

Research papers

Testing the reproducibility of active-distributed temperature sensing for measuring groundwater specific discharge beneath a braided river

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ABSTRACT

Braided rivers are a major contributor of groundwater recharge, yet little is known about how recharge rates vary in time. Existing methods for estimating groundwater recharge from rivers (i.e., river loss) are inadequate for studying highly heterogeneous braided river systems at a sufficient spatiotemporal resolution. To do so, active-distributed temperature sensing (A-DTS) is employed, which combines fibre optic temperature measurements with an active heat source, enabling high-resolution quantification of water fluxes. In this study, twelve successive A-DTS surveys were conducted during a 24-hour experiment on a 100 m horizontal subsurface hybrid fibre optic cable installed at 5 m depth beneath a braided river. The experiment was carried out under conditions where the river stage and flow were relatively stable to demonstrate the reproducibility and effectiveness of the A-DTS method for measuring groundwater specific discharge. This foundational work will provide a high level of confidence in the method for future studies aimed at evaluating temporal variations in groundwater recharge. The median groundwater specific discharge values calculated over the 24-hour period had a very narrow range from 3.5 to 4.0 m d⁻¹ across the wetted footprint of the river, which is within the measurement error of the installation (6 %), indicating relatively stable groundwater recharge during the experiment. This provides confidence in the repeatability of the A-DTS method as an effective technology for quantifying river loss over longer time periods, to understand seasonal variability of groundwater recharge in braided river systems.

1. Introduction

Over the past two decades, surface water–groundwater (SW–GW) interaction has been widely researched and increasingly recognised as having a profound interaction (e.g., Acworth et al., 2015; Brunner et al., 2017; Kalbus et al., 2006; Rau et al., 2010). This includes investigating river loss to underlying aquifers, which is a major source of groundwater recharge at the global scale (Döll & Fiedler, 2008). Despite numerous studies aimed at quantifying river loss to groundwater (e.g., Cook, 2015; Min et al., 2020; Watson et al., 2021), relatively little is known about this process in braided rivers (Coluccio & Morgan, 2019).

Braided rivers are found in diverse regions around the world, including the European Alps, the Himalayas, Alaska, Canada, Russia, Japan and New Zealand (Coluccio & Morgan, 2019; Tockner et al., 2006). They typically occur in areas of high sediment supply, such as alluvial fan deposits, transporting sediment and water down gradient.

They are characterised by a complex network of multiple interconnected channels that meander across a wide braidplain, forming a braided pattern separated by gravel bars (Lane, 1957). Characteristic traits include channel avulsion, margins which expand and contract with varying sediment storage and highly heterogeneous geomorphology (Ashmore, 2013; Hicks et al., 2021; Piégay et al., 2004). Due to the constantly evolving nature of braided rivers and their highly variable flow regime, conducting physical hydrological measurements is extremely challenging (Coluccio & Morgan, 2019).

Braided rivers are crucial freshwater sources in areas where they are situated, contributing to the overall water availability and ecological balance of these regions (Piégay et al., 2006). It is estimated, for example, that the braided Wairau River in the South Island of New Zealand contributes approximately 94 % of groundwater recharge to the underlying Wairau aquifer, which is used extensively for irrigation and drinking water (Wöhling et al., 2018). However, due to limited

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understanding of the underlying mechanisms of SW–GW interaction in braided rivers as well as the challenges in instrumenting these environments, the estimation of river loss to groundwater pose significant uncertainty in water balance models and introduces significant ambiguity in water allocation decisions (Close et al., 2014).

Existing methods to estimate groundwater recharge from rivers fail to characterise SW–GW interactions at sufficient spatial resolution to examine highly heterogeneous braided river settings. Current methods to estimate groundwater recharge from rivers include infiltration experiments, natural tracer methods, water balance models, differential gauging, groundwater dating and using satellite imagery (Coluccio, 2018; Coluccio & Morgan, 2019; Di Ciacca et al., 2023; Kalbus et al., 2006; Shanafield & Cook, 2014; Verhagen et al., 2022). However, these methods often assume uniform groundwater discharge through the riverbed, which does not hold true in braided river systems (Becker, 2004; Xie et al., 2016).

Just as it is challenging to measure spatial variability in groundwater recharge from rivers, temporal variability is hard to quantify due to insufficient long-term data, complexity of interactions, spatial variations, and restrictive technical challenges with long-term monitoring equipment. Many of the existing methods for estimating groundwater recharge only provide reliable recharge rates with a temporal resolution in the order of hours to days, but often accompanied by a trade-off in spatial resolution (e.g., Cranswick & Cook, 2015; Shanafield & Cook, 2014). While hydraulic head measurements and observation well chemistry methods may be able to quantify exchange fluxes over longer time periods on the order of hours to months, measurements are inferred between discrete points, potentially overlooking the heterogeneity inherent to braided river systems. Further, often these estimates are simply long-term averages and are not at sufficient temporal resolution to study seasonal changes in dynamic braided river systems. For braided rivers, in particular, Di Ciacca et al. (2023) proposed that more work needs to be done investigating the spatio-temporal variation of the SW–GW interaction to be able to understand how river loss changes due to variation in river discharge and the wetted footprint of the river.

Heat as a tracer was used originally by Suzuki (1960) and Stallman (1963 & 1965) as a proxy for groundwater flow, using analytical solutions of heat transport equations to calculate groundwater velocity. Recent technological advancements enable the use of fibre optic DTS to collect high resolution data of passive, natural temperature fluctuations over time (Selker et al., 2006; Shanafield et al., 2018; Tyler et al., 2009). Fibre optic cables to conduct DTS can be installed in vertical, open boreholes (Banks et al., 2014; Leaf et al., 2014), in sealed boreholes (Maldaner et al., 2019; Munn et al., 2020) and installed using direct push drilling methods (Bakker et al., 2015; Des Tombe et al., 2019) so the fibres are in contact with the surrounding aquifer material. Temperature profiles from vertical sealed boreholes and direct push installation can be used to monitor dominantly horizontal groundwater flow processes, whereas open boreholes intercept vertical flow. Horizontal fibre optic cables have also been installed just below the sediment–water interface to investigate SW–GW interactions and hyporheic exchange (Gaona et al., 2019; Gilmore et al., 2019; Le Lay et al., 2019; Lowry et al., 2007). Busato et al. (2019) was the first to use a horizontal sub-surface installation of fibre optic cable to conduct DTS measurements 5 m below Vermigliana Creek in Italy to characterise SW–GW interaction.

A-DTS involves combining DTS with a heating source and allows water fluxes to be quantified at a high spatial resolution (Simon et al., 2021; Simon et al., 2022). There have been several studies that used A-DTS to measure groundwater fluxes (i.e., specific discharge), these surveys have been configured differently and in unique hydrogeological settings (Del Val et al., 2021; Selker & Selker, 2018; Simon et al., 2022) or using sandbox experiments and numerical simulations (Simon et al., 2020; Simon et al., 2023b). The sandbox experiments have confirmed that the resultant temperature elevation over time from A-DTS is sensitive to flow conditions and can be used to characterise the temporal dynamic of groundwater flux variations (Simon et al., 2023b).

Consequently, this prompts an inquiry into the prospect of applying A-DTS to quantify groundwater flux variations over time in different hydrogeological settings, such as from braided rivers. Banks et al. (2022) recently described the novel application of two 100 m horizontal, sub-surface A-DTS installations at a depth of around 5 m beneath the braided Waikirikiri-Selwyn River in the South Island of New Zealand. This was a unique installation in a complex braided river system composed of coarse unconsolidated alluvial gravels. The system has been used to estimate losses to groundwater at high spatial resolution. While the work of Banks et al. (2022) provides a snapshot of river loss to groundwater, there is also significant potential for assessing temporal changes in river loss to groundwater, via repeat surveys. However, to address this question it is important to first assess whether the A-DTS equipment has sufficient temporal sensitivity to be able to distinguish between a measurable change in groundwater recharge (i.e., flux) and equipment measurement uncertainty. To date, this type of assessment has not been carried out within a natural field setting and the usefulness of A-DTS for measuring temporal variation in groundwater recharge is unknown.

The main objective of this work is to test the reproducibility (i.e., repeatability) of groundwater flux measurements made in the field using the A-DTS method. We hypothesize that the A-DTS method deployed in a complex field setting has the required sensitivity to measure small changes in groundwater flux. To test the reproducibility of the A-DTS method, we carried out a series of surveys using a 100 m length of hybrid fibre optic cable installed approximately 5 m beneath the braided Waikirikiri-Selwyn River. Specifically, twelve successive A-DTS surveys were conducted over a 24-hour period to calculate groundwater flux. Each 2-hour survey involved applying a heat pulse for one hour and measuring the temperature along the cable every 12 s at a spatial sample interval of 0.25 m. The experiment was carried out under conditions where the river stage and flow were relatively stable to demonstrate the reproducibility of the A-DTS method for measuring groundwater specific discharge. The method provides a framework for future work that aims to quantify temporal changes in river loss over longer time periods, to understand seasonal variability.

