Conceptualising surface water-groundwater exchange in braided river systems

Scott R. Wilson[1], Jo Hoyle[2], Richard Measures[2], Antoine Di Ciacca[1], Leanne K. Morgan[3], Eddie W. Banks[4], Linda Robb[1], and Thomas Wöhling [1,5]

[1]Environmental Research, Lincoln Agritech, Lincoln University, New Zealand
[2]National Institute of Water & Atmospheric Research, Christchurch, New Zealand
[3]Waterways Centre for Freshwater Management, University of Canterbury, Christchurch, New Zealand
[4]National Centre for Groundwater Research & Training & College of Science & Engineering, Flinders University, Australia
[5]Chair of Hydrology, Technische Universität Dresden, Dresden, Germany

Correspondence to: Scott R. Wilson (scott.wilson@lincolnagritech.co.nz)

Abstract

Braided rivers can provide substantial recharge to regional aquifers, with flow exchange between surface water and groundwater occurring at a range of spatial and temporal scales. However, the difficulty of measuring and modelling these complex and dynamic river systems has hampered process understanding and the upscaling necessary to quantify these fluxes. This is due to an incomplete understanding of the hydrogeological structures which control river-groundwater exchange. In this paper we present a new conceptualisation of subsurface processes in braided rivers based on observations of the main losing reaches of three braided rivers in New Zealand.

The conceptual model is based on a range of data including lidar, bathymetry, coring, particle size distribution, groundwater, and temperature monitoring, radon-222, electrical resistivity tomography, and fibre optic cables. The combined results indicate that sediments within the recently active river braidplain are distinctive, with sediments that are poorly consolidated and better sorted compared to adjacent deposits from the historical braidplain which become successively consolidated and intermixed with flood silt deposits due to overbank flow.

A distinct sedimentary unconformity, combined with the presence of geomorphologically distinct lateral boundaries, suggests that a “braidplain aquifer” forms within the active river braidplain through the process of sediment mobilisation during flood events.

This braidplain aquifer concept introduces a shallow storage reservoir to the river system, which is distinct from the regional aquifer system, and mediates the exchange of flow between individual river channels and the regional aquifer. The implication of the new concept is that surface water-groundwater exchange occurs at two spatial scales. The first is hyporheic and parafluvial exchange between the river and braidplain aquifer. The second is exchange between the braidplain aquifer and regional aquifer system. Exchange at both scales is influenced by the state of hydraulic connection between the respective water bodies. This conceptualisation acknowledges braided rivers as whole “river systems”, consisting of channels, and gravel aquifer reservoir.

This work has important implications for understanding how changes in river management (e.g., surface water extraction, bank training and gravel extraction) and morphology may impact groundwater recharge, and potentially on flow, temperature attenuation, and ecological resilience during dry conditions.

1.0 Introduction

This study is motivated by the need to understand processes and quantify losses from braided river systems to alluvial aquifers. In New Zealand more than 150 river systems are thought to have braided reaches (Brower et al., 2024), which provide a substantial component of recharge to alluvial aquifers. For example, in the case of the Wairau Aquifer, it has been estimated recharge is sourced almost exclusively from the Wairau River (Wöhling et al. 2018).

Braided rivers are spatially complex, dynamic hydrologic environments which are not easily measured using field techniques or represented by numerical models. The determination of flow exchanges between the river and groundwater in such a hydrologically complex and morphologically dynamic environment is difficult because of their spatial and temporal flow complexity, numerous and changing number of channels, dynamic bathymetry, coarse substrate, and tendency for loss of field installations during high flow events. The complexity of this challenge demands a collaborative effort that leverages the strengths of a diverse range of disciplines to develop a comprehensive and holistic understanding of the problem.

Many authors have studied surface-aquifer exchanges, and several reviews have been conducted on available techniques for measuring exchanges for alluvial rivers in general (Kalbus et al. 2006, González-Pinzón et al. 2015, Brunner et al. 2017), ephemeral rivers at different spatial scales (Banks et al., 2011, Shanafield and Cook 2014), and braided rivers (Coluccio and Morgan, 2019). Based on these reviews, there is a tendency for previous studies to focus on quantifying exchange fluxes prior to understanding the hydrological processes that influence the exchange. Ward and Packman (2019) suggest that despite the large amount of research on river-groundwater exchanges, an accurate predictive, transferable understanding of the river corridor is lacking. Indeed, a conceptualisation of how braided rivers relate to their underlying groundwater systems is lacking (Coluccio and Morgan, 2019). This has led to ambiguity in what measured exchanges represent, and difficulty in trying to represent braided rivers within aquifer scale models.

The aim of this paper is to formulate a conceptualisation of braided river exchange processes at both the local (reach) and sub-catchment (aquifer) scale. This conceptualisation provides a framework for both providing context for field measurements of flow exchange to be interpreted, and the potential for representation of local scale processes in sub-catchment models. Incorporating these local scale processes is vital to predict how changes in the river system can impact groundwater recharge.

The conceptualisation presented here was developed based on field observations but, for clarity of explanation, we introduce the conceptual framework first and then present the supporting evidence. Our working definition for hyporheic exchange is local bed-scale interaction that occurs within a single channel (e.g., a single riffle), whereas parafluvial exchange occurs between individual channels at larger scales (across a bar or further). We also distinguish between a river as a series of wetted channels, and a “river system”, which consists of wetted channels plus subsurface flow through the associated braidplain gravels. “Braidplain” refers to the lateral extent occupied by the river braids, old bar surfaces and abandoned channels (Warburton, 1996; Gray et al., 2016; Brower et al., 2023). The extent of wetted channels and recently reworked bed material (bare gravel) at a given point in time defines the “active braidplain”, which also has potential to shift laterally. Lateral adjustment of the active braidplain may be limited by hillslope margins, terraces or, in managed rivers, by rock revetments or artificial stop banks (levees) which are typically protected with vegetative buffers. We refer to the extent within which our study rivers can currently adjust as the “contemporary braidplain”, acknowledging that in two of our study rivers (Wairau and Ngaruroro) the contemporary braidplain margins are controlled by engineered flood defences which have narrowed the natural braidplain such that almost the entire contemporary braidplain is active.

2.0 Review of existing concepts

The prevailing conceptualisation for gravel bed rivers in general consists of a surface channel with an associated bed with some hydraulic resistance (Schälchli 1992, Wu and Huang 2000) which exchanges water with an associated fluvial or alluvial aquifer (Stanford and Ward 1993, Poole and Berman 2001). Within the alluvial aquifer lies a hyporheic zone which functions as an interface between groundwater and surface waters (Stanford and Ward 1993, Poole and Berman 2001, Boano et al. 2014). The extent of the hyporheic zone is defined by its function or process of interest, which can be physical, chemical, biological, or a combination of functions (Ward 2015). Accordingly, the vertical or lateral boundaries of the hyporheic zone are transient and flexible, and not easily defined spatially (White 1993, Boulton et al. 1998, Ward 2015). From a hydrological perspective, the hyporheic zone has been defined as the extent to which surface water enters the high porosity subsurface beneath and lateral to a stream and returns to the stream surface farther downstream (Harvey and Wagner, 2000). Authors have suggested the hyporheic zone as extending from the upper few centimetres of sediment (Boulton et al. 1998, Sophocleous 2002) to larger scales (km^3) constituting a hyporheic corridor (Stanford and Ward 1993). Valett at al. (1996) predict the extent of the hyporheic zone to be related to catchment lithology, with interaction being more extensive in sites with higher alluvial hydraulic conductivity, whereas Boano et al. (2008) predict the infiltration depth to be related to bedform for any given hydraulic conductivity.

An additional, ecological, concept is that of a riparian zone (Steiger et al. 2005). This extends river margins beyond the active channel to include the biosphere supported by and including recent fluvial landforms and inundated or saturated by bank discharge (Hupp and Osterkamp 1996). Other authors have considered the presence of a parafluvial zone situated between the hyporheic zone (beneath the river channel) and riparian zone (Holmes et al. 1994) which accommodates longer flow paths within the alluvium adjacent to the stream (Bourke et al. 2014, Cartwright and Hoffmann 2016). Our interpretation of the parafluvial zone is that it constitutes exchange flow within the alluvial aquifer at spatial and temporal scales beyond what is considered hyporheic.

From a hydrological perspective of braided rivers, the framework that emerges from the prevailing concepts is one where river-groundwater exchange occurs within an alluvial aquifer which conveys both hyporheic and parafluvial flowpaths. The proportion of these flowpath components theoretically depends on the degree to which there is a net loss or gain in river channel flow. However, the interpretation of river-groundwater exchanges becomes challenging in a braided river which comprises multiple channels within an alluvial aquifer. Harvey and Gooseff (2015), Barthel and Banzhaf (2016) and Ward and Packman (2019) propose that exchange fluxes be considered at different spatial scales: point, local, sub-catchment, and regional. Following this approach, exchange within individual river braids or channels can be considered to occur at the point-scale, and the sum of all braids within a river reach as a local-scale interaction. While process understanding can be observed and fluxes quantified at the point and local scale, it is imperative to enable an upscaling of observed processes so that fluxes can be estimated for at least the sub-catchment (aquifer) scale.

Previous work on subsurface structure in braided rivers has tended to focus on the role of bed material heterogeneity. A significant body of literature exists to describe braided river deposits via morphology (Huber and Huggenberger 2016), sedimentology (Huggenberger and Regli 2006, Theel et al. 2020), geophysics (Pirot et al. 2019), and modelling approaches (Pirot et al. 2014; 2015, Brunner et al. 2017, Schilling et al. 2022). To date, no conceptual model has been posed for how a braided river and its associated braidplain gravels (alluvial aquifer) relate to those of the underlying regional aquifer. While the structural components of river-groundwater interaction have been identified by previous authors (e.g. Poole and Berman 2001, Steiger et al. 2005), the identification of clear spatial boundaries between structural elements has been missing. From a hydrological perspective of understanding surface water-groundwater interaction, this creates a problem of where the river system ends, and where the regional groundwater system begins. The uncertainty related to this lack of spatial definition transfers to the interpretation of field data, whether a sample is representative of river channel flow, alluvial aquifer (hyporheic or parafluvial zones), or regional groundwater.

Consequently, representation of braided rivers in numerical models is problematic since their complexity is not readily captured by a simple conceptualisation. Water exchanges can potentially be simulated realistically using a fully coupled hydrological model such as HydroGeoSphere (Therrien et al. 2010, Brunner and Simmons 2012). However, the data required to parametrise such a model, and the computational demands of the detailed mesh required to simulate exchanges in braided rivers make this approach only suitable for point and local scale studies. Furthermore, braided river morphology is so dynamic that a new bed morphology would be required following each significant flood event. In recent years, two approaches to simulate the transitions of dynamic bed morphology and sediments on river-groundwater exchanges have been tested. The first approach applied the ensemble Kalman filter and areal imagery to assimilate river bed topography and to update aquifer hydraulic conductivities in a HydroGeoSphere model for a 2-km section of the Emme River in Switzerland (Tang et al. 2018). The data assimilation scheme strongly improved predictions of post-flood hydraulic states of the system. The second approach proposed a pilot point parametrization scheme where both the aquifer properties (hydraulic conductivity) and the location of the pilot points are inferred, e.g. from river-bed training images (Khambhammettu et al. 2020). The corresponding Traveling Pilot points (TRIPS) scheme could potentially be used to describe the transition between discrete states of river morphology. To some extent these approaches enable the application of fully integrated 3D models in dynamic river environments of appropriate scale, although their application in a larger river system or at a larger scale is untested.

The simple river structure offered by the Streamflow-Routing (SFR, Niswonger and Prudic, 2005) and River (Harbaugh, 2005) packages in MODFLOW represent the river as a flux boundary condition with vertical flow impedance in the bed expressed by a lumped parameter termed ‘streambed conductance’. Previous authors have shown that the concept of a streambed resistance concentrating all pressure losses, as implemented in MODFLOW, is questionable in many cases (Anderson 2005, Rushton 2007, Morel-Seytoux et al. 2018, Di Ciacca et al. 2019). Moreover, even if such a streambed exists, a major physical issue with a lumped parameter approach is that streambed conductance values in the field are not homogenous, but vary spatially (Cardenas and Zlotnik 2003, Zhou et al. 2014, Pryshlak et al. 2015, Laube et al. 2018) and temporally (Levy et al., 2011, Wu et al. 2015). In a braided river, bed material typically consists of a heterogeneous mixture of cobbles, gravels, and sands, which can have similar characteristics to alluvial sediments located several metres beneath the bed. Therefore, the streambed conductance concept seems inappropriate to represent surface water-groundwater exchange in braided river systems.

Different modelling approaches have been trialled to represent braided rivers in New Zealand. White et al. (2012) conducted a steady state water balance approach to determine flow losses for a reach of the Waimakariri River. Exchanges between individual channels and the adjacent alluvial aquifer were determined via mass balance, although the subsurface components of the exchange were not explicitly described. An alternative approach by Wöhling et al. (2018, 2020) simulated dynamic Wairau River exchanges at the sub-catchment scale using MODFLOW. In this case, the braided nature of the river was not considered, and the river was represented by the SFR package using a stage-width-flow relationship derived from a representative channel morphology. While this model fitted river flux and groundwater level data well, the approach employed a streambed conductance model, which is difficult to reconcile with the river morphology and bed sediment seen in the field. A particular drawback of the SFR package is its inability to represent the hyporheic or parafluvial exchange fluxes observed at the point or local scale. While the Waimakariri and Wairau modelling studies are relatively comprehensive, in both cases, an understanding of subsurface structure is missing from the river representation. This lack of knowledge about the structural controls on subsurface flow in the braided river environment needs clarification to understand what measured and modelled river-groundwater exchanges represent. In doing this, a more physically realistic method for representing braided rivers in numerical surface water-groundwater flow models may be achieved.

