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DOI: 10.1016/j.advwatres.2017.07.005

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Document Version Peer reviewed version

Citation for published version (Harvard):

Rau, GC, Halloran, LJS, Cuthbert, MO, Andersen, MS, Acworth, RI & Tellam, JH 2017, 'Characterising the dynamics of surface water-groundwater interactions in intermittent and ephemeral streams using streambed thermal signatures', *Advances in Water Resources*, vol. 107, pp. 354-369. https://doi.org/10.1016/j.advwatres.2017.07.005

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PII: S0309-1708(17)30089-1 DOI: 10.1016/j.advwatres.2017.07.005 ADWR 2890 Reference:

article

as:

To appear in: Advances in Water Resources

Received date: 27 January 2017 Revised date: 30 May 2017 Accepted date: 7 July 2017

this

Please

cite



Mark O. Cuthbert, Martin S. Andersen, R.Ian Acworth, John H. Tellam, Characterising the dynamics of surface watergroundwater interactions in intermittent and ephemeral streams using streambed thermal signatures, Advances in Water Resources (2017), doi: 10.1016/j.advwatres.2017.07.005

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Characterising the dynamics of surface water-groundwater interactions in intermittent and ephemeral streams using streambed thermal signatures

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Abstract

Ephemeral and intermittent flow in dryland stream channels infiltrates into sediments, replenishes groundwater resources and underpins riparian ecosystems. However, the spatiotemporal complexity of the transitory flow processes that occur beneath such stream channels are poorly observed and understood. We develop a new approach to characterise the dynamics of surface water-groundwater interactions in dryland streams using a pair of temperature records measured at different depths within the streambed. The approach exploits the fact that the downward propagation of the diel temperature fluctuation from the surface depends on the sediment thermal diffusivity. This is controlled by time-varying fractions of air and water contained in streambed sediments causing a contrast in thermal properties. We demonstrate the usefulness of this method with multi-level temperature and pressure records of a flow event acquired using 12 streambed arrays deployed along a ~ 12 km dryland channel section. Thermal signatures clearly indicate the presence of water and characterise the vertical flow component as

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Preprint submitted to Advances in Water Resources

8 juillet 2017

well as the occurrence of horizontal hyporheic flow. We jointly interpret thermal signatures as well as surface and groundwater levels to distinguish four different hydrological regimes : [A] dry channel, [B] surface run-off, [C] poolriffle sequence, [D] isolated pools. The occurrence and duration of the regimes depends on the rate at which the infiltrated water redistributes in the subsurface which, in turn, is controlled by the hydraulic properties of the variably saturated sediment. Our results have significant implications for understanding how transitory flows recharge alluvial sediments, influence water quality and underpin dryland ecosystems.

Keywords: surface water-groundwater interactions; ephemeral and intermittent streams; heat as a tracer; hydrological characterisation; streambed thermal regimes

1 Highlights

- ² Amplitude ratios of the daily temperature component at two different
- depths in the streambed can be used to distinguish dry from saturated
 sediment
- Multi-level streambed temperature records reveal distinct thermal si gnatures that characterize water flow
- Ephemeral or intermittent surface water-groundwater interactions can
 be categorized into a sequence of hydrological regimes

9 1. Introduction

The spatial and temporal movement of water through dry stream chan-10 nels and the surrounding shallow sediments is highly dynamic. Stream flow 11 cessation and drying occur in more than half of the world's river networks 1 12 with proportions exceeding 80% in dryland regions [2]. Water in otherwise dry 13 channels recharges groundwater through infiltration [e.g., 3, 4, 5, 6, 7] and 14 underpins dryland ecological diversity [e.g., 8, 2]. In fact, shallow groundwa-15 ter is often the only source of freshwater for human and ecosystem activity 16 during periods of dry climate and therefore of critical importance [9, 10, 11]. 17 As groundwater resources are being depleted globally [12], the largest wa-18 ter stresses exist in areas with high population and low surface water availa-19 bility [13] and are intensified by human activity [14]. Because groundwater re-20 charge in dryland regions is predominantly due to infiltration of water during 21 flow events (i.e., 'focused' or 'indirect') [e.g., 9, 5], understanding temporary 22 surface-groundwater interactions is of paramount importance [6, 7]. However, 23 monitoring temporary flow events is challenging and thus observations are 24 scarce [15, 16]. 25

The presence of water in otherwise dry channels is generally referred to as 'ephemeral' or 'intermittent' behaviour depending on the duration of flow [e.g., 17]. When such streams are flowing, the degree of interaction between the surface and groundwater systems depends on complex hydrogeologic controls [18, 19, 20]. The spatiotemporal dynamics of such surface watergroundwater interactions in these contexts are currently poorly understood [7].

It is recognised that streambed temperature data provides useful insight 33 into the flow dynamics of dryland systems especially when complementing 34 pressure data. Daily stream temperature oscillations can cause variations in 35 stream discharge which relate to infiltration caused by the change in water 36 physical properties [3, 21]. Constantz and Thomas [15, 22] found that stream-37 bed temperature can be used as an indicator of streamflow and can provide 38 subsurface water percolation characteristics. Constantz et al. [16] and Blasch 39 et al. 23 determined streamflow frequency and duration using streambed 40 temperature records. Constantz et al. [24] numerically modelled subsurface 41 temperature records and concluded that percolation rates could be constrai-42 ned. While much of this work, summarised in Blasch et al. [25], illustrates 43 the temporal dynamics of transient surface-groundwater interactions, inter-44 pretation is limited by data from discrete spatial locations.

Here, we draw from the large body of heat tracing knowledge developed 46 for surface-groundwater interactions in perennial (saturated) systems [e.g., 47 refer to the reviews of 26, 27, 28] and extend the methodologies to include 48 consideration of dry systems. We exploit the fact that the presence of water in 49 otherwise dry sediments changes the thermal properties [e.g., 15, 29, 30, 31]. 50 In reality, sediments can be variably saturated, i.e. during the wetting 51 and drying stages of a flow event. In fact, streambed sediments may never 52 be entirely dry or fully saturated. However, we limit our analysis to realistic 53 end-members of dry and water saturated conditions as the resulting thermal 54 contrast is large enough to allow reliable detection of water. This simplifica-55 tion also avoids overly complicated saturation measurements and equations 56 that are necessary when coupling the non-linear processes involved in va-57 riably saturated conditions. For details about heat tracing to infer variably 58 saturated processes or properties we refer the interested reader to Halloran 59 et al. [30, 31]. 60

In this paper we demonstrate that (1) streambed temperature data can 61 be interpreted to distinguish reliably between approximately dry and satura-62 ted conditions below dryland streams, thus allowing identifications of stream 63 flow episodes; (2) temperature records, interpreted using this approach, can 64 be used to distinguish between dominantly upward, downward, and horizon-65 tal flow below dryland streams; (3) the qualitative results can be used to 66 constrain conceptual models of temporary surface-groundwater interactions. 67 Our results have significant implications for improving the evaluation of fo-68 cused or indirect groundwater recharge and can underpin further research on 69 water quality and ecohydrology in dryland streams. 70

