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REGIONAL COUNCIL

Heretaunga Springs

Gains and losses of stream flow to groundwater on the Heretaunga Plains

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Environmental Science - Hydrology

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Contents

Executive summary

Springs are the link between groundwater use and stream flows. Managing water use for the protection of stream ecosystems benefits from an understanding of spring inputs, because most water use on the Heretaunga Plains is from groundwater. This groundwater is used for irrigation, industry and drinking water.

Springs of the Heretaunga Plains were investigated to help inform flow-ecology investigations and groundwater modelling. This report is primarily a technical reference, focussing on the location of flow gains and losses, as well as the methods used to quantify them. Readers looking for a general overview are directed to Section 1 (Introduction) and Section 4 (Synthesis). The results (Section 3) provide details for specific sub-catchments.

The location of springs was investigated using a range of methods, starting with aerial photographs to narrow down the areas of interest. Spring locations and property access were then discussed with landowners before walking or kayaking the length of targeted streams. Measuring electrical conductance often revealed spring inputs that were sourced from larger catchments with less dissolved ions (e.g. Ngaruroro water). More quantitative surveys included flow gauging and longitudinal surveys of electrical conductance. Stable isotopes provided further insight into the source of spring water. Loss of flow to groundwater was estimated from the difference in gauged flow between sites.

The investigation identified 64 springs throughout the Heretaunga Plains, which represents the largest number of springs documented to date. This report does not provide a complete list of springs, instead capturing the major gains and major losses for sub-catchments of the Heretaunga plains. Most of the major flow losses to groundwater have been described previously, including losses from the Ngaruroro River that continue through dry summers when the aquifer receives little recharge from local rainfall. During these times, flow losses from the Ngaruroro River are vital for sustaining spring flow to the Raupare, Tutaekuri-Waimate and Waitio streams. Likewise, flow losses from the Tutaekuri River probably sustain springs in Moteo Valley (Tutaekuri-Waimate headwaters).

This investigation located large springs that contribute more than half of the low flow to the Karamu Stream. The Tukituki River was probably the main source of spring inflows direct to the Karamu Stream and to the Mangateretere Stream under low flow conditions on 3/3/2015. This changed in winter (23/8/2017), when isotope results indicated that nearly half of the groundwater originated from the Ngaruroro River.

These investigations also revealed a tufa coating (calcite deposited from flowing water) on the bed of the Paritua Stream. The Paritua Stream has run dry in summers past. This stream loses flow to groundwater where it crosses unconfined alluvial gravels upstream of Bridge Pa. The tufa coating is important in extending the length of flowing stream because it probably reduces the rate of flow loss to groundwater.

An extensive area of shallow Taupo pumice sands contributes groundwater to several streams, including the Karewarewa, Louisa and Awanui. Little is known of the groundwater in this pumice sand layer because it is not used for irrigation or domestic water supply. However, it may be an important source of flow and nutrients for these streams, and hence deserves further investigation.

This report goes beyond describing where streams are, to also describe their location in the past. For example, the Hawke's Bay earthquake of 1931 changed the drainage patterns of the Heretaunga Plains, including shifting the Paritua outflow from Irongate Stream to Karewarewa Stream. On alluvial plains such as the Heretaunga, floods and channel processes generate a naturally dynamic river network. In addition to these natural processes, people have made significant changes to the stream network to reduce flooding and drain soils for horticulture.

1 Introduction

1.1 Scope

Springs link groundwater use to surface water flows. Managing water use for the protection of stream ecosystems benefits from a good understanding of spring contributions to the stream, because most of the water use on the Heretaunga Plains is from groundwater (HBRC, 2014). Loss of river flow to groundwater is important for the same reasons.

This report is primarily a technical reference, focussing on the location of springs and flow losses, as well as the methods used to quantify them. These investigations were initiated for flow-ecology studies. In particular, the spatial oxygen model was based on a hydrogeomorphic template that described flow patterns across the Heretaunga Plains (Wilding, 2016). Additionally, the MODFLOW surface water-groundwater models were constructed in part using the spring information presented here (Rakowski, in prep 2018).

This report is not intended to provide a complete list of springs on the Heretaunga plains. It is intended to capture the major gains and losses for sub-catchments of the Heretaunga plains (sub-catchments as defined by the Section 3 sub-headings). Investigations for this report focused on finding the springs that feed *spring-dominated* streams, with less effort spent on locating springs feeding *runoff-dominated* streams. All streams are fed from groundwater, including those that run dry. All streams also receive some degree of fast rainfall-runoff. The distinction arises between streams where spring inflows dominate the flow for much of the year (*spring-dominated*), and streams where most of the flow arrives as rainfall-runoff, or quickflow (*runoff-dominated*). This simple dichotomy between spring and runoff dominated streams is adequate for this report. However, it is a simplification of the many possible classes of flow regime (Pyne *et al.*, 2016).

Within the spring-dominated streams, more effort was put into finding springs that may be fed from areas outside the stream's surface-water catchment (e.g. Raupare springs originating from Ngaruroro flow losses), as these are more difficult to quantify from conventional investigations (e.g. catchment water-balance models). Tributaries that contributed a larger proportion of mainstem flows were further prioritised. Losses from stream flow to groundwater aquifers were also a priority, including their location and magnitude. Springs emerging offshore, or within estuaries, were not investigated.

The type of information that is presented for each sub-catchment depends on what methods were known and available at the time. Some methods were developed over the course of this investigation, hence were not applied consistently across the sub-catchments.

Information from wells (e.g. lithology, water elevation, chemistry) is critical for understanding how groundwater and surface water interact. Information from springs (location, elevation, flow, chemistry) provides another piece of the puzzle. Springs offer a different perspective on surface water-groundwater interactions, compared to wells. The springs reflect groundwater processes closer to the surface, with a larger outflow that integrates groundwater processes over a larger area of the aquifer, compared to a single well. Well samples offer less spatial representation, but better specificity of results, representing a known location and depth within the aquifer. Patterns in spring location, elevation, flow and chemistry are therefore complementary to groundwater well information for developing conceptual models of surface-groundwater interactions. Information from this report is therefore intended to aid refinements of the conceptual models for surface-groundwater interactions for the Heretaunga Plains.

In describing the pattern of springs and streams, it is also important to understand their history. For example, the Hawke's Bay earthquake of 1931 changed the drainage patterns of the Heretaunga Plains, including shifting the Paritua outflow from Irongate Stream to Karewarewa Stream. On alluvial plains such as the Heretaunga, floods and channel processes generate a naturally dynamic river network (Hauer *et al.*, 2016;

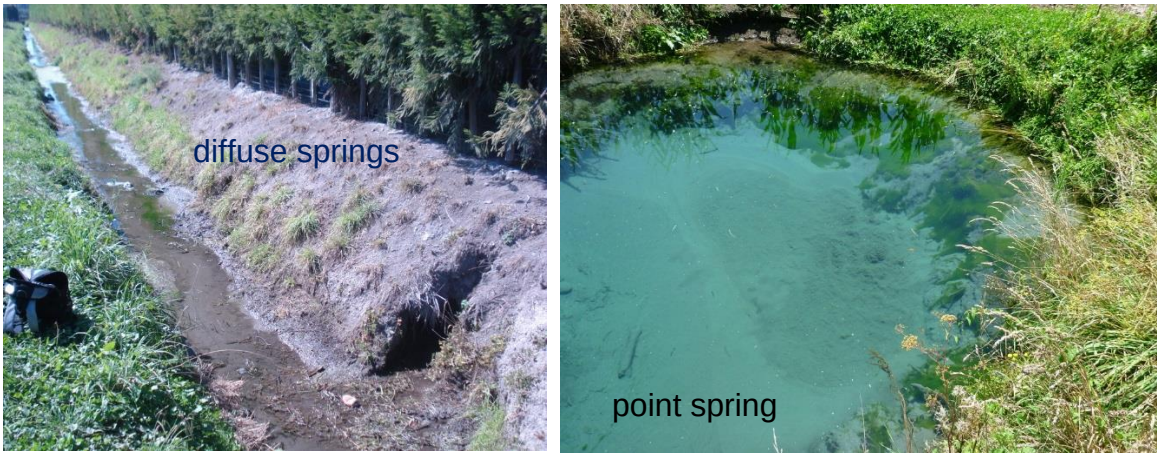
Wakefield *et al.*, 2012). In addition to these natural processes, people have made significant changes to the stream network in order to reduce flooding and drain soils for horticulture (HBRC, 2004). This report therefore goes beyond describing where streams are, to also describe their location in the past.

1.2 Flow gains and losses

Just as streams can gain flow from groundwater, via springs, the reverse can also occur with a net loss of stream flow to groundwater. As groundwater levels drop, spring flow diminishes (Figure 1-1). If groundwater levels continue to drop below the streambed elevation, then the same hole through which spring water emerged can become a hole through which water is lost from the stream to groundwater (when there is surface water to be lost). For many springs, the outflow does not stop, reflecting groundwater levels that remain above stream level (e.g. Raupare spring). Conversely, there are some streams where we have only ever measured a loss (e.g. Ngaruroro upstream of Fernhill).

