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Key Points:

- 28 active-distributed temperature sensing surveys were conducted over a 3-year period on two cables installed up to ~5 m below a braided river
- River flow changed considerably, yet there was a negligible relationship between river flow and specific discharge estimates
- Following a flood event, significant geomorphological changes led to an increase in specific discharge beneath the river margins

Supporting Information:

Supporting Information may be found in the online version of this article.

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Quantifying Interannual Variations in Groundwater Fluxes Within a Braidplain Aquifer Using Active-Distributed Temperature Sensing

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Abstract Braided rivers are a significant contributor to groundwater flow in braidplain aquifer (BPA) systems, yet the spatiotemporal variations in flow exchange remain poorly understood. This study presents an extensive active-distributed temperature sensing data set collected over a 3-year period from two 100-m long fiber optic cables installed up to ~5 m depth below a braided river channel in New Zealand to quantify specific discharge (i.e., parafluvial flow) within the BPA at high spatial resolution. Despite river flow fluctuating by two orders of magnitude, specific discharge within the BPA showed moderate variation, with values ranging from 3.5 to 5.9 m d⁻¹ on HDD1 (20.4% variance relative to the median) and from 2.4 to 4.5 m d⁻¹ on HDD2 (13.3% variance relative to the median) over the study period. Geomorphology influences hydraulic conductivity, thereby affecting the hydraulic gradient. Following a flood event, specific discharge increased beneath the river margins over time, with the largest increase corresponding with areas of greatest geomorphic change, and changes to the subsurface flow dynamics. Surveys conducted after the flood event, where river flow peaked at 225 m³ s⁻¹ (compared to a median of 2.51 m³ s⁻¹), showed a marked increase in specific discharge. These subsurface changes occurred gradually, with no obvious seasonal influence, compared to rapid geomorphic surface alterations. This translates to variable specific discharge within the BPA in both space and time caused by changing bedform and connectivity to preferential flowpaths. Geomorphic changes following a flood event had the largest impact on specific discharge.

1. Introduction

Groundwater recharge from braided rivers plays an important role in sustaining groundwater resources and maintaining river ecosystems globally (Coluccio & Morgan, 2019). These rivers act as key recharge zones where water infiltrates into underlying aquifers under losing conditions or serve as discharge zones in gaining reaches where groundwater contributes to surface flow (Winter et al., 1999). Braided rivers are known for their dynamic flow patterns, influenced by seasonal climatic conditions (Rhoads, 2020). Increasing anthropogenic stresses, principally increased water consumption and land use encroachment on the braidplains (Brower et al., 2023; Hicks et al., 2021), and climate change (Stecca et al., 2019) are having a detrimental impact to the hydrological function of these systems. In New Zealand, there are more than 163 rivers with braided reaches (Brower et al., 2023). In the Canterbury Plains, South Island New Zealand, braided rivers are estimated to contribute more than half of the surface water flow (Canterbury Water, 2009).

A braided river can have multiple individual active river channels; these channels, the bed gravels, and the braidplain aquifers constitute a ‘braided river system’. A ‘contemporary braidplain’ describes the lateral extent of the riverbed, encompassing river braids, active and abandoned river channels, as well as bar surfaces that, in a natural state, the river channels could reoccupy (Brower et al., 2023; Gray et al., 2018). The ‘active braidplain’ is defined by the contemporary braidplain extent occupied by the active river channels during a typical flow regime (Warburton, 1996). Braided rivers have a mobile sediment bed, which acts as a water storage reservoir for exchange with active river channels. A ‘braidplain aquifer’ (BPA) is a recent concept introduced by Wilson et al. (2024), referring to a shallow storage reservoir to the river system, distinct from the regional aquifer. It is formed in the contemporary braidplain by sediment remobilization during flood events, bounded by lateral geomorphological boundaries and a sedimentary unconformity at its base, and forms a shallow aquifer between the active river and the deeper regional aquifer. Wilson et al. (2024) theorizes that the saturated extent of the BPA

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storage reservoir contracts both longitudinally and laterally during prolonged dry periods and reduced river flow conditions. Evolution of geomorphology, such as gravel bar asymmetry impacts river loss rates and residence times (Bray & Dunne, 2017). In braided river systems, surface water–groundwater (SW–GW) interaction occurs at three spatial scales. For this paper, these are defined as: (a) Hyporheic flow between the active river channel and the hyporheic zone, the region beneath and alongside the riverbed where there is mixing of surface water and shallow groundwater (Boano et al., 2014). (b) Parafluvial flow between the active river channel and the BPA, which includes water flow through the sediments adjacent to the river channels. (c) Regional groundwater flow between the BPA and the deeper (>15 m) regional aquifer, encompassing larger-scale groundwater recharge processes. The focus of this study is on parafluvial flow, hereinafter referred to as groundwater flux in the BPA. Previous work at this site suggests that specific discharge within the BPA is predominantly horizontal, with a minor vertical flux component (Banks et al., 2022; Di Ciacca et al., 2024; Sai Louie et al., 2024).

Understanding the temporal changes in specific discharge within the BPA is vital for effective water resource management, especially in regions where braided rivers are the main source of recharge to aquifers. Studies have not examined long-term changes in braided river flow in relation to variations in groundwater flux. Current methods for measuring exchange between river channels and the shallow aquifer encompass a range of techniques, most of which only provide point estimates of river loss, or estimates over a larger distance, but limited to a single temporal snapshot (Coluccio & Morgan, 2019). Assessing changes in water content using seepage meters is one of the few methods to directly measure groundwater fluxes, but these measurements only provide point-scale estimates that have high associated uncertainty (Kalbus et al., 2006). Differential river flow gauging surveys can be used to estimate river channel loss but are challenging to conduct in braided rivers due to large discharge volumes and the high spatial variability of channel flow due to exchange with the braidplain gravels (Wilson et al., 2024). Specific discharge within the BPA is not well understood, indicating a need for further research in this area.

Recent advances in hydrogeological monitoring techniques have significantly improved our ability to monitor subsurface processes. Distributed temperature sensing (DTS) allows for high spatial and temporal resolution temperature measurements along fiber optic (FO) cables (Bense et al., 2016; Selker et al., 2006; Shanfield et al., 2018; Tyler et al., 2009). Heat has been widely used as a groundwater tracer for assessing surface water infiltration and groundwater flow and is now an established hydrological tool (Anderson, 2005; Rau et al., 2014). Further details of various DTS studies including different configurations and heat-tracing work are outlined in Sai Louie et al. (2024). Combined they are used to conduct active-distributed temperature sensing (A-DTS), enabling more precise and continuous assessment of SW–GW interactions by providing high spatiotemporal resolution temperature data sets that can be used to estimate multidirectional groundwater fluxes (Bense et al., 2016; Simon et al., 2022, 2023). Specifically, A-DTS measurements can be used to estimate specific discharge, q (m d^{-1}), the volumetric flux of groundwater per unit area perpendicular to the flow direction (herein specific discharge and groundwater flux are used interchangeably), providing the magnitude of the flux, but not its direction. The use of A-DTS for quantifying groundwater fluxes has been validated using sand tank experiments, which demonstrated high accuracy and associated uncertainties of <5% (Bakx et al., 2019; Simon et al., 2020, 2021, 2023).

While A-DTS provides high-resolution spatiotemporal data, some studies have used passive DTS methods (no active heating source alongside the fiber) in the field to study temporal changes in groundwater recharge from rivers. Vogt et al. (2010) conducted a time series analysis on a vertical DTS installation to estimate seepage rates from a losing stream over 20 days and found that periods with a high watertable resulted in intermittent high infiltration rates. Krause and Blume (2013) conducted three DTS surveys in a streambed to assess seasonal variability in SW–GW exchange, noting that DTS was more reliable for detecting this during winter when diurnal temperature oscillations were lower. More recently, Simon et al. (2022) conducted an 8-month passive DTS experiment, including a 4-hr heating period to conduct A-DTS, supporting the use of passive DTS for long-term monitoring of groundwater discharge into a stream. Simon et al. (2024) further advanced this work by using A-DTS to characterize spatiotemporal variability of hyporheic flows by conducting multiple A-DTS surveys over a 3-year period on a cable buried within the upper 0.1 m of the riverbed, running parallel to river flow direction. The results show that hyporheic fluxes are relatively stable over time (Simon et al., 2024). While that work provides insight into flows within the hyporheic zone, it does not provide information about broader scale groundwater flow below the hyporheic zone, such as in the BPA, where parafluvial flow occurs, which is the focus of this study. Banks et al. (2022) support the use of A-DTS for estimating groundwater fluxes in braidplain settings; however, this study was limited to a single point in time and interannual variations in groundwater fluxes within