2. Study site

The Waikirikiri-Selwyn River is situated in the Canterbury region, which is in the South Island of New Zealand (Fig. 1a). It is considered a benchmark alluvial river system in a relatively natural state and it has been studied extensively (Larned et al., 2008; Larned et al., 2015; McKerchar & Schmidt, 2007; Rupp et al., 2008; Vincent, 2005). More recently research has focused on quantifying river losses to the underlying shallow braidplain and regional aquifer systems using various methods including radon-222 (Robb, 2021; Songola, 2022), A-DTS (Banks et al., 2022), and a combination of satellite imagery and machine learning (Di Ciacca et al., 2023). Groundwater flow in the braidplain gravels occurs subparallel to the river flow direction (Burden, 1984). A detailed description of the study site is provided in Banks et al. (2022), Di Ciacca et al. (2023) and Sai Louie et al., (2023b).

The study site was selected as it is a relatively low energy braided river system, making it an ideal location for long-term instrumentation. It represents an ephemeral losing reach with one or two active channels. Seasonally, a drying front is encountered approximately 1.3 km to 12 km downstream of the study site as detailed by Di Ciacca et al. (2023).

The river contemporary braidplain at the study site is approximately 400 m wide, and is comprised of alluvial gravels of the Springston Formation (Forsyth et al., 2008). The contemporary braidplain is laterally bound by terraces of older Burnham Formation glacial outwash gravels, which are approximately 5 m in height. Both of these deposits are greywacke-derived sediments with distinct hydrogeologic properties (Forsyth et al., 2008).

An analysis of the hydrogeologic properties at the study site provide some insight into the distinct characteristics of the shallow braidplain

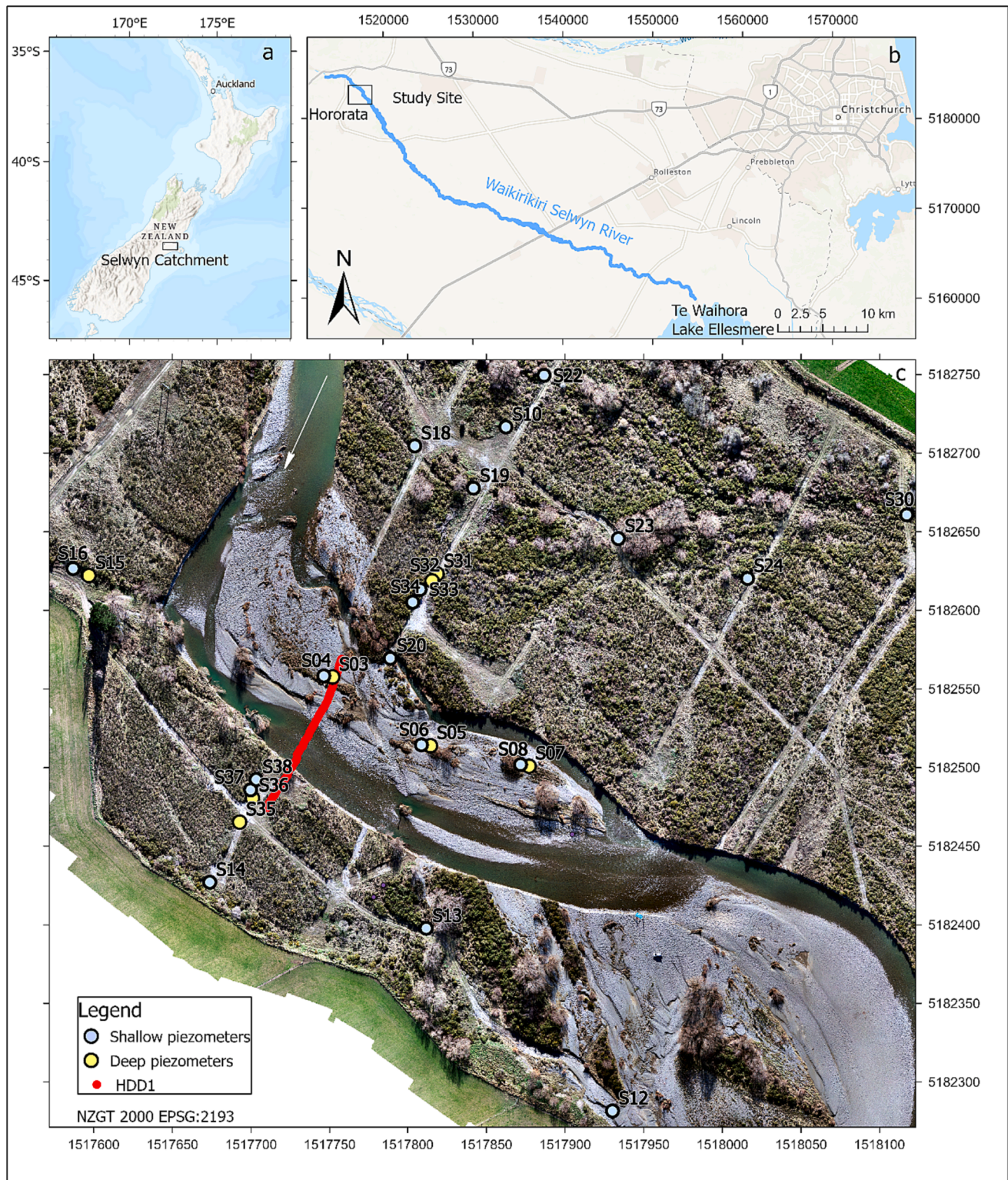


Fig. 1. (a) Location of Waikirikiri-Selwyn River catchment, New Zealand, (b) Waikirikiri-Selwyn River catchment and (c) Study site and the location of the buried A-DTS cable HDD1 and the network of shallow and deep piezometers.

gravels. The hydraulic conductivity of the shallow braidplain gravels estimated from previous A-DTS surveys range from 23 to 33 m d⁻¹ for vertical flow (Banks et al., 2022). On-site aquifer tests suggest hydraulic conductivity values between 75 and 250 m d⁻¹ for horizontal flow (Banks et al., 2022). Based on pumping tests, the transmissivity of the shallow braidplain gravels is estimated to be 1200 m² d⁻¹. A schematic showing the braidplain gravels and monitoring infrastructure (detailed

later) is presented in Fig. 2.

There is a network of infrastructure for monitoring surface water and groundwater across the site (Fig. 1). The network comprises thirty piezometers, including twenty-three shallow piezometers installed in the braidplain gravels (screened between 1 m and 6.8 m depth) and seven deep piezometers (screened between 12.1 m and 35 m depth). Surface water monitoring includes stage height and discharge at the Beattys

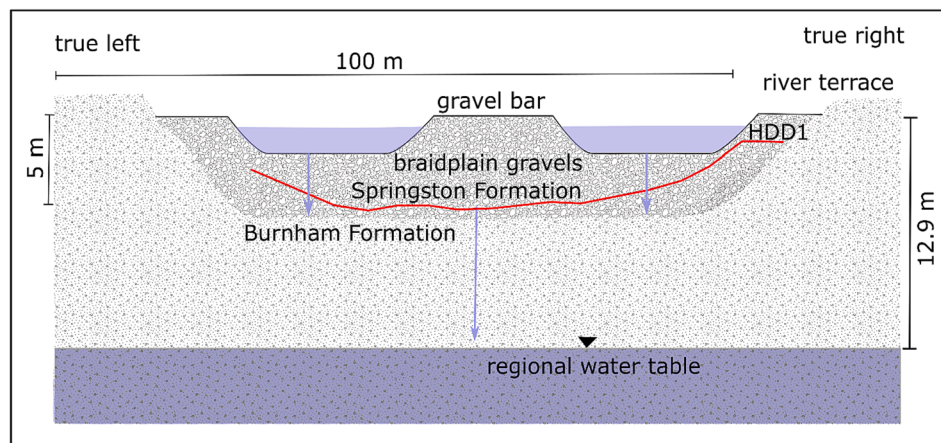


Fig. 2. Schematic showing the approximate location of the A-DTS cable (termed HDD1) beneath the Waikirikiri-Selwyn River.