Considering the full length of a stream, dry sections can transition to gaining sections where the streambed intersects the groundwater level (Figure 1-1). If layers of gravel connect the stream to groundwater, then spring inflows will be more diffuse/spread out. In contrast, the more discrete, point-springs with boiling sands occur where clay confining-layers separate the stream from the groundwater, and the connection to the stream is via discrete holes in the confining layer. Those holes can take the form of gravel and sand interspersed through the confining layer (Harper & Hughes, 2009). The length of diffuse springs can increase during winter when groundwater levels are higher, compared to the fixed location of point-springs.

Between the two extremes of a point spring emerging through a discrete hole, and a diffuse spring emerging through unconfined gravels, exists a gradation of intermediate springs. For example, sand and gravel or tree roots interwoven through the thin edge of the confining clays can create a series of small point springs that together form a relatively diffuse input to stream flow.



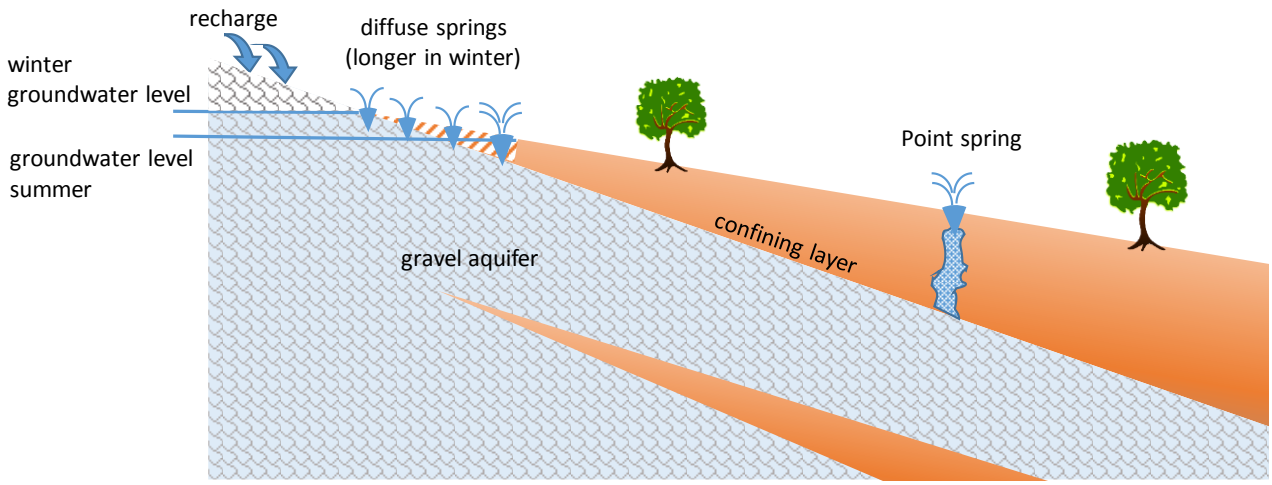


Figure 1-1: Diffuse and point springs. The length of diffuse springs can increase during winter when groundwater levels are higher, compared to the fixed location of point-springs, as demonstrated in this stylised diagram. Example photographs from the Raupare catchment show diffuse springs (left) and a point spring (right).

1.3 Hydrogeology

The Heretaunga Plains have been built up by the outflow of gravel from the three major rivers – Ngaruroro, Tukituki and Tutaekuri (Dravid & Brown, 1997; Harper & Hughes, 2009; Lee *et al.*, 2014). The bedload of gravel that was eroded from the ranges (e.g. Ruahine, Kaweka) exceeded the carrying capacity of these rivers as they crossed the plains. Deposited gravel has pushed the river into new flowpaths and, in doing so, gradually spread the sediment across the plains. Old flowpaths eventually became buried (Figure 1-2). Some of these palaeochannels have larger spaces between the gravel and cobbles, providing a preferential flowpath for faster movement of groundwater.

Sea level rise between glaciations pushed the tidal estuaries further inland, creating depositional environments where the finest river silts were deposited (Lee *et al.*, 2014). Over time, enough fine-silt accumulated to form a thick confining layer of blue clay (Figure 1-2). The series of confining layers do not reach all the way across the plains, instead thinning out at the upper limit of sea level rise (about Twyford to Bridge Pa), (Dravid & Brown, 1997; Harper & Hughes, 2009). The most recently deposited confining-layer has capped groundwater within the Heretaunga aquifer (albeit an imperfect cap). Springs more often arise near

the edge of this confining layer, as well as from holes and fractures close to the edge. Those holes can take the form of gravel and sand interspersed through the confining layer, creating a semi-confined aquifer (Harper & Hughes, 2009). Inland of the confining layers, the uncapped gravels can receive recharge water from the Ngaruroro, Tukituki and Tutaekuri rivers (detailed in this report), in addition to direct rainfall recharge.

The formation of hard pans can inhibit the movement of rainfall recharge down through the soil profile. Three types of pan could be present on the Heretaunga Plains:

- pan cemented by lime (calcium carbonate) that is weathered from adjacent limestone hill-country;
- wind-blown silt (loess);
- duripan with silica from Taupo airfall ash (Griffiths, 2001).

In addition to the airborne ash, the Ngaruroro River delivered a large load of alluvial pumice from more recent Taupo eruptions (e.g. 177 AD). This formed a shallow deposit of pumice sand over extensive areas of the Heretaunga Plains (Grant, 1996; Griffiths, 2001), providing a conduit for shallow groundwater. Another conduit for shallow groundwater is provided by river gravels deposited on top of the confining layer by more recent flowpaths. For example, Rabbitte (2011) proposed that flow losses from the Ngaruroro River could be conveyed to adjacent springs via shallow sand or gravel above the confining layer. Where hard pans are lacking from the soil, these conduits for shallow groundwater are open to rainfall recharge and nutrient leachate.

Earthquakes are also an important physical driver, buckling the hill country around the plains, and fracturing the rock in a way that controls groundwater flow (Cameron *et al.*, 2011). Earthquakes have also tilted the Heretaunga Plains themselves (Lee *et al.*, 2014). The 1931 earthquake lifted large areas sufficient to change the flowpath of some streams, and new springs have emerged after smaller earthquakes (e.g. via fracturing of the confining layer).The complexity of these interacting physical drivers creates a diverse array of habitats both on land and in streams (Hauer *et al.*, 2016; Wakefield *et al.*, 2012).

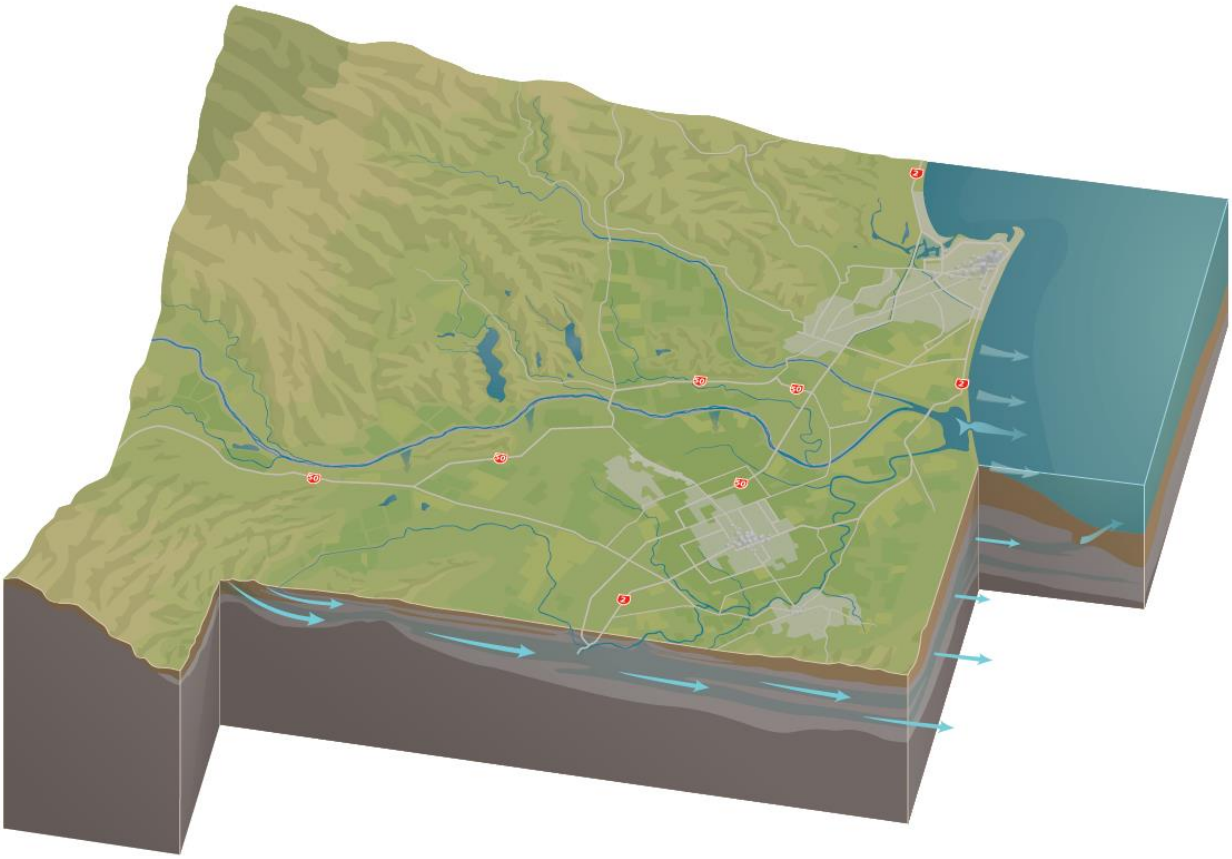


Figure 1-2: Groundwater beneath the Heretaunga Plains. The movement of groundwater through gravel layers (blue layers) is contained within confining layers of clay (brown layers). This figure provides a stylised representation of these elements, rather than a geological map. Artwork by Chris Davidson, 4art.co.nz.