3.0 Proposed conceptualisation

A conceptual framework is proposed which captures the key elements of water exchanges in a braided river system. This conceptualisation builds on the previous work of Fox and Durnford (2003) and Brunner et al. (2009a and b) and recognises that the hydrological controls on river-groundwater exchange occur at two distinct interfaces within a braidplain system. Specifically, these two exchange processes can be summarised as:

1. River channel ↔ braidplain aquifer (hyporheic and/or parafluvial exchange)

2. Braidplain aquifer ↔ regional aquifer (“river system” - groundwater exchange)

The first exchange interface is within the active braidplain, and occurs between individual river channels (braids) and the local shallow water-table in the river bed sediments, and occurs at the point or local scale. We term the water stored within these river bed sediments the “braidplain aquifer” (BPA), which facilitates hyporheic and parafluvial flow. For a perennially flowing river, the BPA will retain some degree of saturation throughout the year, although unsaturated conditions may occur in the case of intermittent or ephemeral rivers if there are prolonged periods with no river flow. Individual river braids can be in hydraulic connection (gaining or losing), disconnection, or a transitional state relative to the BPA. At the active braidplain interface, all possible river exchange processes can occur regardless of the hydraulic relationship between the BPA and surrounding regional water table. The second exchange interface occurs at both the local (reach) and sub-catchment (aquifer) scale, between the BPA and regional water table.

Central to this conceptualisation is the presence of a distinct BPA immediately beneath the river surface which facilitates exchange at these two interfaces. A similar term “braided-river aquifer” has been used by previous authors (Pirot et al. 2015), although we have not found an associated definition. The BPA functions as a storage medium to exchange water between the river and regional aquifer in braided river systems. Direct exchange of water between the river and regional aquifer can only occur if the river channel is in direct contact with the regional aquifer (i.e., where the surface water extends to the boundary of the BPA, for example where a channel follows the lateral margin of the BPA, or there is a connection at the base of a deep scour pool).

A key feature of the BPA is its extremely high transmissivity, which is a product of the highly dynamic nature of braided rivers. Bedload transport during flood events causes braids to form, migrate, and be abandoned; processes that re-work the river bed sediments (Bristow and Best, 1993; Reinfelds and Nanson, 1993). This reworking process strips most fine sediment (silts and clays) from the river bed, as the shear stresses during floods are too high for these size fractions to be deposited except in vegetated areas and backwaters. The result is a braidplain deposit of high transmissivity lag gravels. High transmissivity combined with the relatively large bed slope of braided rivers and the presence of multiple braids for water exchange produces a groundwater flow path which is sub-parallel to the contemporary braidplain, regardless of the regional groundwater flow direction.

Conceptual diagrams have been drawn for situations where the BPA is hydraulically disconnected from (Figure 1a) and connected to (Figure 1b) the regional aquifer. The first case (Figure 1a) consists of three functional lithological layers representing a BPA which is hydraulically disconnected from an underlying regional aquifer by an unsaturated zone. In this case, water moves vertically from the BPA to the regional aquifer under a unit hydraulic gradient. A minimum of three layers is required for an unsaturated zone to develop (Fox and Durnford, 2003), and there must be a sufficient transmissivity contrast between the impeding horizon and underlying sediments (Brunner et al., 2009a and b). It should be noted that braided river deposits are lithologically variable and complex, and in most cases only two functional layers are required due to stratification in the underlying sediments.

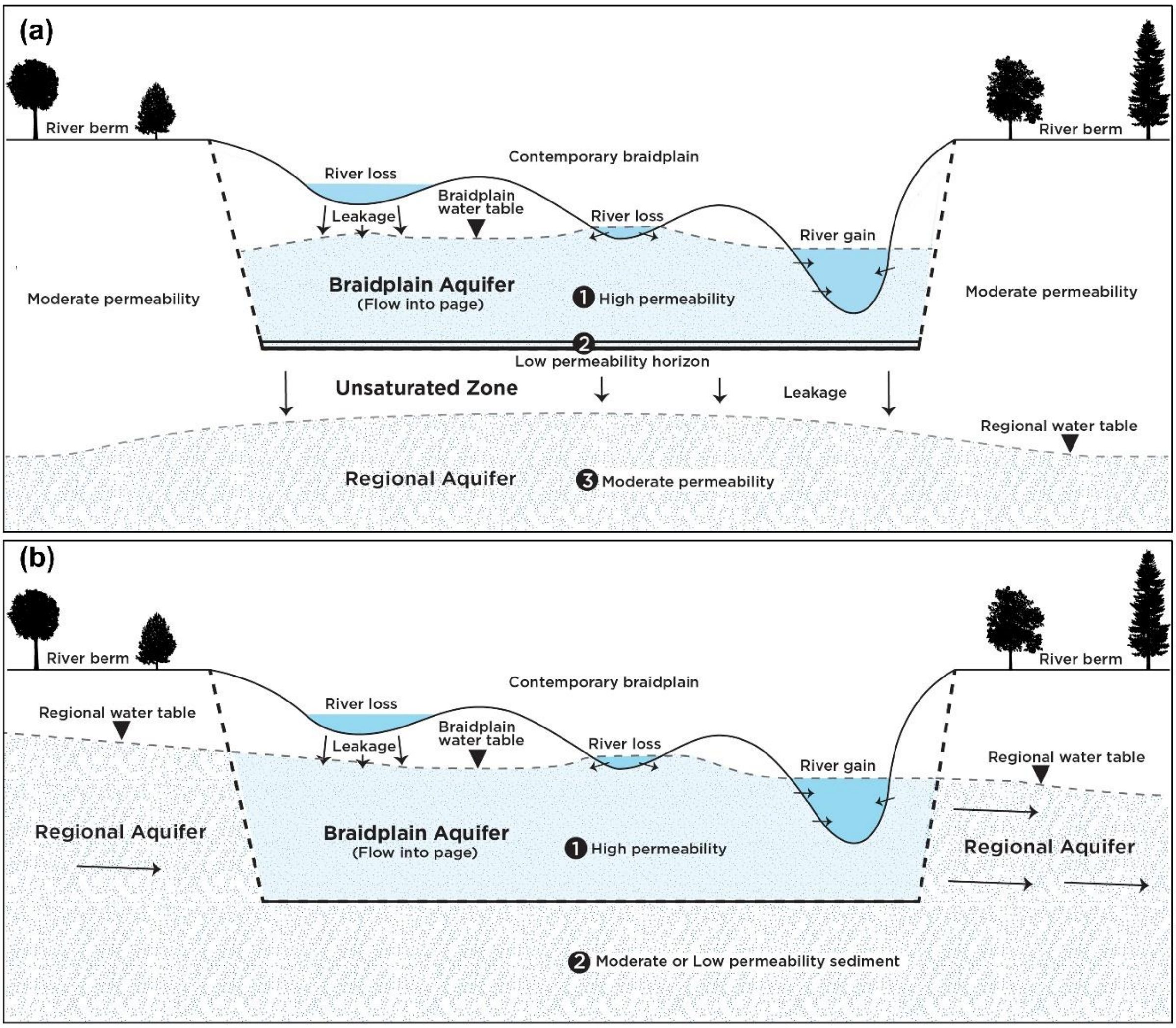

Figure 1. Conceptualisation for a braidplain aquifer which is (a) hydraulically disconnected from or transitional with the regional groundwater system, and (b) hydraulically connected to the regional groundwater system.

A hydraulically disconnected river system setting shares features in common with intermittent or ephemeral rivers (Shanafield et al. 2021). Infiltration to the regional aquifer is regulated by the vertical resistance of lower permeability sediments and the hydraulic head in the BPA. If the BPA is fully saturated across the contemporary braidplain, the rate of infiltration to the regional aquifer will be steady because the braidplain has reached a maximum wetted area. In this condition, some minor temporal infiltration variability will occur due to changing water levels in the BPA. Under ephemeral conditions, the saturated extent of the BPA decreases longitudinally, and locally laterally, during drying phases in response to extended periods of low river flow. The combined reduction in head and saturated area of wetted braidplain result in considerably less infiltration to the regional aquifer (Di Ciacca et al. 2023).

The second case is a setting where the river is hydraulically connected to the regional aquifer system (Figure 1b). A minimum of two lithological units are present, a high transmissivity BPA and regional aquifer (1) overlying less permeable sediments (2) which impede vertical flow. The combination of these two factors creates an anisotropy, with preferential flow in the lateral direction. Hydraulically gaining conditions will occur in situations where regional water levels are elevated by low permeability boundaries (i.e. the presence of bedrock) on one or both sides of the river, or at middle to distal positions on alluvial fans where regional groundwater levels are closer to the land surface.

For a hydraulically connected river system (braids + BPA), the hydraulic gradient is no longer vertical, as it is for the hydraulically disconnected scenario, but variable, with the exchange rate governed by the relative hydraulic gradient between the braidplain and regional aquifers. Thus, groundwater inflow to the BPA or discharge to the regional aquifer can occur laterally along both margins of the braidplain, as well as vertically through BPA base. The setting can also be asymmetric, with inflow on one margin, and outflow on the other (as shown in Figure 1b), with the total river system water balance being gaining or losing, or having no flow along one margin due to the presence of bedrock.

While the hydraulically connected situation is simpler structurally, relative to the disconnected situation, exchange between the braidplain and regional aquifers is more complex. In this case (Figure 1b), water exchange is governed by the hydraulic gradient between the two aquifers, the transmissivity of both aquifers, and the vertical hydraulic conductivity of the underlying sediments. Once water exits the BPA in a losing reach, it will not return unless it is re-routed back to the river system by a reversal of the hydraulic gradient.

The sedimentological features and groundwater-surface water interaction concepts associated within the contemporary braidplain have been identified and detailed by previous authors (e.g., Huggenberger et al. 1998). Regardless of the nature of the relationship between the braidplain and regional aquifers, the braidplain gravels have a higher transmissivity than both the adjacent and underlying sediments because of repeated reworking of the braidplain gravels during high flow events. Hyporheic and parafluvial flow occurs within these highly transmissive BPA sediments, subparallel to the river flow direction, with individual braids acting as recharge or discharge boundaries. As such, the local hydraulic gradient and groundwater flux are largely influenced by river bathymetry.

4.0 Locations and methods for concept validation

4.1 Catchment descriptions

The conceptualisation presented here is based on field observations of the main losing reaches of three braided rivers in the drier eastern part of New Zealand (Figure 2). These are, from north to south, the Ngaruroro (a), Wairau (b), and the Waikirikiri (c, also known as the Selwyn). These study areas were selected to take advantage of the potential for hydrological separation afforded by dominantly losing river reaches. The downward hydraulic gradient, and potential hydrological separation between channel, bed gravels and regional aquifer enable the possibility for different structural and hydrological components to be identified. Summary hydrological and geomorphological statistics for the three study sites and their source catchments are shown in Table 1.

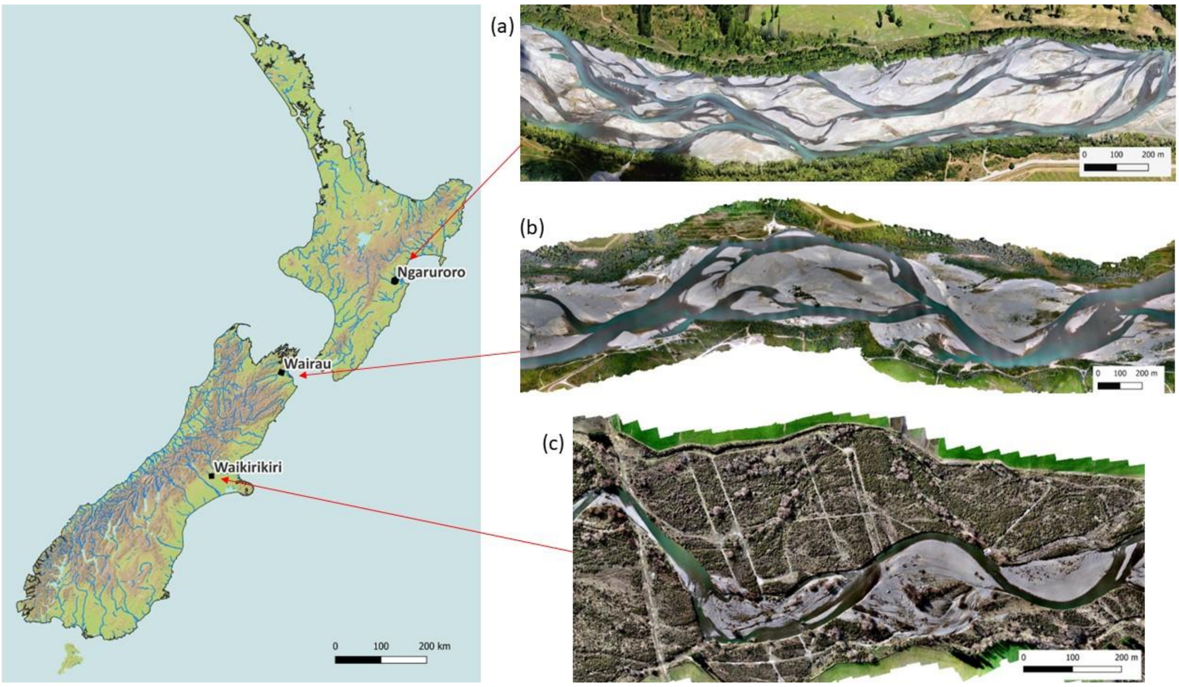

Figure 2. Location of the three study rivers and their aerial images

The Ngaruroro River is 164 km long, and has its headwaters in the Kaweka, Kaimanawa, and Ruahine ranges on the main divide of the North Island. The 3 km long study reach is located at the margin of the Heretaunga plain between Roy's Hill and Fernhill and is the main recharge source for the Heretaunga alluvial aquifer system (Dravid and Brown, 1997). A long-term flow monitoring site at Fernhill (70 years), situated at the lower end of our study reach, has recorded a mean flow of 40 $m^3.s^{-1}$ and mean annual peak flow of 1546 $m^3.s^{-1}$. The Ngaruroro study reach is in a natural depositional zone within the catchment (i.e., bedload into the reach exceeds that leaving the reach). This reach typically has 3 braids, with the active braidplain confined between willow plantings.