71 2. Theoretical background

72 2.1. Propagation of diel temperature fluctuations into shallow sediments

The analysis of heat tracing data utilizes the diel temperature fluctuations that ubiquitously occur at the Earth's surface and propagate vertically downwards into the subsurface where the thermal wave is both damped and delayed over depth [32, 33]. For a 1D vertical section of water saturated (wet) near-surface sediment exposed to sinusoidal temperature forcing at the ⁷⁸ surface, the temperature over depth and time can be described as [33, 34]

$$T^{sat}(z) = T_0 + A \cdot exp\left[\frac{z}{2D}\left(v - \sqrt{\frac{\alpha + v^2}{2}}\right)\right] \cdot \cos\left[\frac{2\pi t}{P} - \frac{z}{2D}\sqrt{\frac{\alpha - v^2}{2}}\right],\tag{1}$$

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where T_0 is the ambient temperature [°C], A is the diel temperature amplitude [°C], z is vertical depth [m] (positive = down), t is time [s], P is the period of the sine wave [s], v is the thermal front velocity linearly related to Darcy flux q. The parameter α is defined as

$$\alpha = \sqrt{v^4 + \left(\frac{8\pi D}{P}\right)^2} \tag{2}$$

and the sediment bulk thermal diffusivity is [35, 26]

$$D = \frac{\kappa}{\rho c} \tag{3}$$

where κ is the thermal conductivity $[Wm^{-1}K^{-1}]$, ρ is the density $[kgm^{-3}]$ and c is the specific heat capacity $[Jkg^{-1}K^{-1}]$ of the sediments; ρc is the thermal capacity $[Jm^{-3}K^{-1}]$ [36]. The thermal parameters depend on the sediment moisture conditions (dry or saturated) and are discussed in Section 2.2. In this investigation we neglect thermal dispersivity as is justified for water fluxes v < 10 m/d [37].

Heat tracing is best conducted using a pair of temperature sensors that are arranged vertically. The advantage is that the sensor spacing, rather than absolute depth, can be targeted or precisely measured. In this case an amplitude ratio can be defined for water saturated streambeds [38]

$$A_r^{sat}\left(\Delta z, D^{sat}, v\right) = \frac{A_2(z_2)}{A_1(z_1)} = exp\left[\frac{\Delta z}{2D^{sat}}\left(v - \sqrt{\frac{\alpha + v^2}{2}}\right)\right]$$
(4)

where A_1 and A_2 are the amplitude of diel temperature fluctuations measured at discrete depths in the sediment $(|z_2| > |z_1|)$.

Analytical heat tracing has been widely used to calculate vertical water fluxes under water saturated conditions [e.g. 27, 28]. We note that in the case of uniform directional flow and in the absence of hydrodynamic thermal dispersion, this approach delivers the vertical flow component of the total flow vector [39].

¹⁰⁵ 2.2. Heat tracing to distinguish between dry and water saturated sediments

Streambed sediments can undergo variably water saturated conditions 106 depending on whether the channel is dry or wet, i.e. the presence of air in 107 the sediments [40]. Consequently, the corresponding difference in thermal 108 parameters must be considered. The bulk thermal diffusivity in Equation 3 109 has a non-linear dependency on saturation [41, 42, 31]. Côté and Konrad 110 [41] presented a generalized thermal conductivity model for variably satura-111 ted sediment which we simplify to its dry and saturated end-members. The 112 thermal conductivity for dry streambeds is [41] 113

$$\kappa^{dry} = \chi \cdot 10^{-\eta n} \tag{5}$$

where χ and η are empirical parameters that depend on the grain size; here, we use $\chi = 1.7$ and $\eta = 1.8$ for rocks and gravels as is most suitable for dryland channels exposed to high energy flows; *n* represents the total porosity [-] of the sediment. In contrast, the saturated thermal conductivity is given as [43, 41, 42]

$$\kappa^{sat} = \kappa_w^n \cdot \kappa_s^{(1-n)} \tag{6}$$

where subscripts w and s represent water and solid matrix, respectively.

The thermal capacity of a sediment with two phases (dry : air and solid matrix, saturated : water and solid matrix) is defined as a porosity weighted volumetric mean [44, 36, 31]

125

$$(c\rho)^{dry} = (1-n)(c\rho)_s \tag{7}$$

126 127

$$(c\rho)^{sat} = n(c\rho)_w + (1-n)(c\rho)_s \tag{8}$$

where subscripts w and s represent water and solid matrix, respectively. The specific heat capacity of air is so small that it can be neglected in our analysis [31].

Thermal diffusivity for water saturated (D^{sat}) and dry (D^{dry}) sediment can be calculated by using Equation 3 in combination with Equations 6 and 8 or Equations 5 and 7, respectively.

¹³⁴ Under the conditions of water saturated streambed sediments, the am-¹³⁵ plitude ratio A_r^{sat} (Equation 4) is a function of the bulk saturated thermal ¹³⁶ diffusivity of the sediment D^{sat} and the thermal front velocity (determined ¹³⁷ by the vertical flow of water), $A_r^{sat}(D^{sat}, v)$. For dry streambed sediments, ¹³⁸ the amplitude ratio will only depend upon the bulk dry sediment thermal ¹³⁹ diffusivity D^{dry} because the absence of water also means that v = 0 (no flow). ¹⁴⁰ Consequently, under dry conditions Equation 4 can be simplified to

$$A_r^{dry}\left(\Delta z, D^{dry}\right) = \frac{A_2(z_2)}{A_1(z_1)} = exp\left[-\Delta z\sqrt{\frac{\pi}{PD^{dry}}}\right]$$

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(9)

This equation can be reformulated to calculate the dry bulk sediment thermal diffusivity D^{dry} from the ratio of the diel temperature amplitudes measured using two sensors located at different depths during a period when the streambed is dry.

In reality, streambed thermal properties and porosity can vary within na-146 tural limits. Significant effort towards additional field measurements would 147 be required to constrain these parameters, as the phase shift of the thermal 148 wave cannot be used to separate the sediment thermal conductivity or spe-149 cific heat capacity from thermal diffusivity. Note also that calculation of the 150 saturated streambed thermal diffusivity is hindered by the degree of freedom 151 introduced through a variable vertical water flux and is therefore impossible 152 to accomplish without independent flow measurements. 153

To determine whether there is always a difference in amplitude ratio for 154 dry and saturated sediments, given the range of natural parameter variabi-155 lity, we evaluated $\Delta A_r^{dry,sat} = A_r^{sat} - A_r^{dry}$ as a function of the respective 156 thermal diffusivity values. Note that for a given location in space, the ther-157 mal properties of the solid matrix, as well as the porosity, remain constant 158 during any change from dry to saturated. While the thermal property values 159 for water are accurately defined (Table 1), the three unknown properties are : 160 The streambed porosity n (which we allow to vary between 0.2 and 0.5), solid 161 thermal conductivity κ_s (low porosity volcanic rocks [46]), and solid thermal 162 capacities $(c\rho)_s$ (rock forming minerals [36]). 163