1.4 Climate

The Heretaunga Plains have a maritime-temperate climate, with approximately 800 mm of rainfall and 2150 sunshine hours each year (Chappell, 2013). The seasonality of rainfall is small compared to the high evaporation rates through summer, which increase water demand for irrigation. The ranges west of the Heretaunga Plains (e.g. Ruahine, Kaweka) capture high rainfall from westerly weather systems. The colder and wetter climates in the ranges generate higher flows in the mainstem rivers (Ngaruroro, Tukituki, Tutaekuri), compared to lowland areas, and this contrast is more pronounced through summer and autumn.

The low rainfall increases the dependence of horticulture on irrigation. Most irrigation water for the Heretaunga Plains is sourced from groundwater (HBRC, 2014). Intensive development of the Heretaunga gravel aquifers provides irrigation water for orchards, vineyards and crops, alongside town and industrial water supplies.

2 Methods

2.1 Locating Springs and Measuring Flow Gains

Aerial photographs were used as a first step to narrow down where each stream started flowing. Talking to landowners was often the next step. People who have lived on the land often have the best local knowledge of springs in the area and their history. Having narrowed down the areas of interest, walking the length of headwater streams provided valuable information on the location of springs, and the start point for the stream network (Wilding, 2016). The location of each spring (NZTM easting and northing) was recorded from a handheld GPS (e.g. Garmin GPSMap64, Garmin eTrex).

The location of a spring will change little over time if it is a point spring that arises through a hole in a confining layer. However, the start location of a diffuse headwater spring will change, with spring flow arising further upstream during the wetter months when groundwater levels are higher. The timing of spring surveys is therefore important. This study targeted dry conditions (e.g. the 2013 drought) for spring surveys because the focus was on perennial streams and their flow requirements (Wilding, 2016).

Only high-priority streams were surveyed on foot. This included streams where flow was spring-dominated; where groundwater was probably fed from outside the surface water catchment, and springs that contributed a large proportion of flow to the Heretaunga streams (Paritua-Karewarewa, Irongate, Mangateretere, Karamu, Tutaekuri-Waimate, Raupare). For other streams, spring mapping relied more heavily on aerial photographs. A time-series of aerial photographs was available from Google Earth. These provided visible indicators, including ponded water in adjacent paddocks during dry periods, more wetland vegetation (e.g. riparian willows, Batelaan *et al.*, 2003), and intensified drainage of adjacent paddocks. A change in soil classification (Griffiths, 2001) was sometimes observed. Often these indicators can only be seen from the air, rather than when standing in the stream. Therefore, aerial photographs are complementary to walking the reach. A time-series of images from Google Street View could distinguish dry from wet channels at road crossings (winter and summer).

Not all spring inputs can be seen with the naked eye. Diffuse spring inputs to flowing streams are not visible, and even the large, discrete springs can be masked within large rivers. Therefore, measurements are needed to detect the start or end of spring inflows in the flowing part of the stream, and these methods are described in Sections 2.1.1 to 2.1.5.

2.1.1 Direct Flow Measurement (Gauging methods)

The flow measurements that were used in the estimation of flow gains and losses span more than 60 years. Mechanical flow meters were mostly replaced with acoustic Doppler velocity sensors in the late 2000s. Most flow measurements undertaken specifically for this study used the Sontek Flowtracker mounted on a wading rod. Velocity and depths were typically sampled from at least 20 offsets/locations across the channel. Velocity was sampled for a period of 40 seconds at each offset, and typically sampled at 60% of depth to approximate mean water column velocity. Site selection and gauging methods otherwise followed guidance from the National Environmental Monitoring Standards in pursuit of a “fair” quality code (i.e. QC 500), (Willsman *et al.*, 2013). The data were checked then stored on Hawke’s Bay Regional Councils hydrometric database (Hilltop Manager).

Other methods were used as and when required. For example, when the stream was not wadable, a boat mounted acoustic Doppler current profiler (ADCP) was used to measure flow (Sontek M9). Between 4 and 8 passes were typically made, and the flow estimates from replicate passes were averaged to provide a flow estimate in pursuit of a “fair” quality code (Willsman *et al.*, 2013).

Conventionally, spring inflows are detected by gauging the flow at points along the stream, then determining if there is a change in flow that cannot be explained by tributary inflows. However, relying on stream gauging imposes limitations on the spatial resolution of the survey, as the number of points along a stream where flow can be accurately measured is limited, and the smaller the flow change that we seek to detect the more accurate those measurements need to be. For instance, in braided rivers, several kilometres can separate suitable gauging locations (i.e. uniform straight runs). For weedy streams, suitable gauging locations may only exist at road crossings (e.g. culvert outlets), where the distortion of the velocity profile by weed beds is minimised. This distance between sites will limit our ability to resolve where springs start and end. Other methods were therefore used to enhance and extend the detection of spring inflows, including the change in electrical conductance and the change in water temperature from spring inflows.

Flow measurements for this study were generally conducted during periods of steady flow. Periods of rainfall produce rapid changes in flow that reduce comparability with other sites gauged on the same day. A lot of the flow data used in this report was not collected specifically for this study. Unless noted otherwise, these flow data were filtered to remove periods of rapid flow change. Flow was considered rapidly changing if the difference between daily maximum and daily minimum flow was more than 30% of minimum flow for the previous 3 days.

2.1.2 Longitudinal Electrical Conductance

The inflow of spring water can be estimated using the downstream change in electrical conductance, if the spring water has a different conductance to the river flow. This situation arises in the Karamu Stream and Tutaekuri-Waimate Stream. These streams are also large enough to be kayaked, with electrical conductivity sensors (standardised to specific conductance at 25°C) mounted on the kayak, together with synchronised GPS tracking to provide a location for each measurement. If combined with flow gaugings at the start and end of the study reach, the downstream change in conductance can then be converted to a flow-profile using Equation 2-1.

Equation 2-1: Estimating flow gain from electrical conductance. Flow at stream location *i* is estimated using the change in EC (Electrical Conductance, µS/cm at 25 °C) between location *i* and the previous location (*i*-1), calculated as a mass flow (EC x Flow L/s). This also requires an estimate of electrical conductance for the spring inflow (EC_{spring}) from Equation 2-2.

$$Flow_i = Flow_{i-1} + \left(Flow_{i-1} \times EC_{i-1} - Flow_{i-1} \times EC_i\right) \div \left(EC_i - EC_{spring}\right)$$

The average conductance of the spring inflow was estimated for each survey, using same-day gaugings upstream and downstream of the gaining section (Equation 2-2). This assumed the conductance of spring water did not change significantly over the gaining reach. For example, if spring conductance was actually lower in downstream reaches, then the relative contribution of downstream springs would be over-estimated. The measured length of gaining sections likely extended beyond the actual spring inputs because of the added distance needed to achieve complete mixing (both vertically and horizontally).

Equation 2-2: Estimating spring electrical conductance. The average EC (Electrical Conductance, µS/cm at 25 °C) of inflowing spring water was estimated using this mass balance equation, using flow (L/s) and electrical conductance measurements downstream ($Flow_{down}$ and EC_{down}, respectively) and upstream ($Flow_{up}$ and EC_{up}, respectively) of the gaining reach.

$$EC_{spring} = \left(Flow_{down} \times EC_{down} - Flow_{up} \times EC_{up}\right) \div \left(Flow_{down} - Flow_{up}\right)$$

The distance between each measurement was estimated from the metric GPS data (NZTM) (Equation 2-3). This was then converted to a distance downstream of the starting point by accumulating the distances between measuring points. The close spacing of measurements (typically <50 m) ensures an accurate representation of meandering stream length, despite using the straight line distance between each pair of points. The location and distance from the GPS was then matched to the conductance data using the synchronised time stamp, with time-matching performed using the VLOOKUP function in Microsoft Excel (after checking the need for a time offset to achieve the closest match). A downstream profile of conductance and flow was then plotted. The start of the spring inflows could then be marked at the distance where steady flow changed to increasing flow.