The Wairau River (Wöhling et al., 2018 and 2020) is 170 km long, and has its headwaters in the alpine Spencer Range on the main divide of the South Island. As such, the Wairau receives considerable source water from rain at higher elevations, producing a mean flow of 100 $m^3.s^{-1}$, and high mean annual peak flow of 1911 $m^3.s^{-1}$ (63 years of record at SH1). The Wairau River is also “flashy”, with over ten events exceeding three times the median flow annually (FRE3, Booker 2013). The 2.2 km long study reach is in the middle section of the Wairau Plain, between Jeffries and Giffords roads. This study reach typically has 2 braids and the active braidplain is confined between engineered rock revetments and groynes. Bedload flux into and out of this reach is approximately equal (bedload transfer zone).

The Waikirikiri River has its headwaters in the foothills of the Southern alps and receives considerably less runoff than the other rivers. The 1 km long study reach is situated near Hororata where the river leaves the foothills and crosses the western margin of the Canterbury Plain. A long-term (58 years) flow monitoring site located upstream of the study reach at Whitecliffs shows a mean annual flow of 4.5 $m^3.s^{-1}$, and peak annual flow of 28 $m^3.s^{-1}$. The Waikirikiri River is also the least “flashy” with an average of only one event each year exceeding a flow of three times the median flow. As a result of its lower flow and high transmission losses, the Waikirikiri River is subject to intermittent flow within our study reach (Larned et al., 2008; Di Ciacca et al., 2023). The Waikirikiri study reach is naturally incised between Pleistocene terraces, with no engineered flood controls. The active braidplain, which typically has 1-2 braids, can freely adjust between these contemporary braidplain margins, however, relatively dense exotic vegetation covers much of the contemporary braidplain.

Table 1. Characteristics of the three study catchments. Mean values are given unless stated otherwise. FRE3 is the annual number of events exceeding three times the median flow. Bed material grain size is near surface but excludes the armour.

	Ngaruroro	Wairau	Waikirikiri
Study reach length (m)	3000	2200	1000
Contemporary braidplain width in study reach (m)	300	400	460
Active braidplain width in study reach (m)	300	400	55
Braiding index in study reach	2.95	1.96	1.56
Reach slope ($m.m^{-1}$)	0.003	0.003	0.006
Catchment Area (km^2)	1930	3320	250
Catchment rainfall ($mm.y^{-1}$)	1414	1543	1045
Catchment potential evapotranspiration ($mm.y^{-1}$)	722	693	750
Annual low flow ($m^3.s^{-1}$)	1.2	9.9	0.97
Median flow ($m^3.s^{-1}$)	23.1	61	2.35
Mean flow ($m^3.s^{-1}$)	40	100	4.5
Annual peak flow ($m^3.s^{-1}$)	1546	1911	28.3
FRE3	6.1	10.7	1.1
River loss at study reach ($m^3.s^{-1}.km^{-1}$)	~0.6	≥0.2	0.25-0.65

All three rivers share a similar bedload lithology dominated by Jurassic greywacke. In the study reaches, the Ngaruroro is bounded to the north by Pliocene siltstone and limestone (Lee et al. 2011), while the Wairau is bounded to the north by schist (Begg and Johnston 2000). These two rivers lose water southwards, the river losing reaches being the primary source of recharge for the alluvial aquifers hosted by Holocene gravels. In the Waikirikiri study reach, the river is bounded by Pleistocene glacial outwash gravels (Forsyth et al. 2008). These gravels form the surface expression of a large, stratified aquifer system hosted by a composite of alluvial fans which underlies the Canterbury Plains. The Ngaruroro and Waikirikiri study reaches are both situated close to the apex of the river system’s alluvial fan (close to where the rivers emerge from the foothills).

Both the Wairau and Ngaruroro are affected by gravel extraction, which has lowered the mean active braidplain bed elevation in the river recharge reaches by approximately one metre since the 1980's (Gardner and Sharma 2016, Measures 2012). The varying physical environments, flow regimes, bed adjustment trajectories, and degree of lateral confinement result in different rates of bed reworking for each river. In particular, the Waikirikiri reworks its bed less frequently than the Ngaruroro and Wairau.

4.2 Field data collection

Investigations in the study reaches involved the collection of a variety of data types (Table 2). Each method and data type has advantages and disadvantages and is scale and process dependant (Gonzalez-Pinzon et al., 2015; Brunner et al., 2017). Stage and temperature recorders were established in the study reaches but were difficult to maintain due to repeated destruction by flood flows and frequent bed movement in the study reaches. Flow rating was conducted at the top of the reach for the Waikirikiri River, and permanent rated flow sites located downstream were used for the Ngaruroro and Wairau. A series of flow loss gauging surveys were undertaken at multiple transects within all three study reaches across a range of discharges to estimate transmission losses. Due to the difficulty of measuring flow in the larger rivers (and the related large measurement uncertainty), the most comprehensive set of flow loss surveys was conducted on the Waikirikiri (Di Ciacca et al., 2023).

Table 2. Type and number of measurements undertaken in the three study reaches.

Measurement type	**Ngaruroro**	**Wairau**	**Waikirikiri**

Differential flow gauging	2	2	14
Local river stage/temperature		3	6
LiDAR and bathy surveys	1	2	2
Piezometers	19	31	43
Cored holes	10	8	21
Particle size distribution	36	38	60
Core porosity	6	12	5
Field porosity	3	10	4
Radon-222 samples	5	53	61
tTEM	Y	Y	Y
DualEM	Y	Y	Y
SkyTEM	Y		
ERT surveys		9	11
Ground penetrating radar (GPR)			5
DTS installations (vertical)		2	3
DTS installations (horizontal)			2

LiDAR data were captured in dry areas of riverbed using a LiDARUSA Snoopy LiDAR scanner deployed on either a UAV or backpack. Bathymetry and water surface elevation were mapped using a kayak or remote controlled jetboat equipped with a paired RTK GPS and echosounder, and wading with an RTK GPS. Interpolation, or (where necessary) optical-bathymetry techniques, were used generate high-resolution bathymetry maps from less-dense echosounder survey data. The dry topography from LiDAR was stitched together with the bathymetry data to provide a complete digital elevation model (DEM) for each reach at a spatial resolution of 1 m or less, and a vertical accuracy of ±0.1 m in dry areas and ±0.2 m in wet areas.

Piezometers were installed at different depths to provide a time series of water levels and temperature, and to enable sampling for radon-222 analysis. The piezometers were installed with 50mm diameter PVC. Screens were a 1-2m length of slotted casing with a geotextile sock and sand pack around the screen, and cement grout around the overlying casing. A sump was used to collect downward percolating water in situations where pore water pressure was below saturation. Drilling methods involved a mix of rotary and sonic drilling, the former being used to install the piezometer network. Sonic drilling with a Geoprobe 8140LC (76.2mm diameter core) was conducted to collect cores for detailed logging and sediment analysis (grainsize and porosity). The sonic drilling method is not ideal as it can be difficult to get good core recovery, particularly in coarser or loose near-surface sediments. As a result, the core record is incomplete, however, there is currently no better method available to extract several meters of core from coarse riverbed deposits. Because core recovery was variable, data from individual drillholes were fragmented, and sediment analysis was only conducted on the most intact cores (0.1-1.2 m in length), with samples taken for grainsize and porosity analysis. Near-surface bed material grain size distribution and porosity were measured using manual excavation sieving combined with a detailed photogrammetry method to calculate excavation volume (Montgomery et al. subm.). Particle size distribution was characterised by fitting Gompertz and Weibull curves (Bayat et al. 2015) to the sieved sediment data, and summary statistics determined via the method of Folk and Ward (1957).

The base of the BPA was identified in two ways. The first method was by mapping surface morphology, since the deepest areas of pools represent the scouring depth that has occurred during flooding. Surface and bed morphology was captured using remote sensing (photogrammetry, LiDAR, and bathymetry). The second method was through observations in drill core with good core recovery, where sediments beneath the BPA are characterised by more cohesion and a finer matrix. To enable data to be compared, contacts between the braidplain gravels and underlying sediments (depositional unconformity) have been expressed as depth below a detrended surface representing cross-section mean contemporary braidplain surface elevation. This datum was chosen because of the spatially variable and temporally dynamic river bed levels.

Hydrogeophysical methods were also used to image the subsurface, including passive (DTS) and active (ADTS) distributed temperature sensing (Banks et al., 2022), ground penetrating radar (GPR), electrical resistivity tomography (ERT), transient electromagnetic (tTEM) and electromagnetic induction (DualEM421). The tracks prepared for tTEM and DualEM surveys are evident in the aerial photos shown in Figure 3. SkyTEM data were also available for the Ngaruroro area (Rawlinson et al. 2021). Of the hydrogeophysical methods employed, DTS/ADTS, and ERT were the most successful methods for delineating sediment structure and saturation associated with the river. The resistivity of New Zealand braided river water (fluid specific conductance ~5 $mS.m^{-1}$) and associated gravel deposits is very high (400-10,000 Ωm). For this reason, we think there was insufficient resistivity contrast for electromagnetic and ERT methods to reveal distinct subsurface features in most of our surveys. SkyTEM data did provide good definition of the basement contact beneath the Ngaruroro River but did not reveal any clear structural features in the near surface (<10 m). GPR surveys that were trialled at the Waikirikiri site clearly revealed the shallow water table but did not reveal any clear structure beneath the water table due to reflection of the signal.

Samples for radon-222 analysis were collected in 250 ml bottles in the Waikirikiri (Songola 2022) and Wairau reaches from riffles, at different depths in pools, and from purged piezometers. Samples were analysed 21-100 hours after sampling for Waikirikiri, and 20-24 hours for the Wairau. Laboratory analysis for radon activity was conducted with a RAD7 (Durridge 2020a), RAD H2O (Durridge 2020b), and active DRYSTIK (Durridge 2021) in a closed loop system, with the results adjusted for decay since the time of sampling (WAT250 method). The radon activity and uncertainty values reported here follow the approach of Durejka et al. (2019), with the mean and standard deviation calculated from five counting cycles, with duplicate samples pooled (ten cycles total). Additional radon measurements were made in the field using the RAD AQUA method (Durridge 2020c) to verify the WAT250 method results, and these returned similar values. For this study we have reported the WAT250 data, which has a larger uncertainty associated with the measurements but enables more samples to be collected in a short time frame from remote field sites. To reduce the uncertainty of the WAT250 results, we increased the aeration time to 10 minutes, and the analysis duration recommended in the Durridge manual to 5 cycles of 10 minutes.

5.0 Field observations

A hydraulically disconnected river-regional aquifer system has been identified from drilling and monitoring data in the Waikirikiri study reach (Banks et al. 2022, Di Ciacca et al. 2023), and at the upper part of the Ngaruroro study reach. A hydraulically connected river-regional aquifer system is observed in the Wairau study reach, and in the lower part of the Ngaruroro study reach. Evidence for the proposed conceptualisation is provided below based on data type. Data referred to in the text and figures is shown spatially in Figure 3. Points where the base of the BPA has been observed in cores, or inferred via deep pools are identified.

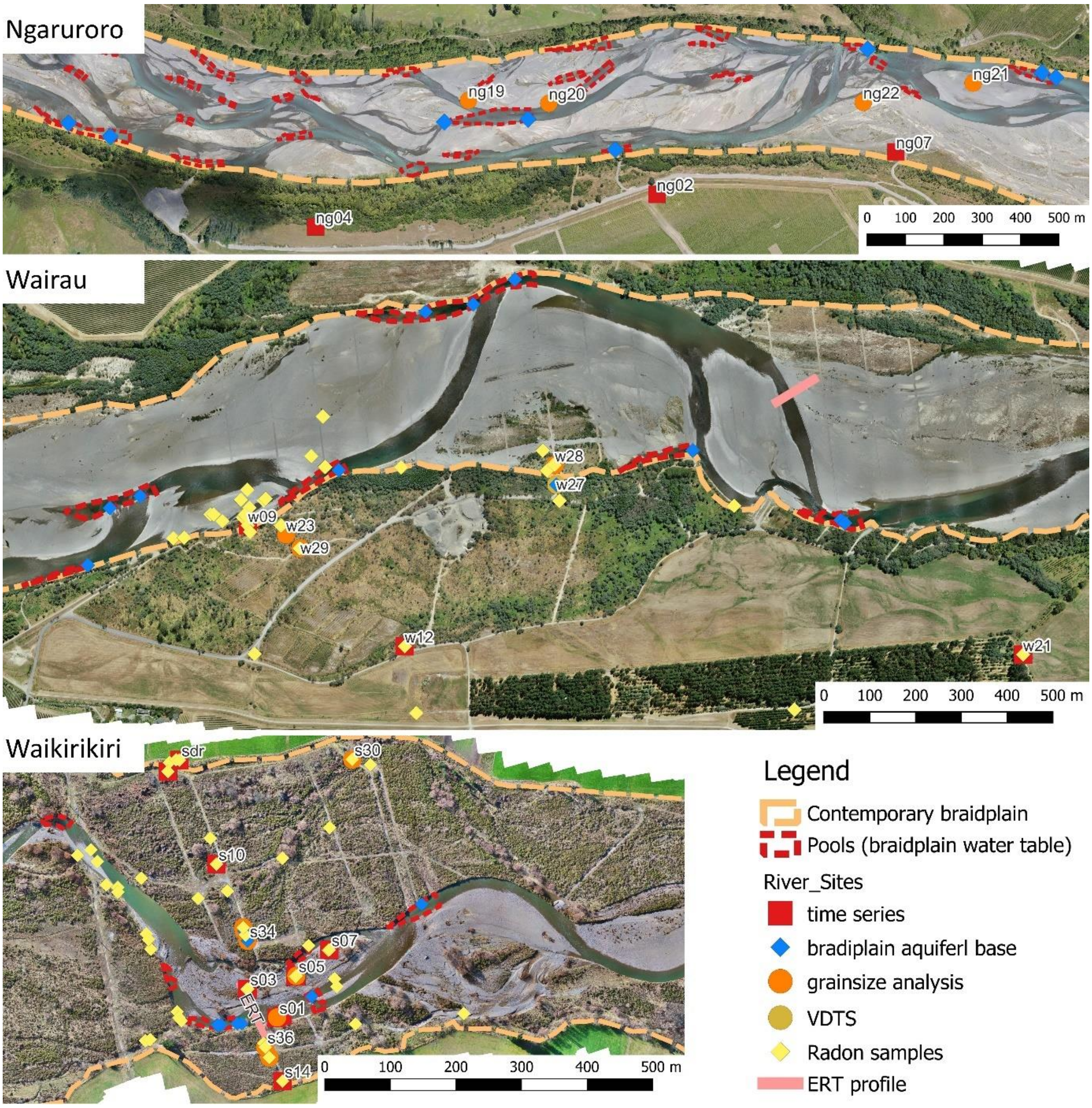

Figure 3. Data sources referred to in the text.

