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Active distributed temperature sensing to assess surface water–groundwater interaction and river loss in braided river systems

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Abstract

Braided river systems are continuously changing environments, comprising heterogeneous gravel-forming meandering channels and bars, and dynamic surface water–groundwater (SW–GW) interactions. In alluvial aquifers associated with braided rivers, river leakage is volumetrically the most important source of groundwater recharge, yet neither the leakage process or recharge rates are well understood. Insights into how braided river systems recharge underlying braidplain aquifers during different hydrological conditions, and how this can be measured, is of critical concern for river management practices. Innovative field techniques using heat as a tracer can be applied to spatially map and quantify the temporal changes of SW–GW interaction and the processes and mechanisms of groundwater recharge for different flow regimes. Horizontal directional drilling was used to construct two 100 m long drillholes at a depth of ~ 5 m beneath and perpendicular to the river channel at a field site on the Waikirikiri Selwyn River, South Island, New Zealand. The two drillholes were completed with a hybrid fibre optic cable, containing four multi-mode fibres and 2x18AWG copper

conductors. A Silixa XT-DTSTM distributed temperature sensor unit combined with a Silixa Heat Pulse System, was used to collect active distributed temperature sensing (A-DTS) measurements along both cables. Results from two A-DTS surveys showed distinct temperature changes across the active river channel indicating spatial variability in river loss and preferential groundwater recharge pathways to the shallow braidplain aquifer. The average groundwater velocity was 14 m d^{-1} . The average calculated river loss at the survey sites was $21 \text{ m}^3 \text{ d}^{-1} \text{ m}^{-1}$ compared to $25 - 26 \text{ m}^3 \text{ d}^{-1} \text{ m}^{-1}$ that was measured using differential flow gauging conducted during the same time period. The horizontal installation of the fibre optic cable and the A-DTS method provided valuable insights into SW–GW exchange between the Waikirikiri Selwyn River and shallow braidplain aquifer and how the river leakage rates varied spatially across the river channel.

1 Introduction

River loss to underlying aquifers (i.e., groundwater recharge) from losing rivers is the quantity of greatest interest in surface water–groundwater (SW–GW) investigations (Shanafield and Cook, 2014). For braided rivers, river leakage is the most important source of groundwater recharge to the connected alluvial aquifers, yet the recharge mechanisms and pathways are not well understood (Coluccio and Morgan, 2019; Rupp et al., 2008). Insights into how these rivers recharge underlying braidplain aquifers and the SW–GW exchange during different hydrological conditions and temporal scales, and how it can be measured, is of critical concern for river management practices (Piégay et al., 2006). Due to the high energy of these environments, there is continuous change with the movement and creation of heterogeneous, gravel-forming, meandering channels and bars, and dynamic SW–GW interactions (Gray and Harding, 2007; Reinfelds and Nanson, 1993). The changing river geomorphology has been identified to impact on the spatial variability in the SW–GW exchange and the recharge processes (Wöhling et al., 2018).

Temperature or heat-based methods are one of the most widely used field techniques to study and characterise SW–GW exchange and to determine the state of connectivity (Anderson, 2005; Constantz, 2008; Kalbus et al., 2006; Rau et al., 2014). Heat is an effective environmental tracer for quantifying SW–GW exchange fluxes because of its strong diurnal fluctuations in surface waters and low cost of application (Hatch et al., 2006; McCallum et al., 2012; Shanafield et al., 2010). A

standard application involves the use of discrete temperature sensors (e.g., i-buttons) at multiple depths to evaluate and quantify the rate of exchange within the hyporheic zone and between the river and the underlying aquifer systems (Rau et al., 2010), and in heterogeneous river environments (Irvine et al., 2017). In the last two decades, higher spatial resolution surveys of river and riverbed settings have been undertaken using fibre optic distributed temperature sensing (FO-DTS) that apply passive-DTS and active-DTS measurements (Selker et al., 2006b; Shanafield et al., 2018; Tyler et al., 2009). Passive-DTS consists of measuring the natural temperature fluctuations with time and the quantification of fluxes is determined together with other distributed temperature measurements and outputs from energy balance models (Selker et al., 2006a). In comparison, active-DTS (A-DTS) consists of measuring temperature changes in response to a continuous heat source along a FO cable, which enables quantification of fluxes at a higher spatial and temporal resolution (Simon et al., 2021a). Laboratory scale studies have been undertaken to validate the A-DTS methodology in using heat as a tracer and to calibrate thermal properties of both the water and sediment to quantify water velocities (Bakx et al., 2019; Maldaner et al., 2019; Simon et al., 2020; Simon et al., 2021b).

At the field and river reach scale, both passive and active-DTS surveys have been utilized to locate SW–GW exchange and quantify groundwater discharge into streams (e.g., González-Pinzón et al., 2015; Simon et al., 2021a). In these applications, the FO cable has been installed at or just below (few centimetres) the riverbed surface (Mohamed et al., 2021; Slater et al., 2010). There are fewer studies that have used DTS to measure seepage rates in losing streams to groundwater (e.g., Vogt et al., 2010). In Trento, Italy, Busato et al. (2019) used horizontal drilling techniques to install a FO-cable beneath the riverbed and applied passive-DTS to understand the interconnection between surface water and groundwater in the Vermigliana Creek catchment. To the authors’ knowledge, this has been the only other study and installation of FO cable beneath a river channel.

One of the key challenges of monitoring SW–GW exchange in braided river systems is the installation of appropriate water monitoring infrastructure, which is at risk of damage due to flooding and riverbed scouring. For the first time, this study used sub-surface horizontal drilling and A-DTS surveys to measure river infiltration loss (i.e., groundwater recharge) in a braided river system in the South Island of New Zealand. The novel approach using the A-DTS survey data and an analytical moving instantaneous line source solution of the heat transport equation, made it possible to quan-

86 tify groundwater velocities and thermal conductivities in the underlying shallow braidplain aquifer.
87 The study demonstrated the advantages of this cutting-edge field technique over other methods
88 to quantify SW–GW exchange. It provided high spatial resolution mapping of river infiltration
89 loss across the width of the active river channel and identified the role of river geomorphology on
90 preferential groundwater recharge pathways in the shallow braidplain aquifer.