Figure 1a shows the resulting $\Delta A_r^{dry,sat}$ as multi-parameter space at dis-164 crete values of porosity over the range of thermal parameters. This illustrates 165 that the diel temperature amplitude is significantly different for a realistic 166 range of dry and water saturated streambed sediments, $A_r^{dry} < A_r^{sat}$. This is 167 because during a flow event the streambed pore space, initially occupied by 168 air, will be replaced with water with significantly different thermal proper-169 ties. A change in A_r can, therefore, be used to distinguish between realistic 170 end-members of water saturation (dry vs. saturated), and therefore acts as 171 an easily measurable proxy for streambed flow processes. 172

Parameter/Phase		Unit		Para	umeter rai	nge		References
			P_{10}	-2σ	π	$+2\sigma$	P_{90}	
Porosity	Total pore space n	I		0.2	0.35	0.5		
	Thermal conductivity κ_w	$Wm^{-1}K^{-1}$			0.6			в
Water	Specific heat capacity c_w	$Jkg^{-1}K^{-1}$			4185			ъ
	Density ρ_w	kgm^{-3}			998			а
-	Thermal conductivity κ_s	$Wm^{-1}K^{-1}$		1.62	3.08	4.54		р
Solid matrix	Thermal capacity $(\rho c)_s$	$MJm^{-3}K^{-1}$		1.8	2.45	3.1		С
	Density ρ_s	kgm *			2650			
Thermal diffusivity	Dry streambed Saturated streambed	$m^2 s^{-1} m^2 s^{-1} m^2$	$1.79 \cdot 10^{-7}$ $1.04 \cdot 10^{-7}$		$2.61 \cdot 10^{-7}$ $5.73 \cdot 10^{-7}$		$3.57 \cdot 10^{-7}$ $7.67 \cdot 10^{-7}$	
TABLE 1: Thermal para ration as a function of s	meters used for the <i>Monte-Carlo</i> treambed thermal diffusivity. Re	analysis to assess ferences : a) NIST	the differenc	e betwee: user [46]	n dry and sa c) Wanles	turated a	amplitude Jes [36]	
A TO TROTACTION AN AM TECTAM	A CONTINUE IN THE PARTY AND ALL AND AL	- ~ · · · · · · · · · · · · · · · · ·		[~+] +Ann	mandmin (a.	4~11 mm	.[^_] .	

ermal parameters used for the Monte-Carlo analysis to assess the difference between dry and saturated amplitude	inction of streambed thermal diffusivity. References : a) NIST [45]. b) Clauser [46]. c) Waples and Waples [36].			
BLE 1: Thermal para	ion as a function of			

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(a) dry streambed

(b) saturated streambed

FIGURE 1: Conceptual model illustrating how to characterize the dynamics of ephemeral surface-groundwater interactions in shallow variably saturated sediments using the diel temperature amplitude ratio (A_r) as a signature : a) The likely range of the diel temperature amplitude ratio for dry and saturated streambeds (resulting from a range of porosity and thermal parameters) is shown for an example sensor spacing $\Delta z = 0.2$ m and thermal front velocities of $v = \pm 1$ m/d. b) The thermal diffusivity of wet streambed sediments is different leading to a change in amplitude ratio during flow. Further, changes in amplitude ratio can indicate the vertical direction of water fluxes in the sediments between the temperature sensors. This can be used to characterise ephemeral surface-groundwater interactions during flow events. c) The difference between dry and saturated (v = 0) amplitude ratio ΔA_r as a function of a range in solid thermal conductivity κ_s and solid thermal capacity (ρc)_s at discrete porosity values. Numbered labels 1-5 are explained in the text.

2.3. Shallow streambed thermal signatures detect water and characterize flow through variably saturated streambed sediments

To estimate the saturated streambed thermal diffusivity $\Delta A_r^{dry,sat}$ can be 175 used. We performed a *Monte-Carlo* analysis (100,000 samples) to establish 176 the most likely values for dry and saturated amplitude ratio as a function 177 of streambed thermal diffusivity. We use the literature derived ranges shown 178 in Table 1 as input assuming that all properties follow a normal distribution 179 and that 95.4% of the existing values fall within these limits (i.e., $\mu \pm 2\sigma$). 180 The resulting mean and percentile $(P_{10} \text{ and } P_{90})$ values for dry and saturated 181 streambed thermal diffusivity are listed in Table 1. These values were used 182 to plot the amplitude-depth relationships in Figure 1b and 1c and visualise 183 the difference between dry and saturated A_r . 184

Figure 1 demonstrates that the A_r can be divided into the following categories (see corresponding labels in Figure 1c):

187 (1) $0 < A_r(t) < A_r^{dry}$: Water saturated sediment and a vertical upward 188 flow component.

(2) $A_r(t) = A_r^{dry}$: Dry end-member of the streambed sediments which can be established from temperature records acquired during dry periods.

(3) $A_r^{dry} < A_r(t) \le A_r^{sat}$: A small range of ambiguity where the exact 191 conditions are unclear, i.e. variable water saturation or fully saturated 192 with a flow component ranging between vertical upward and zero. Here, 193 Monte-Carlo analysis offers a measure of the uncertainty to compare 194 with the difference between A_r^{dry} and A_r^{sat} (0.02 < ΔA_r < 0.175, Figure 195 1a). We note that interpretations can still be made when temperature 196 data are acquired in conjunction with pressure, as values are indicative 197 of the presence of water above the point of measurement. 198

(4) $A_r^{sat} < A_r(t) \le 1$: Water saturated sediment and larger values for an increasing vertical downward flow component.

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(5) $A_r(t) > 1$: Water-saturated sediment and conditions that violate the 1D vertical flow assumption inherent to Equation 1. This has been observed previously [47] and can, in the absence of a daily fluctuating subsurface heat source, only be caused by horizontal hyporheic flow.

To simplify the approach we only consider the end-members of saturation, close to dry and water saturated. In reality, there could be variable saturation in the streambed sediments, particularly during the onset of flow and drying of the channel. During times of variable water saturation, the amplitude ratio will be between A_r^{dry} and A_r^{sat} .

Figure 1 clearly illustrates that under realistic conditions, the saturated amplitude ratio A_r^{sat} (Equation 4) should always be larger than the dry amplitude ratio A_r^{dry} (Equation 9), i.e. $\Delta A_r > 0$. The diel amplitude ratio A_r , therefore, allows detection of the moisture state, i.e. dry or saturated, as well as characterization of vertical water movement through sediments when the system is near the saturated end-member.

In this method we abstain from quantifying infiltration rates because 216 this would require knowledge of the streambed moisture content during flow 217 events as well as the associated thermal diffusivity. In our approach, the zone 218 of A_r ambiguity due to variable moisture content occupies values representa-219 tive of saturated conditions and upward water flow. Given that streams with 220 temporary flow are generally hydraulically disconnected from the ground-221 water table [e.g. 48, 6], water will most likely percolate downwards at least 222 as long as a variably saturated zone remains. Under these conditions, $A_r(t)$ 223 should serve as a novel indicator revealing the streambed processes during 224 ephemeral or intermittent flow. 225

226 2.4. Extraction of the diel amplitudes from temperature measurements

Equation 1 requires that the temperature forcing is a sinusoidal wave. This is not a realistic assumption under real-world conditions. However, we can capitalise on the fact that any signal can be decomposed into a finite sum of sinusoidal components using the *Discrete Fourier Transform*. This is necessary so that the resulting signal component complies with the conditions inherent to Equation 1, and that the amplitude of a single frequency component (e.g., daily) can be used directly with A_r in Equations 4 and 9.