5.1 Geology and geomorphology

The BPA lateral extent in our three sites is identified geomorphologically as being at the contemporary braidplain margins (orange dashed lines in Figure 3). The contemporary braidplain margins in our study reaches are either artificially confined by rock armouring or willow planting for flood protection (Wairau and Ngaruroro), or by terrace boundaries (Waikirikiri). These relatively static lateral controls result in the abutment of relatively loose (recently active) braidplain gravels against older, more poorly sorted sediments. In the Wairau and Ngaruroro rivers the artificial lateral controls are sufficiently narrow that the entire contemporary braidplain is regularly reworked (i.e., the active braidplain margins and contemporary braidplain margins are essentially the same). In the Waikirikiri, the active braidplain is narrower than the contemporary braidplain and the active braidplain is able to adjust laterally, reworking the contemporary braidplain gravels.

In all three braided rivers, deep pools form at the toe of riffles. In the Wairau and Ngaruroro, these are amplified at the braidplain margins where river flow is reflected by rock armouring or willow plantings that limit lateral erosion and enhance bed scour. These pools are zones where the BPA discharges to the river. Parafluvial seeps are commonly observed in these locations, which have higher radon-222 activities and summer temperatures that are cooler than adjacent river braids. The surface expression of the BPA can be seen in abandoned channels where the braidplain surface topography drops below the braidplain water table and exposes groundwater (static pools or flowing springs). These areas are demarked by dashed red lines in Figure 3.

The depositional unconformity at the BPA base in recovered cores was observed at 1.7-2.0 m depth beneath the mean bed level of the contemporary braidplain in the Ngaruroro study reach, 4.3-5.0 m depth in the Wairau, and 2.1-3.7 m depth in the Waikirikiri. The greater depth in the Wairau River is unsurprising, as this river has the highest flood flows and is tightly confined by engineering works, resulting in greater depths of bed reworking.

At the Waikirikiri site, the contact of the depositional unconformity was seen as a change from grey-brown postglacial sandy gravels to yellow-brown clay-rich glacial outwash gravels (Figure 4). During drilling and completion of the groundwater monitoring wells, a drop in the water level with depth below the glacial contact was observed. Six cores were drilled just beneath the unconformity at various locations within the contemporary braidplain margins to investigate the saturation status beneath the braidplain gravels. These holes were drilled through the low permeability contact layer until an apparent increase in permeability was encountered, at which point the water level in the drillhole dropped significantly. Piezometers screened beneath the base of the BPA and above the regional water table showed low and unvarying water levels, indicating unsaturated or variably saturated conditions (Figure 7). Our explanation for the presence of this variably saturated zone is that the glacial outwash gravels are stratified, and vertical infiltration to the regional aquifer is limited by the lower permeability horizons of silt and clay. The presence of higher permeability horizons within the stratified postglacial sequence allows water to move laterally away from the recharge zone at a rate that exceeds vertical infiltration, enabling unsaturated or variably saturated conditions to form.

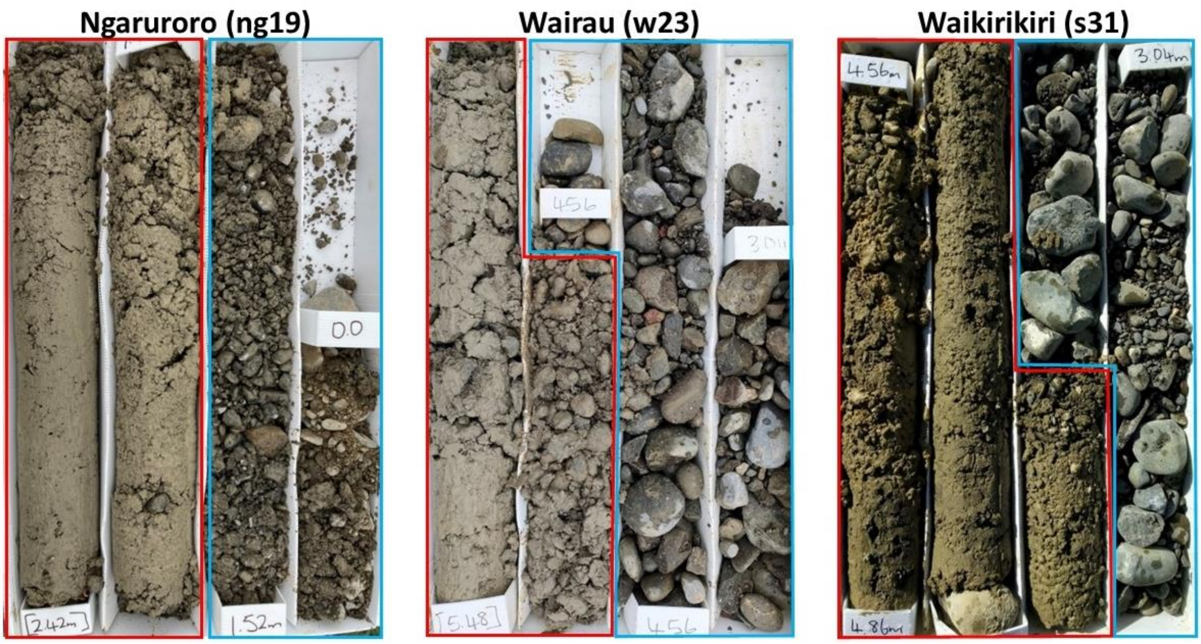

Figure 4. Representative core samples across the unconformity at the three study sites (blue=braidplain gravels, red=underlying very poorly sorted consolidated gravels)

At the Wairau River study site, the base of the BPA is visible as a change from grey-brown sandy clast supported gravel to more cohesive and poorly sorted gravel with increasing proportions of silt and clay. Just beneath this unconformity is a more prominent contact with yellow brown silty clay-bound gravel associated with old outwash fan deposits along the Richmond Range (not shown in Figure 4, but evident in Figure 10). Note that some of the Wairau core holes were positioned on the berms (outside of what we are referring to as the contemporary braidplain), where the active braidplain was located prior to river re-alignment in the 1960s. The river can no longer mobilise sediments on these berms due to rock armouring of the banks leaving a recent remnant braidplain gravel deposit beneath the southern berm. This deposit is both spatially and vertically separate from the contemporary braidplain (the mean bed level of the contemporary braidplain is approximately 2m below the berms).

In the Ngaruroro study site, the unconformity at base of the BPA is more gradational, manifesting as an increase in silt and clay content between 1.7 and 2.0 m depth visible within bore logs (Figure 4) and particle size distribution (Figure 5). As with the Waikirikiri River, a drop in the water table was observed while drilling beneath this lower permeability unconformity. The less-distinct base to the BPA in the Ngaruroro study reach is likely because the reach is depositional, so the underlying gravels were deposited relatively recently compared to the other study reaches.

Particle size distribution summary statistics of core samples collected by sonic drilling and bed excavation are shown in Figure 5, with classifications according to the Folk and Ward (1957) scale. A relationship between sorting and grainsize is apparent at each site, with poorer sorting corresponding to an increase in the finer fraction. Shallow braidplain gravels are overall poorly sorted (1-2, data mean: Ngaruroro 1.98, Wairau 1.90, Waikirikiri 1.96), whereas the underlying gravels tend to be very poorly sorted (2-4, data mean: Ngaruroro 2.88, Wairau 2.28, Waikirikiri 2.60). The grainsize of the braidplain gravels is generally coarser (mean: Ngaruroro -4.76, Wairau -4.23, Waikirikiri -3.55) than the underlying very poorly sorted sediments (mean: Ngaruroro -1.64, Wairau -3.54, Waikirikiri -2.69).

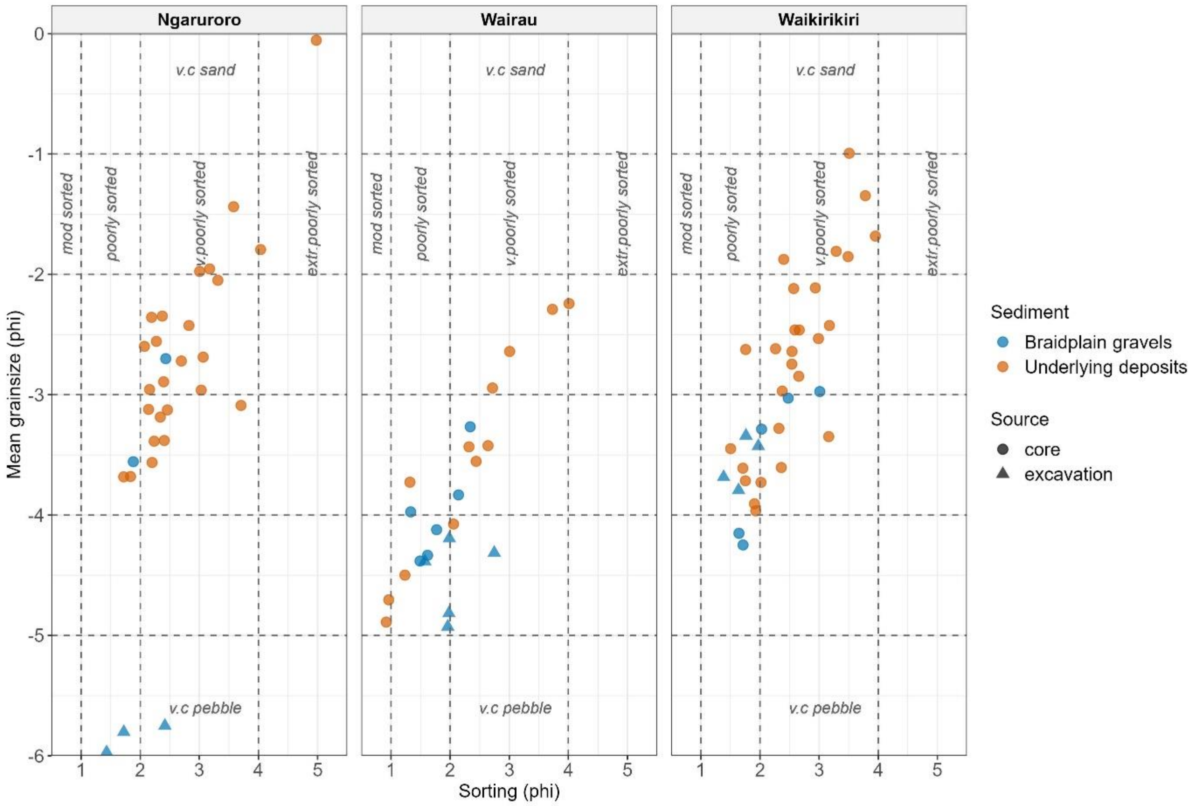

Figure 5. Grainsize analyses and sorting indices from samples beneath the three study reaches. Dashed lines are breaks for grainsize and sorting indices (after Folk and Ward 1957).

Core recovery at all sites was typically poor in the contemporary braidplain gravels, which mostly consist of loose gravel and cobbles, with a high proportion of sand but only a trace of silt and clay. Because of this, some of the braidplain gravel samples have been sourced from samples manually excavated from the braidplain subsurface (Figure 5). These excavated samples exclude the surface armour common in gravel bed rivers. For the Wairau data set there is no clear separation in the particle size distribution between braidplain and underlying gravels. Wairau sediments are coarser than the other two sites, because of the high-energy of this river system. The coarseness of the sediments greatly hampers core recovery, so it is possible that the lack of separation is an artefact of the loss of finer fractions in some samples during drilling.

5.2 LiDAR and bathymetry

LiDAR and bathymetry surveys were conducted in each study area to understand the spatially varying relationship between the river surface, bed levels, and water levels in the braidplain and regional aquifers. Repeat surveys were conducted following significant flood events to capture changes in bed levels. An example of our LiDAR and bed elevation data for the Wairau River is shown in Figure 6. These data were captured on 19 Feb 2020 at relatively low flow conditions, measured at 13.4 $m^3.s^{-1}$ to 11.5 $m^3.s^{-1}$ (±3 %) at the upstream (left) and downstream (right) margins of Figure 6 respectively. The river water surface and bed elevation data within the wetted channel are shown on Figure 6 in relation to a modelled surface of hydraulic head across the river system, represented by piezometric contours which are shown in Figure 6a. This surface was fitted (sum squared error of 5×10^{-6}) to 25 water level observations (yellow points in Fig 6a) located within and outside of the contemporary braidplain by universal kriging with an exponential variogram of anisotropy of 0.9 at 090°, partial sill 0.31 m, and range 670 m. With such a large variogram range, the surface should be considered as indicative of an averaged hydraulic head across the regional and braidplain aquifers. The kriged surface does reveal an inflection of the piezometric contours across the contemporary braidplain margins, indicating that flow within the BPA is largely controlled by river exchange and preferential flow within the BPA, with flow being approximately sub-parallel to the contemporary braidplain longitudinal orientation.

