hyporheic flow path (the extreme low point in Figure 9f). These results are consistent with modeling [*Cardenas et al.*, 2004; *Poole et al.*, 2008] and solute tracer studies [*Haggerty et al.*, 2002; *Gooseff et al.*, 2003, 2007] that show an inverse power law relationship between hyporheic flow path frequency and length in streams, and suggest that most hyporheic exchange in streams arises from short hyporheic flow paths [see also *Kasahara and Wondzell*, 2003].

5.2. Cooled? or Buffered and Lagged?

[34] Our data support the hypothesis that spatial and temporal variation in the temperature of hyporheic discharge creates the observed thermal variation across channel unit types. Temperature cycles in spring channels drive much of the spatial variation in channel water temperature across each site (Figure 6b) and the diel cycle in spring channel water is similar to that of hyporheic discharge (Figures 9c, 9i, and 9o versus all "hyporheic discharge" locations in Figure 9). As is the case with most hyporheic discharge, ΔM is generally small in spring channels (i.e., spring channels have the same 24-h mean temperature as main channel water), while ΔR and ΔP are highly variable. Thus we conclude that much of the substantial variation in channel water temperature observed within stream reaches

of the Umatilla River (Figure 6b) is due to buffered and lagged temperature cycles in hyporheic discharge from short hyporheic flow paths and is not driven by a reduction in the 24-h mean temperature of hyporheic water.

[35] This finding has several important implications for understanding and describing the influences of hyporheic exchange on channel water temperature cycles. First, in order to avoid mischaracterizing the effects of hyporheic exchange on channel water temperature cycles, researchers and managers should adopt and carefully apply language describing associated changes to a river channel's thermal regime. For instance, speaking only in terms of cooling or warming effects implies that hyporheic exchange always exhibits a unidirectional affect on channel water temperature. Second, instantaneous measurement of temperature in streams can yield misinterpretation of dynamics that drive observed temperature patterns. For instance, had we measured temperature of hyporheic discharge only between late morning and early evening, we would have concluded that hyporheic water from short hyporheic flow paths was almost always cooler than channel water and that the cooling effect carried into spring channels along the river [sensu *Fernald et al.*, 2006]. However, continuous recording of hyporheic discharge temperature cycles show that the diel cycles in spring channel water are driven by the buffered and lagged cycles of hyporheic discharge; the daytime-cool condition of hyporheic discharge and spring channel water is transient, and countered 12 h later by a nighttime-warm state (Figure 6).

[36] While advection and dispersion of river-borne heat along hyporheic flow paths are important explanations for the lagged and buffered temperature cycles we observed in hyporheic discharge, we must also underscore the diversity and complexity of heat transfer processes that affect hyporheic discharge. For instance, two data loggers associated with hyporheic discharge recorded increases in daily mean temperature relative to the main channel (highest points in Figure 9f), a dynamic we did not expect along hyporheic flow paths during the summer. Field observations suggested that the water table along these flow paths was at or immediately below the uppermost layer of exposed gravels and cobbles on the floodplain surface. We surmise that these especially shallow flow paths were either: (1) affected by solar heating of the floodplain surface sediments along hyporheic flow paths; or (2) fed by channel water that was heated by solar radiation as it passed through nearly stagnant, shallow habitats at channel margins, prior to entering the hyporheic zone.

[37] Our data show that buffered and lagged diel temperature cycles associated with short hyporheic flow paths can create spatial variation in channel water temperature among channel units, and that this variation is apt to be dynamic over a 24 h cycle (Figure 6b). While this effect on channel water is fundamentally different from a change in the mean temperature of diel cycles in channel water, the observed temperature patterns can still increase the availability of cooler habitat to biota throughout the day. Because diel temperature cycles in spring channels are buffered and lagged relative to the main channel, spring channels provide habitat with lower daily maximum temperatures, and experience the coolest part of their daily cycle at different times of the day, relative to the main channel [*Ebersole et al.*, 2003]. Thus in the presence of such a dynamic mosaic of thermal refugia, mobile biota such as fish can move among habitats during the day, selecting those with the most desirable temperature [*Berman and Quinn*, 1991]. However, for such refugia to be effective in sustaining temperature-sensitive species, the magnitude, size, number, and interconnectivity of refugia must be sufficient to support viable populations during times of heat stress [*Ebersole et al.*, 2001].