To calculate diel temperature amplitudes a *Fast Fourier Transform* (FFT), as implemented in *Python*, can be applied to subsets of the data which span a multiple number of days. The FFT of a signal is defined as

$$\hat{s}(f_k) = \mathcal{F}\{s(t_n)\} = \sum_{n=0}^{N-1} s(t_n) e^{-2\pi i k n/N}$$
(10)

where k and n denote the indices of discretely sampled frequency and time, respectively, which range from 0 to N-1. It is not important to normalize the transform as long as data treatment is consistent and ratios of the amplitudes

237

²⁴¹ are used. The discrete frequencies of the transformed signal are

$$f_k = k f_s / N. \tag{11}$$

(12)

²⁴³ For a window of *i*-multiple days, the absolute value of the *i*-th entry f_i

244

242

$$A(f_i) = |\hat{s}(f_i)| = \sqrt{\mathcal{R}^2(f_i) + \mathcal{I}^2(f_i)}$$

corresponds to the amplitude of the f = 1 cpd (cycles per day) frequency component [30]. This procedure is repeated as a rolling window along the time series whereby $A(f_i)$ is allocated to the time at the center of the window.

Using this approach, a temperature amplitude time series can be extrac-248 ted and used to calculate amplitude ratios from Equation 4. Ephemeral flow 249 events can be characterised using the methodology described earlier. It is 250 important to neglect extracted amplitude values that are below the tem-251 perature resolution of commonly available sensors, i.e. $A > 0.01^{\circ}C$ can be 252 considered valid. Theoretically, the component phases could also be extrac-253 ted and used. However, Rau et al. [49] noted that signal non-stationarity, as 254 inherent in natural temperature oscillations, causes erroneous phase results 255 which significantly decreases the accuracy of any phase-derived calculations. 256

Field example from Middle Creek in the Maules Creek Catch ment, New South Wales, Australia

259 3.1. Catchment context

The Maules Creek catchment is located in the semi-arid northwestern area 260 of New South Wales (NSW), Australia (Figure 2). Middle Creek flows into 261 Horsearm Creek, then Maules Creek and further into the Namoi River which 262 is a tributary of the large Murray-Darling Basin (MDB) (Figure 2). The 263 Nandewar range provides the northern and eastern margin of the catchment 264 and consists of Miocene basaltic mountains peaking at 1,506 m (Mt. Kaputar) 265 Australian Height Datum (AHD). The Namoi River at the western part of 266 the catchment is at approx. 230 m AHD. The difference in topography causes 267 a significant orographic rainfall effect resulting in a long-term average rainfall 268 of 928 mm/a in the mountains (Mt. Kaputar at 1450 m AHD) and 561 mm/a 269 on the floodplain (Narrabri Bowling Club at 229 m AHD and only 35 km 270 west of Mt. Kaputar). 271

A major change in geology separates the Carboniferous and Devonian rocks in the upper catchment from the Permian lower catchment. The Carboniferous and Devonian metasediments and intrusives have been thrust over the Permian Mauls Creek coal measures to the west with the thrust zone occurring at the mountain front between T11 and T10 (Figure 2). The high energy flows from the mountains have cut 10 to 15 m deep channels into the coal measures that are now filled with a very heterogeneous assemblage of boulders, sand and gravels that are substantially reworked by each major flood.

This catchment area has been well instrumented for groundwater moni-281 toring since 2009 through the Australian Government National Collabora-282 tive Infrastructure Strategy (NCRIS). A number of research projects were 283 conducted mainly in the lower part of the catchment : Andersen and Ac-284 worth [50] surveyed the perennial surface-groundwater interactions and no-285 ted the complexity of these processes. Rau et al. [47] successfully quantified 286 the rate of saturated vertical flow in the streambed using heat as a tracer. To 287 evaluate the groundwater resources within the catchment, a comprehensive 288 groundwater model was created and illustrated considerable uncertainty and 289 a lack of information about groundwater recharge through the intermittent 290 stream channels originating at the mountain front [51]. Further research on 291 groundwater resources as well as surface water-groundwater interactions can 292 be found in McCallum et al. [52], Kelly et al. [53] and Cuthbert et al. [7]. 293

3.2. Monitoring of rainfall, groundwater and streambed water levels and tem perature

Middle Creek drains an estimated $106 \, km^2$ of the upper catchment and 296 the discharge point of which is located at the confluence with Horsearm Creek 297 (Figure 2). Rainfall was recorded at weather stations using tipping bucket 298 rain gauges (Campbell Scientific Inc., USA) at three different locations (see 299 abbreviations in Figure 2b) : Mt Kaputar National Park (MK, Australian 300 Government Bureau of Meteorology station #54151), Middle Creek Farm 301 (MCF) and Bellevue Farm (BVF). An additional long-term rainfall dataset 302 is available from the Mount Lindsay Station (ML, Australian Government 303 Bureau of Meteorology station #54021) which has been operational since 304 1886 and located ~ 11 km south-east of the Mt. Kaputar station. The Mount 305 Lindsay Station has an elevation of ~ 870 m but lies in a rain shadow of the 306 higher Mt. Kaputar rain gauge. 307

The loggers used to measure streambed temperature and pressure were a combination of off-the-shelf devices : HOBO temp pro v2 (U22-02), Schlumberger Diver and Solinst Levelogger Gold/Edge. The temperature measured by the loggers was calibrated against a reference (Fluke hand-held 1524) in



FIGURE 2: Map showing (a) the location the Maules Creek catchment in relation to the Murray-Darling Basin (MDB), (b) the state of New South Wales, (c) a catchment elevation map with locations of rain gauges, (d) streambed array installations and piezometers along Middle Creek.

a bucket of well-stirred water at different values. The calibration was appliedas a correction to the temperature field records.



FIGURE 3: Streambed array T3 installed in the dry channel as an example representative of the other locations. Inset plot shows the distance-elevation profile for all arrays as surveyed using differential GPS (Table 2).

Temperature and pressure were recorded at discrete depths in the shallow 314 streambed at a total of 12 different locations along Middle Creek. Multi-level 315 streambed arrays were constructed from 32 mm diameter standard hydraulic 316 PVC pipe. Loggers were placed inside the pipe at defined intervals (multi-317 level monitoring), with the pressure measured at the top and bottom end, 318 and separated by spacers [47, 54]. The effect of this array design on the 319 measured diel amplitudes has been found to be negligible [55]. The length of 320 the streambed arrays depended on the number of loggers used at the different 321 locations of deployment. Table 2 contains the details of the streambed arrays. 322 Because the stream flow events can be high energy, installation of the 323 arrays required the construction of an anchor point. At each location, two 324 star pickets were manually driven into the streambed sediments in an x-325 formation and a small pit was dug around the point of contact between the 326 star pickets. The pit was then filled with quick-set concrete and covered with 327

³²⁸ large cobbles. For an example installation please refer to Figure 3.