Fig. 6a reveals locations in the river system where the river water surface is higher than the braidplain water table (red and orange zones), indicating that the river is losing flow to the BPA in these areas. Areas of the river which are coloured blue in Fig. 6a represent the surface expression of the braidplain water table in pools. These are locations where the river can potentially gain flow. The black areas denoted as “riffles” are identified from a slope raster derived from the digital elevation model (DEM). Locations where maximum potential river water loss occurs can be identified in most cases as being situated at the upstream margins of high elevation riffles.

The bathymetry DEM (Fig. 6b) reveals the presence of scouring along the contemporary braidplain margins, which in the case of the Wairau River is promoted by excessive river narrowing and rock training banks. The corollary of this scouring is the relative mounding of gravel in the middle of the contemporary braidplain. The difference between the river bed level and hydraulic head reveals locations where the river bed is above the braidplain aquifer, and the river braid has the potential to be losing-disconnected at these locations. In most cases these areas also correspond to the upstream margins of high elevation riffles.

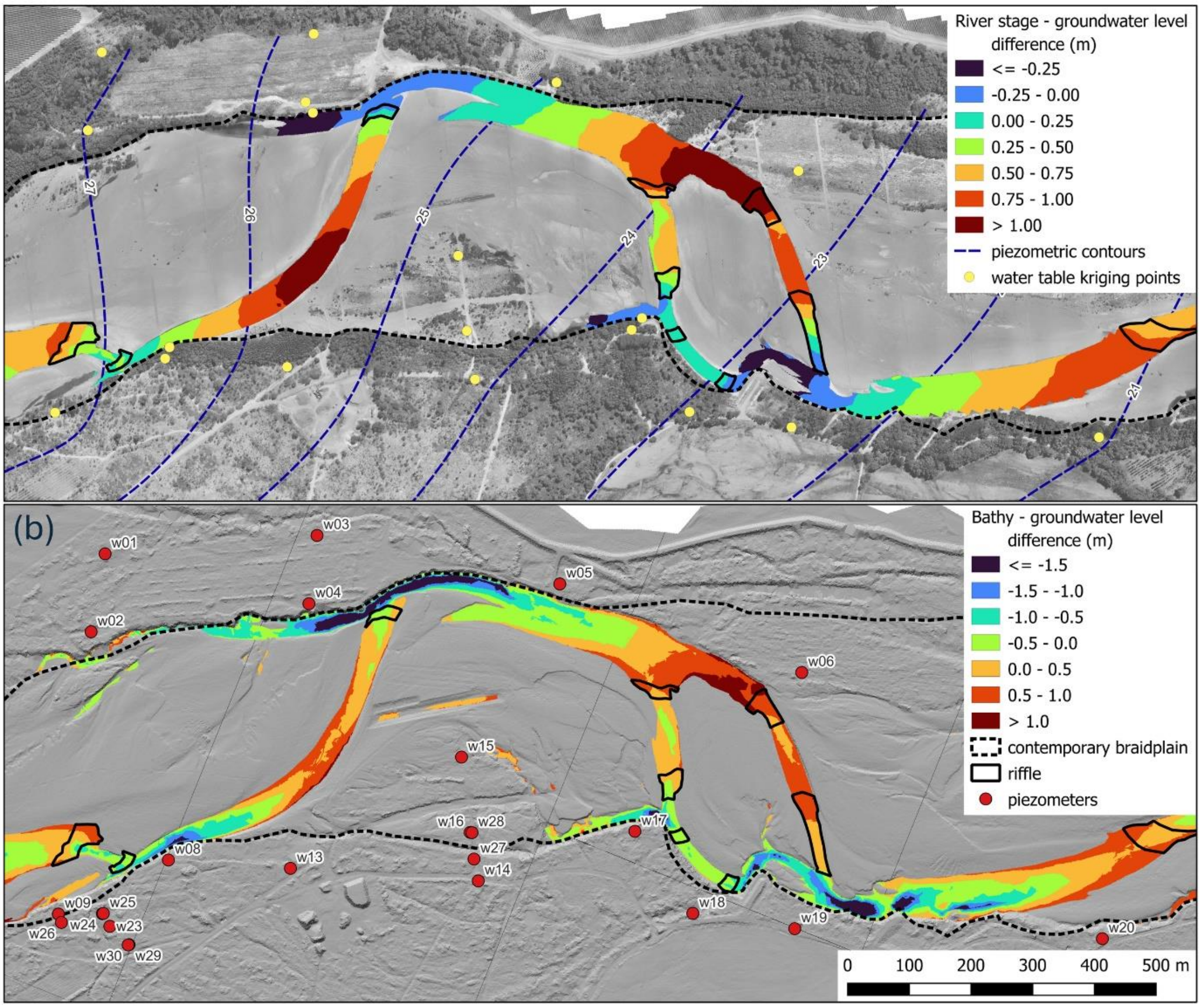

Figure 6. Images of the wetted Wairau River channel showing differences between (a) river water surface and a kriged hydraulic head (overlain on aerial imagery), and (b) bathymetry and kriged hydraulic head (overlain on the DEM). River flow is from left to right.

5.3 Water levels and temperature monitoring

Hydraulic heads within the braidplain aquifer are dynamic and fluctuate in response to changes in river stage. Figure 7 shows heads for the Waikirikiri field site, where the base of the BPA is at approximately 207m elevation, and the regional water table is >13m below the unconformity. The variably saturated zone (pore water pressure below saturation) is at least 10m thick. For a river system that is hydraulically connected to the regional aquifer, the pressure response outside of the BPA is also dynamic and shows a similar response to the

BPA (Figure 7). For a river system with a hydraulic disconnection, the variably saturated zone attenuates the pressure fluctuations in the regional aquifer (Figure 7). The relative water level depth and hydraulic response in the regional aquifer can therefore provide a useful test for hydraulic connectivity between the two aquifer systems.

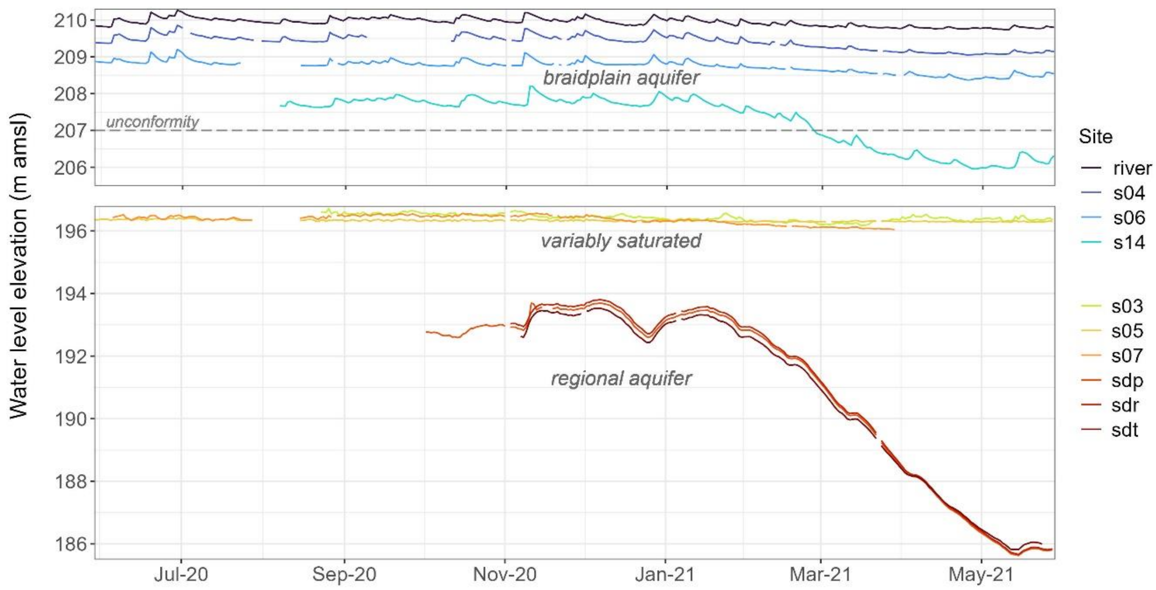

Figure 7. Water level time series data for the Waikirikiri River and braidplain and regional aquifers. Note the vertical offset in the two graphs is due to the thick variably saturated zone beneath the braidplain and regional aquifer. Site locations are shown in Figure 3.

Data from the Ngaruroro reach show characteristics of a regional aquifer that is both hydraulically connected to the BPA and either disconnected or transitional (Figure 8). These data have been normalised by the median water level to highlight relative changes in response. The BPA site (ng07) responds similarly to the river stage but has a slower recession rate due to storage in the gravels. The two sites that are screened in the regional aquifer (ng02, ng04) are both located 150 m adjacent to and downgradient of the contemporary braidplain (see Figure 3). The upstream site, ng04, has the most rapid recession rate and its peak response is slightly delayed compared to ng02 which responds similarly to the river channel stage. This suggests that this upstream section of the study reach may be hydraulically disconnected or transitional from the regional aquifer (assuming similar hydraulic properties throughout the regional aquifer). The water levels in ng04 are up to 2.4 m below the BPA base (mean 1.8 m) assuming a BPA thickness of 1.6 m below mean river bed level at this site. During flood flow events, the regional aquifer water level at ng04 does rise above the river bed elevation, indicating saturated conditions do temporarily occur during flooding.

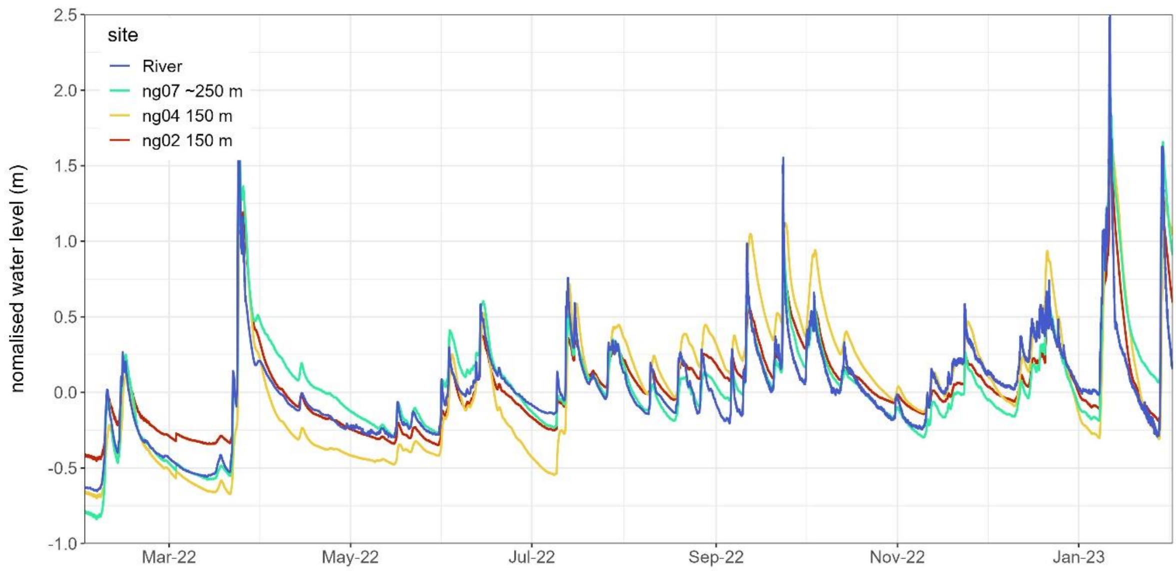

Figure 8. Normalised water level time series data for the Ngaruroro River, braidplain (ng07) and regional aquifers (ng02, ng04). Site locations are shown in Figure 3. Distances in the legend indicate the down gradient distance from the nearest active river channel.

Figure 9 shows daily average temperatures from representative hydrological settings in each study reach, and the lateral distance downgradient of the nearest active river channel. For sites located within the BPA, temperature responses are similar to the river channel, and driven by individual flow events (ng07, w09, s06). However, event-driven responses can be attenuated, and the season response delayed, at large distances from a channel (s25). Note also that s06 becomes more responsive after a flood event on 30 May 2021 which has changed connectivity between the river and groundwater in this piezometer.

For sites within the regional aquifer (ng02, ng04, w12, w21, sdp), the event-driven response is difficult to detect, and the seasonal response depends on the hydraulic relationship between the braidplain and regional aquifers. Sites ng02 and ng04 are equidistant from the Ngaruroro BPA, however ng04 is in a position where the base of the braidplain gravels and regional water table are approximately 1.8m apart, potentially with an intervening variably saturated zone, whereas at ng02 the braidplain and regional aquifers are hydraulically connected. This difference in hydrologic condition may explain the delayed seasonal response in ng04. In the Waikirikiri system, the variably saturated zone is considerably thicker (about 12m), which almost completely attenuates the seasonal temperature response in the regional aquifer.

Figure 9. Representative temperature time series for the three study sites. Piezometer locations are shown in Figure 3, sites ng07, w09, s06 and s25 are screened in the BPA.

In the Waikirikiri and Ngaruroro river systems, the temperature signals propagate efficiently from river channels to the BPA compared to the regional aquifer. This highlights the distinction between these two aquifers, with the BPA acting as an intermediary in groundwater – surface water exchanges.

5.4 Radon-222 sampling

A summary of the radon-222 results and measurement uncertainties for surface water and groundwater sources in the Wairau and Waikirikiri reaches is shown in Table 3. In the Wairau system, samples from riverbed piezometers and riverbed seepages have similar radon activities with ranges 1724-2849 $BQ.m^{-3}$ and 368-2585 $BQ.m^{-3}$ respectively.

Accordingly, samples from seepages and riverbed piezometers are both considered to represent the braidplain aquifer.

The radon data show distinct groupings, with radon-222 activities increasing from river channel to BPA to regional aquifer. At both sites, radon activities in river run samples were significantly lower than those in BPA samples. In the Wairau study reach, there is a notable overlap in radon activities between the braidplain and regional aquifers, indicating a likely hydraulic connection between these two systems. Conversely, in the Waikirikiri study reach, there is a downward increase in radon activities from the BPA to the variably saturated zone and further into the regional aquifer, with no overlapping values. This suggests a hydraulic disconnection between the BPA and the regional aquifer in the Waikirikiri reach.