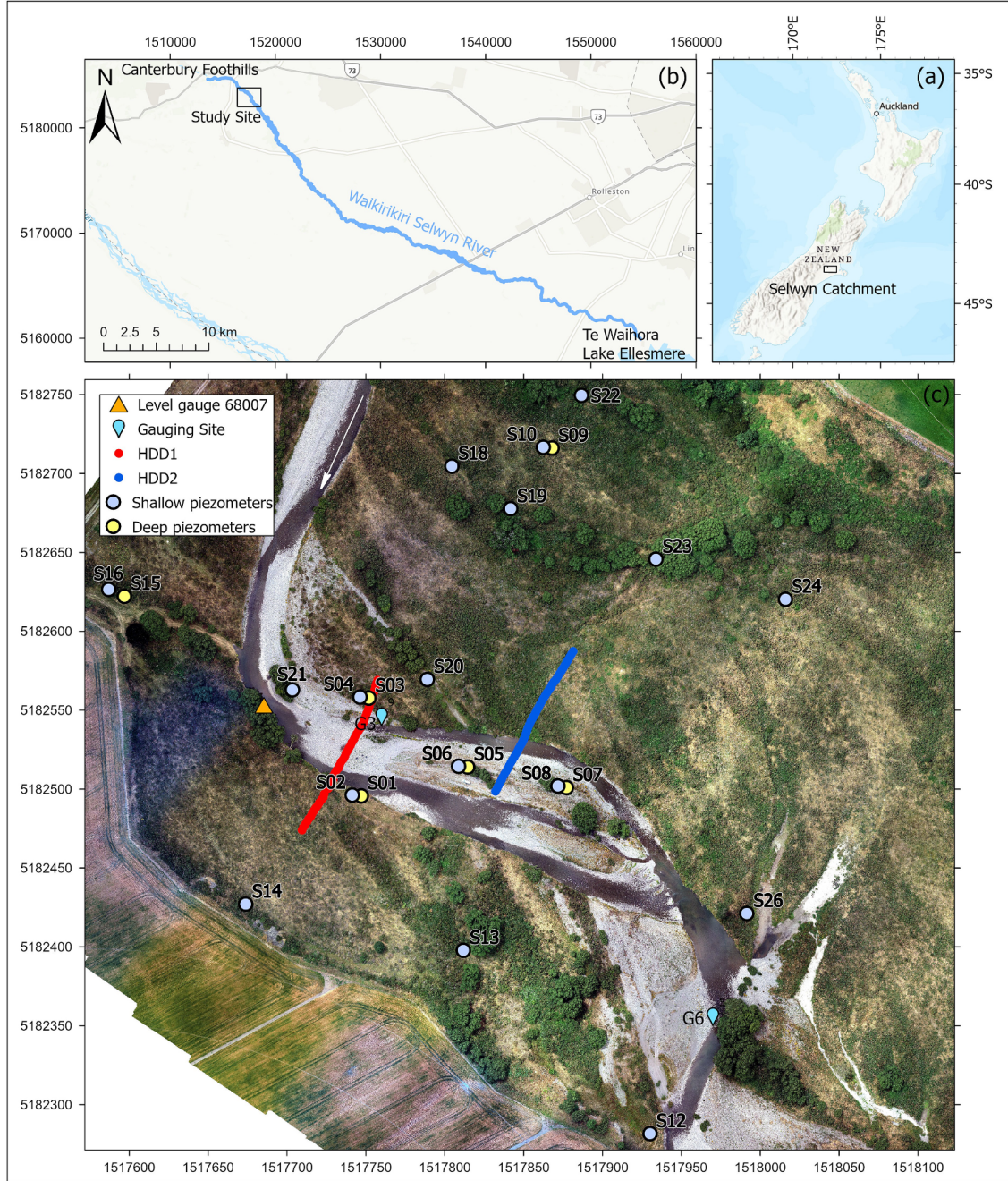


Fig. 1. (a) Location map of the Waikirikiri Selwyn River catchment, New Zealand, (b) Waikirikiri Selwyn River catchment and (c) Study site and the location of the two subsurface horizontal drillholes containing the A-DTS cable and location of river level gauge and the shallow and deep piezometers completed in the braidplain and regional aquifers. Also shown are the manual differential gauging stations G3 and G6.

2 Study site

The Waikirikiri Selwyn River is located on the Canterbury Plains in the South Island of New Zealand (Fig. 1). It flows from the Canterbury foothills southeast across the Canterbury Plains for approximately 80 km, where it joins Te Waihora Lake Ellesmere at the coast. Surface runoff from the hills generates most of the river flow to the upstream section of the Waikirikiri Selwyn River, which has a braided morphology. The downstream section receives groundwater from the Canterbury Plains aquifers and is characterised by a meandering braided river channel. This river is a highly dynamic river system and is characterised by large fluctuations in discharge between summer and winter months (Fig. 2). It has perennial, ephemeral and intermittent reaches that experience both losing and gaining SW–GW conditions. The primary mechanism of aquifer recharge is leakage from the river channels (Larned et al., 2008).

The study site is an active channel of the Waikirikiri Selwyn River (Fig. 1c) and is located approximately 210 m above mean sea level (AMSL). Here, the river comprises one to three braids and is within an ephemeral losing reach. There is a gravel bar within the study site, which has two clusters of piezometers (S05, S06, S07 and S08). The contemporary braidplain is approximately 400 m wide and is bound by river terraces that are up to 5 m high. The active riverbed is up to 60 m wide. There are many paleochannels present at the site, which may act as preferential flowpaths and a recharge mechanism to the underlying braidplain aquifer.

The geology of the area comprises two distinct hydrostratigraphic units of greywacke-derived gravel, known regionally as the Springston and Burnham formations (Forsyth et al., 2008). The Springston Formation was deposited during the Holocene epoch under slightly degrading depositional conditions. As a result, these gravels occur as thin lenses of postglacial fluvial channel and overbank sediments associated with contemporary braided rivers. Deposition during the current postglacial period with a relatively low sediment supply has resulted in a lag deposit of well-sorted gravels, sands and silts (Brown et al., 1988). This unit is approximately 3.5 to 4 m thick in the study area and hosts a high permeability aquifer associated with the Waikirikiri Selwyn River, herein referred to as the ‘braidplain aquifer’. An aquifer test carried out for this study estimated hydraulic conductivity of $75 - 250 \text{ m d}^{-1}$ and a specific yield of 0.03 ± 0.01 for this unit.

The Burnham Formation lies beneath the Springston Formation and hosts the regional aquifer

system. Burnham Formation deposits consist of glacial and periglacial outwash gravels deposited in the last glacial period (Vincent, 2005). Core sediment material recovered during drilling at the study site shows fine to coarse greywacke gravels with occasional cobbles in a fine sand to clay matrix interbedded with silt layers (Fig. S1). The permeability of the Burnham Formation is low near the surface, but generally increases with depth. This vertical change in permeability creates a zone of lower water content between the base of the braidplain aquifer at approximately 4 m depth, and the potentiometric surface of the regional aquifer at approximately 20 m below ground level (bgl). The regional aquifer is not the focus of this study, however, based on water level and temperature data, this aquifer can be considered hydraulically disconnected from the river and shallow braidplain aquifer, and separated by an unsaturated zone of ~ 15 m in thickness.

The calculated porosity values for the Springston and Burnham Formations range between 0.18 and 0.3 from 27 samples of recovered drillhole core material at the study site (Robb, 2021). These values are consistent with other studies on the Canterbury Plains (Dann et al., 2009).

3 Methods

3.1 Horizontal directional drilling

Horizontal Directional Drilling (HDD) was used to construct two 100 m long, 125-mm diameter drillholes at a depth of approximately 5 m beneath and perpendicular to the river channel at the study site on the Waikirikiri Selwyn River. The two HDD drillholes (True Right [HDD1] and True Left [HDD2]; orientation looking down river) were drilled with the assistance of a xanthan gum drilling mud to keep the hole open and provide lubricant during the installation process. Following a successful drillhole, a 125-mm diameter poly pipe conduit was welded to the exposed drill bit. This poly pipe was temporarily installed within the HDD hole via back-reaming. Hybrid FO cable, consisting of multi-mode fibres and 2x18AWG copper conductors (Fig. S2) were then threaded through the temporary conduit. The FO cables were then held in place, while the temporary conduit was removed. This installation method enabled the drillhole to subsequently collapse, allowing the riverbed material to make direct contact with the cable. The cable connector terminals were secured in a buried cable pit on the riverbank to prevent vandalism and damage from river flooding.

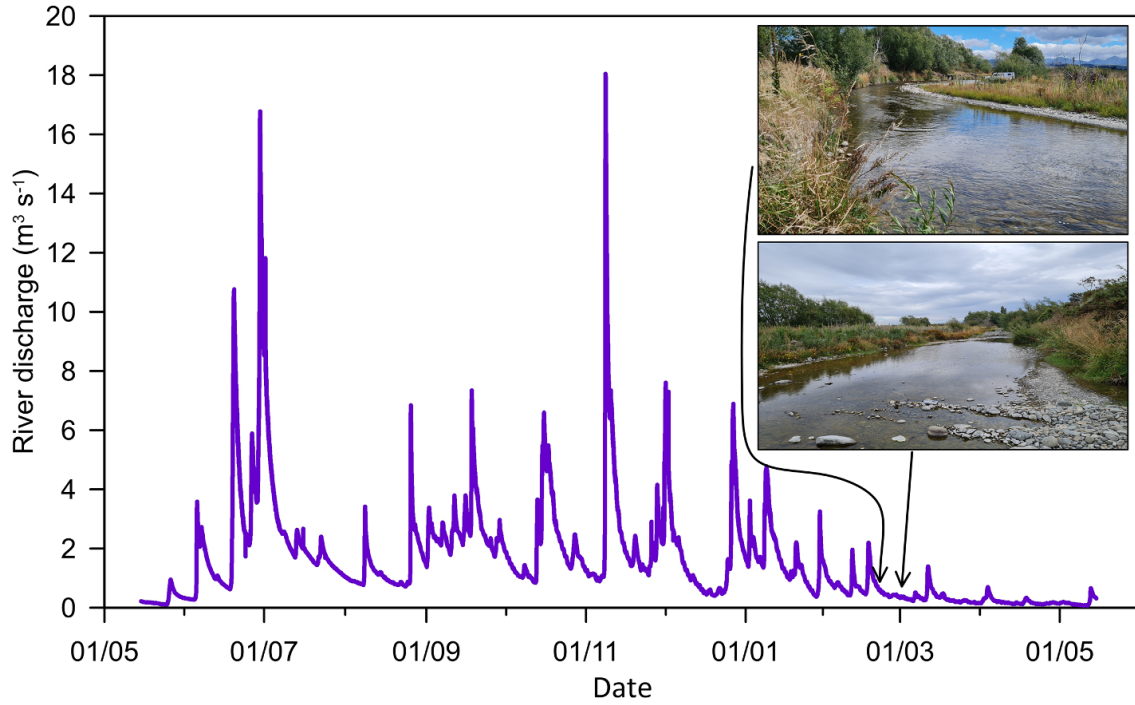


Fig. 2. River discharge of the Waikirikiri Selwyn River between 15 May 2020 to 15 May 2021. Inset photos of the river conditions (looking upstream) for the dates of the A-DTS surveys for the True Right (24/02/2021) and True Left (01/03/2021).