Short arrays were directly attached to the star pickets with the uppermost sensor located at the same vertical level as the streambed. Longer, multilevel arrays were installed with the same method as described by [47] at ~1 m downstream and securely attached to the anchor point. Streambed arrays were installed at the end of July 2013, and loggers were programmed to record pressure and temperature at 15 min intervals. The aim was to capture an entire flow event along the creek.

Geospatial coordinates of all installation points were accurately surveyed 336 using differential GPS equipment (Trimble R10 GNSS). For a summary of 337 streambed monitoring arrays, measured parameters and locations refer to 338 Table 2. An atmospheric pressure record, obtained from the MCF weather 330 station, was used to calculate gauge pressure and hydraulic heads in combi-340 nation with the survey. The approximate flow distance between the first and 341 last monitoring points was traced in ArcMAP based on an identification of 342 the channel from satellite imagery and is reported in Table 2. 343

Multi-level boreholes were installed right next to the ephemeral stream channel (distance within tens of meters) as described by Cuthbert et al. [7]. To determine the hydraulic connectivity between surface flow and groundwater in the sediments along the channel (BH 11, BH 17, BH 18 and BH 20 in Figure 2d), the shallower screens were monitored at 15 min intervals.

3.3. Spatiotemporal surface and groundwater responses to a major rainfall event

Cumulative rainfall of 329 mm, 198 mm and 228 mm was measured at 351 MK, MCF and BVF, respectively, for the 60-day period from 20 March to 18 352 May 2016 (4a). This rainfall occurred as clustered rain events with short per-353 iods of dry weather. The rainfall triggered mountain run-off and led to stream 354 flow along the channel as recorded by the streambed arrays summarised in 355 Figure 4. The rainfall amount was more than double the average long-term 356 (1886-2012) moving 60-day sum of 155 mm (max. 809 mm in February 1971), 357 indicating that it was a sizeable event for this catchment. 358

Figure 4 summarises the dynamics of water movement along Middle Creek, over depth and in time for this event. Note that the array (streambed surface) elevations almost perfectly follow an exponential curve (inset in Figure 3 based on data in Table 2). The run-off moved along the previously dry channel and was captured by the pressure transducers at the streambed as a hydrograph peak with differing heights. Water levels upstream (array T11)

	$oldsymbol{D}^{dry} \ [m^2/s]$	8.35E-07 1.57E-06 8.16E-07 9.44E-07 8.38E-07 8.38E-07 1.39E-06	7.12E-07 5.79E-07 6.00E-07 8.65E-07 4.35E-07	are
	${ m Stdev} A^{dry}_r \ {\ \ [-]}$	$\begin{array}{c} 0.017\\ 0.020\\ 0.033\\ 0.022\\ 0.025\\ 0.019\\ 0.019\end{array}$	$\begin{array}{c} 0.013\\ 0.027\\ 0.011\\ 0.011\\ 0.018\\ 0.015\\ 0.015\end{array}$	pectively.
	$\operatorname{Mean}_{[-]}^{dry}$	$\begin{array}{c} 0.212\\ 0.435\\ 0.281\\ 0.289\\ 0.286\\ 0.413\\ 0.413\end{array}$	0.290 0.222 0.211 0.211 0.211 0.206	direction. Projec
	$\mathbf{\Delta}_{\mathbf{z}}$	$\begin{array}{c} 0.235\\ 0.173\\ 0.190\\ 0.200\\ 0.190\\ 0.173\\ 0.173\end{array}$	$\begin{array}{c} 0.173\\ 0.190\\ 0.200\\ 0.240\\ 0.173\end{array}$	the flow id tempe
	$\begin{array}{c} \mathbf{Length} \\ [m] \end{array}$	$\begin{array}{c} 0.230\\ 0.173\\ 1.129\\ 0.200\\ 1.158\\ 0.173\\ 0.173\end{array}$	$\begin{array}{c} 0.1200\\ 0.173\\ 1.060\\ 0.240\\ 0.200\\ 0.171\end{array}$	tance along ansducer ar
	Parameters	HHH HH 333H33H 200 00	ррррр -22222 -22222	s in order of dis id for pressure t ₁
	Intervals	одородо	100000	ys and location ns p and T stan
	Distance [m]	$\begin{array}{c} 0\\ 2,464\\ 5,934\\ 5,976\\ 6,976\\ 7,00$	$^{1,,12}_{9,92}$ $^{9,92}_{9,979}$ $^{11,125}_{-1}$ $^{-11,903}$	uitoring arra Abbreviatio
	Elevation [m]	371.59 338.57 328.57 320.69 308.36 300.65 293.61	281.37 281.87 274.00 274.94 271.14	streambed mor as in Figure 2
Y	Array	T11 110 178 178 178 178	$^{13}_{ m T2a}$ $^{ m L3}_{ m T2b}$ $^{ m T3}_{ m T1}$	TABLE 2: the same <i>i</i>

peaked on 28 Mar 2014 at 4 :15. The flood took 135 min to move ~ 11.9 km 365 (Figure 2) to the downstream end (array T1) with an average velocity of 366 ~ 1.5 m/s. Note that array T8 and T5 did not contain pressure transducers. 367 The depth to groundwater (thickness of the unsaturated zone) along the 368 stream channel (between BH20 and BH11) was variable before the flow event. 369 and generally decreased in the downstream direction. The shallow ground-370 water responds immediately to stream flow illustrating infiltration of surface 371 water into the alluvial sediments and demonstrating an evolving connection 372 between surface and groundwater [19, 56, 57]. 373

The groundwater hydrograph responses vary at the four locations along 374 the channel. For example, in the downstream locations (from T3 and BH 17 375 to T1 and BH 11) the rapid movement of infiltrating surface water to the 376 water table causes a peak in groundwater levels within days of the flow event 377 followed by a steady decline. This is consistent with the conceptual model 378 of groundwater redistribution beneath transitory streams that has been de-379 veloped by Cuthbert et al. [7] and can be described by the aquifer response 380 time (ART) defined as $t_{ART} = \frac{L^2 S_y}{2T}$, where L is a given length, S_y is specific 381 yield and T is transmissivity. In contrast, the subsurface water mound ups-382 tream (from T9 and BH 20 to T7 and BH 18) increases and redistributes 383 much more slowly as a temporary hydraulic connection to the groundwater 384 is established [19]. Our water level measurements, when interpreted using 385 results from a systematic numerical investigations of variations in ground-386 water head in response to surface flow [57], reveal that hydraulic properties 387 of the alluvium are highly heterogeneous. For example, the responses mea-388 sured upstream (BH18 and BH20) indicate that a low-permeability layer (or 380 clogging layer) may exist beneath the stream and that the average hydraulic 390 conductivity is lower compared to the downstream sites (BH11 and BH17). 391

The slower redistribution of water in the shallow aquifer results in far more prolonged surface flow than in the lower catchment. Note that the initially sharp rise in heads recorded at BH20 during the first few days of the flow event is likely due to a loading effect with the more gradual rise that follows being due to groundwater recharge due to streambed infiltration and lateral movement of groundwater.