Table 3. Measured radon-222 activities and one standard deviation uncertainties (BQ.m^{-3}) for the Wairau and Waikirikiri study reaches.

River	**Wairau**			**Waikirikiri**			
Sample source	River run	Braidplain aquifer	Regional aquifer	River run	Braidplain aquifer	Variably saturated glacial outwash sediments	Regional aquifer
Samples	14	11	21	10	38	7	6
Rn min	260 ± 85	368 ± 150	739 ± 272	384 ± 185	1791 ± 481	7017± 3607	16111 ± 3305
Rn max	604 ± 224	2849 ± 826	5700 ± 629	809 ± 443	9545 ± 1378	14654 ± 4338	22655 ± 2221
Rn mean	395 ± 180	1307 ± 361	3263 ± 666	568 ± 356	4569 ± 1002	11814 ± 2589	19184 ± 2763

The determination of residence times between the river and each aquifer depends on knowing the initial channel condition, representative secular equilibrium for the host gravel deposit, as well as a well-defined flow path length. Our estimate of the initial river channel condition is 260 BQ.m^{-3} for Wairau and 380 BQ.m^{-3} for Waikirikiri, reflecting the lowest measured river radon-222 activities. A secular equilibrium estimate of 5000 BQ.m^{-3} was derived for Wairau aquifer samples by plotting measured groundwater radon activity against distance of the piezometer from the river and fitting the ingrowth equation to the data to determine the 21 day equilibrium value. This exercise indicated that the Wairau BPA activities are too low for samples to reach equilibrium. In the absence of sediment specific data, the Wairau aquifer secular equilibrium was chosen to represent the Wairau BPA equilibrium. The secular equilibrium for the Waikirikiri BPA is estimated at approximately 8500 $BQ.m^{-3}$ based on the lowest activity observed in porewater samples from piezometer sumps in the variably saturated glacial outwash gravels beneath the braidplain aquifer (7017 $BQ.m^{-3}$), and the highest activity observed in the BPA (9545 $BQ.m^{-3}$). Based on the secular equilibrium values chosen, residence times for our study reach samples are estimated to be in the range of 0.1 to 4.4 days for the Wairau BPA in the study reach and 1 to >12 days for the Waikirikiri BPA. Due to the large uncertainties associated with the WAT250 method, these estimates should be considered for comparative purposes only.

5.5 Geophysics

ERT surveys of the contemporary braidplain in the Wairau and Waikirikiri reaches yielded varying degrees of success. The surveys that returned the most consistent subsurface resistivity response were conducted in the river channel itself, due to the presence of water in the near surface, which improved the connection between electrodes and the underlying resistive substrate. Figure 10 shows dipole-dipole array resistivity profiles across a braid of the Wairau (~ 30 m wide), and two braids of the Waikirikiri (~40 m wide). These profiles reveal the contact between the braidplain aquifer and underlying sediments (the unconformities shown in Fig. 4) at elevations consistent with drillhole coring. For comparison, the elevation of the same unconformity, derived by kriging intercepts of the BPA base within the contemporary braidplain (Figure 3.), is shown in red in Figure 10. The two profiles are shown at the same spatial and resistivity scale, and an interpretation has been made based on drilling information. Both profiles show a contrast between resistive (~1000 ohm-m) loose sandy gravels that host the BPA, and the underlying lower resistivity (<800 ohm-m) associated with older, very poorly sorted sediments. The Wairau profile reveals the unconformity at the base of the active braidplain gravels as a less resistive layer at ~19.5 m elevation, and a saturated BPA thickness of ~3.5 m. On the Waikirikiri profile, the unconformity at the base of the braidplain gravels lies at ~207.5 m elevation, and the saturated thickness of the BPA is only 2-2.5 m. The underlying very poorly sorted glacial outwash gravels typically show relatively low resistivities (≤ 600 ohm-m) due to the relative abundance of clay-sized sediment, even when not fully saturated.

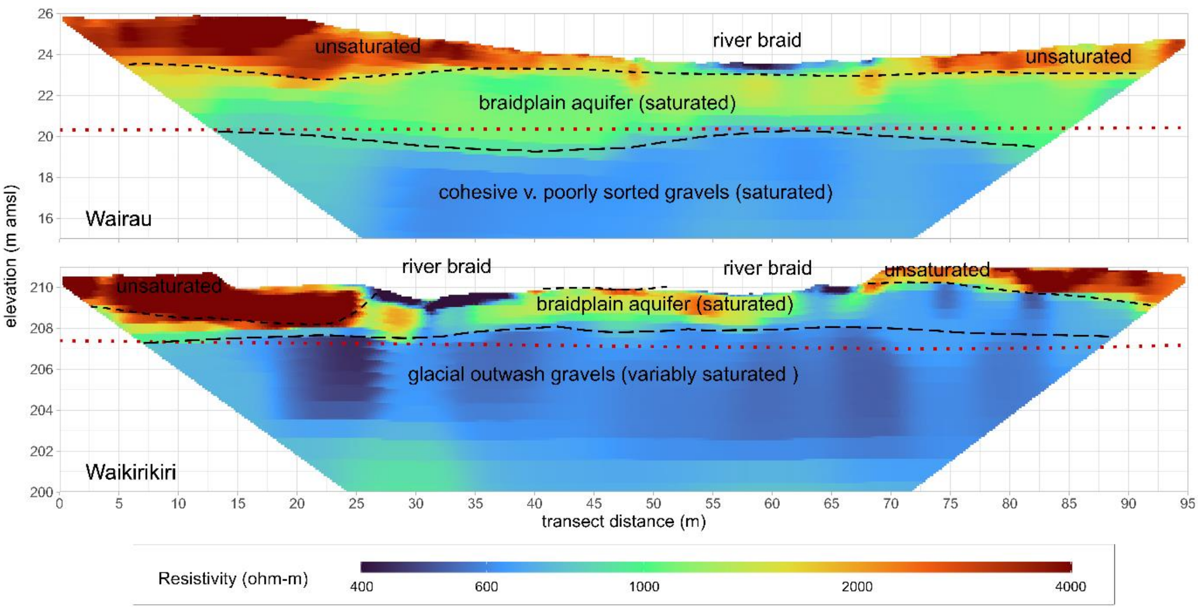

Figure 10. Subsurface resistivity collected by Electrical Resistivity Tomography (ERT) across Wairau and Waikirikiri river braids, looking downstream. Surfaces interpreted from the resistivity data are shown in black. An interpolated surface for the base of the BPA derived from drilling and bed data is shown in red.

5.6 Passive and active distributed temperature sensing

The hydrogeologic structure in the Wairau study reach has been assessed by monthly DTS and ADTS surveys (Figure 11) conducted on a vertically installed fibre optic cable located 20 m from the active braidplain margin (w27 in Figure 3). While w27 lies just outside of the existing engineered contemporary braidplain, this site does contain remnant braidplain sediments deposited prior to stabilisation of the river margins in the 1960s. The timing of the DTS surveys with respect to river temperature is shown in Figure 11c. The survey temperature profiles show consistent inflections with depth due to the attenuation of the river recharge temperature response through saturated sediments of contrasting hydraulic conductivity (Figure 11). These inflections correspond to the depths where a change in sediment characteristics can be seen in the bore log, manifest as progressively increased sediment cohesion, poorer sorting, and increasing silt and clay content with depth. In the DTS profile, the older silty clay-bound and indurated sediments have a large influence on subsurface flow, i.e., a reduction in flow through this material.

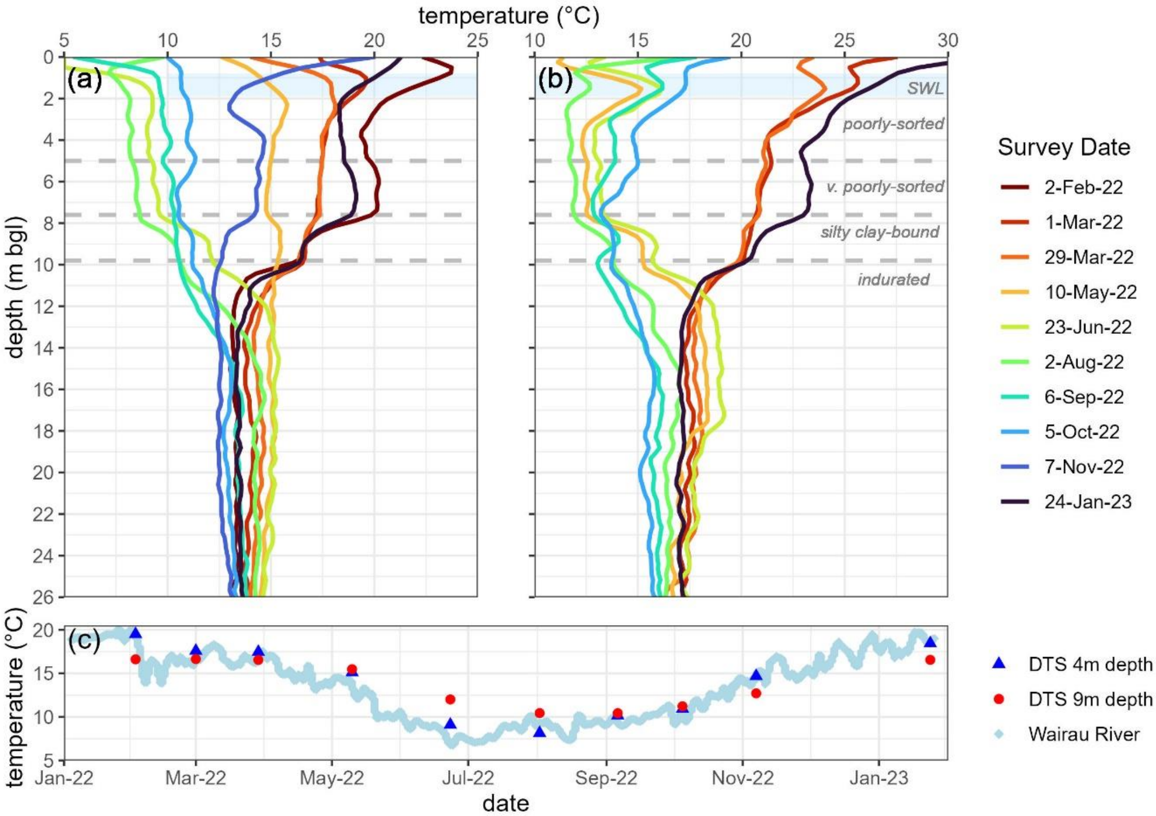

Figure 11. Vertical (a) distributed temperature sensing (DTS) and (b) active-distributed temperature sensing (A-DTS) surveys conducted on the south bank of the Wairau River on w27. The river temperature over the period of the surveys is shown in (c) along with temperatures measured by DTS at 4 and 9 m depth. SWL=static water level measurements over the survey period

6.0 Characteristics of the identified braidplain aquifers

A summary of the approximate dimensions and potential maximum groundwater storage volumes of the three braidplain aquifers investigated is shown in Table 4. In all three reaches, the BPA is laterally extensive, but very thin.

The Wairau River has the greatest BPA storage potential, largely due to its observed saturated thickness of up to 4.1 m. However, the Wairau is also a highly channelised river, and has a large bathymetric range, giving its BPA a large potential variability in saturated thickness in response to river channel stage fluctuations. Of significance for water management in the Wairau Plain is that the very poorly sorted gravels are thin beneath the Wairau contemporary braidplain and underlain by buried fans of silty clay-bound and indurated gravels which act as a vertical flow barrier (see Figure 10). Gravel extraction from the river system has dropped the mean bed level by approximately 1 m within our study reach since the early 1990s (Gardner and Sharma, 2016). This has allowed the river to rework sediment to greater depths, which has thinned and reduced the effective transmissivity of the gravel sequence overlying the buried fan deposits in addition to reducing the hydraulic gradient between the river system and regional aquifer.

Table 4. Dimensions and maximum storage volumes of braidplain aquifers observed in the three study areas.

River	Approximate width (m)	Contact depth (m) min-max (mean)	Porosity min-max (mean)	Storage volume ($m^3.m^{-1}$)
Ngaruroro	300	1.2 - 2.4 (1.6) n=4	0.12-0.35 (0.21) n=9	80-150
Wairau	450	2.4– 5.7 (4.1) n=6	0.09-0.46 (0.22) n=21	210-500
Waikirikiri	400	2.0 - 3.7 (2.8) n=16	0.09-0.22 (0.14) n=9	130-240

7.0 Discussion

The proposed conceptualisation enables a complex braided river system of high complexity to be represented by a few key hydrogeological elements. These are represented in Figure 1 as a contemporary braidplain, the margins of which mark the lateral extent of the braidplain aquifer (1), an underlying unconformity with more consolidated and more poorly sorted sediments (2), and observations of the hydraulic relationship between the braidplain and regional aquifers (a or b). For braided rivers, this approach integrates existing concepts of an alluvial aquifer, and hydrological hyporheic and parafluvial zones into a single conceptualisation of the contemporary braidplain subsurface. By identifying the base and margins of the BPA, and the process which forms it (reworking of bed material), the vertical and lateral extents to which hyporheic and parafluvial exchange occur can be identified by a change in sediment characteristics.

Point-scale features of braided rivers are described well by the existing hydrological framework proposed by Fox and Durnford (2003) and modified by Brunner et al. (2009a and b). Under this framework, individual channels can be considered as hydrologically connected (gaining or losing), disconnected (losing), or transitional (losing). However, this framework starts to break down for braided rivers at local to sub-catchment scales, as all four of these hydrological states can occur along a single cross-section of the braidplain regardless of whether the river has a net gain or loss. However, at the reach or sub-catchment scale, a braided river can be considered a *river system*, which can be described by any one of those hydrological states in relation to the regional groundwater system. For example, a river braidplain reach can be hydrologically disconnected and be losing water to groundwater overall, even though individual channels are hydrologically connected and locally gaining flow from BPA groundwater. The conceptualisation posed by Fox and Durnford (2003) can therefore be applied to different scales within a braided river system, but its application, and therefore interpretation of field measurements, requires knowledge of subsurface structure and saturation.