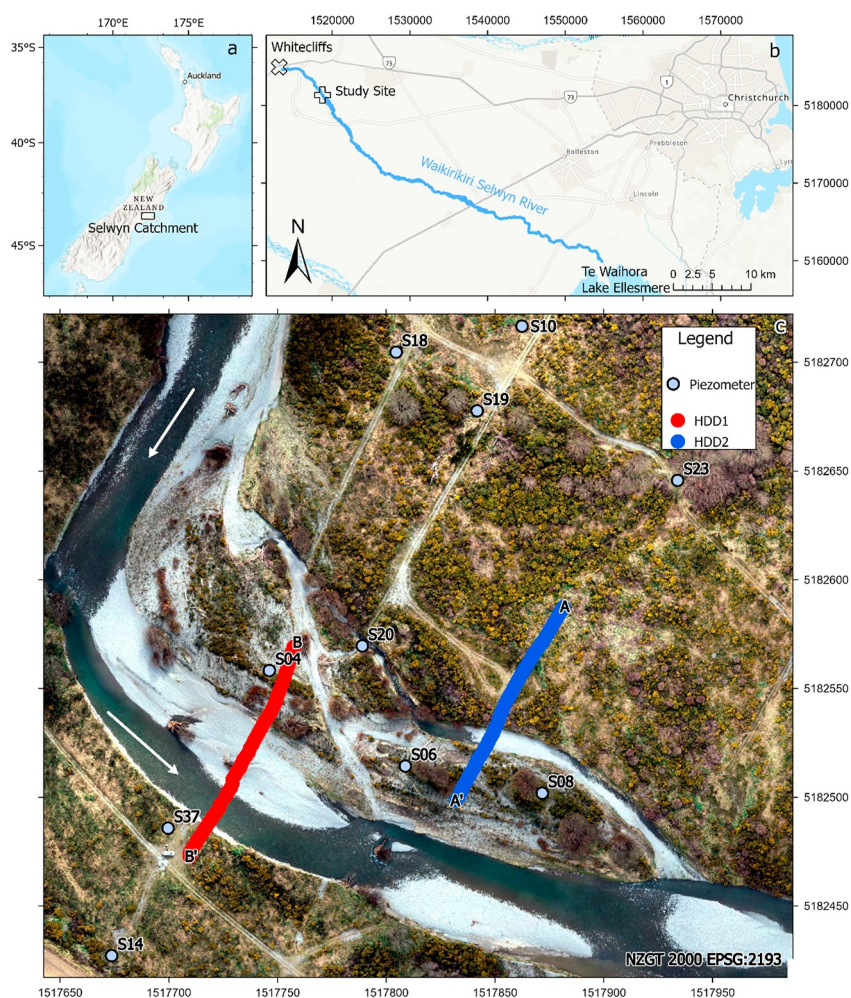


Figure 1. (a) Location of the Waikiri-Selwyn Catchment in the South Island of New Zealand. (b) The study site is in the upper reaches of the Waikiri-Selwyn River. (c) Aerial photograph of the study site taken on 24 August 2023 showing the network of piezometers in the braidplain aquifer at the site and the location of two subsurface horizontal hybrid fiber optic cables, designated HDD1 (BB') and HDD2 (AA'). White arrows indicate the surface water flow direction. Modified from Sai Louie et al. (2023b).

the BPA are not well understood. Sai Louie et al. (2024) demonstrated the reproducibility of A-DTS derived estimates of specific discharge beneath braided rivers by conducting 12 consecutive A-DTS surveys over a 24-hr period during relatively stable river flow conditions. Sai Louie et al. (2024) calculated median specific discharge with uncertainties of 6% for individual surveys, providing confidence in using the A-DTS method. Together, Banks et al. (2022) and Sai Louie et al. (2024) validate the use of A-DTS to examine variation in groundwater flux within the BPA at a high spatial resolution over longer seasonal or interannual periods.

This study addresses a critical knowledge gap in quantifying interannual variation in groundwater flux within a BPA, where the BPA is recharged from a braided river in New Zealand. Multiple (approximately monthly) A-DTS surveys over a 3-year period were carried out and provide an unprecedented data set for exploring surface-water groundwater exchange, capturing channel-scale processes at a high-spatial resolution. Changes in river flow, including two floods, during the study period provides the opportunity to relate groundwater fluxes to river flows and changes in river morphology.

2. Study Site

The Waikiri-Selwyn River, which travels 93 km across New Zealand's Canterbury Plains (Figure 1b) exemplifies a dynamic braided river system characterized by significant fluctuations in river flow (Larned

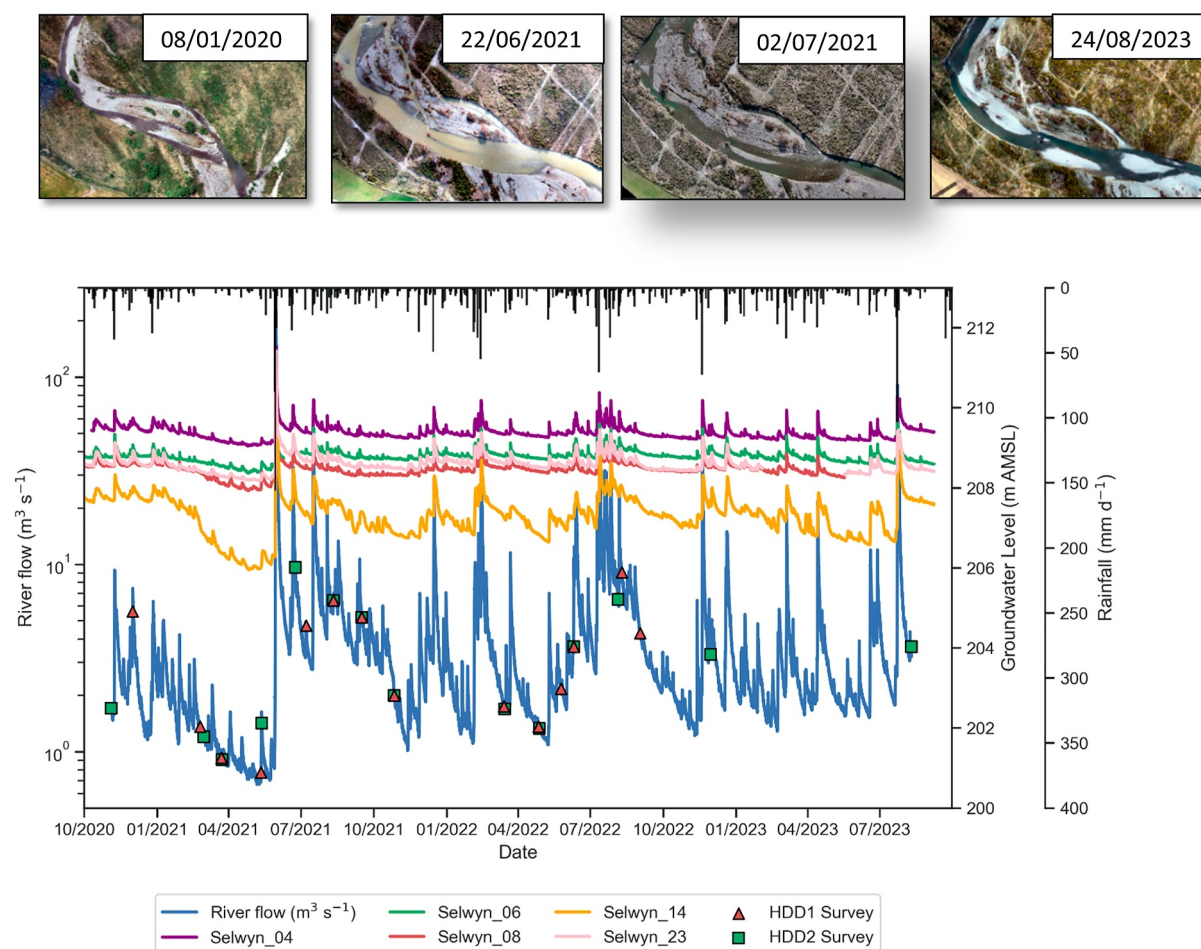


Figure 2. River flow at Whitecliffs from October 2020 to September 2023, plotted on a semi-log scale. Red triangles show when an active-distributed temperature sensing survey was conducted on HDD1 and green squares show when a survey was conducted on HDD2. Groundwater levels (m Above Mean Sea Level) in selected piezometers show hydraulic head within the braidplain aquifer. Daily rainfall measured at the station on the river terrace at the study site for the study period is shown in black.

et al., 2008). The catchment area is approximately 250 km², with an average catchment rainfall of 1,045 mm y⁻¹, and potential evapotranspiration of 750 mm y⁻¹ (Wilson et al., 2024). Due to the relatively low mean river flow and high transmission losses, the river has intermittent flow, with a drying front occurring in summer downstream of the study site (Di Ciacca et al., 2023). Di Ciacca et al. (2023) estimated river transmission losses at the study site between April 2020 and March 2021 ranged from 0.25–0.65 m³ s⁻¹ km⁻¹ (i.e., 21.6–51.2 m d⁻¹), reaching 1.5 m³ s⁻¹ km⁻¹ following a flood event.

The study site is a 1-km reach within the Upper Waikirikiri Selwyn River system that is hydraulically disconnected from the regional aquifer (Figure 1c). A hydraulically disconnected river system is defined by the presence of lower permeability material at the base of the BPA (i.e., which is formed by a sedimentary unconformity), regulating infiltration to the regional aquifer and creating a variably saturated zone between the BPA and regional water table (Wilson et al., 2024). At the study site the contemporary braidplain is 460 m wide and the active braidplain (where sediment deposition and erosion occur beneath the active river channel) is 55 m wide. Within the active braidplain the braiding index (i.e., the average number of channels in a cross-section) is 1.56 and the BPA is confined laterally by the river terraces (Wilson et al., 2024). The base of the BPA ranges from 2.6–4.9 m below ground level (bgl) across the site (Figures 3a and 3b). As per Wilson et al. (2024), the porosity of the sediment which forms the BPA at the study site ranged from 0.09–0.22 (average 0.14) based on porosity measurements from five sonic drill core samples and four field porosity measurements. Field porosity measurements were conducted by manually excavating an area, estimating the value using photogrammetry. The material was

then sieved to determine the grainsize distribution. The sediments comprising the BPA are poorly sorted, hence the low porosity values. Based on the BPA's geometry (i.e., volume) and porosity, the estimated storage volume is $130\text{--}240\text{ m}^3\text{ m}^{-1}$ (Wilson et al., 2024).

Recent research has been conducted at the Waikirikiri Selwyn River study site as part of a wider project on subsurface processes in braided rivers, and further details on the study site can be found in various associated publications (Banks et al., 2022; Di Ciacca et al., 2023, 2024; Robb, 2021; Sai Louie et al., 2024). Data has been compiled to inform a conceptual understanding of SW–GW exchange in braided river systems in New Zealand, as detailed in Wilson et al. (2024). The focus of the current paper is on A-DTS surveys conducted on two subsurface horizontal hybrid FO cables, designated HDD1 and HDD2, installed beneath the active river channels at the study site (Figure 1c). The cables were installed perpendicular to the river flow direction using novel horizontal directional drilling methods as detailed in Banks et al. (2022). The actual cable depth of HDD1 varies from 0.56 to 4.27 m (average 2.71 m), and for HDD2 from 0.91 to 5.55 m (average 2.18 m). Additional information on the structure of the hybrid FO cable and cable installation is given in previous studies (Sai Louie et al., 2023a, 2023b, 2024). These FO cables lie within BPA gravels, and intercept parafluvial flow.