Road gauge (200 m upstream of the site) and at Whitecliffs approximately 9.5 km further upstream. The gauge at Beattys Road was destroyed during a flooding event in May 2021 and subsequent replacement gauging stations were damaged on several occasions causing many gaps in the seasonal flow record. Therefore, an empirical relationship between the Whitecliffs gauge and Beattys Road was developed to create a synthetic flow record for the Beattys Road location. However, we note that the synthetic flow record is only accurate for low flows ($Q < 5 \text{ m}^3 \text{ s}^{-1}$).

3. Methods

3.1. Fibre optic installation and cable structure

The subsurface horizontal fibre optic cable (termed HDD1) was installed using a novel horizontal directional drilling method. A hybrid cable containing fibre optic cables for temperature sensing and copper conductors for heating, of approximately 100 m in length was installed to depths of around 5 m beneath the two main channels of the Waikirikiri-Selwyn River, perpendicular to river flow, as detailed by Banks et al. (2022). The location of HDD1 intercepts river loss to the underlying braidplain gravels (Fig. 2). For reference, the cable location coordinates are provided online via Mendeley Data (Sai Louie et al., 2023a).

Understanding the exact cable structure and especially the radial distance r (m) between the heating and fibre optic cable is important when estimating groundwater velocities using analytical approaches (detailed below). In some A-DTS studies, the heat source is separate from

the optical fibres shrouded in a plastic coating or metal sheath (Bakker et al., 2015; des Tombe et al., 2019), while most other studies have used a hybrid cable comprising optical fibres surrounded by the heating elements inside a plastic jacket or metal sheath (Maldaner, 2021; Munn et al., 2020; Selker & Selker, 2018; Simon et al., 2020) or optical fibres positioned adjacent to heating cables encapsulated within the hybrid cable (Banks et al., 2022; Coleman et al., 2015). To examine the sensitivity of the hybrid cable configuration on calculated velocities, Banks et al. (2022) used the moving instantaneous line source (MILS) model with fixed parameters but varying the r distance from 0.0024 m to 0.0036 m. The resultant average specific discharge ranged from 1.9 to 2.9 m d^{-1} , thus emphasising the importance of understanding the effective distance between the optical fibres and heating cables.

Precise cable dimensions were not provided by the manufacturers (Fig. 3a); therefore, this study dissected twelve short lengths of the hybrid cable perpendicular to the cable orientation and using a microscope to measure the value of r (Fig. 3b). The geometry of the hybrid cable comprises two twisted copper heating cables with a radius of 0.8 mm and a gel filled buffer tube containing four optical fibres. The measured distance r between the heating cables and the optical fibres was 0.7 mm (Fig. 3a). This triple cable bundle is contained within a metal sheath with an inner radius of 2 mm and an outer radius of 3.2 mm, encased by a polypropylene jacket with an outer radius of 5.5 mm (Fig. 3a).

3.2. Active-distributed temperature sensing experimental design

A Silixa XT-DTS was used to collect the temperature measurements

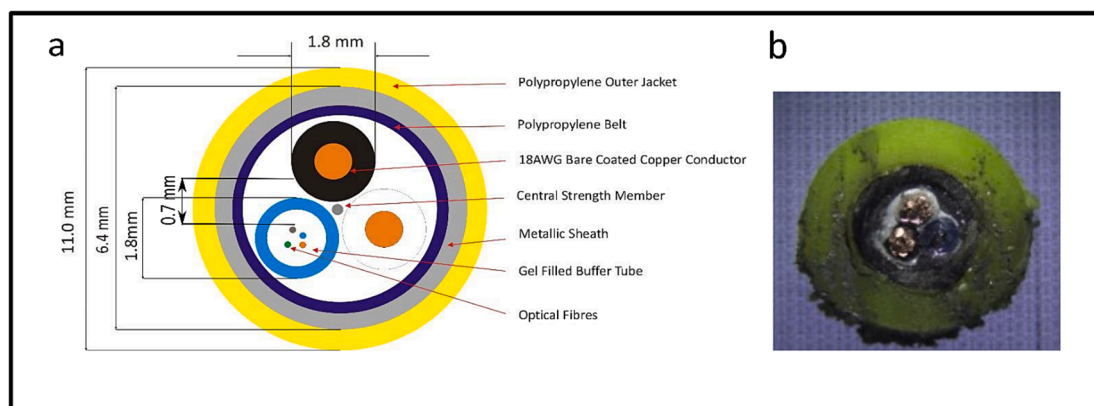


Fig. 3. (a) Manufacturers specification of the hybrid cable modified from Silixa (2020). (b) A section of the hybrid cable used in this study photographed using a microscope to verify cable dimensions and the radial distance r . Photo taken by the author.

with a resolution of 0.01 °C, a spatial sample interval of 0.25 m, and a spatial resolution of 0.54 m (Silixa, 2020). However, the effective spatial resolution in the field setting is likely to be greater than the spatial resolution of the DTS unit (Selker & Selker, 2014; Simon et al., 2020). For this reason, the effective spatial resolution was estimated using the 90 % step change method on the installed hybrid fibre optic cable and with the same DTS unit used in this experiment. This was conducted on 12 May 2021 on HDD1 with instant freeze spray applied at a discrete location on the cable and from this, the effective spatial resolution was estimated to be 0.763 m.

For calibration purposes, a section of flexible fibre optic cable (without heating cables) was left above ground. Two 8 m reference sections of cable were coiled and placed in the two insulated calibration baths, one containing ambient temperature river water and the second containing a slurry of ice water. Each bath had an independent Pt-100 temperature probe connected to the DTS unit and recording the temperature at 12-second intervals. To prevent stratification of water temperature the calibration baths were kept circulating using a small pump.

The DTS was configured with an acquisition time of 10 s, employing a double-ended differential loss correction method consisting of 5 s forward measurement and 5 s reverse measurement (Van De Giesen et al., 2012). To heat the copper conductors within the hybrid cable a Silixa heatpulse unit was used. The heatpulse configuration for each survey comprised 300 s DTS pre-measurements, followed by 3600 s heating with an applied power output of 15 W m⁻¹, and finally a 3300 s cooling duration. Considering the demonstrated thermal equilibrium in 3-hour heating surveys and the relatively high groundwater velocities in this setting, a heating period of 3600 s was deemed appropriate. This was based on a sensitivity analysis conducted whereby data from the 3-hour heating survey was shortened to various durations and the data was processed, to decide upon the optimal heating duration for this A-DTS installation in this particular setting. Based on this a 3600 s duration heating period was selected on the premise that it is sufficient for the temperature to reach thermal equilibrium, yet it is not too long so that it may pick up transient fluxes.

3.3. Surface water and groundwater monitoring infrastructure

A long-term dataset of surface water and groundwater information at the site has also been collected. Concurrently with the 24-hour A-DTS experiment, surface water and groundwater level and temperature were monitored across the site every fifteen minutes. Continuous surface water temperature was recorded using a HOBO pressure and temperature logger deployed on a cinder block located on the edge of the main channel above HDD1. Climate observations including rainfall and air temperature were also recorded at the site in fifteen-minute intervals.

3.4. Measuring thermal properties

Accurate thermal property measurements of the aquifer material from the site were obtained to improve the reliability of subsequent analysis of A-DTS results. While Banks et al. (2022) used estimated thermal properties from the literature, a sensitivity analysis highlighted the importance of accurately defining the bulk thermal conductivity and porosity. In this study, a TEMPOS Thermal Properties Analyser (METER Group Inc., 2018) was used to measure the thermal properties of the finer drill core material that was retrieved from the upper 10 m of drillhole S38 in the braidplain gravels (Fig. 1c). The thermal conductivity and resistivity was measured under both saturated conditions using the TR-3 probe and the SH-3 probe to measure diffusivity and volumetric specific heat capacity and has an accuracy of 10 % (METER Group Inc., 2018). Each unconsolidated sediment sample was tested three times, with the thermal analyser probe reinserted prior to each test. The thermal conductivity values were corrected for water, assuming that all pores were entirely saturated. It is essential to understand the thermal properties of the material surrounding the cable

when interpreting A-DTS results because they affect the heat transfer processes that control the temperature response to heat injection.