Interestingly, the surface water hydrograph after the flood peak behaves differently for each array along the flow path (Figure 4). The upstream arrays show a gradual hydrograph flattening after the initial peak, followed by a stable water level for a period of time which spanned from ~ 3 to 6 weeks



FIGURE 4: Daily rainfall recorded at three stations in the Maules Creek Catchment, hydraulic heads recorded by the streambed arrays installed along Middle Creek, including the nearby groundwater heads where available. Time periods when standing or flowing water was present at the streambed surface are highlighted in grey. Refer to Figure 2 for streambed array and borehole locations. Note that arrays T8 and T5 did not contain pressure transducers. 19

for arrays located at the upper end of the alluvium. During this time surface
water was contained in the stream channel. A steady but significant decline
in water level followed this period of stable water level.

The difference in surface flow behaviour is clearly depicted in Figure 4 405 and is controlled by the rate of groundwater redistribution in the subsurface 406 [7]. It is clear that much of the surface water is retained in the upper part of 407 the channel (upstream from array T6, Figure 2) whereas the lower part of the 408 creek shows short periods of surface run-off consistent with the behaviour of 409 a disconnected ephemeral system [56, 6]. The cause of this behaviour is the 410 subject of ongoing research beyond the scope of this paper, but it is likely 411 controlled by the particle size distribution of the sediment and the general 412 heterogeneity of the channel sediments [58, 20]. 413

414 3.4. Thermal conditions at the streambed surface

Figure 5 illustrates the temperature data recorded by the uppermost pres-415 sure transducer of each array (located at the streambed surface) in individual 416 time colour bars for each location along the channel. Note that the uppermost 417 logger in array T5 failed during deployment and this location is therefore ex-418 cluded from further analysis. The times when surface water was present, as 419 indicated by the sensor measuring values above atmospheric pressure, are 420 indicated as horizontal lines. The air temperature (MCF weather station), is 421 plotted for comparison and varied between -0.7 and $33.5^{\circ}C$ while the sedi-422 ment surface temperatures varied between 2.7 and $45.4^{\circ}C$. 423

A decrease in overall temperature reflects the transition between autumn 424 and winter in the southern hemisphere. While there is an obvious correlation 425 between the air and the streambed surface temperature, the diel tempera-426 ture fluctuations are more pronounced at the streambed surface and vary 427 depending on the array location. Thermal conditions at the streambed sur-428 face were affected by direct insolation during day time and differ depending 429 on location settings caused by variable amounts of shading. The similarity of 430 thermal conditions with low diel variability during the flow event is apparent. 431 The streambed surface temperatures clearly contain diel temperature os-432 cillations modulated by mesoscale weather events (Figure 5). Figure 6 shows 433 the diel amplitudes extracted from the air and streambed surface tempera-434 ture records using FFT analysis. The range of air temperature amplitudes 435 was between 1.1 and $9.7^{\circ}C$, whereas the range of streambed surface tem-436 perature amplitudes ranged between 0 and $10^{\circ}C$. A correlation between air 437 and streambed surface temperature amplitudes is clearly visible in Figure 6 438



FIGURE 5: Temperatures recorded in the air and at the streambed surface along Middle Creek. Black lines indicate saturated conditions at the surface, i.e. the time during which the sensor was submerged in water. Note that the air temperature was not recorded during a small period in May 2014, that array T8 did not contain a pressure transducer, and that array T5 probe failed during deployment.



FIGURE 6: Amplitudes of the diel component of recorded temperature variations in the air and at the streambed surface along Middle Creek. Black lines indicate saturated conditions at the surface, i.e. the time during which the sensor was submerged in water.

for periods when the streambed surface was dry. Diel amplitudes show significant damping during the flow event when ponded or flowing water was
present at the streambed sediment surface.

As observed by Constantz et al. [16], the onset of flow is preceded by lower absolute temperatures and smoothed diel amplitudes associated with the mesoscale low-pressure system. Our measurements confirm that flow cannot be deduced from temperature measurements and extracted amplitudes alone.

3.5. Streambed thermal signatures can detect the presence of water and cha racterise vertical water movement

If amplitude ratios for dry and saturated conditions can be calculated, 448 then the vertical amplitude ratio time series in shallow streambed sediments 449 (Figure 6) can be used to detect both the presence of water and to characte-450 rise the flow regimes according to the theory developed above. While A_r^{dry} can 451 be evaluated from measurements during dry periods, A_r^{sat} requires estima-452 tion based on the likely values established from Monte-Carlo analysis. Note 453 that the difference between both values is relatively small ($\Delta A_r^{dry,sat} < 0.12$). 454 Both values constrain a narrow range between them where the interpretation 455 of vertical flow is ambiguous. However, as explained in Section 2.3, A_r values 456 outside that range are directly indicative of the direction and magnitude of 457 vertical water flow. 458

The amplitude ratio A_r^{dry} for dry streambed sediments at each location 459 was calculated using the diel amplitudes extracted from temperature records 460 using FFT analysis between 8-15 March 2014, and values are summarised in 461 Table 2. While thermal diffusivity results comply with those calculated from 462 the *Monte-Carlo* analysis, they are higher than expected which indicates the 463 presence of large sized grains. Visual inspection of the streambed sediments 464 confirms this inference and many large cobbles can be seen in the foreground 465 of Figure 3 [41]. 466

During flow events (wet streambed conditions) the amplitude ratio will 467 depend on the vertical streambed water flux (see Equation 4). Theoretically, 468 the A_r could be used to quantify this vertical flux [38, 59] and, provided that 469 phases of the diel frequency components are also extracted, the saturated 470 thermal diffusivity of the streambed could also be quantified [52, 60]. Howe-471 ver, Rau et al. [49] demonstrated that analytical heat tracing methods fail to 472 provide accurate results when the diel component in the temperature signal 473 is non-stationary. This includes highly transient infiltration as is expected 474



FIGURE 7: Diel temperature amplitude ratios A_r between the uppermost pair of sensors in the streambed. The colour map is adapted for each location to correctly reflect : A_r^{dry} as established from measurements during a dry period, and $A_r^{sat} = A_r^{dry} + \Delta A_r$ calculated using thermal diffusivity values from *Monte-Carlo* results as well as site-specific sensor spacings. The colours reflect saturated conditions, where increasing blue represents an increasing vertical upward flow component (1) and colours increasing towards red represent increasing vertical downward flow component (4). Red reflects periods during which the $A_r > 1$ and indicates horizontal hyporheic flow (5). Black lines indicate wet conditions at the surface (top) and at depth (bottom) in the streambed, i.e. the times during which the loggers were submerged in water. The numbers along the colour bar correspond to the thermal signature characterizations defined in Section 2.3 and Figure 1. The daily rain is plotted to show the influence on the streambed thermal regime.