5.3. Indirect Riparian Controls on River Temperature

[38] Floodplain vegetation is the primary source of large wood (e.g., tree boles and root wads) to rivers. In alluvial rivers, large wood provides roughness elements that deflect flows, scour depressions, create gravel bars, facilitate channel braiding, [*Abbe and Montgomery*, 1996; *Gurnell et al.*, 2002], and encourage channel avulsions that lead to the formation of anabranched channel patterns [*Nanson and Knighton*, 1996]. Associated complex bed topography and channel patterns enhance hydraulic gradients within the hyporheic zone and substantially increase hyporheic exchange [*Dent et al.*, 2001; *Cardenas et al.*, 2004; *Lautz et al.*, 2006; *Wondzell*, 2006] while creating features such as spring channels that express resulting surface water variation in diel temperature cycles. Therefore despite the fact that the low-flow main channel of the Umatilla River is largely unshaded (Figure 3), riparian vegetation is apt to influence temperature cycles in the main stem Umatilla River by creating geomorphic features that enhance hyporheic exchange and support dynamic temperature mosaics within stream reaches (Figure 6b).

5.4. Site- and Scale-Dependence of Study Results

[39] Our work has three important limitations directly related to site characteristics and the fine spatial scale of our observations (i.e., temperature diversity within stream reaches) relative to hyporheic flow path lengths in the river. First, site conditions resulted in rapid transport of hyporheic water. The coarse grained gravel- and cobble-dominated floodplain aquifer combined with steep hydraulic gradients across gravel bars allowed shallow hyporheic water to move at velocities on the order of ones to 100s of m d^{-1} (W. Woessner and B. Boer, unpublished data, 2003). Under different site conditions with lower hydraulic conductivities (and therefore lower subsurface water velocities), diel temperature patterns may attenuate over much shorter distances along subsurface flow paths, and conduction of heat between the channel and alluvial aquifer may be far more important relative to advective heat transfer. [*Cardenas and Wilson*, 2007].

[40] Second, our results are compiled only from examining the thermal regime of small (<100 m in length) spring channels that occur within the scour zone of the bankfull main channel. The hyporheic flow distance from low-flow main channel to each of the spring channels is short (~10–50 m). In contrast, larger (greater than ~500 m in length) spring channels can be formed by channel avulsions, which cause sections of the main channel to be abandoned as the river's flow suddenly moves to a new location on the floodplain. These large spring channels exist outside the scour zone of the active bankfull channel, and therefore are farther away from the low-flow main channel than the spring channels we studied. Such a channel ("Minthorn Spring

Channel, " which is 1.7 km in length) exists on the Minthorn study site [see *Jones et al.*, 2007] and is fed by long (often >500 m) hyporheic flow paths [*Poole et al.*, 2008]. Consistent with the expected temperature cycles of the long hyporheic flow paths that feed it, both the diel and annual temperature cycles in Minthorn Spring Channel are buffered relative to the main channel (G. Poole, unpublished data, 2003). Thus results shown in Figure 9 may be applicable only to surface habitats near the main channel and fed by relative short hyporheic flow paths. Experiments in surface water habitats farther from the main channel may yield different results, because hyporheic discharge from longer hyporheic flow paths can impart buffered and lagged annual temperature cycles to hyporheic discharge (Figure 5). Such buffered annual temperature cycles have the potential to lower the mean of summertime diel cycles and raise the mean of wintertime diel cycles in channels that receive hyporheic discharge from long flow paths.

[41] Third, an uncritical evaluation of our results might suggest that: (1) only short hyporheic flow paths affect channel water temperature; and (2) that lateral channel habitats are affected, but not the main channel (Figure 9). However, our data do not support such conclusions because our observations were not made at sufficiently coarse spatial scales. Effects of hyporheic exchange (including effects of long flow paths) on main channel temperature cycles would be cumulative as water moves downstream. Addressing the cumulative influences of hyporheic exchange would require comparisons of temperature cycles among long (several kilometers or more) stream segments with varying magnitudes of hyporheic exchange.