3.5. Mathematical model for estimating groundwater flux

The Hantush-Jacob analytical solution solves for energy conservation in a porous media subject to heat conduction in the effective media and advection of water through the pore space. It was introduced by Ren et al. (2000) for the problem of groundwater velocity analysis and later re-examined by Bakker et al. (2015). The governing equation in 2D is:

$$\frac{\partial T}{\partial t} = D \left(\frac{\partial^2 T}{\partial x^2} + \frac{\partial^2 T}{\partial y^2} \right) - \theta V_w \frac{(\rho c)_l}{\rho c} \frac{\partial T}{\partial x}, \quad (1)$$

Where D (m² s⁻¹) is the thermal diffusivity coefficient, T (°C) is temperature, t (s) is time, ρc (J m⁻³ °C⁻¹) is the thermal energy density, which is the product of rock density, ρ (kg m⁻³) and specific heat capacity, c (J kg⁻¹ °C⁻¹). Subscript l denotes a property of the water, θ (-) is porosity, and V_w (m s⁻¹) is the pore water velocity in the x direction. The thermal diffusivity, D , depends on the thermal conductivity, K (W m⁻¹ °C⁻¹) and energy density:

$$D = \frac{K}{\rho c} \quad (2)$$

The specific discharge, q (m s⁻¹) is defined as:

$$q = \theta V_w \quad (3)$$

The ratio $(\rho c)_l / \rho c$ in Eq. (1) accounts for water as the heat carrier during advection. It is the reciprocal of the thermal retardation coefficient, R , given by Bakker et al. (2015), i.e., $1/R = (\rho c)_l / \rho c$.

The analytical moving instantaneous line source (MILS) solution of Eq. (1), also known as the Hantush-Jacob solution, was introduced by Carslaw and Jaeger (1959) and then later adapted by Zubair and Chaudhry (1996) to solve the advection – conduction equation. It has been re-examined by others for the problem of groundwater flow (Ren et al., 2000; Bakker et al., 2015). The MILS model assumes an infinitely long constant heat source with strength Q (W m⁻¹) at the origin beginning at time t_0 (s). It describes how temperature changes at different locations (x, y) and times t (s):

$$T(x, y, t) = \frac{Q}{4\pi K} \exp\left(\frac{qx}{2DR}\right) W\left(\frac{x^2 + y^2}{4D(t - t_0)}, \frac{q\sqrt{x^2 + y^2}}{2DR}\right) \quad (4)$$

Where $W(\cdot)$ is the Hantush Well function defined:

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

The A-DTS temperature data is processed using the MILS solution to estimate the q and K along the entire cable. Parameter estimates are obtained by fitting the MILS model to A-DTS temperature data using the curve-fit function from the scipy.optimize library, which returns a locally optimal least-squares fit function, while allowing us to constrain physical quantities to be non-negative (Virtanen et al., 2020).

To remove observed linear trends in the long-term temperature signal, in addition to the MILS model, an additional fitting parameter has been introduced, dT/dt , modifying the MILS temperature model as:

$$T(r, t) = T_{MILS(r, t)} + \frac{dT}{dt} (t - t_0) \quad (6)$$

Where t_0 is the time at the beginning of the heat pulse.

The ‘skin effect’ caused by the heat transfer to the cable material is discussed in detail by Del Val et al. (2021) and Simon et al. (2021). This is initially dominated by the skin effect and conduction, followed by advection dominant heat transfer processes. The inflection point between the conduction dominant and advection dominant heat transfer

phases can be used to calculate the velocity using the MILS model (Simon et al., 2021). The ‘skin effect’ (ΔT_{sk}) was defined by Del Val et al. (2021) and described by Simon et al. (2021) and denotes the rise in temperature resulting from low thermal conductivity between the energy source and the surrounding porous media around the cable. The MILS model, Eq. (4), is used to estimate thermal conductivity and specific discharge along the A-DTS cable by finding the best fit solution to cable temperature data recorded at the 0.25 m sampling intervals. The analytical solution was also refitted for each of the successive twelve heat pulses from the A-DTS surveys. The model was designed to use the first survey as a reference to estimate independent K and D values at all locations along the length of the cable, and then these values were fixed for the subsequent eleven surveys to identify any changes in or temperature. The assumption is that the thermal properties, K and D , remain constant between surveys. Therefore, any differences observed between surveys are assumed to be due to changes in the specific discharge.

4. Results

4.1. Surface water and groundwater time series

During the 24-hour A-DTS experiment, the river discharge and groundwater level remained relatively stable. The river stage height at the Whitecliffs rated flow site ranged from 0.297 to 0.309 m (4 % variation during the experiment period) with an accuracy of 8 %. The discharge ranged from 4.02 to 4.29 $\text{m}^3 \text{s}^{-1}$ (5 % variation during the experiment period) with an accuracy of 10 %. In the shallow braidplain gravels, specifically piezometers S04, S06 and S08, the groundwater levels remained stable, with less than 0.006 m change occurring over this period (Fig. 4b). This consistency in groundwater levels reflects the relatively stable conditions observed in the river, as they are highly interconnected. At the time of the experiment, the regional groundwater depth was measured at 12.90 m in piezometer Selwyn 35. Groundwater levels one week prior to the experiment, and rainfall are shown in

Fig. 4a.

Over the course of the 24-hour experiment on HDD1, the river temperature in the main channel displayed diurnal variation with the temperature above the cable ranging from 8.5 °C to 11.0 °C, indicating a diurnal range of 2.5 °C (Fig. 4d). The lowest temperature was recorded at 07:30, while the maximum temperature occurred at 13:15. Concurrently, the air temperature ranged from 5.0 °C to 22.0 °C. Notably, a diurnal temperature signal was also observed in the shallow piezometer S08 (Fig. 4c). Conversely, the groundwater temperatures remained mostly constant, apart from S06 and S37, which increased by approximately 0.5 °C over the course of 24-hours possibly reflecting the increase in river temperature (Fig. 4d). Groundwater temperature and river temperature one week prior to the experiment provide insight into the baseline conditions prior to the experiment (Fig. 4c).

4.2. Active-distributed temperature sensing

The A-DTS results provide high temporal resolution data, indicating that there was very little change in temperature responses during the 24-hour experiment. Measured temperature increase (i.e. peak heat pulse temperature minus the ambient temperature) from the 24-hour experiment are shown in Fig. 5a, for the section of the cable that is within the saturated zone of the braidplain gravels between 31.65 m and 103.59 m. Within the saturated zone the temperature increases by up to 4.7 °C during a heating pulse; both conductive and advective flow processes contribute to heat transfer in the saturated zone. Fig. 5a illustrates the small temporal variation in temperature that occurred during the 24-hour experiment. This is evident by the small range in temperatures at each location along the cable, as shown in Fig. 5a. For reference, the raw A-DTS dataset is available online (Sai Louie et al., 2023b).

The measured temperature at nominal distances along the cable for the duration of the 24-hour experiment is shown in Fig. 5b. The observed changes in temperature are generally consistent across the twelve surveys at any given location along the cable (Fig. 5b). An exception to this

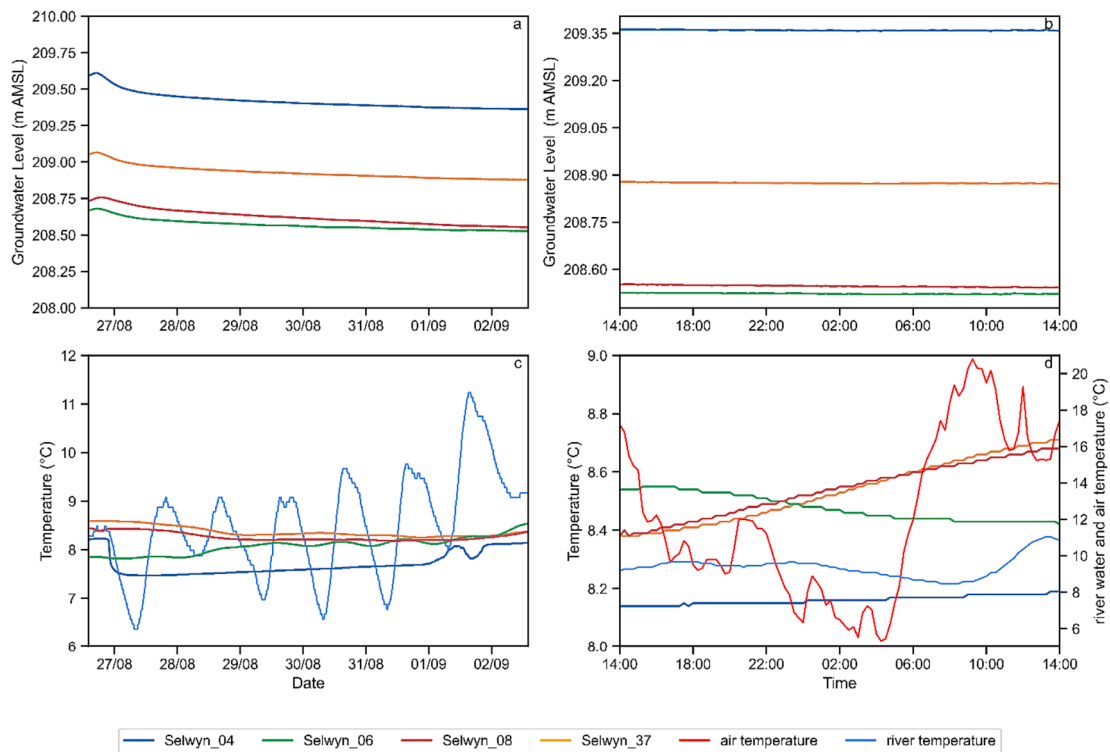


Fig. 4. (a) Groundwater level data from one week prior to the experiment. (b) Groundwater level data for the duration of the 24-hour experiment between 14:00 02/09/2022 and 14:00 03/09/2022. (c) Groundwater and river temperature from one week prior to the experiment. (d) Groundwater, river and air temperature for the duration of the experiment. Groundwater level and temperature data is from selected shallow piezometers located close to HDD1.