3. Methods

3.1. Active-Distributed Temperature Sensing

A Silixa XT-DTS was used to measure the temperature along the cable with a manufacturer specified temperature resolution of 0.01°C , and a spatial sample interval of 0.25 m with a spatial resolution of 0.54 m (Silixa Ltd, 2022). The actual measurement resolution under field conditions was calculated as the standard deviation of the temperature values when the heat pulse temperature had stabilized and detrended of any linear component. This resulted in a noise level under field conditions of approximately 0.04°C . Sai Louie et al. (2024) estimated the effective spatial resolution of the cable to be 0.763 m using the 90% step change method as detailed by Simon et al. (2020). The DTS was configured using a double-ended differential loss correction with a 10 s sampling rate comprising 5 s forward measurements followed by 5 s reverse measurements (Van De Giesen et al., 2012). The calibration method used for these experiments is detailed in Banks et al. (2022) and Sai Louie et al. (2024). To heat the copper conductors, a Silixa heat pulse unit was used with an applied power output of 15 W m^{-1} .

Experiments involved monitoring temperature changes within the cable during heating and cooling cycles. For experiments conducted prior to 27 October 2021, the sequence consisted of a 300 s period during which the ambient temperature was measured, followed by a 1,200 s heating period and then a 1,200 s cooling period. For experiments after this date, a heating period of 3,600 s and a cooling period of 1,800 s was used. The modified heating period was selected to provide more time for the cable temperature to reach equilibrium during a heat pulse, yet not too long so that it captured changing groundwater fluxes. A comparison between the two different heating periods (1,200 and 3,600 s) was conducted for the experiment on 27 October 2021 and the estimated specific discharge using the 1,200 s heating period was 3.90 m d^{-1} whereas for the 3,600 s heating period the estimated specific discharge was 3.88 m d^{-1} . These two heating periods are both deemed sufficient for estimating specific discharge because of the relatively high groundwater velocities at the site, causing prompt temperature stabilization and plateauing of the change in temperature, characterized by the advection dominant phase shown in the temperature breakthrough curves (Sai Louie et al., 2024).

3.2. Delineating the Cable Locations Relative to the Base of the BPA

The cable position relative to the base of the BPA was determined using the drilling logs and surveying methods (Real Time Kinematic Global Navigation Satellite System, RTK GNSS). The network of piezometers was installed using sonic drilling methods, which enabled sediment core to be extracted and logged in detail. The location of the contact that forms the base of the BPA was recorded in the core log and the coordinates were recorded in NZGD 2000 New Zealand Transverse Mercator coordinate system. A surface of the BPA base (i.e., sedimentary unconformity) was estimated using the 3D Ordinary Kriging method in the `scipy.interpolate` Python library and for comparison the Empirical Bayesian Kriging tool in ArcGISpro was also used.

During the horizontal directional drilling to install the two subsurface hybrid FO cables, the cable location coordinates (eastings, northings and elevation i.e., x, y and z) were recorded at 1 m intervals with a vertical accuracy of $\pm 0.25\text{ m}$. To understand the spatial context of the A-DTS data, the cable location coordinates were interpolated

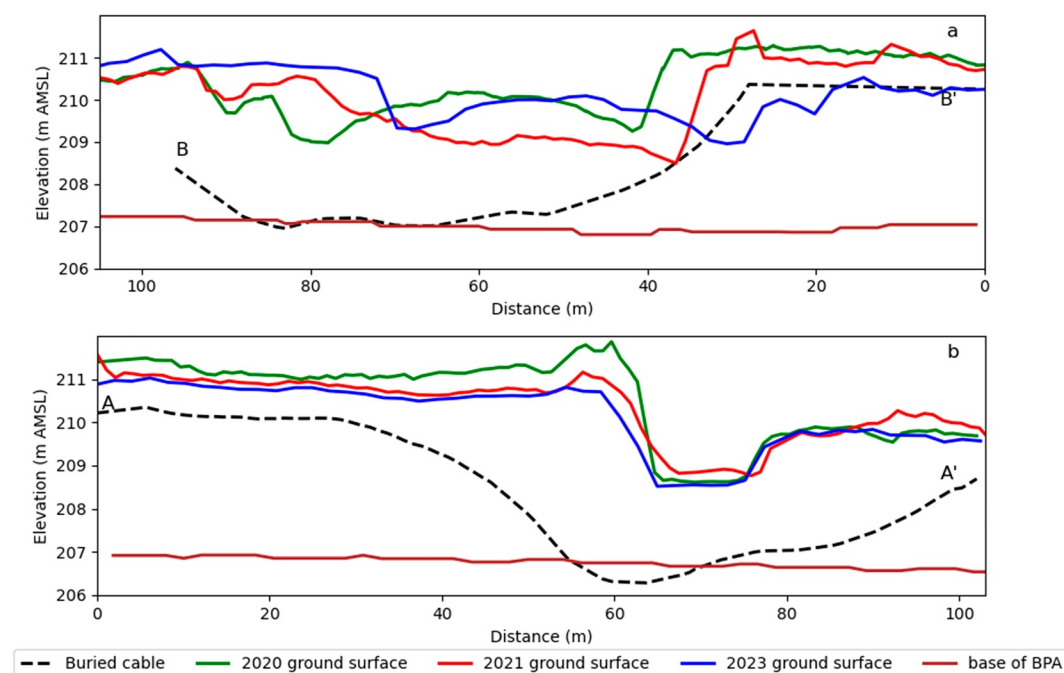


Figure 3. Cross-section depth profile of the buried cables, relative to the original ground surface, the ground surface following the May 2021 flood, the ground surface following the June 2023 flood and the interpolated braidplain aquifer contact surface based on drilling lithological logs. (a) Buried cable HDD1 (BB'). (b) Buried cable HDD2 (AA'). Both profiles are looking in the downstream direction.

into cumulative distances of every 0.254 m, starting from 0.115 m, to get coordinates for each measuring point along the cable. Additionally, the bathymetry above the cable was surveyed annually to assess changes to the thickness of overlying material. The location of the cables relative to the BPA surface were then compared.

Aerial photography was collected four times over the 3-year period to survey geomorphological changes at the site. These were flown using a DJI Matrice 210 drone equipped with a X4S camera and a Klau PPK system for corrections, flying at approximately 60 m above the ground. An RTK GNSS base station and ground control point were used to correct the images to produce an orthomosaic image (Figure 2).

3.3. Hydrological Monitoring Infrastructure

The study site is equipped with a network of hydrological monitoring equipment described in previous studies and shown in Figure 1 (Banks et al., 2022; Sai Louie et al., 2024). Surface water measurements including river stage and temperature were recorded by a long-term monitoring station at Whitecliffs, located 9.5 km upstream. To establish a flow rating curve for calculating river flow, regular manual flow measurements were conducted.

The initial groundwater monitoring network comprised 27 piezometers logging groundwater pressure head and temperature at 15-min intervals. These operated from August 2020 and were screened at depths ranging from 1 to 35 m. Seven of these piezometers were destroyed during a flood in late May 2021, and subsequently replaced with 10 new piezometers. To assess the groundwater level in the BPA, 10 piezometers screened from depths of 1.02–4.56 m were monitored (Figure 1c). Weather data including rainfall was monitored from a weather station at the study site located on the true left river terrace.

3.4. Mathematical Model for Estimating Specific Discharge

The key equation used to process the A-DTS data is the analytical moving infinite line source (MILS) solution Equation 1, commonly referred to as the Hantush-Jacob solution, initially introduced by Carslaw et al. (1962). Subsequently, Zubair and Chaudhry (1996) adapted this solution to address the advection-conduction equation. The MILS model has since undergone further investigation in the context of groundwater flow by Ren et al. (2000) and Bakker et al. (2015):

$$T(x, y, t) = \frac{Q}{4\pi K} \exp\left(\frac{qx}{2DR}\right) W\left(\frac{x^2 + y^2}{4D(t - t_0)}, \frac{q\sqrt{x^2 + y^2}}{2DR}\right) \quad (1)$$

Where T (°C) is temperature response, during time t (s), at different locations (x, y) relative to a heat source at the origin. The heat source, Q (W m^{-1}) starts at time t_0 (s) and is constant. The thermal properties include thermal diffusivity, D ($\text{m}^2 \text{s}^{-1}$), thermal conductivity, K ($\text{W m}^{-1} \text{°C}^{-1}$) and the thermal retardation coefficient, R . For full details describing the mathematical model used, see Sai Louie et al. (2024).

Because the heat transfer equations are susceptible to parameter compensation, where different combinations of parameter values can produce similar outputs, it can be challenging to uniquely identify the values of individual parameters from temperature data alone, as different combinations of parameters can produce similar thermal responses (Bakker et al., 2015; Ren et al., 2000). Initially, only thermal conductivity was fitted, and the results indicated a clear distinction between the vadose and the phreatic zone. For subsequent analysis, the focus was on the fully saturated phreatic zone, where there is no saturation variability. In particular, the evaluation was restricted to between 31–104 m on HDD1 and 45–98 m on HDD2, where the sediment was fully saturated. Therefore, K was fixed at $1.33 \text{ W m}^{-1} \text{°C}^{-1}$ and D was fixed at $0.06 \text{ m}^2 \text{d}^{-1}$, these values were determined based on a sensitivity analysis. Then, automatic model calibration was used to solve for the best-fitting value of specific discharge, denoted by q (m d^{-1}), when matching measured and modeled temperatures. For the distance between the heating cable and the FO cable, r (m), a value of 0.0018 m was used as this was determined to be the optimal r value by Sai Louie et al. (2024). Due to the cable's asymmetric construction, there is likely to be a skin effect that has not been quantified. Although this may impact absolute estimates of groundwater specific discharge, the method is still sufficient to identify relative changes in discharge along the cable length (Banks et al., 2022) or over time (Sai Louie et al., 2024). Furthermore, previously inferred values of specific discharge are consistent with groundwater velocities estimated from Di Ciacca et al. (2024) who used a simplified 1D analytical model to simulate groundwater recharge from the site and compared it against more complex 2D and 3D numerical models, which offers confidence in the approach.