Equation 2-3: Estimating distance from GPS location. This equation was used to estimate distance between sequential GPS locations (from Pythagorean theorem). This relies on a metric grid, in this case New Zealand Transverse Mercator grid.

$$Displacement = \sqrt{(easting_i - easting_{i-1})^2 + (northing_i - northing_{i-1})^2}$$

The Karamu was kayaked from Havelock North to Floodgates (about 11 km) towing a water quality meter (YSI ProPlus) that sampled every minute (measured conductance, temperature, dissolved oxygen and pH). Most locations named in this section are mapped in Figure 3-24. A separate handheld GPS (Garmin GPSmap 60) provided locations that were matched by time stamp. The 1-minute spacing provided measurement-spacing in the order of 50 to 100 metres. The probe was placed in a stilling well located on the keel line of the boat ("Mission Flow" kayak for ADCP), and so measured chemistry at 10 to 20 cm below the water surface. The water quality meter (YSI Proplus) and GPS were synchronised within 3 seconds of each other. On one occasion, a second water quality meter (YSI Proplus) was left stationary at the downstream end of the reach (Karamu at floodgates) to confirm that conductance did not change markedly over the day. Avoiding rain, floods and small freshets is important for ensuring steady electrical conductance throughout the day. Relative change in conductance is of prime importance here, with weekly calibration of the conductance probes undertaken to improve the estimates of inflow conductance for comparison between dates. To check for calibration drift through the day, measurements from a conductance standard solution (e.g. 1413 µS/cm KCl buffer solution) taken before and after the survey was added to the protocol for more recent surveys.

For the 13/3/2014 survey, flow was measured at selected sampling points, including the survey start point (Karamu at Havelock Rd) and the survey end point (Karamu at floodgates). These two sites have concrete flumes that are less affected by aquatic plant growth. Major ion samples were also collected at the gauging sites, in addition to samples from two springs flowing directly into the Karamu (upstream of SH2 and adjacent Golflands golf course). The conductance of tributaries was measured by paddling the kayak with portable into these streams.

The longitudinal conductance survey was repeated on 1/4/2014, this time focussing on a shorter section of the Karamu (2.7 km reach between Flanders Rd and SH2 bridge; Figure 3-24). The same water quality meter was used (YSI ProPlus). A Garmin eTrex was used for GPS locations. The sampling interval was every 30 seconds this time, providing a closer spacing of measurements (typically 35 to 45 metres). While the downstream kayak survey was being completed, two people walked the true right bank looking for surface springs and seepage areas, taking photographs and measuring electrical conductance.

The Karamu at floodgates was gauged on the same day, with three additional sites gauged the next day (2/4/2014) in an effort to quantify flow changes over the gaining section (end of Watson Road – 1.1 km upstream of SH2; Flanders Rd; Havelock Rd). This section of stream is difficult to gauge because it is mostly

too deep to wade and has too much aquatic plant growth for a boat-mounted flow-meter (ADCP). Only two suitable locations, with less plants and a more stable bed (exposed cobble at Watson Rd; coarse sand at Flanders Rd), were identified during the kayak survey. Considerable time was still required to clear weed, and the deeper areas were marginal for wading, even under low-flow conditions. A good flow estimate was obtained at Flanders Rd, with four passes by the M9 ADCP (mounted on the hydroboard) providing flow estimates within 5% of the mean. The flow measurements at Watson Rd were more variable (within 22% of mean, from 8 passes).

A third kayak conductance survey was completed 5/2/2015 between Flanders Road and the Raupare confluence (Hach HQ40d, Garmin GPSMap60). The Karamu was gauged the day before, both at floodgates and at Havelock Road. Flow from rated stage was used to estimate inflow from the Mangateretere Stream. Data from a fourth run at higher flows (10/11/2016) could not be used because the GPS memory card filled up (1 second readings generated too much data) and because a large unaccounted flow was entering the Karamu via the Karituwhenua Stream (Brookvale municipal well testing).

A similar longitudinal conductance survey was completed for the Tutaekuri-Waimate Stream. Conductance was surveyed between Swamp Rd and the Goods Bridge monitoring site (6.2 km, Figure 3-35), and this was intended to delineate gains downstream of the reach surveyed by GNS using Distributed Temperature Sensing (Moridnejad, 2015). The electrical conductance survey was completed on 4/12/2015, with a sampling interval of 10 seconds for conductance and temperature (Hach HQ40d) and every 5 seconds for GPS location (Garmin GPSMap64, with GPS & GLONASS receiver). The conductance clock was 2 seconds faster than the GPS clock, and this offset was removed in post-processing. The conductance sensor was calibrated against a 1413 µS/cm standard on 2/12/2015, and was checked against the same standard after the survey (measured 1430 in the 1413 µS/cm standard, after 1 hour for temperature stabilisation). Flow gaugings were completed 2 days prior to the kayak survey on 2/12/2015 both upstream (Roadside Drain plus Repokai te Rotoroa) and downstream (Goods Bridge) of the survey reach. Stage monitoring at Goods Bridge confirmed negligible flow change between flow and conductance measurements (0.1% change in rated flow). The flow magnitude from each tributary through the survey reach was calculated from dilution of their measured conductance (Paherumanihi, Korokipo, Waipiropiro, Waima).

2.1.3 Spot Conductance

The use of electrical conductance is not limited to intensive studies. It can also be used as a screening tool, before committing to more intensive investigations. Spot measurements of electrical conductance at easily accessible locations (or archived data) can quickly identify large changes in chemistry. For example, a single visit to Waitio Stream revealed a large drop in conductance between the hill country and Ohiti. This was then used to define the section of stream targeted for closely spaced measurements of conductance and flow. Prior to this investigation commencing, spot measurements of electrical conductivity were used to investigate spring inputs to the Tutaekuri-Waimate Stream (Rabbitte, 2011).

The electrical conductance method will only reveal spring inflows if there is sufficient contrast in conductance between the stream inflows and the groundwater springs. The greater the spring input (as a proportion of stream flow), the smaller the contrast in conductance required to detect the spring input.

2.1.4 Longitudinal Temperature

Groundwater temperatures are typically stable (e.g. 15 ± 0.5 °C across seasons), compared to stream temperatures that can increase 10 °C from night to day. This provides an opportunity for detecting spring inflows using temperature, in much the same way as electrical conductance. Compared to using electrical conductance, the benefit of using temperature is that it does not depend on different sources feeding the stream and groundwater (e.g. to achieve a contrast in major ions). However, the difficulty with using

temperature is that stream water temperature increases through the day, and along the length of the stream, making interpretation of temperature changes between spot measurements difficult. GNS (Institute for Geological and Nuclear Sciences) overcame this problem by using a 1 km long fibre optic cable that was laid on the bed of the river. This measures temperature at hundreds of points along its length at the same time, and repeats all these measurements day and night to capture the largest temperature contrast. This specialised DTS (distributed temperature sensing) equipment was deployed by GNS in the Tutaekuri-Waimate to quantify flow gains upstream of State Highway 50. See the GNS reports for more detailed methods (Moridnejad, 2015). Deployments were also made by GNS in the Ngaruroro River, though the results are not yet available.

2.1.5 Stable Isotopes of Water

The origins of groundwater, and the springs fed by groundwater, are concealed from view. In large alluvial-valleys, like the Heretaunga Plains, flowpaths can cross surface catchment boundaries via underground gravel channels deposited by prehistoric rivers (i.e. palaeochannels). Measuring the isotope ratios of a given spring can offer clues to where the water has come from, especially if each potential source has a unique isotope ratio (Taylor *et al.*, 1989).

Stable isotopes of water are naturally occurring, do not degrade in groundwater, and are inexpensive to sample and analyse (Stewart & Taylor, 1981). These isotopes are useful for spring investigations because each river can have a unique isotope ratio (or isotopic signature). For example, a river draining cool mountain ranges will have more negative values of $\delta^{18}O$ and $\delta^{2}H$, compared to a stream draining warm lowland areas (Stewart & Taylor, 1981).

The heavy water isotopes are less likely to evaporate from the ocean, and more likely to fall as rain. Cold temperatures affect the isotopes' movement from ocean, to clouds, to rain - more so than normal water. The net effect of this discriminatory movement is measured using $\delta^{18}O$ and $\delta^{2}H$, with ocean water producing values close to zero, compared to rain that produces more negative values. Rainfall in colder areas, and areas further inland, produce more negative values again. For a more detailed description of these physical processes, and other applications for isotope tracers, see Stewart and Morgenstern (2001). The heavy isotopes of water were used as tracers for this report, including:

- $\delta^{18}O$ – the isotope ratio of heavy oxygen ^{18}O to normal oxygen ^{16}O, divided by the standard isotope ratio for seawater (Vienna Standard Mean Ocean Water). Units are ‰ (parts per thousand).
- $\delta^{2}H$ – the isotope ratio of heavy hydrogen ^{2}H (deuterium) to normal hydrogen ^{1}H, divided by the standard isotope ratio for seawater (Vienna Standard Mean Ocean Water). Units are ‰.