⁴⁷⁵ during the dynamic flow events which are characteristic of Middle Creek ⁴⁷⁶ (Figure 4). We therefore abstain from using phase results in our analysis.

Figure 7 shows the amplitude ratio time series for all arrays along Middle 477 Creek translated into colours that reflect the different categories explained 478 in Figure 1. It is clear that A_r can be used to distinguish between dry and 479 saturated streambed conditions as confirmed by the pressure transducers de-480 tecting water (compare the black line with the coloured pattern representing 481 A_r variation). The influence of rainfall prior to the arrival of the surface 482 run-off is also detected. Further, most arrays show variable downward water 483 movement throughout the flow event (red colour corresponding to range 4 in 484 Figure 1) as is expected for an intermittent system. The only exception is T7 485 which indicates upward movement during the period of surface run-off and is 486 discussed later. Here, water is retained within the alluvium for a time period 487 that exceeds all other locations, as indicated by the hydrograph measured by 488 the sensor at the bottom of the streambed array (Figures 4). 489

The results in Figure 7 contain a wealth of information that could be 490 attributed to processes that have been found to influence transitory SW-GW 491 interactions. For example, it is widely accepted that the hydraulic properties 492 of alluvial sediments are strongly heterogeneous which can lead to zones of 493 variable saturation beneath the stream [61, 62]. A field investigation using 494 moisture sensors to measure the temporal behaviour of infiltration has repor-495 ted localised preferential flow which contributes to a rising water mound that 496 can saturate the streambed from the bottom upwards [18]. An increase in 497 saturation in the alluvial sediments due to infiltration may be considerably 498 delayed after the onset of flow due to variability in sediment properties such 499 as grain size [18, 63]. Moreover, certain combinations of channel geometry 500 and stream water level can induce water saturation beneath the stream but 501 without a saturated connection to the groundwater (inverted water table) 502 [64]503

We note that all these processes could affect the shallow streambed ther-504 mal diffusivity and therefore also the derived temperature amplitude ratios. 505 As an example, T11 illustrates a thermal signature indicative of variably sa-506 turated sediment at the beginning of the flow event (Figure 7) during the 507 same time as the pressure transducer clearly indicates the presence of sur-508 face water (Figure 4). This observation is in agreement with the previous 509 findings of delayed saturation or rising water mound and illustrates that 510 thermal signatures can enhance interpretation of the complexity of dryland 511 SW-GW interactions, even more so when combined with water level measu-512

rements. We further note that thermal signatures and water levels acquired during multiple flow events can be used to reveal the temporal dynamics of infiltration over longer time scales which could enhance the interpretation of transience in streambed conductance[65]. This could further improve our understanding of the complex water flow dynamics at the variably saturated stream-aquifer interface.

519 3.6. Streambed thermal regimes and spatio-temporal flow behaviour

To characterise the thermal conditions during flow events, the hydraulic 520 head and temperature records for two representative multi-level arrays were 521 plotted for T9 in Figure 8 and for T7 in Figure 9. These plots include the 522 temperature data measured at multiple levels within the topmost meter of 523 the channel sediment and diel temperature amplitudes as extracted from 524 the measurements using FFT analysis. Both streambed arrays contain the 525 thermal signatures which are found in all other locations (Figure 7) and are 526 therefore worthy of detailed inspection. 527

Figure 8a clearly shows the temporal character of flow events measured 528 at the location of streambed array T9. T7 shows a similar hydrograph mea-529 sured by the pressure transducer at the bottom, but the one at the top only 530 captured the peak of the flow event whereas the bottom logger remained 531 submerged in water contained in the streambed for a period of time. From 532 Figure 4 it is clear that all hydrographs which captured more than the initial 533 peak illustrate a similar shape but with differing duration of the stable or 534 receding water level (intermittent stream behaviour). 535

The following flow regimes can be derived from the observed hydrograph shapes, and are categorised below and illustrated in a conceptual model of transitory surface-groundwater interactions (Figure 10, colours refer to Figures 8 and 9) :

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[A] Dry channel (red) as a default for dryland streams : The dry sediments are characterised by large temperature amplitudes at the surface that is rapidly damped with depth for both T9 (Figure 8b) and T7 (Figure 9b). The large amplitudes at the boundary are a result of insolation and indicate dry conditions (absence of water). The A_r -depth profile for a location, as shown in Figures 8d and 9d, can be used to benchmark the thermal conditions in the dry streambed.

[B] Rapid surface run-off (green) : Surface run-off and infiltration along the channel may result in a spatially heterogeneous distribution of alluvium



FIGURE 8: Streambed array T9 : a) Hydraulic head at the top and bottom of the array. The grey band indicates the depth interval in which temperature data is interpreted in Figure 7. b) Multi-level temperature records. b) Multi-level temperature records. c) Amplitude ratio time series $A_r(t)$ of the diel temperature component for 3 depths (same legend as panel b). d) Depth profiles of diel temperature amplitude ratios averaged over the time period corresponding to the colour coded flow regimes A-D labelled at the top of panel (a) and which are sketched in Figure 10

water saturation beneath the channel. Upon arrival of the water in the dry channel, the temperature rapidly changes over depth with an associated increase in the diel temperature amplitude (Figures 8b and 9b). This reflects the highly transient infiltration of water which carries a contrasting temperature downwards [24]. Further, this marks a period of highly transient infiltration [29, 66] in particular for locations that show ephemeral behaviour (T4-T1 in Figure 7). The streambed saturation may be significantly delayed compared to the arrival of surface water (T11 in Figure 7).



FIGURE 9: Streambed array T7 : a) Hydraulic head at the top and bottom of the array. The grey band indicates the depth interval in which temperature data is interpreted in Figure 7. b) Multi-level temperature records. c) Amplitude ratio time series $A_r(t)$ of the diel temperature component for 3 depths (same legend as panel b). d) Depth profiles of diel temperature amplitude ratios averaged over the time period corresponding to the colour coded flow regimes A-D labelled at the top of panel (a) and which are sketched in Figure 10

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[C] Pool-riffle sequence (blue) : This regime is characterised by water flow through pool-riffle sequences including varying proportions of both subsurface (hyporheic) and surface flow that is predominantly horizontal. It only occurs if the infiltrated water is not redistributed fast enough so that the groundwater table rises above the streambed surface thereby intersecting the channel topography. The duration of this regime varies depending on the lateral aquifer response time (ART), the rate at which the subsurface water mound redistributes [7]. Consequently, this regime is much shorter or may never be reached in locations that have a low ART. Further, the timing of the transition to the next flow



FIGURE 10: Conceptual model of the different hydrological regimes that occur during transitory surface water-groundwater interactions in ephemeral or intermittent streams. Note that while the regions of longer and shorter aquifer response time (ART, a measure for the redistribution rate of infiltrated water [7]) greatly simplify realistic conditions, it is reflective of our field conditions and provides a range of conditions which may be relevant to other studies. A variable ART also explains the potential occurrence of regime C and D. Note further that streambed arrays T9 and T7 are located to qualitatively reflect the measured water levels and thermal signatures (Figures 8 and 9). The hydrological and thermal conditions of this sequence is detailed in the discussion.

regime depends on the local streambed morphology and is therefore somewhat ambiguous. The shallow subsurface temperatures during this regime are similar to those observed in perennial systems dominated by hyporheic exchange [67, 68].