4. Results

4.1. Background Hydrological Conditions

The study period spanned nearly three years, from 4 November 2020 to 10 August 2023 and during this time the hydrologic conditions at the study site were highly variable. This is apparent from the shape of the hydrograph of the Waikirikiri Selwyn River at the Whitecliffs gauge station located approximately 9.5 km upstream of the site (Figure 2). The river exhibits short lag times between rainfall events and changes in river level, that rise steeply and peak rapidly. The study period coincided with three consecutive years of La Niña (El Niño Southern Oscillation (ENSO) influence) between June 2020 to March 2023, which brought above-average rainfall and flooding to much of New Zealand (NIWA, 2023). El Niño Southern Oscillation climatic drivers influence the magnitude and frequency of flooding in New Zealand (Jenkins, 2010).

There was a significant flood event on 31 May 2021 with up to 150 mm rainfall recorded. The river flow at Whitecliffs gauging station during this event peaked at $225 \text{ m}^3 \text{s}^{-1}$, compared to a median (during the study period) of $2.51 \text{ m}^3 \text{s}^{-1}$ (Figure 2). During this flood event, there was a rapid, synchronous rise in groundwater levels within the network of shallow piezometers (Figure 2), indicating the close hydraulic connection between the river and BPA. The underlying regional groundwater level, measured in meters Above Mean Sea Level (m AMSL), at the study site increased by up to 10 m. The aerial photograph taken on 22 June 2021 shows the elevated floodwaters. In a subsequent photograph taken on 02 July 2021, the floodwaters have dissipated, making the geomorphological changes apparent. These changes include the dominant braid of the river migrating to the south with the true right bank eroding and the main channel becoming approximately 11 m wider. This was evidenced by several mature willow trees on the true right bank that, after this event, ended up in the middle of the main river channel. Additionally, following this event, the secondary channel became connected in the upper reaches of the study site and became deeper.

Another flood event on 23 July 2023 caused river flows to reach $91 \text{ m}^3 \text{s}^{-1}$ and led to the southwest bank of the main channel collapsing and snapping the A-DTS cable at HDD1. During the final survey on HDD2 on 10 August

2023, the true left channel was empty; however, the A-DTS survey indicated that the cable intercepted saturated material between 46 and 99 m along the cable. Based on the aerial photograph flown at the same time as the A-DTS survey (Figure 2), there is a small pool in the true left river channel upstream of the cable, but directly above the cable there is no surface water. However, the measurements along HDD2 showed that the sediments are saturated at this location and there is likely to be parafluvial flow.

A total of 28 A-DTS surveys (14 on HDD1 and 14 on HDD2) were conducted at the study site over the 3-year period (Figure 2). During this 3-year period, the river flow varied by two and a half orders of magnitude ($0.67\text{--}225\text{ m}^3\text{ s}^{-1}$). However, if river flows during the specific times of the A-DTS surveys are considered, the variation in river flow is one order of magnitude (it was not possible to carry out surveys during flood periods).

4.2. Geomorphological Changes

The high flow event on 23 July 2023, despite being smaller in magnitude, caused the most geomorphological changes to the site over the 3-year period, presumably because the riverbanks had already been destabilized from the previous flood event. The true right bank eroded an additional 17 m to the southwest. The 2023 ground surface (Figure 3a) was measured in August 2023, and during this survey the hybrid cable HDD1 was observed protruding out of the bank and had been severed during the last flood event. Nonetheless, this observation confirmed the assumption of Sai Louie et al. (2023a) that, following installation, surrounding sediments collapsed during installation to provide direct contact between sediment and the cable. Geomorphological changes to the true left channel, overlying HDD2 were minimal over the 3-year period (Figure 3b).

The location of the buried cables relative to the base of the BPA was determined using a topographic survey, using a high-precision GPS instrument. To obtain the cable depth, a DigiTrak HDD cable locating system was used to determine the position of the drill bit (at $\pm 0.25\text{ m}$ accuracy). This involves tracking the drill bit using a GNSS receiver, operated by a technician who follows the path of the drill bit with the receiver. This provided the depth of the cable below ground surface (z component). A Trimble RTK was then used to accurately measure the x , y and z coordinates of the ground surface directly above the drillhole position, which had been located by the drill. From this information we determined the location (x , y and z coordinates) of the cable below ground and its relative position in the BPA. Figure 3a shows that cable HDD1 is close to the base of the BPA between 65 and 70 m, and the cable is beneath the BPA between 80 and 87 m. Figure 3b indicates HDD2 is up to 0.5 m below the base of the BPA between 55 and 70 m. As such, the majority of the 100 m long A-DTS cable is positioned within the BPA.

4.3. Interannual Active-Distributed Temperature Sensing Surveys

Results from the 28 A-DTS surveys conducted on HDD1 and HDD2 over the nearly 3-year study period are presented in this section. The nearly monthly A-DTS surveys allow for comparison across monthly, seasonal, episodic (i.e., pre and post flood) and interannual periods. Here, to assist with orientation when reviewing the data presented in figures, the term ‘right margin’ is used to refer to the region near the true right bank while the term ‘left margin’ refers to the region near the true left bank of the river. This is looking downstream and noting that at this location there are two channels and HDD1 is beneath the main channel and HDD2 is beneath the smaller secondary channel, see Figure 1c. For HDD1 the left margin occurs at approximately 85 m and the right margin refers to 40 m along the cable. The midpoint is at 65 m along the cable. For HDD2 the left margin is at approximately 45 m, the right margin is at 90 m and the midpoint is at 70 m along the cable.

The heat pulse temperature increase (ΔT) generally followed the same spatial pattern along the length of the cable for all surveys. For example, along the saturated portion of HDD1 (between 31 and 104 m) ΔT remained relatively consistent between 2.2° and 4.9°C , this is beneath the main river channel. For HDD2 surveys, ΔT in the saturated zone from 45 to 98 m along the cable beneath the left margin, ranged from 2.7° to 5.4°C . Detailed plots showing ΔT for each A-DTS survey are provided in Supporting Information S1 (Figures S1a and S1b, respectively).

Estimates of specific discharge along the saturated length of the cable for all 28 surveys are presented in Figure 4. In general, there was significant spatial variability for surveys conducted on HDD1, with values ranging from $2.2\text{--}15\text{ m d}^{-1}$. All surveys on HDD1 showed higher specific discharge values between 60 and 63.5 m along the cable (Figure 4). At 87 m along the HDD1 cable a maximum specific discharge of 16.1 m d^{-1} was estimated on 15 March 2022. Specific discharge estimates displayed less spatial variability along HDD2, with values ranging from

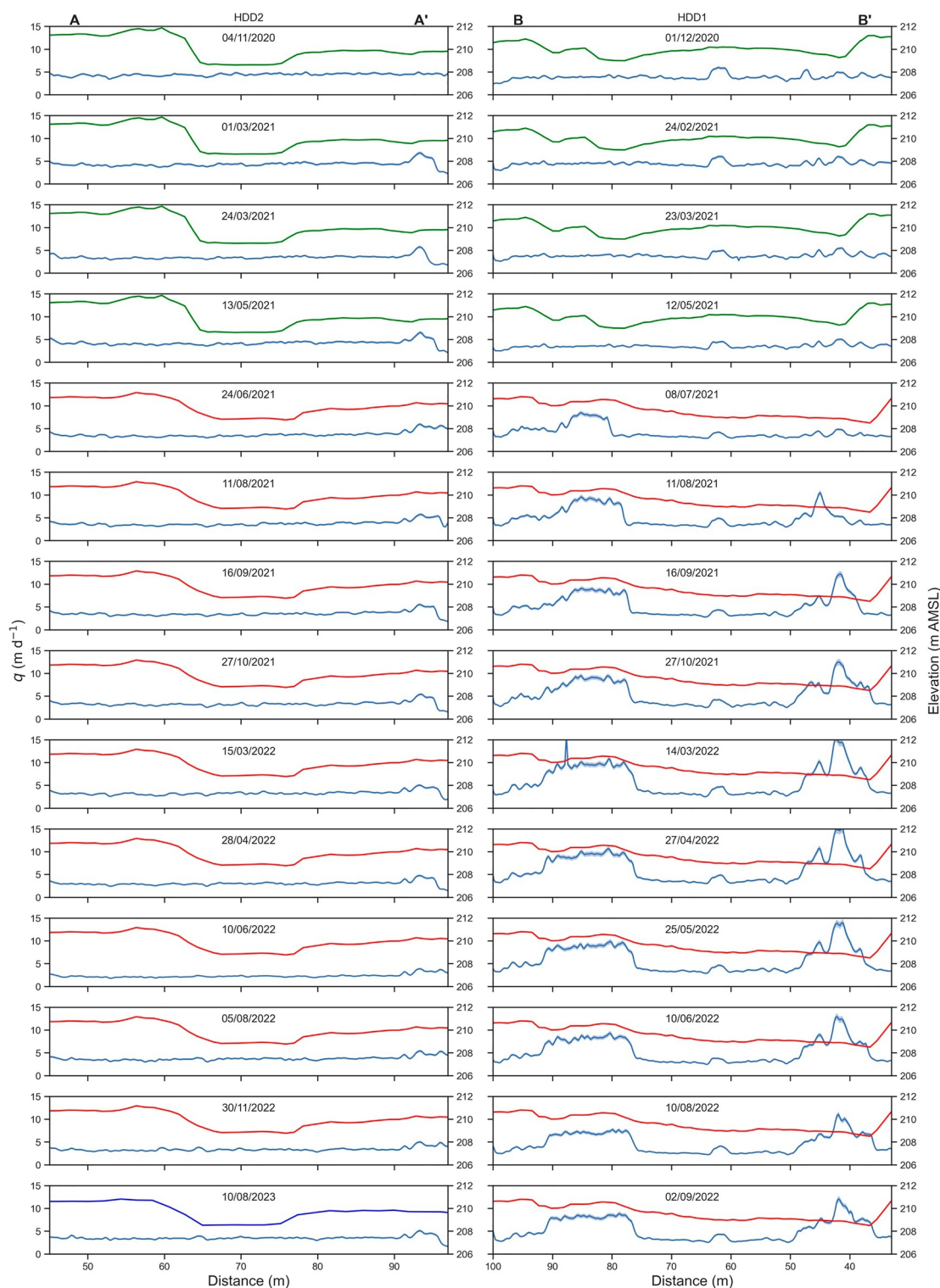


Figure 4. Seasonal variation in the average specific discharge along the cable for HDD2 (AA') shown on the left and HDD1 (BB') shown on the right, graphs are oriented facing downstream. Calculated groundwater flux in the braidplain aquifer (i.e., specific discharge) is plotted in blue on the y-axis, with error shading representing 6% error associated with each survey. Overlying river morphology is shown on the secondary y-axis, the green line is the pre-flood profile from 2020 and red is post-flood morphology. The blue line is the river morphology following another high flow event which occurred in July 2023. Survey dates are shown, noting that surveys were not always conducted on the same date on both cables HDD2 (AA') and HDD1 (BB').