Isotope sampling for this spring investigation was an add-on to water tracing and aging investigations initiated for groundwater modelling of the Heretaunga Plains (Morgenstern *et al.*, 2018). Some additional springs, streams and wells were sampled for this report, focussing on the spring inflows to the Karamu and Mangateretere streams whose origin was in question. Most warm-season sampling was conducted 3 and 4 March 2015 (Appendix B), when river levels were extremely low (e.g. flow for Ngaruroro at Fernhill was less than half the mean annual low flow). This sampling run covered the Karamu Stream, Mangateretere Stream, and the potential recharge rivers (see Appendix B for sites and results). Sampling of some additional springs and shallow wells (well number 2580 and 16202) was conducted 6 and 12 March 2016, again under extreme low-flow conditions. A second round of sampling for stable isotopes was completed in winter to help understand seasonal changes in groundwater origin. This was completed 23/8/2017, targeting winter baseflow conditions. A subset of sites was chosen to represent the lower Karamu Stream and upper Tutaekuri-Waimate. Additional sites were sampled to represent potential source water (Appendix B).

The stable isotopes reported here did not require special sample handling. Wells were purged to flush stagnant water from the casing, using casing volume and pumping rate to estimate the pumping time required. Electrical conductance was also monitored using a hand held probe to confirm the purged water had reached static levels before sampling. Samples were collected in bottles supplied by the GNS Stable Isotope Laboratory, who analysed the samples using isotope ratio mass spectrometry with reported precisions of 0.1 ‰ for $\delta^{18}O$ and 1.0 ‰ for $\delta^{2}H$ (Morgenstern *et al.*, 2018). Chemistry samples were collected at the same time, and sent to Hill Laboratories for analysis of major ions and nutrients.

For rivers and streams, flow measurements were made within 24 hours of sampling, under prevailing stable-flow conditions (see Section 2.1.1 for gauging methods). The two exceptions were warm-season sampling of the Mangateretere Stream and Tukituki River, for which flow was calculated from rated water level, which is not as accurate as gauged flow. Where isotope samples were collected directly from the spring head, flow was not measured.

Rain falling directly onto the Heretaunga Plains is a potential source of the groundwater that feeds springs, in addition to the river recharge sources described in the results (Sections 3.1, 3.2 and 3.3). The isotope signature of local rainfall was not sampled for this study because the high variability of heavy-water isotopes in rainfall (between seasons and rainfall events) necessitates long-term monitoring to adequately characterise this source (Stewart & Taylor, 1981). Data were obtained from GNS (Baisden *et al.*, 2016) for two locations at low-elevations close to the Heretaunga Plains (station 29509, Lat. -39.425 Long. 176.825, n=33, station 30050 Lat. -39.725 Long. 176.975, n=30). These were sampled monthly between 2007 and 2010, revealing the more negative $\delta^{18}O$ from winter rainfall compared to summer (Figure 2-1). Bias towards months that were sampled more often over the three year period was removed by first averaging all January samples across years, and so on for other months, before averaging over the 12 calendar months. The long-term mean $\delta^{18}O$ was -5.9‰ ($\delta^{2}H$ -36.0‰) for site 30050. The mean $\delta^{18}O$ was more negative for the other site (station 29509), with $\delta^{18}O$ of -6.7‰ ($\delta^{2}H$ -42.4‰). However, an outlier in April (Figure 2-1) had high leverage on this mean (excluding outlier $\delta^{18}O$ -6.3‰, $\delta^{2}H$ -39.5‰). An estimate for rainfall for the unconfined recharge area of the Heretaunga Aquifer was calculated using the isoscape model developed by Baisden *et al.* (2016). This predicted a long-term mean $\delta^{18}O$ of -5.0‰ ($\delta^{2}H$ -36.0‰) for the selected Virtual Climate Station (station 28640, Lat. -39.625, Long. 176.775, elevation 19 m, period July 2013 to June 2015).

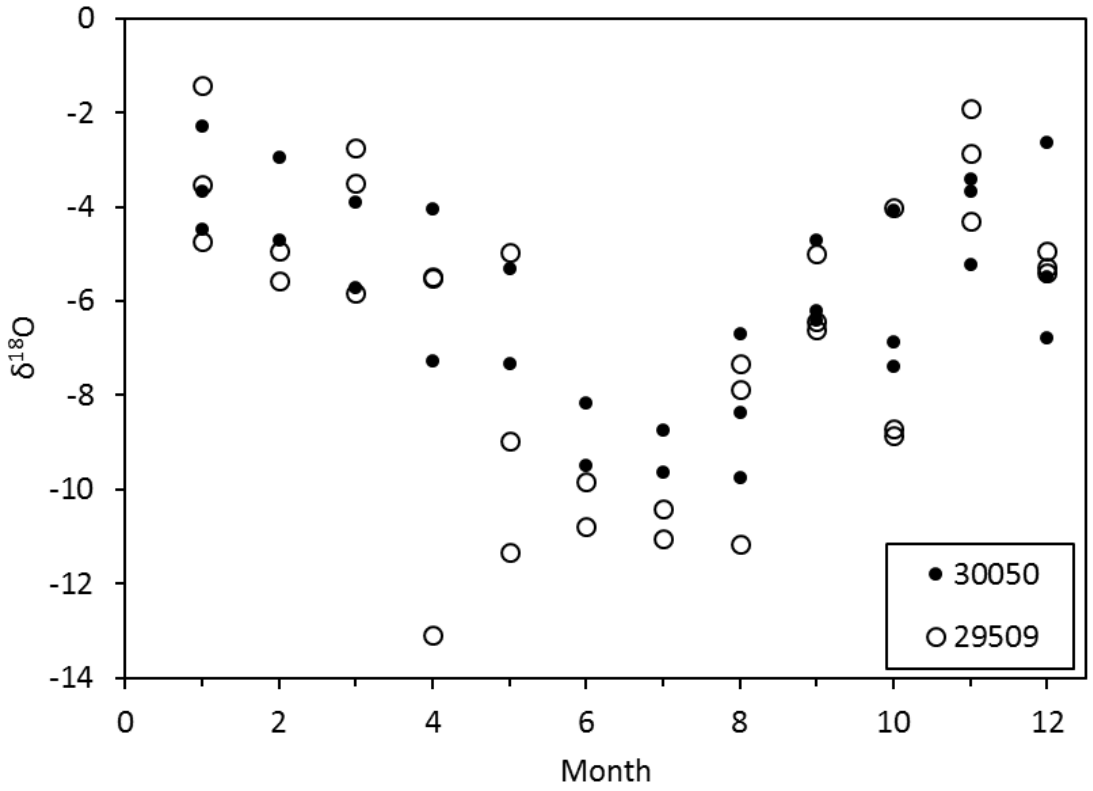


Figure 2-1: Stable isotopes in rainfall by month. The $\delta^{18}O$ for rainfall samples collected at two sites (30050 and 29509) close to the Heretaunga Plains. Samples were collected for a separate study (Baisden *et al.*, 2016) between 2007 and 2010, and are over-plotted by calendar month.

Understanding the isotopic signature of rainwater is the first step in understanding the isotopic signature of groundwater originating from local rainfall recharge (Gat, 1995; Stewart & Taylor, 1981). Small rain events falling on dry soils can evaporate or run off before reaching groundwater, producing a seasonal bias to recharge in winter and spring. But, in addition to seasonal bias, the water that is recharged has been fractionated by evaporation of lighter isotopes from plant and soil surfaces (Gat, 1995). Hence, the bias toward more negative $\delta^{18}O$ from winter recharge is countered by a bias to less negative $\delta^{18}O$ as a consequence of evaporative fractionation (Gat, 1995). In the absence of lysimeter isotope data, an appreciation of the net effect of these, and other, physical processes on the isotopic signature of rainfall recharge (i.e. the isotopic transfer function) is provided by streams draining low-elevation sub-catchments, particularly those lacking inter-basin groundwater connections and lakes. The least negative isotope signatures were observed in such catchments, with $\delta^{18}O$ in the range of -5.8‰ to -6.0‰ and $\delta^{2}H$ of -35‰ to -38‰ (e.g. Paritua at water wheel 4/3/2015, Kaikora upstream of Papanui confluence 17/4/2015, synthesized Awanui upstream of Karewarewa 3/3/2015). Not all low-elevation catchments were this high. For example, the stream draining the highest elevation sub-catchment had a $\delta^{18}O$ of -6.8‰ and $\delta^{2}H$ of -41‰ on 5/3/2015 (Te Waikaha Stream at Mutiny Road drains Mt Erin). In the absence of direct measurements from lysimeters, this report used interim $\delta^{18}O$ of -6.0‰ and $\delta^{2}H$ of -39‰ to characterise rainfall recharge on the Heretaunga Plains (average Awanui, Karewarewa, Paritua and Kaikora streams during low flow conditions).