During this flow regime, the locations show differing behaviour : T9 fea-572 tures an A_r -depth profile that is significantly different from dry condi-573 tions and indicates a downward flow component (Figure 8). In contrast, 574 the shallower part of T7 indicates an upward flow component whereas 575 the deeper part shows increasingly downward flow (Figure 9). The dif-576 ference between T9 and T7 are indicative of their different locations 577 within the pool-riffle sequence and in relative elevation of water table 578 relative to the ground surface (Figure 10). T7 was located at the end 579 of a gravel bar with up-welling hyporheic flow at the top of the array 580 throughout the short duration of the surface run-off. T9 was located at 581 the downstream end of a pool. 582

Note that the array locations relative to the pool-riffle system will 583 change as the water level recedes, and also due to potential erosion 584 during surface run-off. It is noteworthy that during this flow regime 585 the diel amplitude propagates to the lowest sensor in the sediment and 586 can cause an amplitude ratio that is larger than unity $(A_r > 1)$ thus vio-587 lating the conditions required to apply vertical analytical heat tracing. 588 In the absence of a subsurface thermal source, $A_r > 1$ is an indicator 589 for hyporheic flow with a significant horizontal component [47, 69]. 590

⁵⁹¹ [D] Cessation of riffle flow and drying of the isolated pools and sediments ⁵⁹² (yellow) : A steady decrease in hydraulic head indicates that water ⁵⁹³ is redistributing in the subsurface leaving the channel sediments to ⁵⁹⁴ dry out. Similar to (C), this regime may be bypassed under certain ⁵⁹⁵ conditions. The increase of the diel temperature amplitude, particularly ⁵⁹⁶ at the lower sensors, is an indication of a significant downward water ⁵⁹⁷ flux.

⁵⁹⁸Our conceptual model is supported by the fact that surface flow exists at ⁵⁹⁹locations when surface water further upstream has disappeared (Figure 4). ⁶⁰⁰Consequently, water contained in the shallow alluvium must move downs-⁶⁰¹tream and sideways as the overall water table elevation slowly falls below the ⁶⁰²lowest elevations of the streambed surface. We further note that the existence ⁶⁰³of these regimes was verified by visual observations made during numerous ⁶⁰⁴field trips throughout the hydrological sequence. This is further verified by time lapse images captured using a camera mounted beside the stream near BH20/T9, as described in a previous study [7].

607 4. Conclusions

We have shown how amplitude ratios of the diel component in tempera-608 ture time series measured at two vertical locations in shallow streambeds can 609 be used to detect saturation conditions and to characterise transitory flow 610 conditions. This is an advantage over head measurements due to the lower 611 cost involved and ease of installation which allows the possibility of a wider 612 spatial deployment of sensors. Amplitude ratios depend on the sediment ther-613 mal diffusivity, which is a function of the different thermal properties of air or 614 water occupying the pore space. While the dry streambed thermal diffusivity 615 can be determined from temperature records acquired during dry periods, 616 the saturated thermal diffusivity is always higher depending on the sediment 617 properties. The likely difference between dry and saturated amplitude ra-618 tios does not exceed ~ 0.175 as illustrated using a *Monte-Carlo* analysis with 619 probable ranges in matrix thermal properties available in the literature. 620

A small range of amplitude ratios exists for which interpretation of the 621 state of saturation is ambiguous, i.e. either variably saturated sediments or 622 full saturation with upward flow. The range of ambiguity is determined by the 623 difference between dry and saturated streambed thermal diffusivity, which 624 depends both on porosity and matrix thermal properties. However, when 625 interpreted in combination with pressure data, which is indicative of whether 626 or not water is present above the point of measurement, this range can still 627 be used to reveal streambed processes. 628

We have applied this new approach to multi-level temperature data from 629 streambed arrays deployed along a ~ 12 km channel section. Hydraulic heads 630 were measured simultaneously by the arrays as well as at co-located shallow 631 piezometers. The data demonstrate that intermittent surface water-groundwater 632 interactions are highly variable in space and time. The interpreted tempera-633 ture and pressure data enable categorization of these interactions into four 634 generic hydrological regimes that can occur sequentially in time : (A) dry 635 636 channel, (B) rapid surface run-off along the channel, (C) pool-riffle sequence with horizontal hyporheic flow, (D) isolated pools. The duration of each re-63 gime will depend on the channel morphology as well as the lateral aquifer 638 response time (ART) which controls the rate of groundwater redistribution. 639 Our results illustrate that sequence C and D may not be reached in the case 640

that the infiltrated water is redistributed fast enough so that the groundwater
level does not rise above the streambed surface for a significant duration.

Such analysis enables determination of the intricate dynamics inherent to 643 the connectivity between intermittent surface flow and groundwater and is 644 directly relevant to other semi-arid and arid regions of the world [1]. Unders-645 tanding such hydrological behaviour is imperative to conjunctive resource 646 management in water-limited environments [2]. Furthermore, thermal condi-647 tions in the shallow streambed influence water quality through hydrochemical 648 and biological processing and determine the ecological habitat [70, 1]. Our 649 approach to monitoring, understanding and interpreting thermal regimes in 650 intermittent and ephemeral streams can, therefore, improve spatiotemporal 651 understandings of hyporheic processes and associated water quality dyna-652 mics, groundwater recharge, and when and how dryland streams support 653 riparian ecosystems. 654

655 Acknowledgements

We are grateful for technical and field support provided by Evan Jensen, Ed-656 wina Davison, Calvin Li and to landowners in Mauls Creek (Philip Laird, Alistair 657 Todd, and Steve Bradshaw) for giving access to the field sites. The figures in this 658 manuscript were made with the help of *Matplotlib* [71]. The data used in this ana-659 lysis was collected with equipment provided by the Australian Federal Government 660 financed National Collaborative Research Infrastructure Scheme (NCRIS, data 661 available at : http://groundwater.anu.edu.au/fieldsite/maules-creek). We 662 would like to thank the Cotton Catchment Communities CRC for their financial 663 support (Cotton CRC projects 2.02.03 and 2.02.21). The elevation data in Figure 2 664 is courtesy of Geoscience Australia http://www.ga.gov.au. MOC was suppor-665 ted by the European Community's Seventh Framework Program (FP7/2007-2013) 666 under grant agreement 299091. GCR was supported by the National Centre for 667 Groundwater Research and Training (NCGRT), an Australian Government ini-668 tiative supported by the Australian Research Council (ARC) and the National 669 Water Commission (NWC). LJSH was supported by the NSW State Govern-670 ment's Research Acceleration and Attraction Program (RAAP) in the year 2016. 671 We further thank the Editor Paolo D'Odorico as well as the AE and 2 reviewers (all 672 anonymous) for handling and suggesting improvements to this manuscript during 673 peer-review. 674

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