3.2 Active distributed temperature sensing

A Silixa XT-DTS[™] distributed temperature sensor combined with a Silixa Heat Pulse System was used to collect A-DTS measurements along both FO cables. The reported manufacturer specifications of the XT-DTS details a spatial sample interval of 0.25 m, spatial resolution of 0.54 m and a temperature resolution of 0.01 °C (Silixa Ltd, UK). The spatial resolution is larger than the sample interval and is typically estimated by applying the 90% step change method where the true temperature change occurs over a step-wise shift in distance along the FO cable (Selker et al., 2014; Simon et al., 2020; Tyler et al., 2009). A double-ended configuration was used where the instrument makes measurements from both directions of a looped FO cable that has its two ends connected to the XT-DTS. The non-uniform differential loss along the length of the fibre is corrected by aligning the forward and reverse signals that are measured from both ends of the fibre. In addition, 8 m of cable reference sections were positioned in two calibration baths of different temperatures (ambient river water collected at the time of survey and ice water: 0 °C) that were monitored using Pt-100

temperature probes connected to the XT-DTS (Fig. S3). The temperature measurement error for the ice calibration bath for the True Right was 0.068 °C and 0.099 °C for the True Left, over the duration of the measurement period. The measurement error for the river calibration bath for the True Right was 0.100 °C and 0.11 °C for the True Left. Each measurement cycle used the ice calibration bath temperature to correct the reported temperature value along the length of the cable at a sample interval of 0.25 m. The double-ended configuration together with the calibration baths greatly improves the temperature accuracy and precision along the length of the fibre (Hausner et al., 2011; van de Giesen et al., 2012).

The measurement sequence, acquisition sampling rate, and heat-pulse duration for the surveys, using the double-ended configuration, involved the following: An acquisition sampling rate of 10 s (sampled at 5 s on channel 1-forward measurement) and 5 s on channel 2- reverse measurement) for a measurement period of 300 s (i.e., baseline temperature conditions) prior to heating the cable for 1200 s at 15 W m⁻¹ of cable, followed by a post-heating measurement period of 1200 s. This sampling routine was sufficient to capture subtle temperature changes along the fibre and a significant increase in temperature from the baseline conditions.

3.3 Surface water and groundwater monitoring infrastructure

There is an extensive network of surface water and groundwater monitoring infrastructure at the study site (Fig. 1). It includes a river level gauge recorder (station: 68007) on the western edge of the main channel and a network of 27 shallow piezometers (e.g., S01, S02 etc) completed in the unconfined shallow braidplain aquifer (screen intervals between 1 and 4 m below ground) and deep piezometers completed in the semi-confined regional aquifer (screen intervals between 19 and 35 m below ground). River stage and surface water temperature are recorded at 15-min intervals. Regular manual flow gauging has been undertaken to create a flow rating curve to calculate surface water discharge at the level gauge and differential flow gauging to determine water loss along the river reach within the study site. Water levels and temperature in the shallow and deep aquifers are recorded at 15-min intervals.

3.4 Mathematical model for estimating thermal conductivity and groundwater velocity

The evolution of temperature in saturated porous media due to advection and conduction is described by the 2D heat transport equation (Bakker et al., 2015; Domenico and Schwartz, 1998):

$$\frac{\partial T}{\partial t} = D \left(\frac{\partial^2 T}{\partial x^2} + \frac{\partial^2 T}{\partial y^2} \right) - \frac{u}{R} \frac{\partial T}{\partial x} \quad (1)$$

where T ($^{\circ}\text{C}$) is temperature, t (s) is time, D (m^2/s) is thermal diffusivity, u is groundwater velocity (m/s) in the positive x -direction and R (-) is a thermal retardation coefficient.

The porosity-weighted thermal diffusivity and thermal retardation coefficients are defined as:

$$D = \frac{\lambda}{c\rho} = \frac{n\lambda_w + (1-n)\lambda_s}{nc_w\rho_w + (1-n)c_s\rho_s} \quad (2)$$

$$R = \frac{c\rho}{nc_w\rho_w} = 1 + \frac{(1-n)c_s\rho_s}{nc_w\rho_w} \quad (3)$$

where n (-) is porosity, λ ($\text{W/m } ^{\circ}\text{C}^{-1}$) is thermal conductivity, c is specific heat capacity ($\text{J kg}^{-1} ^{\circ}\text{C}^{-1}$) and ρ (kg m^{-3}) is density. Subscript w refers to water, s refers to solids and no subscript indicates the combined rock-fluid matrix. It is assumed that the temperature of water and solids are at instantaneous equilibrium and the effect of temperature on density and viscosity is negligible (Anderson, 2005).

In this study, A-DTS survey temperature data and the analytic moving instantaneous line source (MILS) solution of Zubair and Chaudhry (1996), as detailed in Bakker et al. (2015), are used to estimate thermal conductivity and groundwater velocity (i.e., average linear or pore fluid velocity; u) along the A-DTS cable at a spatial sample interval of 0.25 m. It is assumed that the estimated groundwater velocity represents the magnitude of the dominant groundwater flow vector across the cable, although the direction of flow cannot be determined by the method and remains unknown. In the MILS model, an infinitely long source of constant heat, Q (W/m) is located at the origin, normal to the plane of constant groundwater flow. The change in temperature ΔT ($^{\circ}\text{C}$) since the

207 start of heating t (s) at a radial distance r (m) from the line source is given by:

$$\Delta T = \frac{Q}{4\pi\lambda} \exp\left(\frac{ux}{2DR}\right) W\left(\frac{r^2}{4Dt}, \frac{ur}{2DR}\right) \quad (4)$$

208 where W is the Hantush Well function defined as:

$$W(a, b) = \int_a^\infty \frac{1}{s} \exp\left(-s - \frac{b^2}{4s}\right) ds \quad (5)$$

209 Observations of temperature change during the heating period were used to fit Equation (4) using
 210 a non-linear least squares algorithm implemented in Python. The output was best-fitting values of
 211 thermal conductivity, groundwater velocity and thermal diffusivity optimised using the temperature
 212 change at each individual measurement point along the cable. The root mean square error (RMSE)
 213 was calculated from the residuals between the best-fitting model and the data, which gives a measure
 214 of the goodness of fit. Additionally, a covariance matrix was calculated, and the parameter standard
 215 deviations used to construct error bars. In this context, a small error bar indicates a good fit was
 216 achieved between measured and modelled data over a small range of parameter values during the
 217 non-linear least squares estimation procedure.

218 The thermal retardation value, R , is calculated as 3.1. This is calculated using a porosity of 0.2
 219 as measured by Robb (2021) in the vicinity of the A-DTS cable. Values of density and specific heat
 220 capacity for water were taken from Bakker et al. (2015) as follows: c_w of $4,100 \text{ J kg}^{-1} \text{ }^\circ\text{C}^{-1}$, ρ_s of
 221 $2,650 \text{ kg m}^{-3}$ and ρ_w of $1,000 \text{ kg m}^{-3}$. Specific heat capacity for the solid, which is a greywacke
 222 derived material, was taken from average values reported by Mielke et al. (2016) as c_s of 800
 223 $\text{J kg}^{-1} \text{ }^\circ\text{C}^{-1}$. This value is slightly lower than the $860 \text{ J kg}^{-1} \text{ }^\circ\text{C}^{-1}$ used by Bakker et al. (2015).

224 The copper heating conductor and optical fibres are contained within a composite FO cable, as
 225 shown in Fig. S2. The manufacturer specifications indicate that the distance between the copper
 226 heating conductor and optical fibre is within the range of 1.8 mm and 3.7 mm. In order to estimate
 227 this distance, we used Equation (4) with $u = 0$ (and other values as listed above) and iteratively
 228 altered the value of λ and r to optimise the match (i.e., minimise the RMSE) between the modelled
 229 and measured heating curves along the cable during the conduction dominant period of the heating
 230 curve. Using this approach, r values of 2.9 mm and 3.1 mm were estimated for the true right and true

left cables, respectively. For the modelling, 3.0 mm was used for the base case scenario. The impact of uncertainty associated with this parameter is explored as part of a sensitivity analysis. A key assumption in our interpretation of the data is that the drillhole annulus has collapsed around the FO cable. This assumption is needed to consider that the A-DTS measurements are representative of perpendicular flow across the cable from the surrounding sediments.