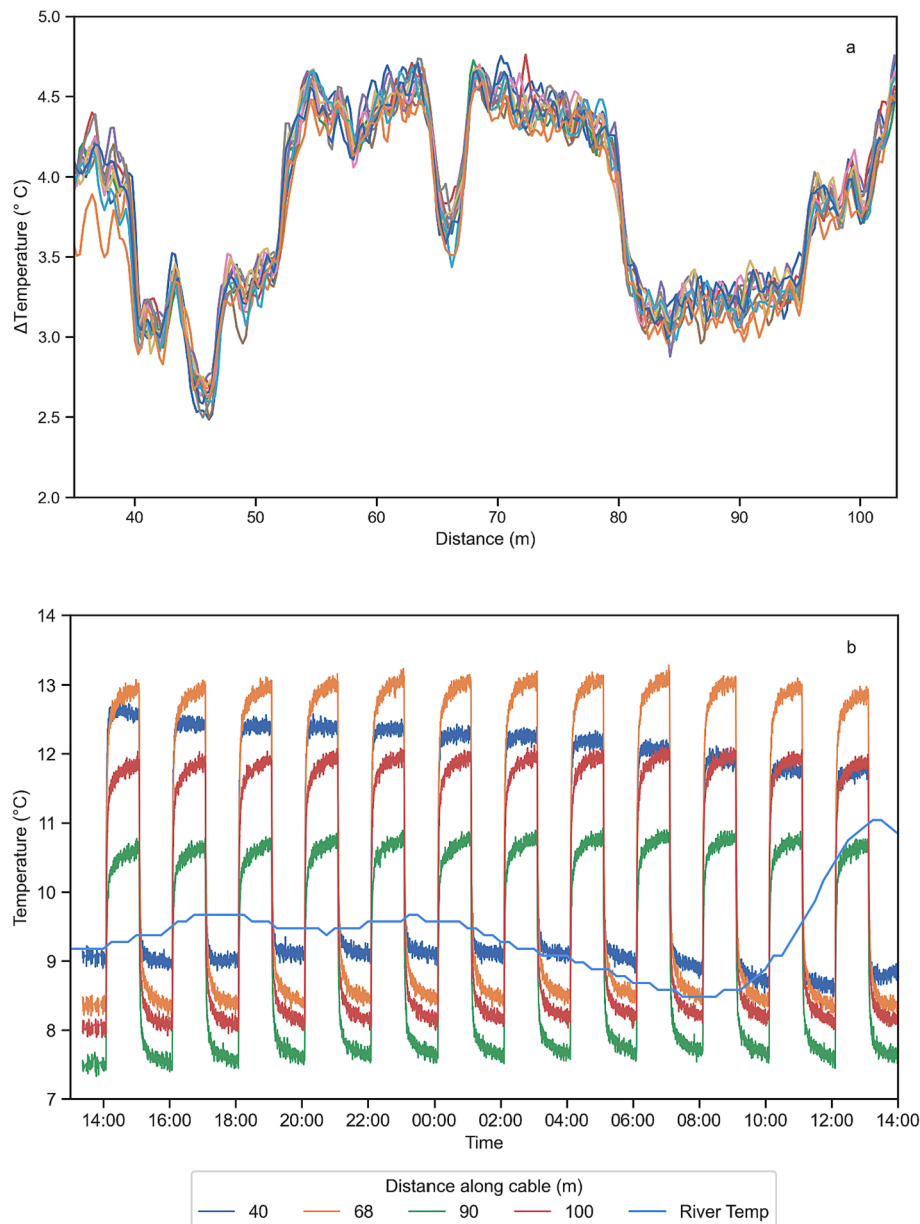


Fig. 5. (a) Measured ΔT for each of the twelve surveys conducted during the 24-hour experiment on HDD1 along the saturated portion of the cable from 31.65 to 103.59 m. (b) A-DTS temperature response during the 24-hour experiment at nominal distances along the cable: 40, 68, 90 and 100 m. The blue line represents the measured river temperature in the true right river channel, and directly above the cable at a distance of approximately 40 m along HDD1. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

occurs at 40 m along the cable where, over the course of the 24-hour experiment the maximum heat pulse temperatures reached at 40 m along the cable gradually decreased, despite the same amount of power being applied (15 W m^{-1}). For example, during the first heating cycle the temperature reaches 12.7°C whereas during the final heating cycle the temperature only reaches 11.9°C (Fig. 5b). This may suggest a modest increase of the specific flux at this location over the 24-hour period. Alternatively, this could be caused by a lag between the decreasing river temperatures, between 00:00 and 08:30, and the water intercepting the cable at 40 m along the cable, which appears to be an approximately 6-hour delay.

4.3. Analytical model results

The MILS solution was fitted to each temperature breakthrough curve at 0.25 m spacing along the cable and for each of the twelve A-DTS

surveys. The main output of this modelling exercise was a series of estimates of specific discharge, varying in time and space. Fig. 6a and 6b presents examples of the fitted data from survey 6, which was conducted in the middle of the 24-hour experiment, running from 00:00 to 02:00 on 03/09/2022, when the temperature contrast between the surface water and groundwater was at its peak. The black dots show the measured change in temperature since the beginning of the heat pulse at 40 m (Fig. 6a) and at 80 m (Fig. 6b) along the cable. The best fitting MILS model is shown in green and the de-noised data are shown by the red dashed line (following the procedure of Ramos et al. (2017)). Once the temperature stabilises, there is a good fit between the measured temperature data and the model.

The temperature increase observed in the first 600 s is likely due to the skin effect. This is evident from the characteristic bell curve shape of the $dT/d\ln t$ shown in the semi-log plots presented in Fig. 6a and 6b. We identified that for longer heating periods, initially there was an

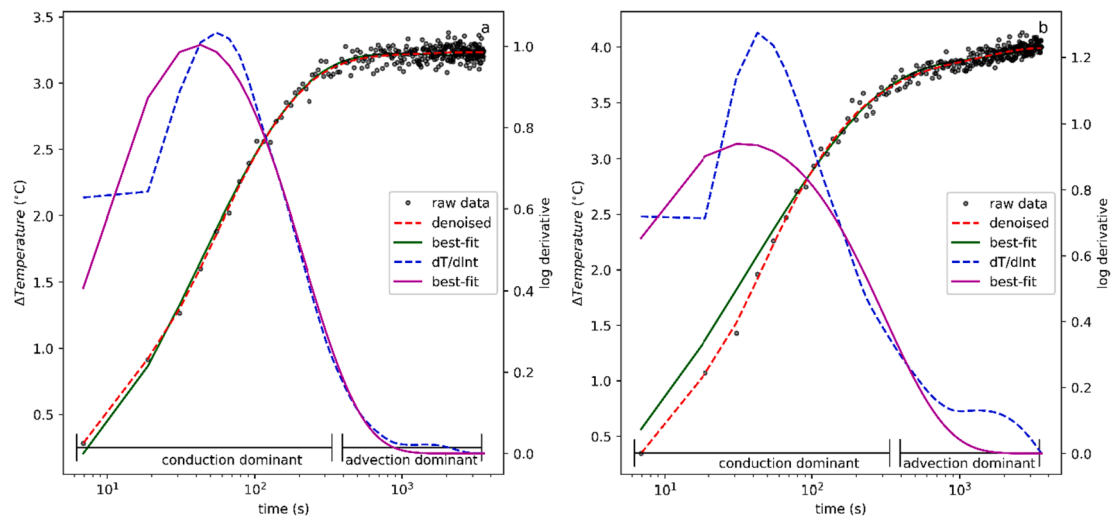


Fig. 6. Fitting the MILS solution to the temperature breakthrough data and the thermal response curve for survey 6, which was conducted from 00:00 – 02:00 03/09/2022 (a) measured at 40 m along the cable and (b) at 80 m along the cable.

approximately logarithmic temperature increase, which likely corresponds to the expected conductive heat transfer response. Subsequently, the temperature increase rapidly slows as the effect of groundwater flow stripping the heat begins to dominate. However, at many locations along the cable, the observed temperature did not stabilize at an equilibrium

value. Instead, a consistent linear trend was observed, either positive or negative. For times greater than 800 s, the heat transfer is advection dominant.