2.2 Measuring Flow Losses

The physical processes by which spring inflows from groundwater enter streams are the same, in many respects, to flow losses from streams to groundwater. The change from a gain to a loss occurs when the groundwater level drops below the stream level. However, there are fewer options for measuring a loss of flow from a stream, compared to measuring a gain. For example, the downstream change in electrical conductance cannot be used to measure a flow loss because the concentration of major ions is not affected by a flow loss (i.e. major ions are lost along with the flow). For this investigation, measuring a flow loss required direct measurement of the stream flow at discrete locations along the river (the change in flow equating to the net loss). As described in Section 2.1, the number of points along a stream where flow can be accurately measured is limited. This distance between suitable sites will limit our ability to resolve where losses start and end.

In some cases, it was possible to refine the end of a losing section using an estimate of groundwater level (i.e. where stream level was the same as the adjacent groundwater level). Indicators of adjacent groundwater level include recorded water level from nearby wells, or the elevation of adjacent springs. These do not give an exact representation of the groundwater level directly below the riverbed. However, this information can improve the delineation compared to gaugings spaced several kilometres apart. Other evidence that was useful for locating springs (e.g. wetland vegetation), were sometimes also useful in narrowing down the end of a losing reach (see Section 2.1).

3 Results

In total, 64 spring locations were mapped for this report (Appendix A). These are described separately for each sub-catchment.

3.1 Ngaruroro

Flow losses to groundwater

The lower Ngaruroro River typically loses more than 4000 L/s to groundwater. This recharges the Heretaunga aquifer which, in turn, feeds many of the springs on the Heretaunga Plains. The Ngaruroro is the most intensively gauged river on the Heretaunga Plains (Dravid & Brown, 1997; Grant, 1965), with flow losses measured from 336 concurrent gaugings dating back to 1952. This is in addition to the 1392 gaugings at monitoring sites on the Ngaruroro River (Whanawhana, Fernhill and Chesterhope). Flow gaugings are the only way to measure flow losses at present (Section 2.2), because the loss of water has little effect on the chemical characteristics of the water (compared to spring inputs).

Most of the flow loss from the Ngaruroro occurs between Roys Hill and Fernhill (Dravid & Brown, 1997; Grant, 1965), with a median loss of 4250 L/s (107 concurrent gaugings at median gauged flow of 7020 L/s at Fernhill). This 5 km length of river is termed the “major loss” reach (Figure 3-4). The material underlying this reach includes unconfined gravels (see Section 1.3). The concurrent gaugings demonstrate some variability in the amount of water lost from the river (interquartile range 3900 to 4600 L/s loss). Some of this variability is attributable to measurement error, and the L/s of measurement error increases with flow (e.g. 10% error equates to ±100 L/s error at 1,000 L/s river flow, or ±10,000 L/s error at 100,000 L/s river flow). In addition, the true loss from the river can change over time. When groundwater levels are lower, more water may be lost from the Ngaruroro River (Dravid & Brown, 1997). Higher river flows can also increase the amount of flow lost, by increasing the pressure head and the area of wetted gravel through the losing reaches. However, there was a lack of correlation between flow loss and the total river flow, or groundwater level (R^2 = 0.02 and

0.01 respectively). Likewise, lower groundwater levels did not increase the likelihood of more loss (odds ratio 0.8 for flow loss >4250 L/s at groundwater levels <20 m at substation well 10371). It is still possible that loss of river flow increases at higher flows, but this increase is swamped by the increasing absolute error at flows sufficient to inundate the gravel bed. Factors unrelated to river level and groundwater level can also affect flow-loss, including clogging of the riverbed by silt.

Studying the variability in the flow loss is useful in understanding its physical drivers. But it is the consistency of this loss of river flow that is most vital to ecosystems and water users of the Heretaunga Plains. This steady loss keeps the aquifer recharged and the springs flowing through the dry Heretaunga summers. In contrast, less than 1 mm of total recharge depth was recorded for the 2013/14 summer, averaged across Bridge Pa, Maraekakaho and Substation lysimeters. The sustained flow of the Ngaruroro through summer is a product of rainfall in the ranges, which catch the westerly weather systems that often do not reach the Heretaunga Plains (Chappell, 2013).

An additional flow-loss was often detected downstream of Fernhill, which was smaller and more variable than the major-loss (median 120 L/s loss, interquartile range 410 L/s loss to 110 L/s gain, from 46 concurrent gaugings at a median gauged flow of 5010 L/s at Fernhill). This is termed the "variable loss" reach, and extends up to 3 km downstream of Fernhill (Figure 3-4). That 3 km length was estimated from the subset of concurrent gaugings with closely spaced sites between Fernhill and the Hawke's Bay Expressway (Figure 3-1). Harper and Hughes (2009) described the groundwater aquifer as relatively unconfined through the variable loss reach, with loss of flow enabled by groundwater levels that decline away from the river. No correlation was detected between the rate of flow loss and either the river flow or the groundwater level (R^2 = 0.007 and 0.06 respectively). However, the more sensitive odds-ratio indicated a flow loss was 9 times more likely to be measured at groundwater levels less than 20 m AMSL, compared to higher groundwater levels (using substation well 10371).

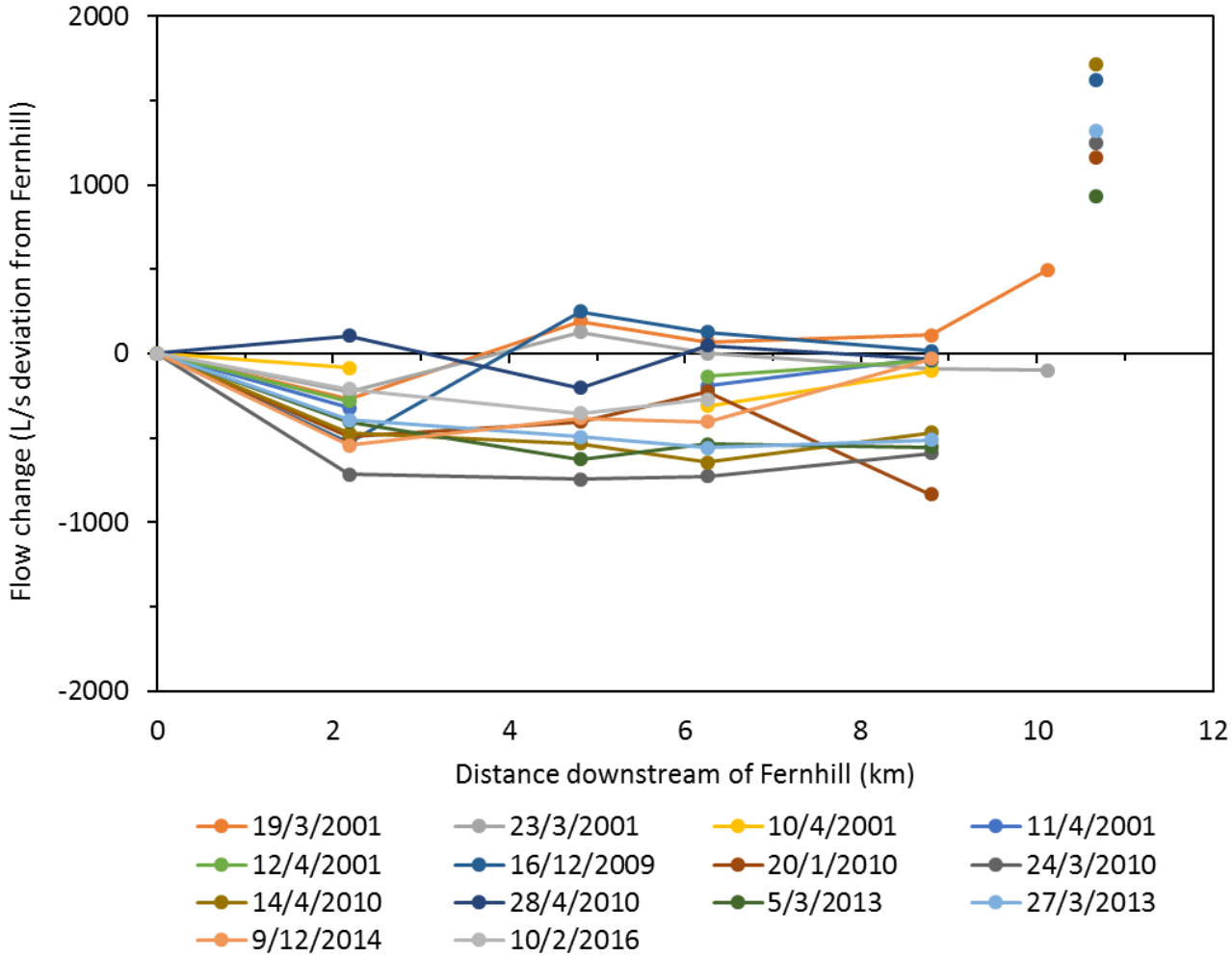


Figure 3-1: Ngaruroro flow-loss downstream of Fernhill. The magnitude of flow loss was variable and, when there was a loss, it typically occurred within 3 km downstream of Fernhill. To demonstrate this, Fernhill flow was subtracted from each gauging on the same day. Therefore, Fernhill is plotted as zero flow change, and negative values represent a flow loss. Each site is represented on the x-axis by its distance downstream of Fernhill, so Fernhill Bridge is at 0 km and Chesterhope Bridge is at 10.7 km. These dates were selected for this plot because multiple sites were gauged downstream of Fernhill (excluded gauging dates covered fewer sites). The flow increase at 10.7 km (Chesterhope) reflects the inflow from the Tutaekuri-Waimate Stream (see map, Figure 3-4).