4 Results

4.1 Surface water and groundwater time series

Groundwater levels in the shallow aquifer (piezometers: S02, S04, S06, S08 and S21) show very similar temporal trends to river stage (Fig. 3a). The direct response in the groundwater to changes in river stage indicates that the river and shallow braidplain aquifer are intimately connected. Although the True Right and True Left surveys were conducted 1-week apart, the river level and the shallow groundwater level conditions were similar. The hydraulic head gradient between the river and the observed shallow piezometers ranged from -0.15 to -0.76 (average = -0.5) confirming that the local flow field at the cable locations was from the river to groundwater (Fig. 3b).

Similar temporal trends are observed in the temperature time series data between the river and shallow groundwater (Fig. 3c). At the daily timescale, there is a clear diurnal temperature response in the river, where the surface water temperature increases from early morning and peaking mid-afternoon before decreasing again in the evening (Fig. 3d and Fig. 3e). The diurnal temperature signal in the shallow groundwater is dampened compared to the river. An exception is piezometer S21, which is located on the inside of the river meander bend and shows a similar diurnal temperature signal to the river but there is a clear phase shift (lag), and its amplitude is also slightly dampened. The dampened amplitude and phase lag signals can be used to infer the losing river conditions. We assume that these diurnal temperature variations do not have a significant effect on the ADTS measurements given the overall duration of the experiment (approximately 45-min). The surveys were also undertaken before the highest daily temperatures that are typically experienced by mid-afternoon.

4.2 Active distributed temperature sensing

The baseline and peak heat-pulse temperature measurements for the True Right (24/02/2021) and True Left (01/03/2021) surveys are shown in Fig. 4 and Fig. 5, respectively. For the True Right survey from 0 to 32.1 m, the cable is located within the unsaturated zone and above the water level of shallow piezometer S02. This is also the case at 100 to 105 m, where the cable is above the water level of piezometer S04 (Fig. 4a). The river temperature (21 °C) at the time of the True Right survey was higher than the measured temperature of the shallow groundwater in piezometers S02 (17.1 °C), S04 (16.6 °C) and S21 (15.1 °C), and the measured baseline temperature along the A-DTS cable that was located below the shallow watertable (Fig. 4b).

For the True Left survey from 0 to 46.1 m and from 98.7 to 106 m, the cable is within the unsaturated zone and above the water levels of shallow piezometers S06 and S08, which are located on the gravel bar between two of the braided river channels (Fig. 5a). The river temperature (17.6 °C) at the time of the True Left survey was higher than the measured temperature of the shallow groundwater in piezometers S06 (15.8 °C) and S08 (15.6 °C), and the measured baseline temperature along the A-DTS cable that was located below the shallow watertable (Fig. 5b).

For the A-DTS cable that is located below the watertable in both surveys, there is some visible spatial variability in the measured baseline and peak heat-pulse temperature measurements that suggest zones beneath the river that have greater or less advective flow. For example, at a distance between 62 and 68 m for the True Right there is an abrupt change in the temperature profile. The measured baseline temperature at this low-point is similar to the measured groundwater temperature of S021 at the time of the survey (Fig. 4b) and the river temperature measured early that morning on the 24/02/2021 (Fig. 3d). Similar comparisons can be made for the True Left survey where the measured early morning river temperature on the 01/03/2021 is close to the baseline A-DTS survey and the shallow groundwater.

Temperature breakthrough curves at selected locations along the saturated sections of the FO cable are shown for the True Right survey in Fig. 6 and for the True Left survey in Fig. 7. When the heating cable is turned on, there is a temperature increase from the baseline, of about 4 °C, where it reaches an asymptotic plateau. Once the heating is turned off, the temperature returns to the baseline (Fig. 6a and Fig. 7a). The sections of the cable that are located within the unsaturated

zone and above the watertable of the shallow braidplain aquifer (True Right: 0–32.1 m; 100–105 m and True Left: 0–46.1 m; 98.7–106 m), show greater increases in temperature (8–10 °C) immediately after the heating is turned on (Fig. 6b and Fig. 7b). For the groundwater velocity calculations these measurements are not considered due to the partially saturated conditions of the unsaturated zone.

Other noticeable features in the temperature heat maps (Fig. 6c and Fig. 7c) are locations along the FO cable where there is more rapid cooling (or where the temperature increase is not as high) compared to adjacent locations. These features of rapid cooling are likely to be locations of higher groundwater velocities created by the spatial variability in the hydrogeological conditions of the braidplain aquifer.

For the True Right survey on 24/02/2021, fitting Equation (5) to the temperature data suggests an average groundwater velocity of 12.0 m d⁻¹, with values along the length of cable (at a spatial sample interval of 0.25 m) ranging between 10.8 and 13.6 m d⁻¹ (Fig. 8a). The average thermal conductivity is 0.56 W m⁻¹ °C⁻¹ with a range in values from 0.48 to 0.62 W m⁻¹ °C⁻¹. The clear change in the baseline temperature at a distance between 62 and 68 m is also the location of the calculated highest velocity.

For the True Left survey on 01/03/2021, the average groundwater velocity is 15.8 m d⁻¹ and along the length of cable, the velocities range from 12.4 to 19.4 m d⁻¹ (Fig. 8b). The average thermal conductivity is 0.74 W m⁻¹ °C⁻¹ with a range in values from 0.66 to 0.80 W m⁻¹ °C⁻¹. RMSE for the surveys conducted on the 24/02/2021 and 01/03/2021 are less than 0.07 °C, indicating a good fit between modelled and measured temperature change along the cable.