1.7–4.2 m d⁻¹. However, at the right margin relatively high specific discharge estimates of up to 6.8 m d⁻¹ were determined (Figure 4).

Specific discharge estimates along HDD1 and HDD2 showed different temporal patterns over the 3-year study period. Specific discharge estimates (averaged across the saturated portion of the cable) on HDD1 ranges from 3.5 to 5.9 m d⁻¹ and on HDD2 ranged from 2.4 to 4.5 m d⁻¹. As a measure of how much the estimated specific discharge varies relative to its magnitude, the percentage variance relative to the median was calculated as $(q_{75}-q_{25})/q_{\text{median}}$. For HDD1 this is 20.4% and for HDD2 this is 13.3%. Another measure of variance is to compare the absolute range of specific discharge estimates to the minimum specific discharge estimate at each site, that is, $(q_{\text{max}}-q_{\text{min}})/q_{\text{min}}$, and this gives 66% for HDD1 and 86% for HDD2.

On HDD1 there was greater variation in specific discharge at the margins (up to 10 m d⁻¹ higher), compared to HDD2. This increase in specific discharge beneath the margins on HDD1 could have implications for understanding the spatial heterogeneity within the BPA, and is explored in the discussion.

4.4. Specific Discharge and River Flow

Specific discharge estimates, averaged over the saturated portion of both cables, ranged from 2.4 to 5.9 m d⁻¹ over the entire 3-year period (with an average of 4.2 m d⁻¹). Overall, there was a negligible relationship between specific discharge estimates and river flow for surveys conducted on HDD1, and a moderate negative relationship for surveys conducted on HDD2 (Figure 5). Specific discharge estimates from surveys carried out before and after the flood event in May 2021 also indicated no relationship between specific discharge and river flow (Figure 5).

4.5. Seasonal Variability of Specific Discharge

To assess seasonal variability in specific discharge, estimates of specific discharge were grouped by survey calendar month and categorized by the season: summer (December–February), autumn (March–May), winter (June–August) and spring (September–November). The most noticeable seasonal relationship was present in surveys conducted in winter, which showed slightly higher than average specific discharge values (i.e., 4.2 m d⁻¹) and coincided with moderate to high river flows (Figure 5). Only two surveys were conducted in summer, more sampling during summer would be necessary to establish any patterns. Despite Autumn surveys occurring during lower-than-median river flows ($Q < 2 \text{ m}^3 \text{ s}^{-1}$), average specific discharge estimates across the saturated portion of the cable ranged between 3.5 and 5.9 m d⁻¹ for HDD1 and between 2.4 and 4.5 m d⁻¹ for HDD2 (Figure 5). Spring surveys occurred during low to moderate river flows and specific discharge estimates ranged between 5.0 and 5.2 m d⁻¹ for HDD1 and between 3.4 and 4.5 m d⁻¹ for HDD2 (Figure 5). Notable outliers include the surveys conducted on 10 June 2022, which showed contrasting results: a high average specific discharge value of 5.1 m d⁻¹ on HDD1, yet a low specific discharge of 2.4 m d⁻¹ on HDD2. This survey was conducted as river flow at Whitecliffs reached $10.7 \text{ m}^3 \text{ s}^{-1}$, significantly higher than the median river flow during the study period of $2.51 \text{ m}^3 \text{ s}^{-1}$. These findings suggest that there is no strong seasonal influence on specific discharge values averaged over the saturated portion of the cable (Figure 5).

4.6. Analysis of Hydraulic Gradients in the BPA

Groundwater levels measured using piezometers screened in the BPA showed a consistent trend, including a clear relationship to river levels (Figure 2). Most noticeably, groundwater levels rose rapidly following the flood events in May 2021 and between May and June 2022 (Figure 2). Piezometer S04 consistently shows the highest groundwater levels ranging between approximately 209 and 210.5 m AMSL and exhibits a pronounced rapid response to river flow change (Figure 2). Piezometers S06, S08 and S23 all display similar patterns and demonstrate moderate response to high river flow events (Figure 2). Piezometer S14 exhibited the greatest variability in groundwater level (Figure 2). S14, situated on the outside of a bend in the main river channel and approximately 90 m from the active channel, exhibited significant drops in groundwater levels during low river flow. Horizontal hydraulic gradients within the BPA were calculated using the daily average groundwater level to reduce noise. The hydraulic gradient between S04 (representative of the river level) and S14 (parallel to HDD1) has the largest magnitude and increased in summer 2021 and 2023, during low river flows. Additionally, the fluctuations in hydraulic gradient highlight the connection between the river and BPA at this location (Figure 6a). In contrast, S23 on the inside of the river bend but still within the BPA, has limited fluctuations in groundwater levels, suggesting that in this location fluxes are steady over time. The hydraulic gradient between S04 and S23 is

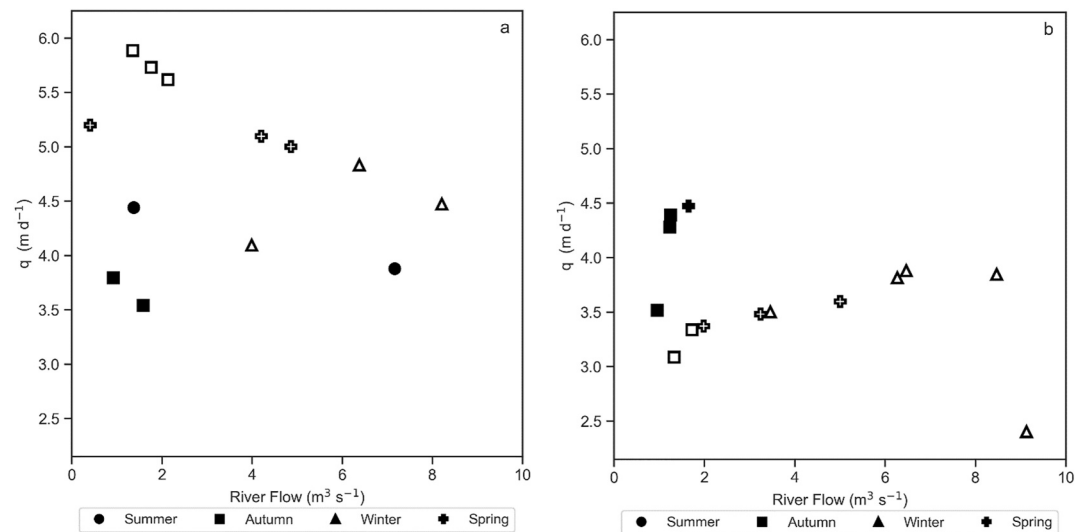


Figure 5. Average specific discharge within the braidplain aquifer versus river flow at Whitecliffs for each survey conducted on (a) HDD1 (BB') and (b) HDD2 (AA'). Circles represent surveys conducted in summer (December, January, and February). Squares represent surveys conducted in autumn (March, April, and May). Triangles represent surveys conducted in winter (June, July, and August) and crosses represent surveys conducted in spring (September, October, and November). Specific discharge values are averaged over the saturated portion of the cable. Surveys conducted prior to the flood event are shown by solid symbols and post-flood surveys are shown as outlined symbols.

relatively small, which typically suggests less flow and connection to the river at this location within the BPA. However, the flux could still be high if the river bed in this location has a high hydraulic conductivity.

Piezometers S04, S06, and S08 are all orientated in a direction parallel to the river flow and are within the active braidplain. The hydraulic gradients between S06–S04 and S08–S06 are relatively similar throughout the study period, with some sharp declines followed by recovery (Figure 6a). The hydraulic gradients have opposite trends in the active braidplain (S06–S04 and S08–S06) compared to those across the BPA (S04–S14 and S04–S23). The hydraulic gradients within the active braidplain exhibit less variability, supporting the concept that the BPA acts as a buffer, smoothing hydraulic responses.

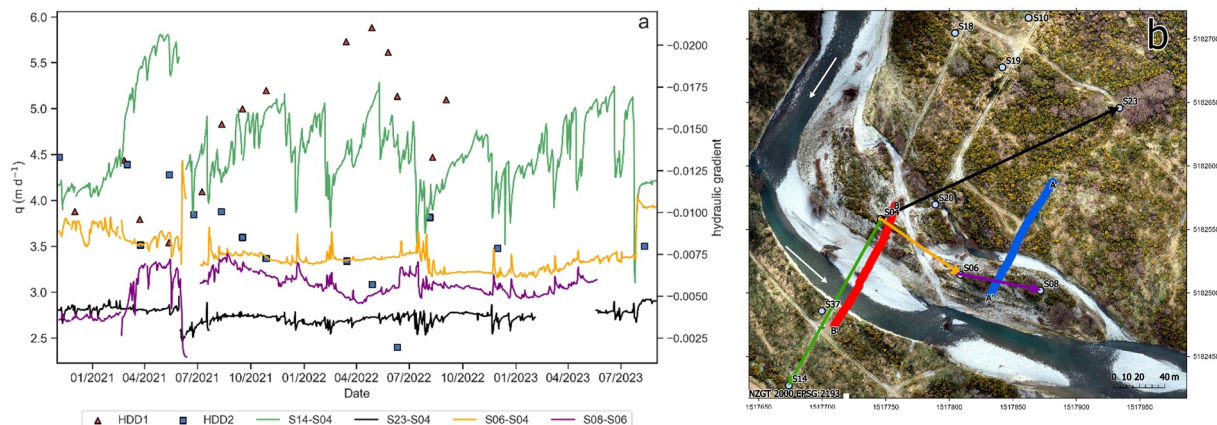


Figure 6. (a) Seasonal variation in average specific discharge in the braidplain aquifer (BPA) interpreted from the active-distributed temperature sensing surveys on HDD1, shown by the red triangles, and HDD2, shown by the blue squares. Specific discharge values are averaged over the saturated portion of the cable. For comparison, hydraulic gradient between the piezometers in the BPA are plotted on the secondary y-axis. (b) The location of piezometers used to calculate the hydraulic gradient S14–S04 is shown in green, S23–S04 is shown in black, S06–S04 is shown in orange and S08–S06 is shown in purple. The hydraulic gradient of piezometer S04 is comparative to the river level at the study site.