The average specific discharge values over the 24-hour experiment period ranged from 3.2 to 3.4 m d^{-1} at a nominal distance of 54 m along

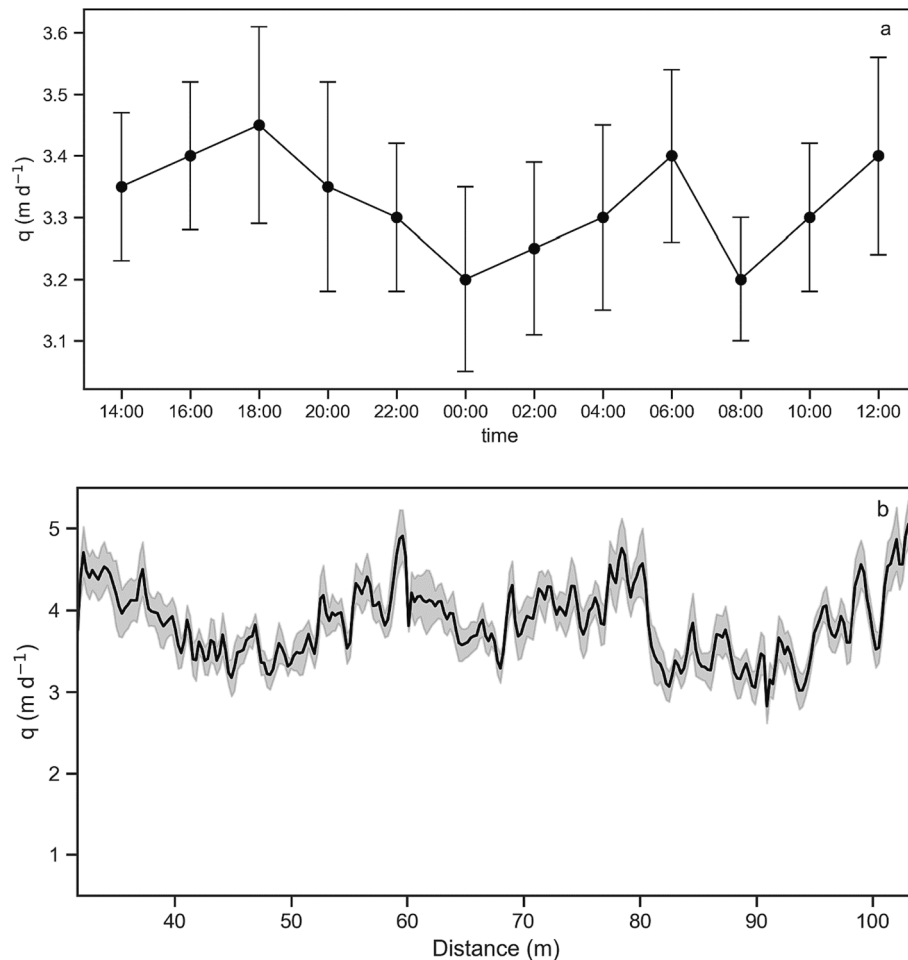


Fig. 7. (a) Calculated specific discharge for surveys S1 to S12 at 54 m along the cable conducted over the 24-hour experiment using the Hantush-Jacob solution. (b) Calculated average specific discharge along the saturated length of cable between 31.65 m and 103.59 m for survey S6 from 00:00 to 02:00 03/09/2022.

the cable (Fig. 7a). Error bars showing one standard deviation of variability were obtained from the covariance matrix for the fitted model parameters assuming an error in the temperature measurement of 0.1 °C. The small change in magnitude of the calculated specific discharge over the 24-hour period when the river and groundwater were relatively stable demonstrates the reproducibility of A-DTS derived estimates of river loss in braided river systems. Therefore, the level of uncertainty expected from this method equates to 6 % of the derived specific discharge values (Fig. 7a). To calculate the uncertainty, first, the relative error during each survey, at each point along the cable was calculated. Then, this was multiplied by the relevant specific discharge value to calculate the absolute error.

There was notable spatial variability in the calculated specific discharge beneath the river (Fig. 7b). For survey S6, specific discharge ranged from 3.1 to 5.0 m d⁻¹ and the lowest specific discharge values occurred between 81.2 m and 94.7 m along the cable, where the cable was beneath the gravel bar (Fig. 7b). Error is shown in grey shading and represents the absolute error. The low error is because of a good fit between the model and the measured temperature data.

4.4. Thermal properties

The thermal properties of the stony glacial sediment material were measured in the lab on core material taken from the upper 10 m of S38. Based on the average of 24-measurements, this material had a saturated K of 2.47 W m⁻¹ °C⁻¹. Additional thermal properties that were measured in the lab include specific heat capacity, c of 1.52 kJ kg⁻¹ °C⁻¹ and a density, ρ of 1,770 kg m⁻³, which equates to a sediment heat capacity, ρ of 2.69 MJ m⁻³ °C⁻¹.

Thermal conductivity values were also estimated from the 24-hour A-DTS experiment by fitting the MILS solution to invert the thermal conductivity values. The K values estimated from the 24-hour A-DTS experiment ranged from 0.40 to 2.23 W m⁻¹ °C⁻¹ (average 1.27 W m⁻¹ °C⁻¹). Although the thermal conductivity estimated from the A-DTS surveys varies along the cable, when refitting an independent value of K for each pulse in the experiment, we observed minimal temporal variation in the values over the 24-hour period. This is expected, because there is no obvious mechanism for thermal conductivity to change over such short time intervals. However, a difficulty with using the MILS solution to estimate K is that the values it returns represent some unknown combination of the thermal conductivity of the cable material, the skin region around the cable and the pristine porous media. The cable that was used is encased in polypropylene, which has a low K of approximately 0.12 W m⁻¹ °C⁻¹ (Cengel, 2015) and this would be likely to contribute to the lower inferred K values from the A-DTS experiment compared to those measured on core samples.

5. Discussion

The average specific discharge for the 24-hour experiment within the saturated zone was 3.8 m d⁻¹, over a wetted footprint of 72 m (31.6 m to 103.59 m along the cable). This equates to a vertical discharge of 274 m² d⁻¹ (or 274,000 m³ d⁻¹ km⁻¹) and a total river infiltration loss of 242,957 m³ d⁻¹ across the 888 m long study site. Based on raw data from Banks et al. (2022) using an average specific discharge value of 2.8 m d⁻¹ from the survey on 24 February 2021 on HDD1, over a wetted footprint of 60 m, results in a vertical discharge through the riverbed (adjacent to the cable for 1 m³) of 168 m² d⁻¹. For comparison this equates to a total river infiltration loss of 149,184 m³ d⁻¹ across the 888 m long study site, which is approximately 38 % less than the river infiltration loss estimated from the 24-hour experiment. The difference in these results likely reflects the seasonal variation in river loss. Di Ciacca et al. (2023) estimated a total river infiltration loss of 21,600 to 56,160 m³ d⁻¹ km⁻¹ for the period March 2020 to May 2021 which is about five times smaller than the calculated river loss derived from the A-DTS survey conducted in September 2022. Arguably, the satellite

photograph interpretation method used by Di Ciacca et al. (2023) only captures vertical fluxes whereas the A-DTS cable intercepts both vertical and lateral fluxes. This variation could also be explained by seasonal variations in groundwater recharge as the period of March 2020 to May 2021 was a period of low streamflow conditions.

The measured temperatures along the cable as well as the temperature difference for the duration of the experiment provide insight into the spatial variability of river loss. Notably, certain sections of the cable exhibit relatively low temperature changes during cable heating. These sections occur between 39 m and 52 m, and again between 80 m and 95 m, corresponding to likely river loss from the true right and true left river channels above the cable. In these locations, we hypothesize that advective heat stripping from the cable is greater. Additionally, there is a clear temperature anomaly around 68 m along the cable where groundwater flow is particularly effective at stripping away heat from the cable. The cable at this location is situated beneath the gravel bar between the two channels. Possible causes for this temperature anomaly could be erosion of the sediments around the cable in this location or the cable is transiting a paleochannel. Banks et al. (2022) also noted an abrupt temperature change at this location.