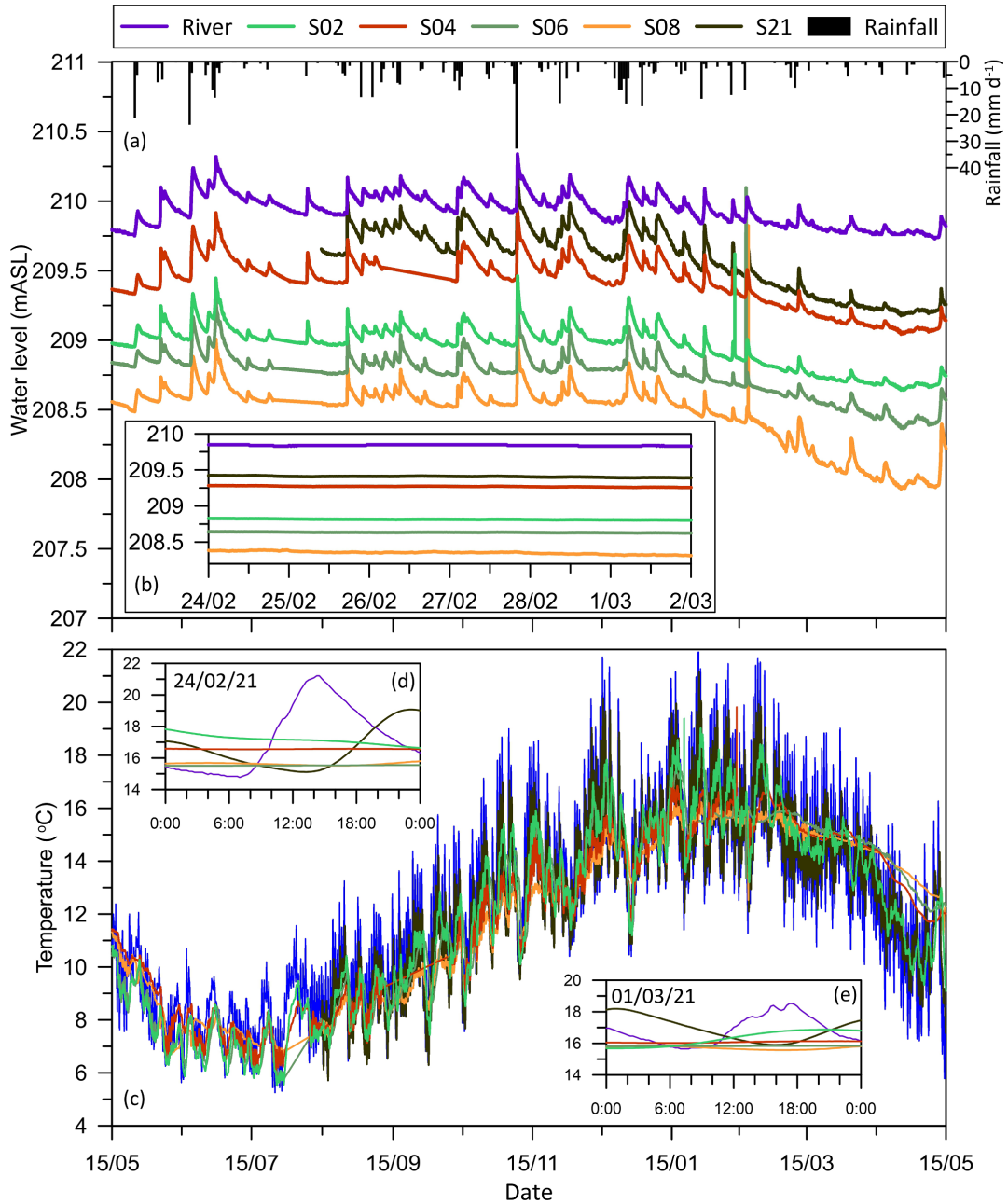


Fig. 3. (a) Selected water level time series data of the Waikirikiri Selwyn River and monitored piezometers in the shallow braidplain aquifer (S02, S04, S06, S08 and S21). Daily rainfall is also shown. (b) Water level time series data at time of surveys. (c) Selected temperature time series data of the Waikirikiri Selwyn River and shallow braidplain aquifer (15 May 2020–15 May 2021). (d) Temperature at time of True Right survey 24/02/2021. (e) Temperature at time of True Left survey 01/03/2021.

A simple sensitivity analysis of the input parameters that were used in the MILS flow and heat transport model was undertaken on the TR survey data. The derived groundwater velocity and

bulk thermal conductivity values as a result of increasing and decreasing the input parameters of porosity, specific heat capacity and r by 20 % are shown in Table 1. Changing the porosity from the base case of 0.2 to 0.16 increased the groundwater velocity from an average of 12 to 14.4 m d⁻¹. Whilst modifying the distance between the line heat source and the fibre (r) from 2.4 to 3.6 mm changed the estimated groundwater velocity from 9.6 to 14.4 m d⁻¹. There was no change to the estimated bulk conductivity with the range of input parameters that were tested.

5 Discussion

Assuming that the measured groundwater velocity is in the vertical direction, the average specific discharge (q), i.e., river loss to the shallow braidplain aquifer, is 2.8 m d⁻¹ (ranged from 2.3 to 3.3 m d⁻¹). This was determined using the calculated average groundwater velocity (u) of 13.9 m d⁻¹ (ranged from 11.6 to 16.5 m d⁻¹) for the True Right and True Left surveys and a measured porosity of 0.2 from the drillcore material. The groundwater velocities do not vary significantly beneath the width of the active river channel (as defined by the wetted footprint of the channel). In a previous study by Vincent (2005) of the upper reaches of the Waikirikiri Selwyn River, it was suggested that the groundwater movement within the shallow braidplain aquifer would be influenced by remnant paleochannels that have formed by progressive shifting and abandonment of river channels on the alluvial fan surfaces. Furthermore, preferential groundwater flow would occur where there is a complex network of interconnected channels of more permeable gravels surrounded by less permeable sediments and would also be influenced by the heterogeneity of the aquifer material. This does not appear to be a significant controlling factor on the groundwater velocities beneath the active river channel at the study site. However, the measurement scale (0.25 m sample interval) compared to the potential scale of macropore flow ($\sim 1\text{--}2$ cm) needs to be taken into consideration.

The calculated groundwater velocities are dependent on the derived thermal conductivities. In this study, there was not a large variation in the values along the length of the cable for the two surveys and the uncertainty in thermal conductivity was relatively low (Fig. 8). However, we note that Simon et al. (2021b) and del Val et al. (2021) found that calculated thermal conductivity uncertainty increases when the conduction dominant period is short, as occurs under high flux conditions, such as those found in our study. The effects of homogeneous and heterogeneous dis-

tributions of sediment thermal conductivity on vertical groundwater flux estimates in a high-flux stream and low-flux lagoon was studied by Sebok and Müller (2019). Their study found that the vertical heterogeneous distribution of thermal conductivity did not significantly change the modelled versus observed temperature data compared to the homogeneous distribution. However, it was critical that spatially representative thermal properties are used to obtain more precise flux estimates rather than using standard textbook values. In our study, the best-fit values (lowest RMSE, which was less than $0.07\text{ }^{\circ}\text{C}$ for both surveys) for the bulk thermal conductivity of the saturated fluid-rock matrix for the True Right and True Left A-DTS transects are contained within a narrow range from 0.48 to $0.8\text{ W m}^{-1}\text{ }^{\circ}\text{C}^{-1}$. Reported thermal conductivity for greywacke basement rock (porosity = 0.03) is $1.7\text{ W m}^{-1}\text{ }^{\circ}\text{C}^{-1}$ (Mielke et al., 2016). The variations to the geometrical arrangement of layers, mineral grains, and pores will also affect the bulk conductivity, in particular, changes in porosity (Clauser, 2011). Therefore, the derived low bulk thermal conductivities from the two surveys are reasonable for the saturated fluid-rock matrix of the greywacke aquifer material that has an average porosity of 0.2 . When the porosity values were changed from 0.16 to 0.24 compared to the base case ($n = 0.2$), the calculated groundwater velocities decreased from 14.4 to 6.3 m d^{-1} .

Another consideration as to why the derived bulk thermal conductivities are low for the saturated aquifer material in this study is how the subsurface gravels and fines have collapsed and repacked around the FO cable after horizontal drilling and cable installation. This re-packing, settling of heavier gravels and infill of finer material would change the proportion of water filled voids to solid material. Unfortunately, there is no way to physically measure this change in-situ but repeat A-DTS surveys may be able to identify if there is any notable change to the derived bulk thermal conductivities along the length of the cable with time. Physical measurements of drillcore material using a thermal properties analyser would also help constrain the range in plausible values for the aquifer material under different saturated conditions.