An additional "minor loss" reach may also occur upstream of the major-loss reach (Figure 3-4). Dravid and Brown (1997) estimated the minor loss at 800 L/s. A similar magnitude of loss was estimated for this report, using different methods. The inflows located upstream of Ohiti were measured, including from the mainstem (Ngaruroro at Whanawhana), in addition to tributaries between Whanawhana and Ohiti (Poporangi, Otamauri, Mangatahi, Kikowhero, Maraekakaho, Figure 3-4). Using eight concurrent gaugings that included sites on all tributaries, the median loss was 450 L/s (interquartile range 350 to 750 L/s), (Figure 3-2). The actual loss would be higher if there were additional seepages from valleys draining directly to the river mainstem. Conversely, the actual loss would be lower if some of the flow that is lost subsequently returns to the Ngaruroro downstream of Ohiti (e.g. via Waitio Stream).

The location of this minor loss is uncertain. For example, Dravid and Brown (1997) defined the minor loss zone as the reach between Maraekakaho to Roys Hill (Figure 3-4), despite using concurrent gaugings from more than 10 km further upstream (Mangatahi) in estimating the magnitude of loss. The small number of

concurrent gaugings between Whanawhana and Ohiti, together with the small proportion of flow that is lost or gained, prevents better spatial definition of where the gains and losses occur.

Another outstanding issue is the potentially large subsurface flow proposed by Grant (1965). It is reasonable to expect some water to travel as groundwater through the alluvial gravels, in addition to the surface flow. But the magnitude of subsurface flow proposed by Grant (1965) is surprisingly high (30% of flow). Grant (1965) observed that the flows for the Ngaruroro at Mangatahi exceeded the sum of inflows, and concluded that additional flow was missed at upstream sites (e.g. Whanawhana) because it travelled as subsurface flow. It is unclear why this subsurface flow would resurface at Mangatahi, given this sites alluvial valley setting, compared to Whanawhana, which is confined within a bedrock canyon. A closer look at the concurrent gaugings indicates that the higher flows observed at Mangatahi might be attributable to gauging error. Grant (1965) based his conclusions on two concurrent gaugings (28/11/1961 and 13/2/1964). Subsequent gaugings instead indicate that, at most sites on most occasions, the measured Ngaruroro flow is equivalent to, or less than, the summed inflows (Ngaruroro at Whanawhana plus gauged tributaries). Of 73 gaugings along the Ngaruroro, 68 measurements were equivalent to, or less than, the estimated sum of inflows, allowing for a 10% margin of gauging error (Figure 3-3). It is possible that those five (out of 73) gaugings with higher than expected flows reveal a large sub-surface flow. However, it is a weak evidence in its own right.

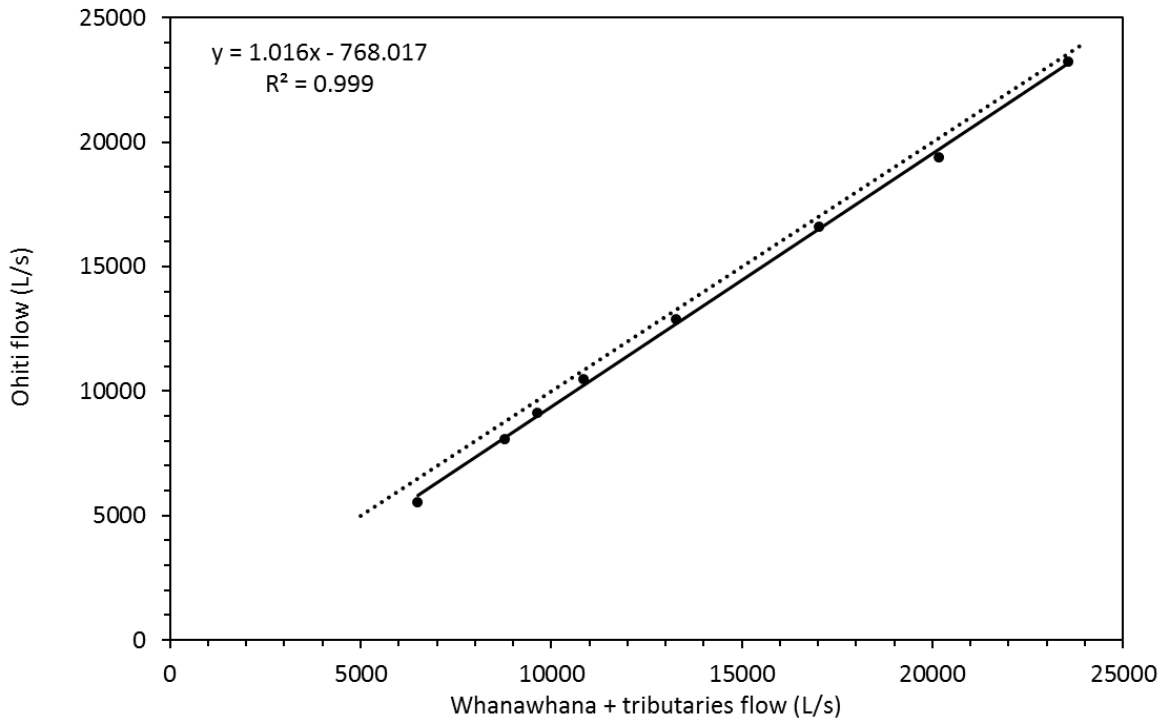


Figure 3-2: Ngaruroro flow losses upstream of Ohiti. The flow measured at Ohiti is plotted against the flow expected if there were no losses, which was estimated from gaugings at Whanawhana, plus all major tributaries (Poporangi at Ohara Station, Otamauri at Whanawhana Rd, Mangatahi at Aorangi Rd, Kikowhero at Crownthorpe Rd, Maraekakaho downstream of Tait Rd). Each point is a concurrent gauging run, with all major tributaries gauged on eight occasions (28/11/1961, 13/2/1964, 11/11/2009, 16/12/2009, 20/1/2010, 24/3/2010, 14/4/2010, 28/4/2010, 5/3/2013). A linear trend line is fitted to all eight dates (solid line, with equation). Points below the dotted-line (1:1) indicate a net flow loss to groundwater.