The difference in sediment characteristics above and below the unconformity at the base of the BPA indicates that the process of BPA formation is controlled by the mobilisation of bed material during flood flows, which loosen and sort the braidplain gravels, and winnow the finer fraction. This process of gravel mobility associated with flood events is supported by bathymetric observations of the depth of river channel scouring in deeper pools which agrees with the depths of the unconformity in core data. In the absence of drill core and particle size distribution data, we suggest that the elevation of pool depths measured soon after a flood event can be used to approximate the base of the BPA. This will only provide a minimum depth of the BPA base since the river is expected to deposit some sediment in scoured areas during the flow recession. The thickness of a river's BPA is likely to depend on the inter-relationship between several factors, including contemporary braidplain width, sediment characteristics and the balance of sediment supply, the frequency and magnitude of peak flow events, and the use of "hard engineering" to control a river's position. While some of these factors are natural, the factors related to width and depth are influenced by river engineering applied at each river. The Wairau River has the thickest BPA and has the highest hydrological energy of the reaches studied. This river is considered by some river engineers to be excessively narrowed, and a wider contemporary braidplain may result in a dispersion of energy during flood flows, and subsequent thinning of the BPA. Interestingly, the Ngaruroro River is also subject to high peak flows, although bed mobilising flows occur less frequently than the Wairau. It is likely that the Ngaruroro BPA is being thinned by the large volume of gravel extraction occurring within the study reach.

The LiDAR and bathymetric data gathered from our three study sites indicate that individual river channels locally merge with and diverge from the water table surface of the BPA.

Water exchange across the bed is determined by bathymetry and hydraulic properties, where the river can be forced above the water table by gravel lobes or dropped into the water table in locations of scouring. Areas of relatively still water (pools) are therefore the surface expression of the BPA, with seeps representing locations where the bedform drops below the level of the BPA water table. This explains the occurrence of higher radon-222 activities in pools compared to flowing river channels, as well as differences in seasonal water temperature.

For the interpretation of field data, the proposed conceptualisation highlights the importance of identifying and knowing the nature of the relationship between these three potential water sources to interpret observations. For example, to understand the nature and magnitude of river-aquifer interaction by only sampling radon-222 in river channels can be misleading. This is because the radon activity measured in the river depends both on the river setting (run or pool), and the nature of the interaction between the BPA and regional aquifer.

From a modelling perspective, it is questionable if a streambed conductance term is an appropriate physical mechanism for representing braided river-aquifer exchange at the local or catchment scale. The role of bed conductance, if significant, is to regulate exchange between individual river channels and the BPA. Due to the relatively high transmissivity of the bed materials, hyporheic exchange is an integral process of braided river flow, and water can be seen to freely exchange between the surface and the bed. It follows that consideration of water storage in the bed sediments (BPA) with a conductance term to impede flow beneath those sediments would be a more appropriate approach for simulating braided rivers at larger scales than to simulate individual channels with bed conductance.

The BPA concept resolves vague definitions of “groundwater” in groundwater – surface water interactions in braided rivers by considering the river as a whole system with an associated subsurface storage component distinct from the regional groundwater system. This provides specificity to the concept of a “river corridor” (Harvey and Gooseff, 2015), where a river comprises not just the river channels, but the surrounding, fluvial deposits, riparian zones, and floodplains between which hydrologic exchange occurs.

Braided river systems are spatially and temporally variable, which introduces heterogeneity both within a BPA, and the adjacent older sediments. This heterogeneity can manifest as preferential flowpaths, which can influence exchange fluxes at a local scale, as evident in spatial variability of temperature and radon data. While the BPA consists of high transmissivity sediments, and can itself be considered a preferential flow path at the regional scale, the presence of preferential flow within the BPA at the local scale is not captured by the conceptualisation presented here. We therefore recommend application of the BPA concept at the regional scale, and to provide a hydrogeological context for local scale studies. An additional consideration for applying the BPA concept is the volume of reworked material associated with the river. In braided river environments, the volume of gravel associated with the BPA is large, and significantly greater than the wetted channel volume at average flow conditions. However, in some gravel bed rivers, the volume of sediment mobilised by flooding events could potentially be very thin, and the relevance of these mobile sediments on the exchange between the river and regional groundwater system will depend on the scale of the study.

At the regional scale, the BPA concept is best applied to braided rivers that have stable or actively degrading beds, or have had some form of bank stabilisation, which is common where flooding is considered a risk to adjacent land. Stabilisation of river margins serves to increase the frequency of gravel remobilisation within the active braidplain and prevent reworking of adjacent bed material that may be part of the historical braidplain. Fine sediment can percolate through the gravels or be deposited on the surface, and if these gravels are not subsequently reworked, this can gradually consolidate and potentially clog the pore spaces, accentuating the difference between contemporary and adjacent historical braidplain sediments. Narrowing the area of braidplain that is reworked can thereby narrow the BPA. The conceptualisation may be less appropriate, or the BPA boundaries may be less distinct, where the contemporary braidplain margins are much wider than the active braidplain and reworked over longer timeframes. This type of braided river behaviour is typically seen in mountainous areas with low land use pressure, for example in the Southern Alps of New Zealand, but would have occurred on lowland plains prior to widespread engineering of river margins in New Zealand during the 1960s. If river channels can adjust laterally over a wider area, it is expected that sediments within the contemporary braidplain will be more heterogeneous, with channels of permeable recently mobilised gravel intermingled with islands/areas of varying age and permeability. In these cases, the extent of the BPA beyond the active braidplain across the contemporary braidplain may depend on the frequency of contemporary braidplain reworking, or the connectivity with the active braidplain. Similarly, it would be difficult to detect the base of the BPA in situations where the river bed is rapidly aggrading.

8.0 Conclusions

By investigating the surface and subsurface sediment and saturation in three study reaches, we have developed a conceptualisation of how braided rivers exchange water at the local (reach) and sub-catchment (aquifer) scale. The interaction between the river system and groundwater can be considered to occur between (a) individual river channels and the BPA (hyporheic and parafluvial exchange), or (b) the braidplain and regional aquifers. Central to this conceptualisation is the presence of a braidplain aquifer (BPA), a thin (2-5 m) layer of loose, poorly sorted gravel which is formed via the process of bed mobilisation during flood-flows. The base of the BPA can be identified in drill core as an unconformity between poorly sorted unconsolidated gravels overlying more consolidated very poorly sorted gravels. Individual river channels can be hydraulically connected, transitional, or disconnected from the BPA, depending on the relationship between the braidplain water table and bathymetry. The nature of the hydraulic relationship between the braidplain and regional aquifers can also be hydraulically connected, transitional, or disconnected, depending on the relationship between the regional water table and base of the BPA gravels. Approaching the braided river as a (whole) system (river and BPA combined) enables field data to be interpreted within the context of its water source.

From a modelling perspective, the conceptualisation enables rivers to be appropriately represented at a local scale (river-BPA), or a regional scale (BPA/river system-regional aquifer), depending on the modelling objective. A parallel investigation not presented here is focussed on the implementation of the proposed conceptualisation into sub-catchment scale models to quantify how changes in river morphology and BPA storage influence recharge to the regional groundwater system. A key research gap is to understand the relationship between BPA dimensions, bathymetry, and river flow dynamics, and this is a subject for future research.

Acknowledgements

The authors would like to thank the New Zealand Ministry of Business, Innovation and Employment for funding this research through the project "Subsurface processes in braided rivers - hyporheic exchange and leakage to groundwater". We are also grateful to additional support from Hawkes Bay Regional Council, Marlborough District Council and Environment Canterbury. We also acknowledge the contribution of the Hydrogeophysics Group, Aarhus University who carried out the EM surveys on our rivers, and Christy Songola who did the radon sampling and analysis at our Selwyn study site. Lastly, we would like to express our gratitude to the reviewers for their valuable comments, which helped us improve the quality of this article.

Financial support

This research has been supported by the Ministry of Business, Innovation and Employment (grant no. LVLX1901).

References

Anderson, E.I. (2005) Modeling groundwater–surface water interactions using the Dupuit approximation. *Advances in Water Resources* 28(4), 315-327.

Banks, E.W., Simmons, C.T., Love, A.J., and Shand, P. (2011). Assessing spatial and temporal connectivity between surface water and groundwater in a regional catchment: Implications for regional scale water quantity and quality. *Journal of Hydrology*, 404(1-2): p. 30-49, doi:10.1016/j.jhydrol.2011.04.017.

Banks, E.W., Morgan, L.K., Sai Louie, A.J., Dempsey, D., Wilson, S.R. (2022). Active distributed temperature sensing to assess surface water-groundwater interaction and river loss in braided river systems. *Journal of Hydrology 615*, 128667, doi:10.1016/j.jhydrol.2022.128667

Barthel, R., Banzhaf, S. (2016). Groundwater and Surface Water Interaction at the Regional-scale - A Review with Focus on Regional Integrated Models. *Water Resources Management 30*: 1–32, doi:10.1007/s11269-015-1163-z

Bayat, H., Rastgo, M., Mansouri Zadeh, M., Vereecken, H. (2015). Particle size distribution models, their characteristics and fitting capability. *Journal of Hydrology* 529, 872-889, doi:10.1016/j.jhydrol.2015.08.067

Begg, J.G, Johnston, M.R. (compilers) (2000). Geology of the Wellington area. Institute of Geological & Nuclear Sciences 1:250 000 geological map 10. 1 sheet + 64p. Lower Hutt, New Zealand. GNS Science.

Boano, F., Harvey, J.W., Marion, A. Packman, A.I., Revelli, R., Ridolfi, L., Wörman, A. (2014). Hyporheic flow and transport processes: Mechanisms, models, and biogeochemical implications. *Reviews of Geophysics 52*: 603–679, doi:10.5194/hess-22-1917-201810.1002/2012RG000417

Boano, F., Revelli, R., Ridolfi, L. (2008). Reduction of the hyporheic zone volume due to the stream-aquifer interaction. *Geophysical Research Letters 35*: L09401, doi:10.1029/2008GL033554.

Booker, D. J. (2013). Spatial and temporal patterns in the frequency of events exceeding three times the median flow (FRE3) across New Zealand. *Journal of Hydrology (NZ) 52*, 15-39.

Boulton, A.J., Findlay, S., Marmonier, P., Stanley, E.H., Valett, H.M. (1998). The functional significance of the hyporheic zone in streams and rivers. *Annual Review of Ecology & Systematics 29*: 59–81, doi: 10.1146/annurev.ecolsys.29.1.59

Bourke, S.A., Cook, P.G., Shanafield, M., Dogramaci, S., Clark, J.F. (2014). Characterisation of hyporheic exchange in a losing stream using radon-222. *Journal of Hydrology 519*: 94–105, doi:10.1016/j.jhydrol.2014.06.057.

Bristow, C.S., Best, J.L. (1993) Braided rivers: perspectives and problems. From: Best, J.L. and Bristow, C.S. (eds) *Braided Rivers*, Geological Society, London, Special Publications. Volume 75 Pages 1-11, doi:10.1144/GSL.SP.1993.075.01

Brower, A., Hoyle, J., Gray, D., Buelow, F., Calkin, A., Fuller, I., Gabrielsson, R., Grove, P., Brierley, G., Sai-Louie, A.J., Rogers, J., Shulmeister, J., Uetz, K., Worthington, S., Vosloo, R. (2024). New Zealand's braided rivers: The land the law forgot. *Earth Surface Processes and Landforms 49 (1)*: 10–14, doi: 10.1002/esp.5728

Brunner, P., Cook, P.G., Simmons, C.T., (2009a), Hydrogeologic controls on disconnection between surface water and groundwater, *Water Resource. Res*., 45(1), 1–13, W01422, doi:10.1029/2008WR006953

Brunner, P., Simmons, C.T., Cook, P.G. (2009b). Spatial and temporal aspects of the transition from connection to disconnection between rivers, lakes and groundwater. *Journal of Hydrology 376* (1–2): 159–169, doi:10.1016/ j.jhydrol.2009.07.023.

Brunner, P., Simmons, C.T. (2012). Hydrogeosphere: A Fully Integrated, Physically Based Hydrological Model. *Groundwater 50*: 170-176, doi:10.1111/j.1745-6584.2011.00882

Brunner, P., Therrien, R., Renard, P., Simmons, C.T., Hendricks Franssen, H.-J. (2017). Advances in understanding river-groundwater interactions. *Reviews of Geophysics 55*: 818-854 doi:10.1002/2017RG000556

Cardenas, M.B., Zlotnik, V.A. 2003. Three-dimensional model of modern channel bend deposits. *Water Resources Research 39*: 1141, doi:10.1029/2002WR001383

Cartwright, I., Hoffmann, H. 2016. Using radon to understand parafluvial flows and the changing locations of groundwater inflows in the Avon River, southeast Australia. *Hydrology & Earth Systems Sciences 20*: 3581–3600, doi:10.5194/hess-20-3581-2016

Coluccio, K., Morgan, L. K. (2019). A review of methods for measuring groundwater-surface water exchange in braided rivers. *Hydrology & Earth Systems Sciences 23*: 4397–4417, doi:10.5194/hess-23-4397-2019

Di Ciacca, A., Leterme, B., Laloy, E., Diederik, D., Vanderborght, J. (2019). Scale-dependent parameterization of groundwater–surface water interactions in a regional hydrogeological model. *Journal of Hydrology 576*: 494-507, doi:10.1016/j.jhydrol.2019.06.072.

Di Ciacca, A., Wilson, S., Kang, J., Wöhling, T. (2023). Deriving transmission losses in ephemeral rivers using satellite imagery and machine learning. *Hydrology & Earth Systems Sciences 27*: 703-722, doi:10.5194/hess-27-703-2023

Dravid P.N., Brown L.J., 1997. Heretaunga Plain Groundwater Study - Vol 1: Findings. Hawke's Bay Regional Council, 254p.