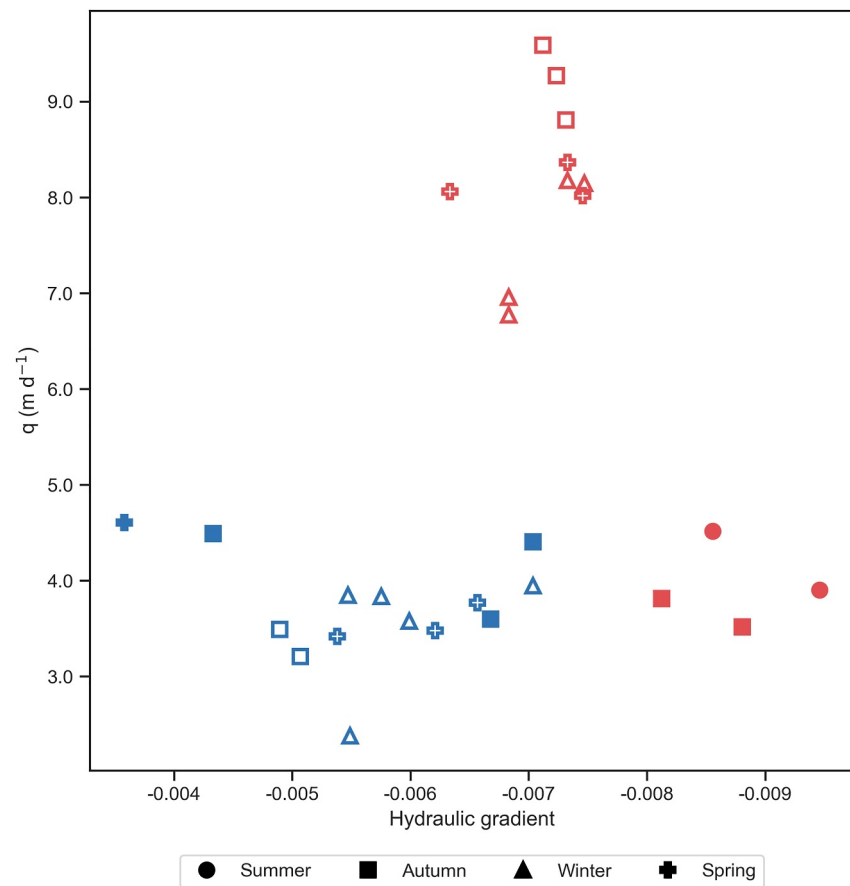


Figure 7. Specific discharge versus hydraulic gradient. Specific discharge was averaged across a 10 m cable length at the location where the hydraulic gradient transects intercept the cables. Red symbols represent values for the S04–S06 transect (which traverse HDD1) and blue symbols represent values for the S06–S08 transect (which traverse HDD2). Circles represent surveys conducted in summer (December, January, and February). Squares represent surveys conducted in autumn (March, April, and May). Triangles represent surveys conducted in winter (June, July, and August) and crosses represent surveys conducted in spring (September, October, and November). Surveys conducted prior to the flood event are shown by solid symbols and post-flood surveys are shown as outlined symbols.

Analysis was conducted to assess the relationship between hydraulic gradient and specific discharge in the BPA before and after the flood event. A subset of the specific discharge estimates was used, specifically those within a 10 m zone of the cable where the hydraulic gradient transects intercepted the cables. Overall, three distinct clusters were found (Figure 7): HDD1 pre-flood, HDD1 post-flood and HDD2 surveys. For HDD1, there is a distinct increase in specific discharge values following the flood event, accompanied by an increase in hydraulic gradient (Figure 7). Seasonal patterns were hard to identify, however hydraulic gradients were higher in summer which coincided with low specific discharge values (Figure 7). In contrast, HDD2 exhibited little change pre and post flood, and also showed no distinguishable seasonality (Figure 7).

Hydraulic conductivity, K (m d^{-1}), was calculated by dividing specific discharge estimated using A-DTS by the horizontal hydraulic gradient of the BPA. For HDD1 the K ranged from 399 to 1,347 m d^{-1} (average 953 m d^{-1}), while for HDD2 the K ranged from 434 to 1,286 m d^{-1} (average 682 m d^{-1}). These values are slightly higher than previous estimates of K at the site from aquifer tests of 75 to 250 m d^{-1} .

5. Discussion

5.1. Interannual Trends in Specific Discharge

The interannual results show modest temporal change in specific discharge beneath the braided river, with 20.4% variance relative to the median on HDD1 (Figure 8a) and 13.3% variance relative to the median on HDD2 over the

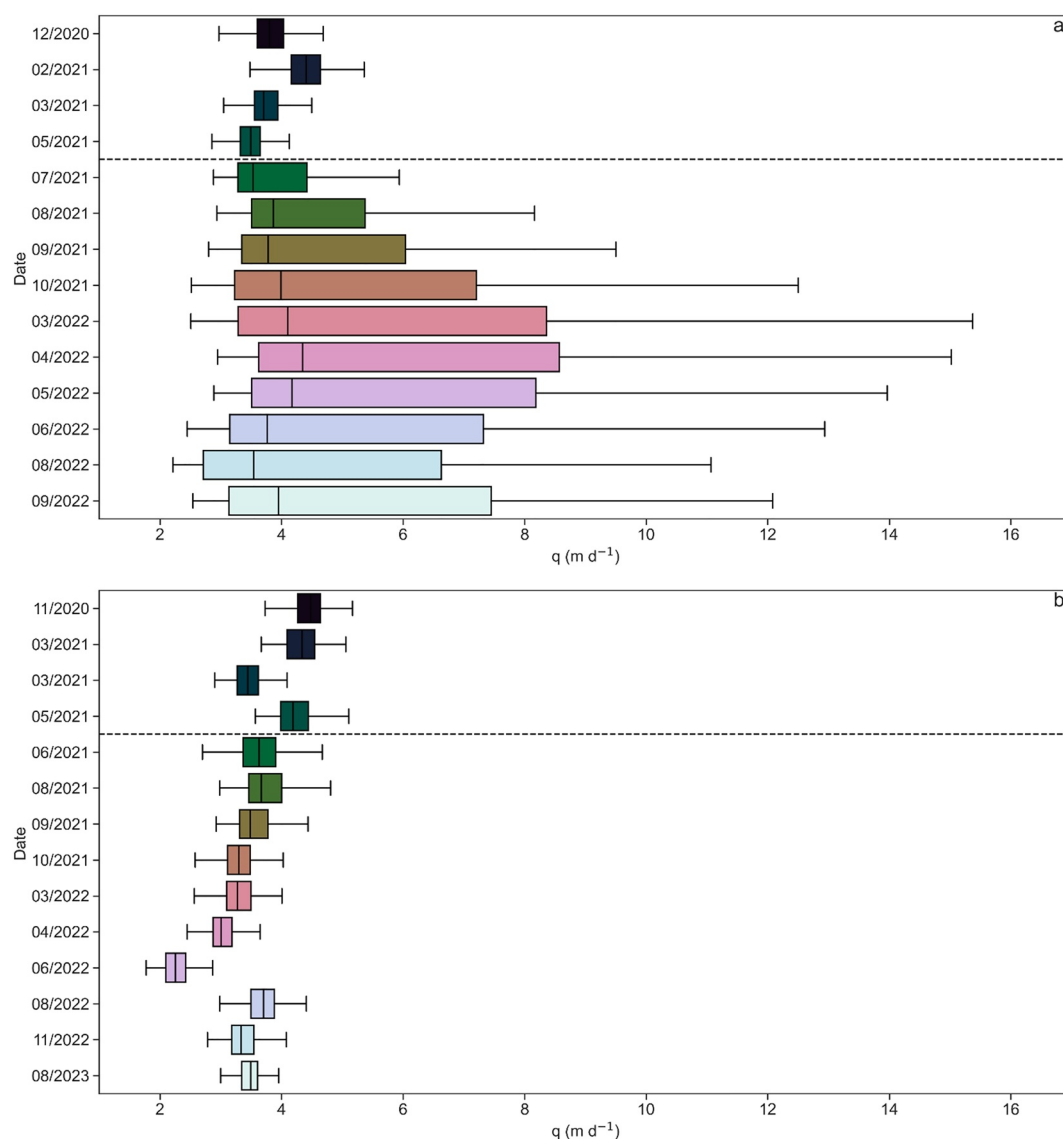


Figure 8. Box and whisker plot illustrating the average specific discharge estimated from all surveys conducted on (a) HDD1 (BB') and (b) HDD2 (AA') across the saturated portion of the cable. Dashed line indicates the date of the flood.

nearly 3-year study period (Figure 8b). Despite the larger absolute range of average specific discharge values for surveys on HDD2, the relative dispersion within the interquartile range was lower compared to HDD1. The surveys conducted on HDD1 showed increased variability in specific discharge over time, with wider interquartile ranges and numerous outliers, suggesting more fluctuations in specific discharge values for post-flood surveys. In contrast, surveys conducted on HDD2 had a more stable distribution of specific discharge values, with a smaller and more consistent interquartile range. Additionally, there is an increasing trend in specific discharge values on HDD1 post-flood for a 1-year period (Figure 8a), but a decreasing trend in HDD2 values during the same period (Figure 8b). These substantial variations in specific discharge highlight the dynamic nature of the BPA system over longer time scales.