To investigate the repeatability of measuring specific discharge at each location along the cable, the relative error inherent to each measurement location along the cable has been assessed. This evaluation across the multiple surveys proved that the spatial variability in q significantly exceeds the measurement error across the cable. The average relative error of q was calculated as 6.0 % and 6.1 % for an $r = 0.0018$ m and $r = 0.0007$ m, respectively, as shown in Fig. 8a. This indicates that relative error is stable and independent, regardless of r . Therefore, the A-DTS methodology has sufficient spatial sensitivity to reliably quantify changes in specific discharge. Noteworthy this, the true uncertainty from repeated A-DTS surveys appears to be lower than the associated uncertainty for a single A-DTS survey. For instance, there is a lower accuracy of q derived from a single A-DTS survey, as presented in Fig. 8b with a level of uncertainty of up to 0.44 m d⁻¹ for survey S6. Conversely, the level of precision from the twelve repeated surveys within the 24-hour experiment have an absolute error of 0.09 m d⁻¹, as presented in Fig. 8b. This suggests that there is a decreased level of uncertainty of q associated with the repeated A-DTS surveys.

Some spatial variability in specific discharge between all twelve surveys was observed. The minimum range in specific discharge at the same location for all surveys was 0.2 m d⁻¹ at 49.9 m along the cable and the maximum range was 1.6 m d⁻¹ at 61.6 m along the cable. Interestingly, both locations are below the true right channel, which may be associated with the river geomorphology and heterogeneous nature of the shallow braidplain gravels. The standard deviation of the calculated specific discharge for each location along the saturated portion of the cable ranges from 0.06 to 0.44 m d⁻¹. This suggests that there is some change in the specific discharge at each point along the cable, this could be caused by the heterogeneity of sediments surrounding the cable. The median values of specific discharge range from 3.5 to 4.0 m d⁻¹ across the saturated portion only, between 31.65 m and 103.59 m along the cable (Fig. 9). The advection dominant process clearly occurs after 800 s of heating. Similar to Simon et al., (2021), this temperature stabilization time equates to an estimated groundwater flux of 3 to 4 m d⁻¹ and further supports the field survey results.

The probability density of specific discharge estimates is visualised using violin plots shown in Fig. 9, which display the distribution characteristics across the twelve surveys. The violin plots show a unimodal distribution with a single dominant peak in the specific discharge values, indicating consistency in the measurements across the twelve surveys of the 24-hour experiment (Fig. 9). The specific discharge distribution exhibits a slight negative skewness, indicating a tendency for lower discharge values. The presence of outliers with lower q values (but within the saturated zone) are possibly areas along the cable where the sediments surrounding the cable are packed differently.

Understanding the initial skin effect of the fibre optic cable on the

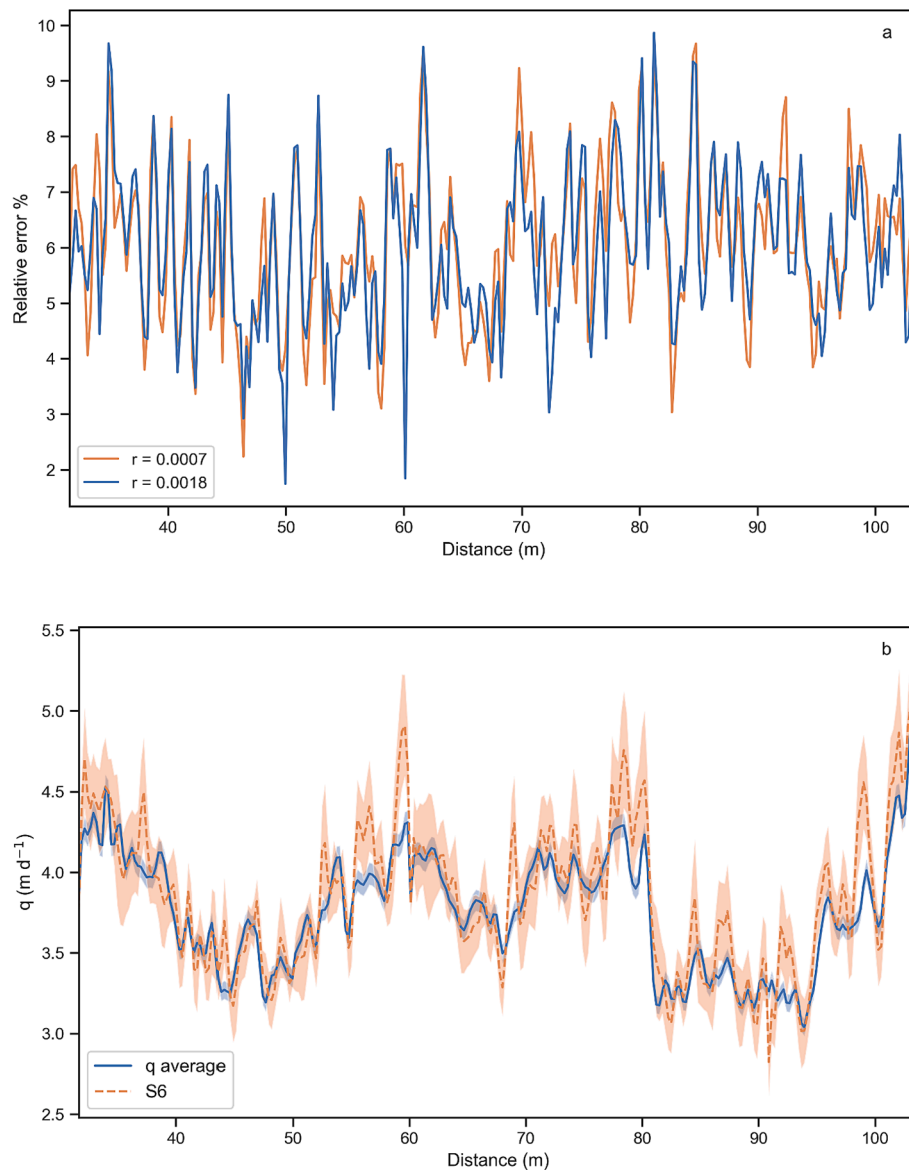


Fig. 8. (a) Calculated relative error along the cable, with a varied radius of 0.0007 m and 0.0018 m. (b) Calculated average specific discharge along the saturated length of cable between 31.65 m and 103.59 m for all twelve surveys is shown in blue, with blue shading representing standard error. Calculated specific discharge for survey S6 from 00:00 to 02:00 03/09/2022 is shown in orange dashed line, with orange shading representing the calculated absolute error along the cable associated with this survey. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

heat transfer processes is important when interpreting the initial temperature increase measured using hybrid fibre optic cable arrangements. It is not possible to quantify the skin effect in this field setting, due to the transfer of noise related error and the unknown impact of our specified cable geometry. However, it is acknowledged that the skin effect will have some effect on the initial temperature evolution. Del Val et al. (2021) and Simon et al. (2021) found the initial stage of the thermal response where temperature conduction is dominated by the skin effect of the cable to last around 110 s, using a cable with a 3.8 mm diameter. Given the cable we used had a different geometry and a greater diameter, we are unable to quantify this. However, this does not affect our objective which is to prove reproducibility of the A-DTS installation to measure small changes in specific discharge, which has been achieved demonstrating the repeatability of A-DTS derived estimates of river loss.

It was noted that at many locations along the cable, the observed temperature did not stabilise at an equilibrium value. Instead, a consistent linear trend occurred, either positive or negative. The stabilised temperature drift could be measuring a rate of change of specific

discharge which in turn controls the long-term stable temperature in the MILS model. This was previously explored by Simon et al. (2023b) using sandbox laboratory experiments; however, this is the first time that this has been observed in a field setting. Alternatively, the temperature drift might be caused by changes in the background river temperature.

It is important to note that the cable is measuring groundwater flux in the braidplain gravels beneath the riverbed. This could represent, to varying degrees, recharge from the river to the gravels, flow within the gravels with no exchange or discharge from the gravels to the river. Additionally, the derived specific discharge values from the MILS model are assumed to be one-dimensional, in the vertical direction perpendicular to the cable direction. However, the actual subsurface SW–GW interactions beneath braided rivers are likely more complex. The derived fluxes appear to follow bathymetry, indicating that bathymetry plays a key role in groundwater flow beneath the riverbed. For example, the northeast end of HDD1 the cable is located in the inside bend of the river and is potentially intercepting parafluvial flow which is discharging to the river.