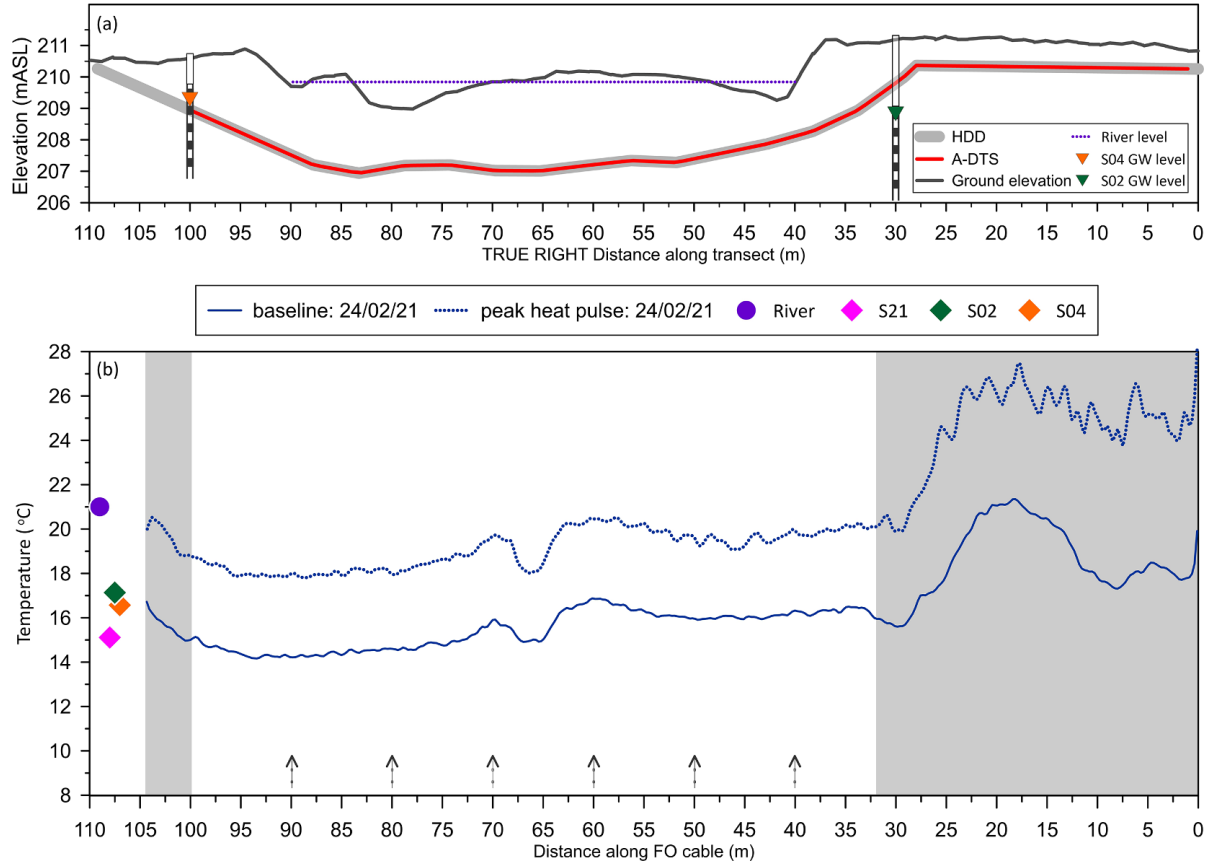


Fig. 4. (a) True Right FO cable transect and measured river and shallow groundwater levels. (b) Measured baseline and peak heat pulse temperatures on 24/02/2021. The coloured symbols are the observed temperature at the time of the survey from the river and shallow groundwater from piezometer S02, S04 and S21. Vertical arrows show locations along the FO cable where temperature breakthrough curves are presented in Fig. 6. Grey shaded area is where the FO cable is located within the unsaturated zone.

One of the more sensitive parameters in the MILS flow and heat transport model was associated with knowing the exact start time of the heating input (s) and the value of r , as the distance between the heating line source and the observing fibre (refer to equation (5)). A simple sensitivity analysis of the TR survey showed that the effect of changing the start time from 240 to 252 s changed the calculated groundwater velocity from 12.0 to 15.4 m d⁻¹ and the bulk thermal conductivity from 0.56 to 0.69 W m⁻¹ °C⁻¹. The results presented here indicate that a higher temporal resolution in the early time measurements is important to capture the rapid temperature change in response to the line source heating. Whilst changing the distance between the line heat source and the fibre (r) by 0.8 mm was very sensitive to the derived groundwater velocities, implying that knowing the

cable specifications is critical to the survey results.

The main challenge with interpreting the A-DTS data is that the inferred direction of water flow is assumed to be 1D and does not account for flow in the two other directions. Multi-direction, 3D-flow fields in the shallow streambed have been evaluated for hyporheic exchange using other active heat pulse temperature methods (e.g., Banks et al., 2018), however, there are no known studies that have been undertaken at depths greater than a few metres within river channels. It is assumed that the longitudinal flow component in the near surface of the riverbed in the direction of river flow would be greater than the vertical and lateral flow components because of the permeability structure in riverbed environments.

The installation depth of the horizontal FO cable at the study site is several metres below the influence of hyporheic zone exchange (Lewandowski et al., 2019), and therefore the derived groundwater velocities represent the dominant direction of flow within the shallow braidplain aquifer. Based on the observed hydraulic head gradient between the river and water level of the shallow braidplain aquifer as well as the measured differential river flow gauging, we can be confident that the calculated specific discharge over this river reach represents river loss.

The hydraulic conductivity of the aquifer material of the shallow braidplain aquifer was estimated to be in the range of 23 to 33 m d⁻¹. This was determined based on the range of calculated groundwater velocities (11.6 and 16.5 m d⁻¹) and the measured average hydraulic head gradient between the river and shallow braidplain aquifer, which at the time of the surveys was 0.5 m. These hydraulic conductivities are less than the values based on aquifer tests that were done at the site, which ranged from 75 to 250 m d⁻¹. The permeability of riverbed materials are known to vary over eight orders of magnitude (9×10^{-6} to 9×10^2 m d⁻¹) (Calver, 2001), and because of the significant spatio-temporal variability, this can strongly control river loss to groundwater. Aquifer test results primarily represent horizontal flow, whereas river leakage is controlled by vertical hydraulic conductivity, which is expected to be lower because of stratification and imbrication.

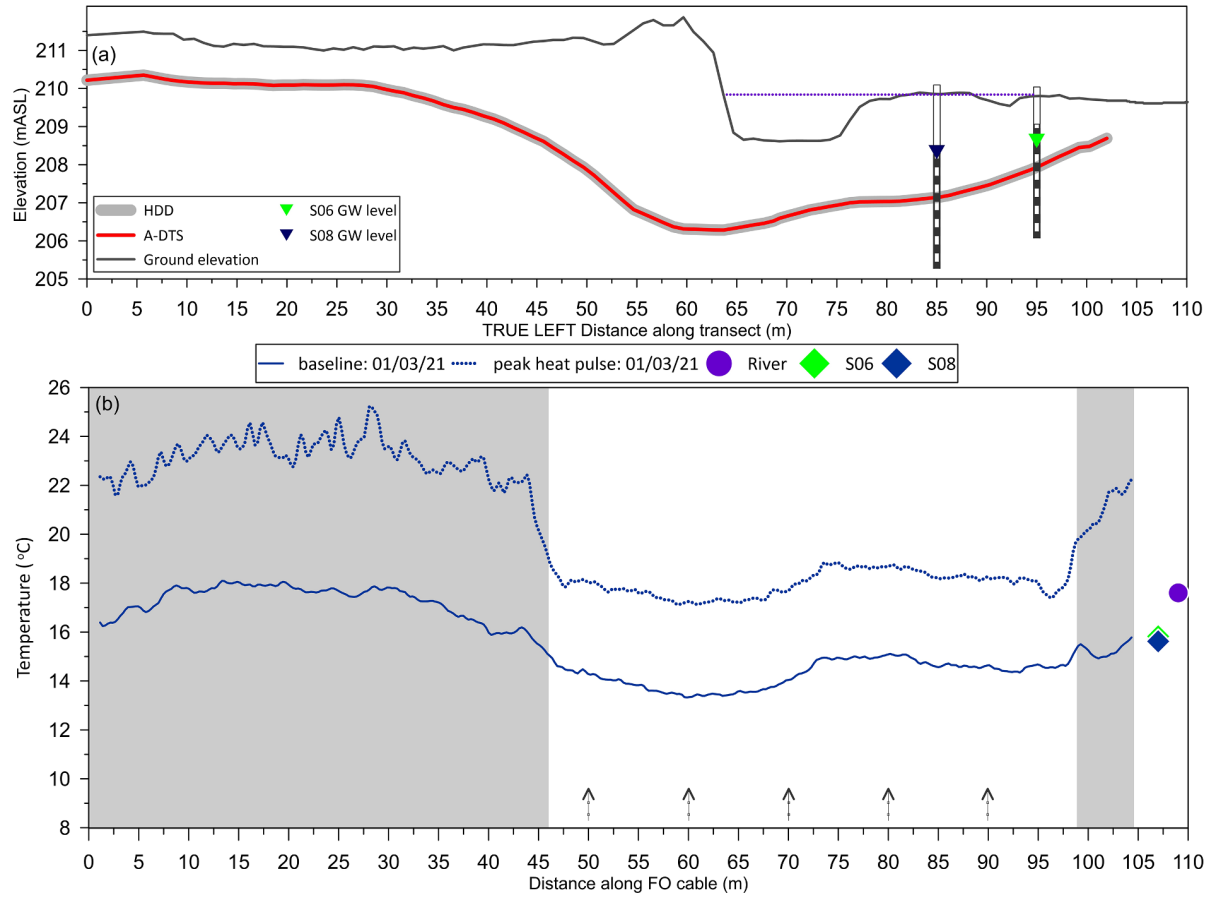


Fig. 5. (a) True Left FO cable transect and measured river and shallow groundwater levels. (b) Measured baseline and peak heat pulse temperatures on 01/03/2021. The coloured symbols are the observed temperature at the time of the survey from the river and shallow groundwater from piezometer S06 and S08. Vertical arrows show locations along the FO cable where temperature breakthrough curves are presented in Fig. 7. Grey shaded area is where the FO cable is located within the unsaturated zone.