We postulate that the decreasing trend in specific discharge values on HDD2 (Figure 8b) could be due to the secondary river channel, which HDD2 travels under, transitioning from a losing to a gaining channel. The secondary channel is intermittent in nature; this is supported by photographic evidence (Figures S2 and S3 in Supporting Information S1) showing the secondary channel changing from empty to flowing after the flood event. Unfortunately, the flood destroyed the river monitoring equipment, limiting our ability to quantitatively verify

this change. However, this transition could indicate a reversal in the predominant direction of groundwater fluxes intercepting the cable in this location, as floods have been shown to cause temporary reversal of hydraulic gradients (Dudley-Southern & Binley, 2015).

The interannual surveys highlight the complexity of flow within the BPA, and the value and insights that can be gained from high-resolution measurements to characterize hydrological processes, capturing channel-scale processes at a high-spatial resolution. Overall, the divergence in trends between HDD1 and HDD2 reinforces the spatial heterogeneity of groundwater flow between river channels and more broadly within the BPA.

5.2. Association Between Specific Discharge and River Flow

The comparison between specific discharge and river flow revealed an unexpected lack of association. During the 3-year study period, river flow at Whitecliffs was highly dynamic exhibiting an episodic temporal regime. e.g., during the May 2021 flood, river flow reached $225 \text{ m}^3 \text{ s}^{-1}$, approximately 90 times greater than the median river flow. Despite this large variation in river flow, the response (i.e., changes in specific discharge) in the BPA was muted and complex. For example, there was episodic flood-related variation in specific discharge following the May 2021 flood and a 30.5% increase in specific discharge for surveys conducted on HDD1 after the flood (Figure 8a) and a 17.6% decrease in specific discharge for surveys on HDD2 after the flood (Figure 8b). While the predominant flow direction in the BPA is likely to be horizontal, there is insufficient evidence to conclusively determine the flow direction from the horizontal A-DTS cables.

In the study by Simon et al. (2024) they found that there was limited variation in hyporheic fluxes, despite large variation in river flow. In that study, which also used A-DTS, the focus was quantifying flow within the hyporheic zone, in the upper 0.1 m of streambed sediments, longitudinally with the river. As such it differs from this study because here, we are quantifying parafluvial flow within the BPA at depths of up to 5 m, laterally beneath a braided river. By comparison at the Rio Grande changes in groundwater flux resulting from river stage were <20% (Rankin et al., 2013). Nevertheless, our results also showed a limited relationship between river flows and groundwater specific discharge within the BPA.

5.3. Evolution of Spatial Variation in Specific Discharge

Specific discharge values measured beneath the left margin, midpoint and right margin of the river channels using the HDD1 and HDD2 cables are shown for the 3-year study period in Figure 9. It can be seen that the HDD1 cable showed relatively large spatial variation compared to that measured on HDD2. Spatial variation of specific discharge measured with HDD1 increased following the May 2021 flood event, with increases occurring under the left and right margins of the main channel.

It is interesting to note that the first post-flood survey conducted on HDD1 on 8/07/2021, approximately five weeks after the flood, showed an increase in specific discharge beneath the left margin (Figure 9a). However, it was not until the following survey on HDD1 on 11/08/2021 that an increase in specific discharge was measured beneath the right margin. These increases in specific discharge remained elevated for the remainder of the surveys.

The pre-flood spatial patterns of specific discharge appear to have been amplified post-flood (Figure 10). This is particularly evident for the left margin. The spike at 87 m along the cable for the survey conducted on 14/03/2022 is a particular outlier and could potentially be the activation of a preferential flow path. To assess how these zones of increased flow beneath the margins of the main channel developed over time, the inset of Figures 10a and 10b shows the integral under the curve of specific discharge versus distance along the cable, which quantifies the volumetric flux within a given zone. This indicates an increase post-flood both on the left and right margins. The last four to five surveys show a decrease in specific discharge along the cable of volumetric fluxes beneath the left and right margins. However, these values do not return to pre-flood levels.

Following the May 2021 flood event there was an abrupt change in bathymetry that is, the active river channel expanded in depth and width in the south-west direction. There was also a slight change observed in the saturation boundary along the cable, which is an indicator of the lateral expansion of the BPA. This fits with the conceptual models of Wilson et al. (2024) who found a direct connection between the water levels in the BPA and the Waikirikiri Selwyn river.

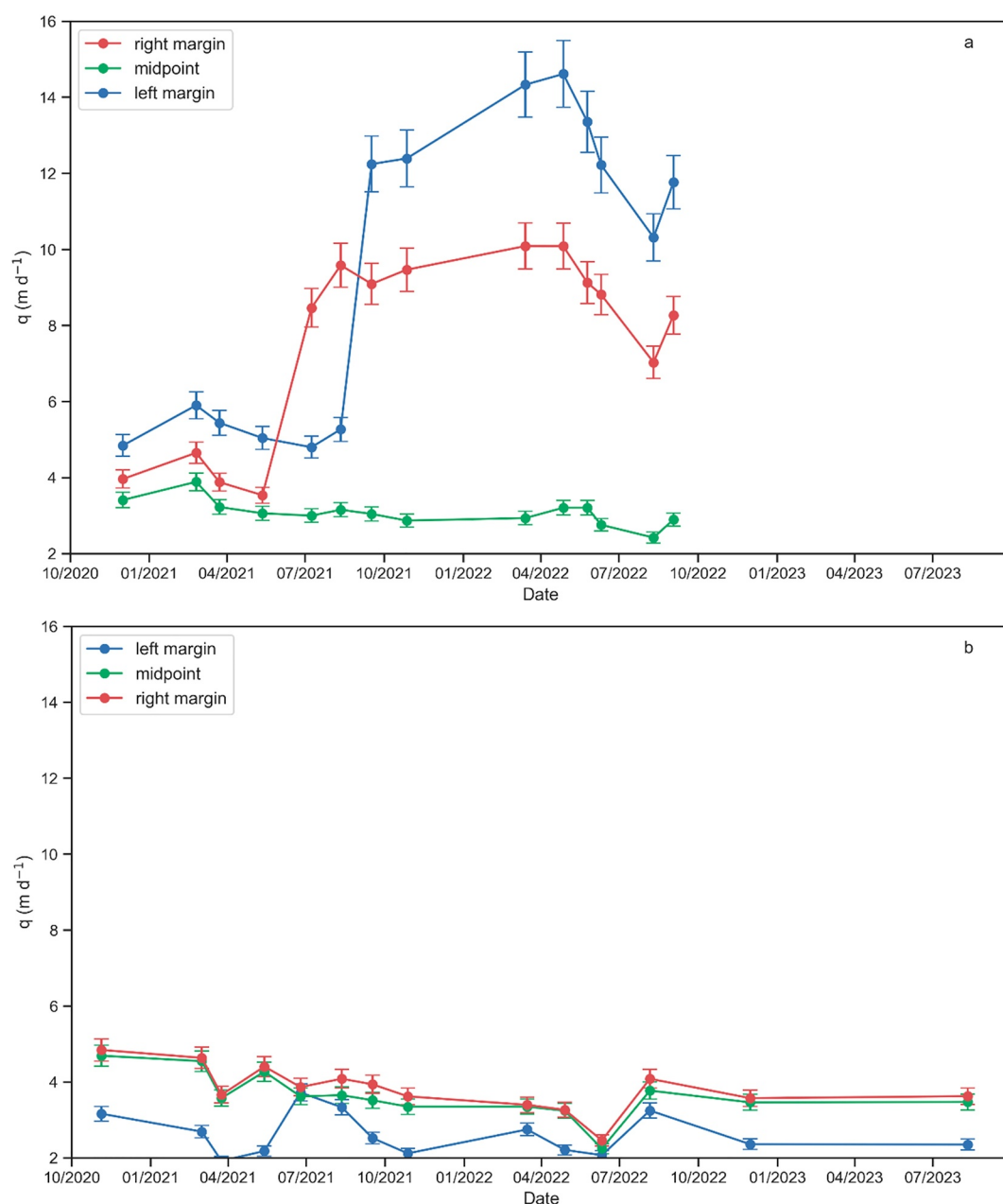


Figure 9. Specific discharge over the 3-year study period at selected locations along (a) HDD1 (BB') beneath the left margin in blue, midpoint in green and the right margin in red and (b) HDD2 (AA') beneath the left margin in blue, midpoint in green and the right margin in red. Values are averaged over the saturated portion of the cable. 6% error bars (as per Sai Louie et al., 2024) are shown.

5.4. Geomorphological and Hydraulic Drivers of Change in Specific Discharge

The changes in specific discharge observed over the 3-year period are likely caused by geomorphological changes occurring at the site. Several studies have found geomorphological changes to be key drivers of changes to hyporheic exchange (Dudley-Southern & Binley, 2015; Simon et al., 2024; Wu et al., 2018). However, to our knowledge this is the first study to show the relationship between geomorphological changes and parafluvial flow over an extended period of time. The most significant increases in specific discharge appear to occur beneath the locations of greatest geomorphic change, particularly where there was a reduction in the thickness of overlying sediments or the presence of remnant paleochannels. For example, there is evidence of an increase in specific