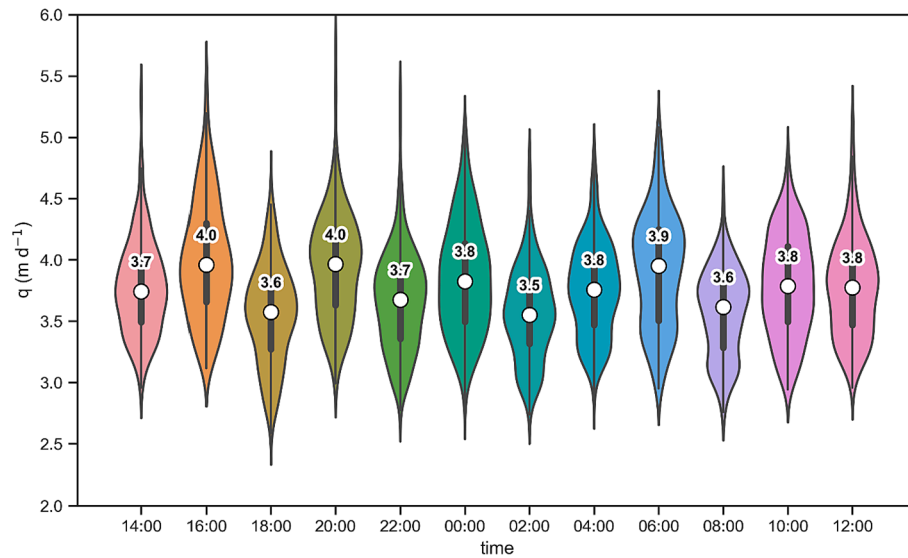


Fig. 9. Violin plot showing the probability distribution of specific discharge variability for the twelve surveys (S1 – S12). The thick grey bar represents the inter-quartile range and the white dot represents the median value.

A sensitivity analysis, presented in Table 1, was conducted using the MILS model by testing a range of credible parameter values to identify suitable ranges. The analysis focused on survey S6, with variations in the parameters K and r . The complex cable geometry compromises a key assumption of the MILS model so that, although a good fit to the data can generally be obtained, an effective r value must be assumed in doing so and this may not correspond to the measured r value. There is some ambiguity in which value to select for r , hence three values of r were examined: the measured r , a high r value used by Banks et al. (2022), and an optimal r value determined from this analysis (0.0018 m). Additionally, K was fixed at various values (ranging from 0.56 to 2.43 W m⁻¹ °C⁻¹) or optionally allowed to vary during the model fitting procedure. It was found that K remained relatively consistent between surveys. Therefore, as discussed in Section 3.5, K was estimated from the first survey, and then fixed at this value for the remainder of the surveys to assess the variability of q . Based on the results, the specific discharge values are relatively stable; however, there is some uncertainty in the

absolute value of q because of the uncertainty in the appropriate value to use for r .

Processing the 24-hour A-DTS experiment data with various r and K values, the calculated values of q vary by about one order of magnitude. The expected level of uncertainty from this method (6 %) using this particular A-DTS installation could be reduced in future. One possible approach could involve using sand tank experiments where q can be controlled, to determine an optimal value of r . The unknown nature of r has implications for inverted q in an absolute sense. Nevertheless, given the main objective of gauging q fluctuations in a relative capacity to comprehend temporal dynamics, the uncertainty is acceptable and does not impact the conclusion. Furthermore, future investigations, oriented towards evaluating seasonal dynamics of q from a relative perspective remain feasible.

In previous studies, A-DTS has been used to obtain spatially continuous temperature profiles over large distances, providing insights into various phenomena, notably SW–GW interaction. Despite its evident potential, previous field installations employing A-DTS surveys such as Banks et al., 2022; Del Val et al., 2021; Simon et al., 2021 have been limited to a single heat pulse survey. Consequently, only fluxes for a discrete period were derived. Simon et al. (2023b) recently proved that A-DTS is efficient in quantifying the temporal variations of groundwater fluxes with low uncertainties, based on sandbox experiments and numerical simulations. This is the first study of its kind to use A-DTS to conduct multiple, successive heat pulses to demonstrate the reproducibility of A-DTS to quantify groundwater fluxes enabling numerous fluxes to be quantified over extended periods in different hydrological settings. The small change in magnitude of the calculated specific discharge over the 24-hour period when the river and groundwater were relatively stable demonstrates the reproducibility of A-DTS derived estimates of river loss in braided river systems.

6. Summary

This study tested the hypothesis that the A-DTS method deployed in a complex field setting has the required sensitivity to measure small changes in groundwater flux. This method was developed to evaluate if changes in braided river leakage remains constant under stable river stage height and flow conditions. In doing so, this study demonstrates the reproducibility of groundwater flux measurements under similar river flow and stage conditions using the A-DTS method in a complex field setting. This will allow for subsequent research endeavours

Table 1
Sensitivity analysis of selected parameters tested in the MILS model for the sixth survey conducted at 00:00 – 02:00 on 03/09/2023.

Setting	r (m)	K (W m ⁻¹ °C ⁻¹) average (range)	q (m d ⁻¹) average (range)
Base case	0.0018	1.33 (0.40 – 1.89)*	3.8 (1.3 – 5.1)
Measured r	0.0007	1.32 (0.51 – 1.87)*	1.5 (1.1 – 2.0)
Moderate K	0.0018	0.90	3.0 (2.5E-13 – 3.9)
Measured K , measured r	0.0007	2.38	2.4 (1.3 – 5.1)
Measured K , base r	0.0018	2.38	6.3 (3.3 – 13.1)
Measured K , Banks et al (2022) r	0.003	2.38	10.4 (5.5 – 21.8)
Banks et al. (2022) r	0.003	1.32 (0.51 – 1.86)*	0.6 (0.5 – 0.8)
Banks et al. (2022) K	0.003	0.56	18.1 (1.5E-13 – 43.0)
Banks et al. (2022) raw data, updated parameters	0.0018	0.49 (0.38 – 2.23)*	3.0 (2.2E-8 – 3.3)

* K estimated from the MILS model. Note K estimates inverted from the model are compromised by the presence of cable material and are therefore not entirely representative of the surrounding porous material. See Section 4.4 for further explanation.

assessing temporal changes in groundwater recharge, to assess seasonality of groundwater recharge from braided river systems.

Based on the results from the 24-hour experiment, the median groundwater specific discharge values calculated along the cable had a small range of 3.5 to 4.0 m d⁻¹ across the wetted footprint of the river, indicating relatively stable groundwater recharge during the 24-hour experiment. The absolute error from the twelve repeated surveys was 0.09 m d⁻¹, which is within the expected measurement error of the installation the level of uncertainty expected from this method (6 %), demonstrating this method. The findings of this study enable seasonal changes in groundwater recharge from braided rivers to be quantified.

6.1. Future recommendations

Further work is required to improve the understanding of SW–GW interactions at this site, which will contribute to greater knowledge of subsurface processes in braided rivers. This would include quantifying the lateral river loss component and independent estimates of specific discharge based on hydraulic head and aquifer measurements using the extensive groundwater piezometer network at the study site. Conducting repeat A-DTS surveys under varying hydrological conditions, ideally with concurrent river flow gaugings, would provide complimentary information and a validation method to quantify the seasonal variation of groundwater recharge and how this relates to river stage height conditions.

Through the experimental design of the study and the data analysis of the A-DTS surveys, uncertainties in the cable material and thermal properties were identified as an important knowledge gap, which has also been identified in other studies by Del Val et al. (2021) and Simon et al. (2021). One important unknown in the data analysis routine is quantifying the skin-effect of different types of cable designs and specifications on heat transfer in a heterogeneous environment. This could be addressed through high-resolution numerical modelling of the cable geometry, its surrounds, and heat flow dynamics.

CRedit authorship contribution statement

Alice J. Sai Louie: Conceptualization, Methodology, Software, Validation, Formal analysis, Investigation, Writing – original draft, Visualization. **Leanne K. Morgan:** Writing – review & editing, Supervision. **Eddie W. Banks:** Writing – review & editing, Supervision. **David Dempsey:** Writing – review & editing, Visualization, Supervision, Software, Formal analysis. **Scott Wilson:** Writing – review & editing, Supervision, Funding acquisition, Conceptualization.

Declaration of competing interest

The authors declare the following financial interests/personal relationships which may be considered as potential competing interests: Scott Wilson reports financial support was provided by New Zealand Ministry of Business Innovation and Employment.

Data availability

Data is available from Sai Louie, A.J., Morgan, L.K., Banks, E.W., Dempsey, D., Wilson, S., 2023a. Active-distributed temperature sensing data collected beneath a braided river for quantification of groundwater recharge. Mendeley Data V2. doi: 10.17632/w2kr9m6s5h.2.

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