Manual differential flow gauging that was undertaken on the 24/02/2021, using an Acoustic Doppler Velocimeter by hydrographers from NIWA, measured a river infiltration loss of $8,640 \text{ m}^3 \text{ d}^{-1}$ (average point measurement uncertainty of 3%) over a river reach length of 333 m (equivalent to $26 \text{ m}^3 \text{ d}^{-1} \text{ m}^{-1}$) that encompassed the True Right and True Left A-DTS transect locations (Fig. 1). In comparison, the measured river infiltration loss over the entire 888 m long river reach of the study area was $22,464 \text{ m}^3 \text{ d}^{-1}$ (equivalent to $25 \text{ m}^3 \text{ d}^{-1} \text{ m}^{-1}$). Using a wetted footprint width of the active river channel of 60 m and an effective distance over which the ADTS measurement can be applied that is equivalent to the diameter of the drillhole (0.125 m) and combined with the

calculated specific discharge from the A-DTS surveys (2.8 m d^{-1}), the river infiltration loss at the transect locations is $21 \text{ m}^3 \text{ d}^{-1} \text{ m}^{-1}$.

The calculated river losses using the A-DTS survey data and differential flow gauging are similar, which is an encouraging result. The scale of investigation between the two techniques is the most likely contributing factor to the different river loss calculations. Despite the low measurement uncertainty, the differential flow gauging provides a net loss of surface water over a river reach that is 100's of metres long. Whilst the A-DTS survey is measuring the groundwater velocities within the braidplain aquifer at a local scale, directly beneath the active river channel but also at a high spatial resolution along the length of the transect that is perpendicular to the river channel.

6 Summary

The novel approach that has been used in this study provides an improved understanding of the complex SW–GW exchange processes in braided river systems. The FO cable installation beneath the riverbed has demonstrated the value of A-DTS to quantify river infiltration losses at a high spatial sample interval (0.25 m) and over the entire width of the active river channel. There was some spatial variation in the calculated groundwater velocities beneath the active river channel, which ranged from 11.6 to 16.5 m d^{-1} (average = 13.9 m d^{-1}). These findings suggest that changes in the riverbed geomorphology or the presence of more transmissive paleochannels are, to a small degree, influencing preferential groundwater flowpaths in the shallow braidplain aquifer at the survey sites. The calculated average infiltration loss of $21 \text{ m}^3 \text{ d}^{-1} \text{ m}^{-1}$ from the A-DTS surveys was similar to values derived from differential flow gauging across the study site (25 to $26 \text{ m}^3 \text{ d}^{-1} \text{ m}^{-1}$).

The buried permanent installation of the FO cable allows for repeat surveys at different river stage and aquifer conditions to be conducted more frequently to evaluate how hydrologic conditions may affect groundwater velocities and recharge rates. A higher frequency of monitoring has potential to be used to identify critical river flow conditions when the highest groundwater recharge occurs, thus providing valuable information to water management decisions.

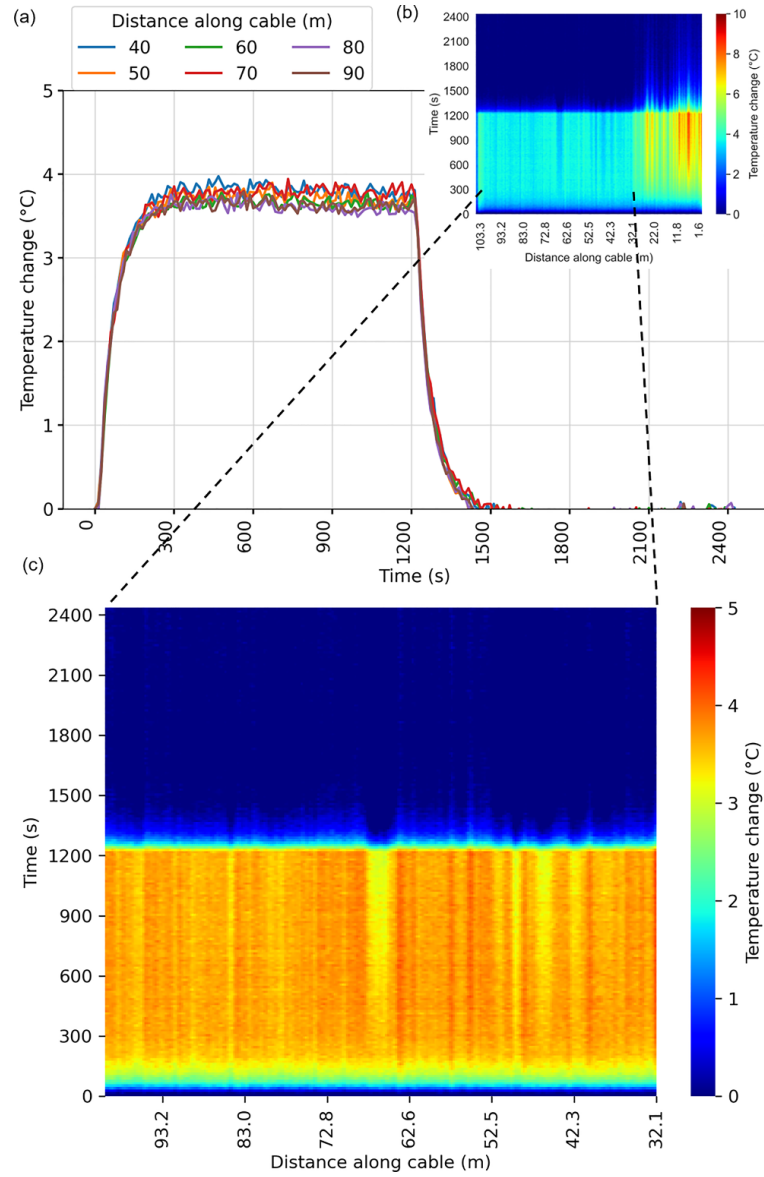


Fig. 6. (a) Temperature breakthrough curves from FO cable locations: 40, 50, 60, 70, 80 and 90 m of the True Right transect for the survey conducted on 24/02/2021. (b) Temperature change heat map along the full length of the FO cable. (c) Temperature change heat map between 32.1 and 100.0 m where the FO is below the standing water level of the shallow unconfined aquifer. Note the different temperature scale in b and c.

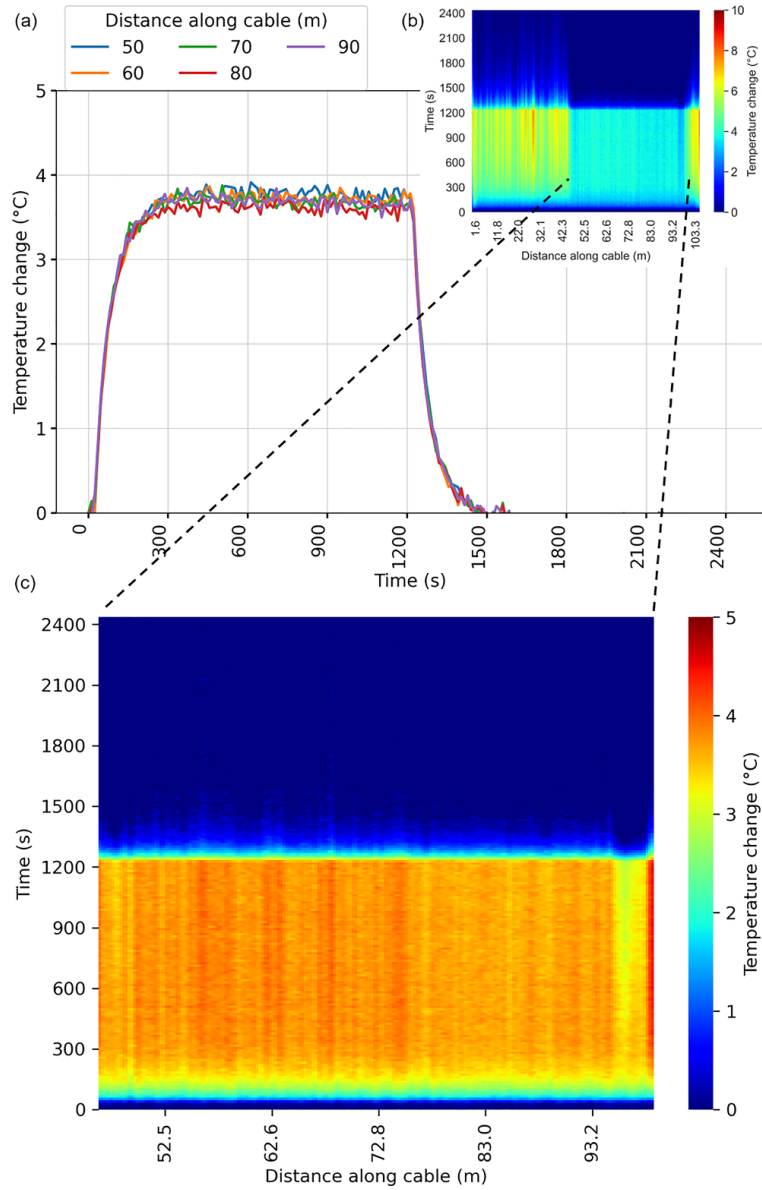


Fig. 7. (a) Temperature breakthrough curves from FO cable locations: 50, 60, 70, 80 and 90 m of the True Left transect for the survey conducted on 01/03/2021. (b) Temperature change heat map along the full length of the FO cable. (c) Temperature change heat map between 46.1 and 98.7 m where the FO is below the standing water level of the shallow unconfined aquifer. Note the different temperature scale in b and c.