Durejka, S., Gilfedder, B., Frei, S. (2019). A method for long-term high resolution 222Radon measurements using a new hydrophobic capillary membrane system. *Journal of Environmental Radioactivity 208-209*: 105980, doi:10.1016/j.jenvrad.2019.05.012.

Durridge (2020a). RAD7 electronic radon detector user manual.

Durridge (2020b). RAD H2O Radon in Water accessory for the RAD7 User Manual.

Durridge (2020c). Continuous Radon in Water Accessory for the RAD7 user manual.

Durridge. (2021). DRYSTIK Models ADS-3 andADS-3R Active Moisture Exchanger Accessory for the RAD7 User Manual

Folk R.L., Ward W.C., 1957. Brazos River bar: a study in the significance of grain size parameters. *Journal of Sedimentary Petrology 27*: 3–26.

Forsyth, P.J., Barrell, D.J.A., Jongens, R. (compilers) (2008). Geology of the Christchurch area. Institute of Geological & Nuclear Sciences 1:250 000 geological map 16. 1 sheet + 67p. Lower Hutt, New Zealand. GNS Science.

Fox, G.A., Durnford, D.S., 2003. Unsaturated hyporheic zone flow in stream conjunctive systems. *Advances in Water Resources* 26 (9), 989–1000.

Gardner, M., Sharma, N., 2016. Wairau River Mean Bed Level and Volumetric Analysis: 2010 to 2016. Report to Marlborough District Council, 57p.

González-Pinzón, R., Ward, A.S., Hatch, C.E., Wlostowski, A.N., Singha, K., Gooseff, M.N., Haggerty, R., Harvey, J.W., Cirpka, O.A., Brock, J.T. (2015). A field comparison of multiple techniques to quantify groundwater-surface-water interactions. *Freshwater Science 34*: 139-160, doi:10.1086/679738

Gray DP, Hicks M, Greenwood M 2016. Advances in geomorphology and aquatic ecology of braided rivers. In Jellyman PG, Davie TJA, Pearson CP, Harding JS. (Eds) Advances in New Zealand freshwater science. New Zealand Freshwater Sciences Society, Christchurch, New Zealand. Pp 357-378.

Harbaugh, A.W. (2005). MODFLOW-2005, the U.S. Geological Survey modular ground-water model -- the Ground-Water Flow Process: U.S. Geological Survey Techniques & Methods 6-A16.

Harvey, J., Gooseff, M. (2015). River corridor science: Hydrologic exchange and ecological consequences from bedforms to basins. *Water Resources Research 51*: 6893–6922, doi:10.1002/2015WR017617.

Harvey, J.W., Wagner, B.J. (2000). Quantifying Hydrologic Interactions between Streams and Their Subsurface Hyporheic Zones. In: Jones, J.B. and Mulholland, P.J. (eds) Streams and ground waters. San Diego, CA: Academic Press, pp. 3–44, doi: 10.1016/B978-012389845-6/50002-8.

Holmes, R. M., Fisher, S. G., & Grimm, N. B. (1994). Parafluvial Nitrogen Dynamics in a Desert Stream Ecosystem. *Journal of the North American Benthological Society, 13 (4)*: 468–478, doi:10.2307/1467844

Huber, E., Huggenberger, P. (2016). Subsurface flow mixing in coarse, braided river deposits. *Hydrology & Earth Systems Sciences 20*: 2035–2046, doi:10.5194/hess-20-2035-2016

Huggenberger, P., Hoehn, E., Beschta, R., Woessner, W. (1998). Abiotic aspects of channels and floodplains in riparian ecology. *Freshwater Biology 40*: 407-425, doi:10.1046/j.1365-2427.1998.00371.x

Huggenberger P., Regli, C. (2006). A sedimentological model to characterise Braided River deposits for hydrogeological applications. In: Sambrook Smith GH (ed) Braided rivers: process, deposits, ecology and management. Blackwell, Malden, MA, pp 51–74.

Hupp, C.R., Osterkamp, W.R. 1996. Riparian vegetation and fluvial geomorphic processes. *Geomorphology 14*: 277-295.

Kalbus, E., Reinstorf, F., Schirmer, M., 2006. Measuring methods for groundwater-surface water interactions: a review. *Hydrology & Earth System Sciences 10*: 873-887 doi:10.5194/hess-10-873-2006

Khambhammettu, P., Renard, P., & Doherty, J. (2020). The traveling pilot point method. A novel approach to parameterize the inverse problem for categorical fields. *Advances in Water Resources 138*: 103556, doi:10.1016/j.advwatres.2020.103556.

Larned, S. T., Hicks, D. M., Schmidt, J., Davey, A. J. H., Dey, K., Scarsbrook, M., Arscott, D. B., Woods, R. A. (2008). The Selwyn River of New Zealand: A benchmark system for alluvial plain rivers. *River Research & Applications 24*, 1–21, doi:10.1002/rra.1054

Laube, G., Schmidt, C., Fleckenstein, J.H. (2018). The systematic effect of streambed conductivity heterogeneity on hyporheic flux and residence time. *Advances in Water Resources 122*: 60-69, doi:10.1016/j.advwatres.2018.10.003

Lee, J.M, Bland, K.J., Townsend, D.B., Kamp, P.J. J. (compilers) (2011). Geology of the Hawkes Bay area. Institute of Geological & Nuclear Sciences 1:250 000 geological map 8. 1 sheet + 93p. Lower Hutt, New Zealand. GNS Science.

Levy, J., Birck, M. D., Mutiti, S., Kilroy, K. C., Windeler, B., Idris, O., & Allen, L. N. (2011). The impact of storm events on a riverbed system and its hydraulic conductivity at a site of induced infiltration. *Journal of Environmental Management, 92*: 1960–1971, doi:10.1016/j.jenvman.2011.03.017

Measures, R. (2012). Modelling gravel transport, extraction and bed level change in the Ngaruroro River. NIWA Client Report CHC2012-121 for Hawkes Bay Regional Council.

Montgomery, J., Bind, J., Measures, R., Hoyle, J. (subm.). Novel methods for measuring porosity in gravel bed rivers. *Water Resources Research*.

Morel-Seytoux, H.J., Miller, C.D., Mehl, S. and Miracapillo, C. (2018) Achilles' heel of integrated hydrologic models: The stream-aquifer flow exchange, and proposed alternative. *Journal of Hydrology* 564, 900-908.

Niswonger, R.G., Prudic, D.E. (2005). Documentation of the Streamflow-Routing (SFR2) Package to include unsaturated flow beneath streams - A modification to SFR1: U.S. Geological Survey Techniques and Methods 6-A13, 50 p.

Pirot, G., Huber, E., Irving, J., Linde, N. (2019). Reduction of conceptual model uncertainty using ground-penetrating radar profiles: Field-demonstration for a braided-river aquifer. *Journal of Hydrology 571*: 254-264, doi:10.1016/j.jhydrol.2019.01.047

Pirot, G., Straubhaar, J., and Renard, P. (2014). Simulation of braided river elevation model time series with multiple-point statistics. *Geomorphology 214*: 148–156, doi:10.1016/j.geomorph.2014.01.022

Pirot, G., Straubhaar, J., Renard, P. (2015). A pseudo genetic model of coarse braided-river deposits. *Water Resources Research 51*: 9595-9611, doi:10.1002/2015WR017078.

Poole, G.C., Berman, C.H. (2001). An ecological perspective on in-stream temperature: natural heat dynamics and mechanisms of human-caused thermal degradation. *Environmental Management 27*: 787-802, doi:10.1007/s002670010188

Pryshlak, T.T., Sawyer, A.H., Stonedahl, S.H., Soltanian, M.R. (2015). Multiscale hyporheic exchange through strongly heterogeneous sediments. *Water Resources Research 51*: 9127–9140, doi:10.1002/2015WR017293.

Rawlinson, Z.J, Westerhoff, R.S., Foged, N., Kellett, R.L. (2021). Hawke's Bay 3D Aquifer Mapping Project: Heretaunga Plains SkyTEM data processing and resistivity models. GNS Science Consultancy Report 2021/93

Reinfelds, I., Nanson, G. (1993), Formation of braided river floodplains, Waimakariri River, New Zealand. *Sedimentology 40*: 1113-1127. https://doi.org/10.1111/j.1365-3091.1993.tb01382.x

Rushton, K. (2007) Representation in regional models of saturated river–aquifer interaction for gaining/losing rivers. Journal of Hydrology 334(1), 262-281.

Schälchli, U. (1992). The clogging of coarse gravel river beds by fine sediment. *Hydrobiologia 235/236(1)*:189–197

Schilling, O. S., Partington, D. J., Doherty, J., Kipfer, R., Hunkeler, D., & Brunner, P. (2022). Buried paleo-channel detection with a groundwater model, tracer-based observations, and spatially varying, preferred anisotropy pilot point calibration. *Geophysical Research Letters, 49 (14)*: e2022GL098944. https://doi.org/10.1029/2022GL098944

Shanafield, M., Cook, P.G. (2014). Transmission losses, infiltration and groundwater recharge through ephemeral and intermittent streambeds: A review of applied methods. *Journal of Hydrology 511*: 518–529, doi:10.1016/j.jhydrol.2014.01.068

Shanafield, M., Bourke, S.A., Zimmer, M.A., Costigan, K.H. (2021). An overview of the hydrology of non-perennial rivers and streams. *WIREs Water 8*: e1504, doi:10.1002/wat2.1504Fsongol

Songola, C. (2022). Characterising Surface Water and Groundwater Interactions in Braided Rivers Using Hydraulics and Environmental Tracers: The Waikirikiri Selwyn River. MSc thesis, Lincoln University.

Sophocleous, M. (2002). Interactions between groundwater and surface water: The state of the science. *Hydrogeology Journal 10 (1)*: 52-67, doi:10.1007/s10040-001-0170-8

Stanford, J.A. & Ward, J. V. (1993). An Ecosystem Perspective of Alluvial Rivers: Connectivity and the Hyporheic Corridor. *Journal of the North American Benthological Society 12 (1)*: 48-60.

Steiger, J., Tabacchi, E., Dufour, S., Corenblit, D., Peiry, J.-L. (2005), Hydrogeomorphic processes affecting riparian habitat within alluvial channel–floodplain river systems: a review for the temperate zone. *River Research and Applications 21*: 719-737, doi:10.1002/rra.879

Tang, Q., Schilling, O. S., Kurtz, W., Brunner, P., Vereecken, H., & Hendricks Franssen, H.-J. (2018). Simulating flood induced riverbed transience using unmanned aerial vehicles, physically-based hydrological modelling and the ensemble Kalman filter. *Water Resources Research 54*: 9342–9363, doi:10.1029/2018WR023067.

Theel, M., Huggenberger, P., Zosseder, K. (2020). Assessment of the heterogeneity of hydraulic properties in gravelly outwash plains: a regionally scaled sedimentological analysis in the Munich gravel plain, Germany. *Hydrogeology Journal 28*: 2657–2674, doi:10.1007/s10040-020-02205-y

Therrien, R., McLaren, R. G., Sudicky, E. A., Panday, S. M. (2010). *HydroGeoSphere: A Three-dimensional Numerical Model Describing Fully-integrated Subsurface and Surface Flow and Solute Transport*. Groundwater Simulations Group, University of Waterloo. 443 p.

Valett, H.M, Morrice, J.A., Dahm, C.N., Campana, M.E. (1996). Parent lithology, surface–groundwater exchange, and nitrate retention in headwater streams. *Limnology and Oceanography 41*: doi: 10.4319/lo.1996.41.2.0333.

Warburton, J. (1996). Active braidplain width, bed load transport and channel morphology in a model braided river. *Journal of Hydrology (New Zealand)*, 35(2), 259–285. http://www.jstor.org/stable/43944775

Ward, A.S. (2015). The evolution and state of interdisciplinary hyporheic research. *WIREs Water 3*: 83-103, doi:10.1002/wat2.1120

Ward, A.S., Packman, A.I. (2019). Advancing our predictive understanding of river corridor exchange. *WIREs Water 6*: doi:10.1002/wat2.1327

White, D.S. (1993). Perspectives on defining and delineating hyporheic zones. *Journal of the North American Benthological Society 12*: pp. 61– 69.

White, P.A., Kovacova, E., Zemansky, G., Jebbour, N., Moreau-Fournier, M. (2012). Groundwater-surface water interaction in the Waimakariri River, New Zealand, and groundwater outflow from the river bed. *Journal of Hydrology (NZ) 51*: 1-23.

Wöhling, T., Gosses, M.J., Wilson, S.R., Davidson, P. (2018). Quantifying river–groundwater interactions of New Zealand's gravel-bed rivers: The Wairau Plain. *Groundwater, 56*: 647–666, doi:10.1111/gwat.12625

Wöhling, T., Wilson, S.R., Wadsworth, V., Davidson, P. (2020). Detecting the cause of change using uncertain data: Natural and anthropogenic factors contributing to declining groundwater levels and flows of the Wairau Plain aquifer, New Zealand. *Journal of Hydrology: Regional Studies 31*:100715, doi:10.1016/j.ejrh.2020.100715

Wu, G.D., Shu, L.C., Lu, C.P., Chen, X.H., Zhang, X. Appiah-Adjei, E.K., Zhu, J.S. (2015). Variations of streambed vertical hydraulic conductivity before and after a flood season. *Hydrogeology Journal 23*: 1603-1615 doi:10.1007/s10040-015-1275-9

Wu, F.-C & Huang, H.-T. (2000). Hydraulic Resistance Induced by Deposition of Sediment in Porous Medium. *Journal of Hydraulic Engineering 126 (7)*: 547-551

Zhou, Y., Ritzi, R.W., Soltanian, M.R., Dominic, D.F. (2014). The influence of streambed heterogeneity on Hyporheic flow in gravelly Rivers. *Groundwater 52*: 206–216, doi:10.1111/gwat.12048