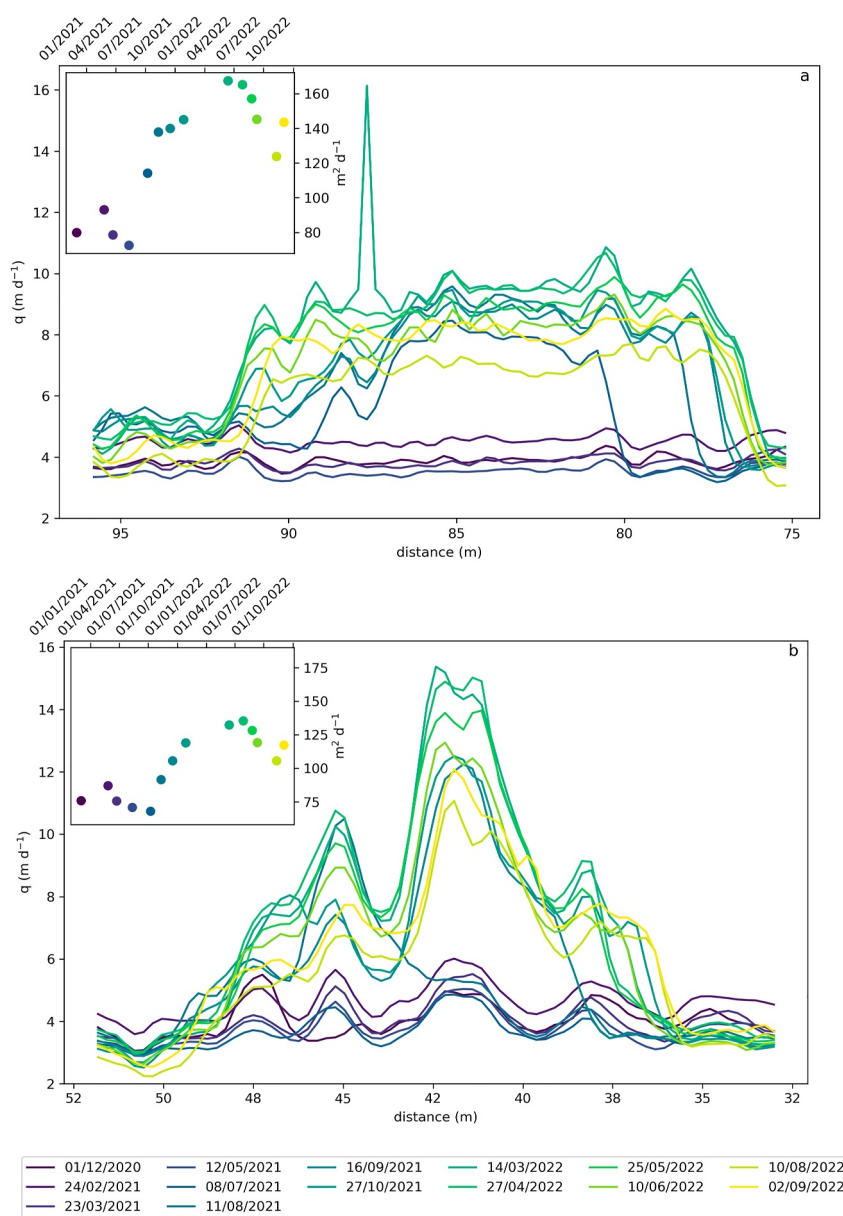


Figure 10. Evolution of the specific discharge values through time beneath the (a) right margin and (b) left margin on HDD1. Graphs are oriented facing downstream. The insets show the integral of the specific discharge with respect to the distance (i.e., the cumulative discharge over each section for each survey), shown in colors corresponding with the survey days shown in the legend, where darker colors indicate the start of the 3-year study period, and lighter colors depict surveys conducted toward the end of the survey period.

discharge at the midpoint of the main channel (between 64 and 68 m along cable HDD1), which may be representative of an old or remnant braid with higher hydraulic conductivity sediments. Additionally, when the main river channel migrated south-west following the flood, a remnant braid appears to have been connected to the active river channel. This, we theorize, contributed to the increase in specific discharge observed beneath the left margin. Also, there was a thinning of overlying sediments above the cable at the margins of the main active river channel post-flood, which likely also contributed to the measured increases in specific discharge. In contrast, the secondary channel experienced little geomorphic change and there was minimal change in measured specific discharge. Based on the twelve back-to-back A-DTS surveys carried out over 24-hr, Sai Louie et al. (2024) found that specific discharge patterns followed river morphology. The findings from this study are consistent with that observation. While we have no direct evidence from this study, it is possible that rearrangement of the riverbed

sediments and especially the removal of the surface armor layer (which occurs in the first tens of centimeters of the riverbed) post-flood caused changes to porosity and hydraulic conductivity, leading to the observed changes in specific discharge. However, there is little understanding of this process in braided rivers (Wilson et al., 2024).

It would be expected that an increase in specific discharge post-flood would be accompanied by an increased in hydraulic gradient, however this was not observed. The results showed that specific discharge under the main channel increased post-flood and this was accompanied by a decrease in hydraulic gradient. This showed that the flood induced significant changes affecting the subsurface flow dynamics. Under the secondary channel, there was limited change in specific discharge and also hydraulic gradient. The exception is one survey in Winter, which had the highest hydraulic gradient.

Independent Darcy flux estimates were calculated using the measured K values from aquifer tests (75–250 m d⁻¹) multiplied by the horizontal hydraulic gradient of the BPA, with the resultant fluxes ranging from 0.6 to 1.9 m d⁻¹. The values are lower than the specific discharge estimates derived from the A-DTS surveys (HDD1 average q from all surveys 4.8 m d⁻¹ and for HDD2 average q 3.6 m d⁻¹). This discrepancy between the K values derived from aquifer tests and A-DTS derived K values highlight anisotropy in the BPA, providing evidence that the groundwater flux is predominantly horizontal. To further evaluate this, a Darcy flux was calculated by multiplying the average K values calculated in Section 4.6 (953 and 682 m d⁻¹ for HDD1 and HDD2, respectively) by the horizontal hydraulic gradient of the BPA. The resultant flux for HDD1 was 7.2 m d⁻¹, significantly higher than the A-DTS derived flux of 4.7 m d⁻¹, suggesting vertical groundwater fluxes in this location. Whereas for HDD2 the Darcy estimate was 3.9 m d⁻¹, which is comparable to the A-DTS derived q of 3.6 m d⁻¹, indicating predominantly horizontal flow in this location. Antecedent moisture conditions are known to significantly influence subsurface hydrological dynamics. Bennett et al. (2018) demonstrated that flood volumes varied by up to 50% depending on antecedent moisture conditions prior to an event. In our study, the antecedent moisture conditions prior to the flood from October 2020 to May 2021 were very dry, as indicated by the low groundwater levels in Figure 5. The May 2021 flood event was followed by a predominantly wet period, which appeared to influence specific discharge within the BPA.

5.5. Limitations and Sources of Uncertainty

The specific discharge values estimated in this study represent flow within the BPA and constitute parafluvial flow, capturing channel-scale processes at a high-spatial resolution. The approach used in this study assumes that specific discharge estimates from A-DTS surveys are one-dimensional. However, in reality, subsurface flow beneath the river is multidirectional and is spatially complex that is, the presence and reactivation of remnant paleochannels (Sai Louie et al., 2024; Wilson et al., 2024). Estimates from a numerical model indicate that the horizontal flux component is approximately one order of magnitude greater than the vertical flux component within the BPA at the study site (Di Ciacca et al., 2024). Therefore, the specific discharge estimates within the BPA presented in this study likely have a dominant horizontal flux component, as well as a vertical flux component contributing to the measured values.

Another source of uncertainty is the exact cable locations, and the base of the BPA. The elevation of the buried cables have an accuracy of ± 0.25 m and is therefore reasonably well constrained. The base of the BPA has been interpolated using 15 well logs and additional drilling would improve the accuracy of this surface. We are confident nonetheless that the cables are primarily within the BPA.

In this study there is a range of measured specific discharge values over the 3-year period, primarily reflecting changes that occurred due to flooding. We are confident that these variations are above measurement error, which was evaluated in an earlier study by Sai Louie et al. (2024) to be 6%. Short-term riverbed variability was not quantified in this study; however, the high flow events captured in the study were observed to mobilize the streambed and sediment deposition was observed during low flow events (Simpson & Meixner, 2012). Sediment deposition could contribute to changes in hydraulic conductivity, which in turn influence specific discharge. Caution must be exercised when generalizing the insights from this study to other study sites that have different geomorphology, subsurface conditions, or hydrological history. The character and controls of BPA processes we have identified here may elsewhere be stronger, weaker or entirely absent. Opportunities for further work is possible however there was a high cost to install cables horizontally below the river. Installing vertical or sub-vertical cables would be more economical and could help upscale the findings of this study to the reach scale.

6. Conclusions

This study presents a unique 3-year data set of specific discharge measured within the BPA, beneath the braided Waikirikiri Selwyn river. The data was collected during 28 surveys, conducted approximately monthly, on two A-DTS cables installed up to ~5 m depth below the river channel. Cable HDD1 was beneath the main channel of the braided river, HDD2 was below the secondary channel.

Specific discharge estimates, averaged over the entire saturated length of the cable, showed different interannual trends for the 3-year study period. Specific discharge estimates on HDD1 varied between 3.5 and 5.9 m d⁻¹ (20.4% variance relative to the median). In comparison, specific discharge estimates on HDD2 ranged from 2.4 to 4.5 m d⁻¹ (13.3% variance relative to the median).

The river flow changed considerably over the 3-year period, varying by 2.5 orders of magnitude and demonstrating a flashy response to two flood events. These changes in river flow caused substantial geomorphological changes at the site. Despite, at times, dramatic change in river flow and levels, there was negligible relationship with specific discharge estimates. Specific discharge also showed no clear seasonal trends.

Spatial variation in specific discharge was relatively large on HDD1 compared to HDD2 over the 3-year period. The spatial variation of specific discharge measured on HDD1 increased post-flood with increases measured beneath the left and right margin of the main active river channel. These increases in specific discharge were, somewhat surprisingly, accompanied by a small decrease in hydraulic gradient between the river and the BPA groundwater levels. The cause of the increase in specific discharge is likely related to a post-flood thinning of the sediments at the margins of the main active river channel.

Overall, the data reveal complex and often counterintuitive trends in specific discharge in the BPA. Potential drivers of specific discharge changes in the BPA include river flow, geomorphological alteration and hydraulic gradient. Geomorphological changes following a flood event were shown, in this study, to be of greatest impact on specific discharge. The greatest changes in specific discharge occurred beneath the river margins.

Data Availability Statement

The A-DTS data used for quantifying specific discharge within the BPA in this study are available at <https://figshare.com/articles/dataset/Dataset/27328710> with license CC by 4.0. This link is being made available for the editor and reviewers for review purposes only and we kindly ask that it is not distributed. The data repository will be made publicly available upon publication.

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