6.1 Future recommendations

To improve the conceptualisation of river infiltration loss in braided river systems, the significance of 1D versus 2D flow components beneath the river needs to be addressed. To address the uncertainty in the derived groundwater velocities and to explore the phenomena of multi-directional flow, two

vertical A-DTS FO cables have recently been installed on the riverbank adjacent to the horizontal drillholes to a depth of 30 m. Planned future surveys will hopefully provide greater insight into the complex flow fields beneath the river and resolve whether the vertical or lateral flow component is more dominant at this study site.

Further work is required to compare the different analytical approaches of Bakker et al. (2015) to those of other studies, including the ADTS Toolbox by Simon and Bour (2022) and how they align to different fibre and heating cable configurations. Furthermore, how this approach relates to the analytical solutions described by others, including Diao et al. (2004) and del Val et al. (2021).

The installation of the horizontal FO cables was a very costly field investigation method. However, this type of installation provided very high spatial resolution of groundwater velocities beneath a complex braided river system. It has also allowed for multiple surveys to be conducted during different flow conditions with a reduced risk of losing equipment due to changing river flow conditions. This was demonstrated in June 2021, when the installation survived a significant 1/50-year flood event where the river level gauge station and eight piezometers were damaged and lost. Time series data from a high-resolution vertical DTS profile used in the study by Vogt et al., (2010) observed sporadic high infiltration rates at times of high water tables. The installation described in our study is of considerable benefit to being able to measure the dynamic behaviour of groundwater recharge beneath complex braided river systems and improve modelling efforts of SW–GW exchange. Future planned research activities will provide new insights to how the controlling factors of river stage and channel elevation relative to watertable elevation of the shallow braidplain aquifer changes river infiltration loss prior to and after a large flood event.

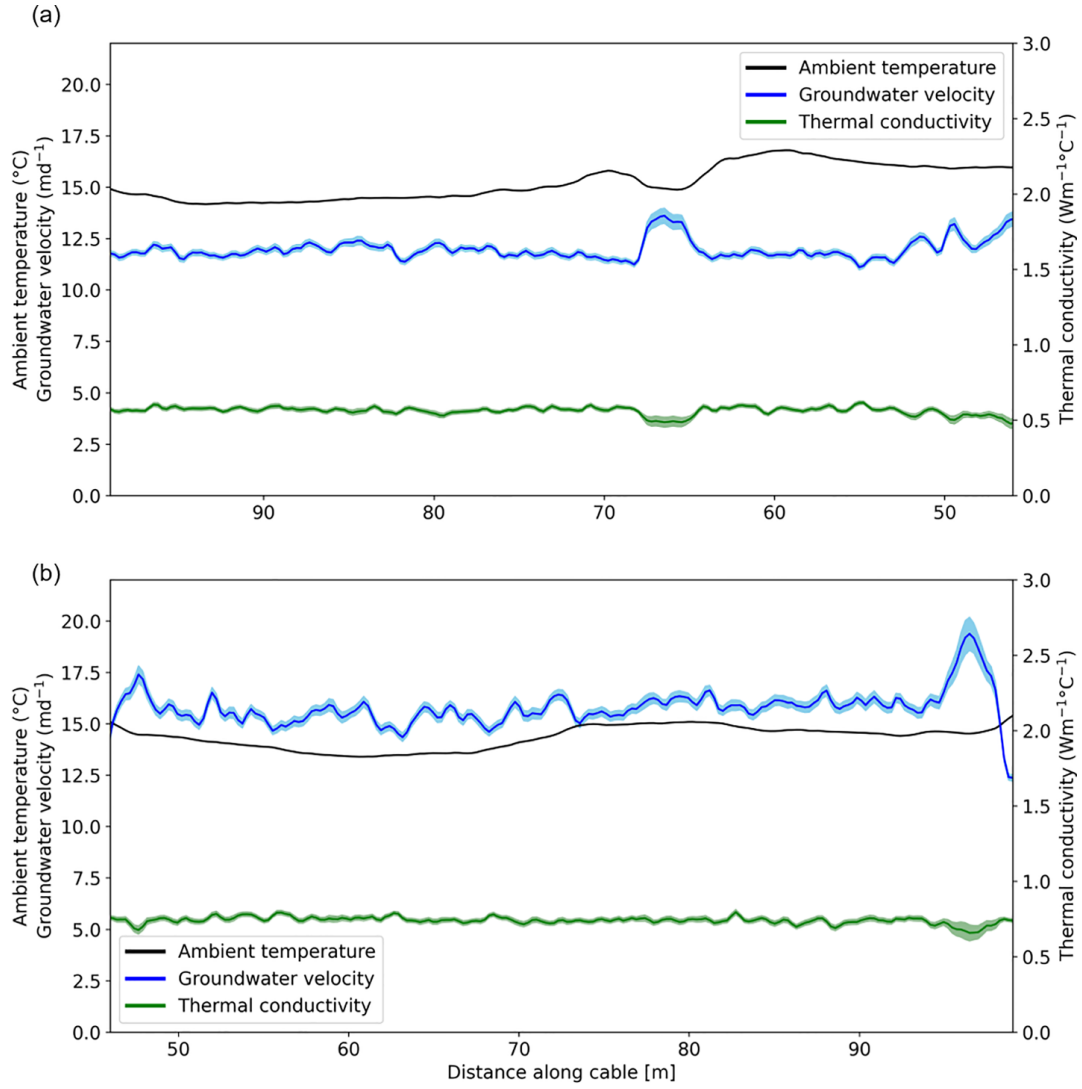


Fig. 8. Calculated groundwater velocities and thermal conductivity values for the (a) True Right on 24/02/2021 and (b) True Left on 01/03/2021. Error shading is based on standard deviation of the fitted parameter values determined by the linear least-squares fitting algorithm.

Table 1

Sensitivity analysis of input parameters used in the MILS model for the True Right survey only.

Scenario	Input parameters						Results	
	r	n	c_s	ρ_s	c_w	ρ_w	u (m d ⁻¹)	λ (W ⁻¹ m ⁻¹ °C)
	(m)	(-)	(J ⁻¹ kg ⁻¹ °C)	(kg m ⁻³)	(J ⁻¹ kg ⁻¹ °C)	(kg m ⁻³)	average (range)	average (range)
Base case	0.003	0.2	800	2650	4100	1000	12.0 (10.8, 13.6)	0.56 (0.48, 0.62)
Low n	0.003	0.16	800	2650	4100	1000	14.4 (13.0, 16.3)	0.56 (0.48, 0.62)
High n	0.003	0.24	800	2650	4100	1000	6.3 (5.6, 7.1)	0.56 (0.48, 0.62)
Low c_s	0.003	0.2	640	2650	4100	1000	10.3 (9.3, 11.7)	0.56 (0.48, 0.62)
High c_s	0.003	0.2	960	2650	4100	1000	13.5 (12.2, 15.3)	0.56 (0.48, 0.62)
Low r	0.0024	0.2	800	2650	4100	1000	9.6 (8.7, 10.9)	0.56 (0.48, 0.62)
High r	0.0036	0.2	800	2650	4100	1000	14.4 (13.0, 16.3)	0.56 (0.48, 0.62)

Other types of installation of FO cable or individual temperature sensors at an oblique angle beneath the river channel are also being considered to provide a simpler solution to be able to measure river infiltration losses during different flow events at several locations and at a reduced cost.

Declaration of Competing Interest

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

Data availability

Data will be made available on request.

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Appendix A. Supplementary data

Supplementary data to this article can be found online at <https://doi.org/10.1016/j.jhydrol.2022.128667>.

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