5.5. Future Research Needs and Ecological Significance

[42] Mechanistic hydrologic simulations would be invaluable for assessing the local and cumulative effects of hyporheic discharge on whole-river temperature cycles. However, because of the interdependency of temperature cycles in channel water and hyporheic water, a model would need to treat the river channel and hyporheic zone as a single, integrated hydrologic, and thermal system. To fully address the problem, a model would simulate: (1) dynamic, floodplain-wide patterns of surface water distribution, ground/surface water exchanges, and subsurface water movement as a function of channel geomorphology and flow regime under transient conditions [*Poole et al.*, 2006]; (2) heat transport within the alluvial aquifer and channel as a function of advection and conduction; (3) heat exchange between the atmosphere and channel surface; (4) heat exchange between the alluvial aquifer and atmosphere as mediated by the vadose zone and floodplain vegetation; (5) heat exchange between the alluvial aquifer and underlying geologic features that confine and bound the aquifer; and (6) a sufficient channel length to encompass the expected distribution of hydrologic spiraling lengths (i.e., the downstream distance a water molecule travels to complete a cycle of hyporheic recharge, flux through the hyporheic zone, hyporheic discharge, and flow in the channel to a new point of recharge [*Poole et al.*, 2008]) in the modeled system.

[43] Our conclusion that hyporheic temperature cycles are seldom cooler than channel water cycles, but are instead buffered and lagged, should not be misinterpreted to suggest that hyporheic influences on channel water temperatures are somehow ecologically insignificant. On the contrary, hyporheic exchange enhances temperature diversity in surface and subsurface habitats [*Malard et al.*, 2001; *Johnson*, 2004; *Schmidt et al.*, 2006], moderates both diel and annual temperature cycles, and therefore induces responses from biota including fish [*Baxter and Hauer*, 2000; *Geist et al.*, 2002; *Hanrahan*, 2007] and aquatic macroinvertebrate communities [*Olsen and Townsend*, 2003; *Brown et al.*, 2005]. We highlight the distinction between cooler cycles and buffered and lagged cycles simply to promote a more critical evaluation of the underlying dynamics driving hyporheic influences on channel water temperatures. Such knowledge should improve our ability to predict how the thermal regime of whole river systems will respond to changes in channel morphology and associated changes in the magnitude of hyporheic exchange.

6. Summary

[44] Our results show the mean of channel water temperature cycles is not altered by water discharged from short (<50–100 m) hyporheic flow paths in the Umatilla River. However, the phases of channel water diel cycles can be altered, the daily maxima reduced, and minima increased (Figure 9). These effects are most pronounced in spring channel habitats, although the potential cumulative effects on main and side channels remain uninvestigated. These lagged and buffered hyporheic temperature cycles create dynamic reach-scale mosaics of channel water temperatures observed across channel habitats in the scour zone (Figure 6b).

[45] We identified the role of buffered and lagged cycles only because we monitored and analyzed the 24-h temperature cycles in channel water and hyporheic discharge, and the annual temperature cycle along longer flow paths through this highly conductive aquifer. Had we taken only point measurements during the day or analyzed only daily maxima, we would have erroneously concluded that water is simply cooled as it flows along hyporheic flow paths.

[46] In contrast to short flow paths, hyporheic water following longer hyporheic flow paths (>100–500 m) does emerge cooler than the mean summertime diel temperature cycle in channel water (Figure 5a), but these relatively cool temperatures are transient across seasons. In winter, hyporheic discharge from long flow paths is warmer than the mean diel cycle in main channel water (Figure 5b). Ultimately, however, any consistent summer-cool and winter-warm patterns in hyporheic discharge are caused simply by buffered and lagged annual temperature cycles.

[47] Our data suggest that the influence of hyporheic exchange on Umatilla River channel water temperature cycles may be best described in terms of temporary heat storage within the aquifer. Obviously, in the summertime, hyporheic exchange has the potential to remove a substantial amount of heat from a river channel and store that heat in the alluvial aquifer. Yet much of that stored heat appears to be advected back to the river channel along with the reemerging hyporheic water. Along short hyporheic flow paths, the duration of heat storage may be just minutes to days, while the duration of heat storage along long hyporheic flow paths may be weeks or months. The duration and magnitude of hyporheic heat storage, then, determine the extent to which diel and annual temperature cycles are

buffered and lagged along hyporheic flow paths, and ultimately control the magnitude and patterns of hyporheic influence on channel water temperature.

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