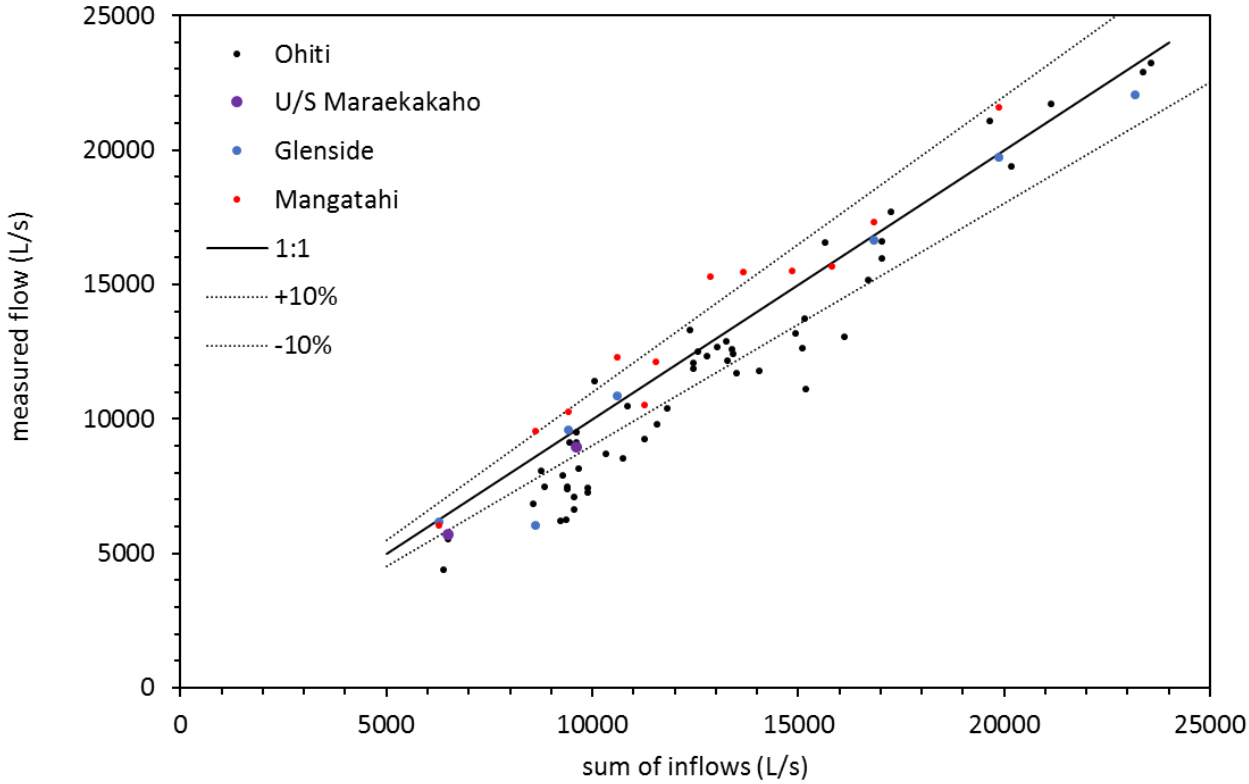


Figure 3-3: Ngaruroro measured flow versus summed inflows. Flows for Ngaruroro at Mangatahi (red dots) exceeded the sum of inflows (x-axis) on several occasions (red points above solid line), leading Grant (1965) to conclude that large subsurface flows were reappearing at this site. The alternative explanation is that the higher flows recorded at Mangatahi were attributable to measurement error. To investigate the alternative, this plot demonstrates the small number of flow measurements that exceeded summed inflows by more than a 10% margin of error. The inflows were gauged on 8 occasions, and inflows for other occasions were estimated from correlations of tributary inflows with Whanawhana. Points above the solid black line had measured flow that exceeded the sum of inflows. The dotted-lines delineate a nominal 10% margin of measurement error (QC500, Willsman *et al.*, 2013).

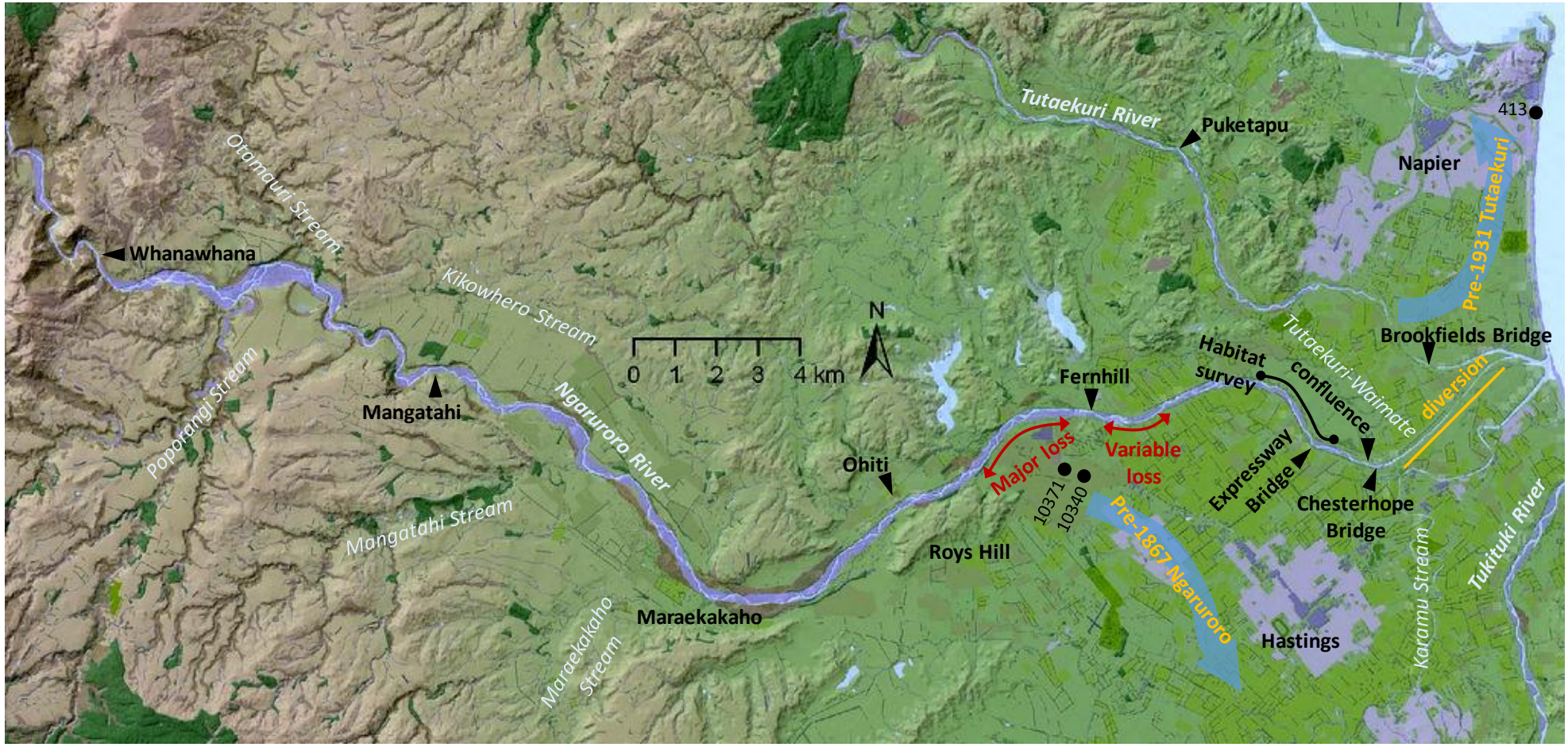


Figure 3-4: Ngaruroro River. Map of the Ngaruroro River showing the location of flow monitoring sites (e.g. Whanawhana, Fernhill and Chesterhope Bridge), reaches that lose flow to groundwater (Major loss and Variable loss) and the fish-habitat survey reach (Johnson, 2011a). The location of the "Minor loss" reach upstream of Roys Hill is uncertain (see text). The section of river that was diverted to a constructed channel is also labelled ("diversion"), in addition to the approximate Ngaruroro flowpath prior to the 1867 flood. Selected monitoring wells (black dots) are labelled with the well number.

Stable isotopes of water

The major loss of flow from the Ngaruroro River to the Heretaunga aquifer makes it a likely source for many of the groundwater springs in neighbouring catchments. Stable isotopes can be used to trace groundwater originating from the Ngaruroro, provided it has an isotope signature that is distinct from other sources (Section 2.1.5). For example, Taylor noted more negative values of $\delta^{18}O$ for Ngaruroro water (Section 7.8 in (Dravid & Brown, 1997). Stable isotope ratios for river water can vary seasonally, and in response to rainfall (Stewart & Taylor, 1981). Wells that are located close to the river recharge source typically provide a less variable isotope ratios that better typify the groundwater originating from that recharge source, compared to a single river sample (Scott, 2014).

To determine the isotope signature of Ngaruroro sourced groundwater, samples of groundwater were collected from a shallow well closer to the Ngaruroro major loss reach (well no. 10340, 17 m deep, 2.5 km from river, Figure 3-4). This groundwater had a mean $\delta^{18}O$ value of -7.6‰ (-7.5‰ on 23/6/2014 and -7.7‰ on 23/8/2017). Groundwater wells with a $\delta^{18}O$ value of -7.5‰ to -7.7‰ were widespread across the Heretaunga Plains (e.g. -7.6‰ from well 1459 at 53 m, -7.6‰ from well 15003 at 55 m, -7.7‰ from well 16361 at 23 m). However, other wells that are also likely to be sourced from the Ngaruroro had more negative $\delta^{18}O$ between -7.8‰ and -8.2‰, and these occurred at a range of depths (e.g. -7.8‰ from well 16360 at 65 m, -8.0‰ from well 5915 at 38 m, -8.2‰ from well 1674 at 38 m).

Results from the river itself help shed light on some this variability (Figure 3-5). Samples collected for this study ranged from -7.4‰ in March (mean of two sites sampled 3/3/2015) to -8.1‰ in August (mean of four sites sampled 23/8/2017). Previous monitoring of isotopes from the Ngaruroro River (Morgenstern *et al.*, 2018) produced a wider range of $\delta^{18}O$ values (Figure 3-5). To maximise the generality of sampling to date, the relationship between $\delta^{18}O$ and ocean temperature was used to predict mean $\delta^{18}O$ for a longer period of time (Figure 3-5). This relationship was based on the mean sea-surface temperature for the Tasman Sea for the month prior to isotope sampling (www.bom.gov.au/cgi-bin, data accessed 31/10/2017). In addition to affecting evaporation from the sea surface, the temperature of the ocean also affects the weather patterns that drive local rainfall and evaporation. These predictions from ocean temperature demonstrate the inter-year climate variability that contribute to the spatial variability of $\delta^{18}O$ in groundwater (Figure 3-6). For example, well 413 in Napier (Figure 3-4) was estimated to have a mean residence time of 110 years (Morgenstern *et al.*, 2018). The more negative $\delta^{18}O$ of -8.08‰ for this well corresponds with more negative river $\delta^{18}O$ of -8.16‰ predicted for 110 years prior to sampling. The climate was generally cooler then, and hence river water from recent years is predicted to have a less negative $\delta^{18}O$ of -7.9‰ (-7.90‰ for 2013-2017, -7.92‰ for 2008-2017, -7.93‰ for 1998-2017).

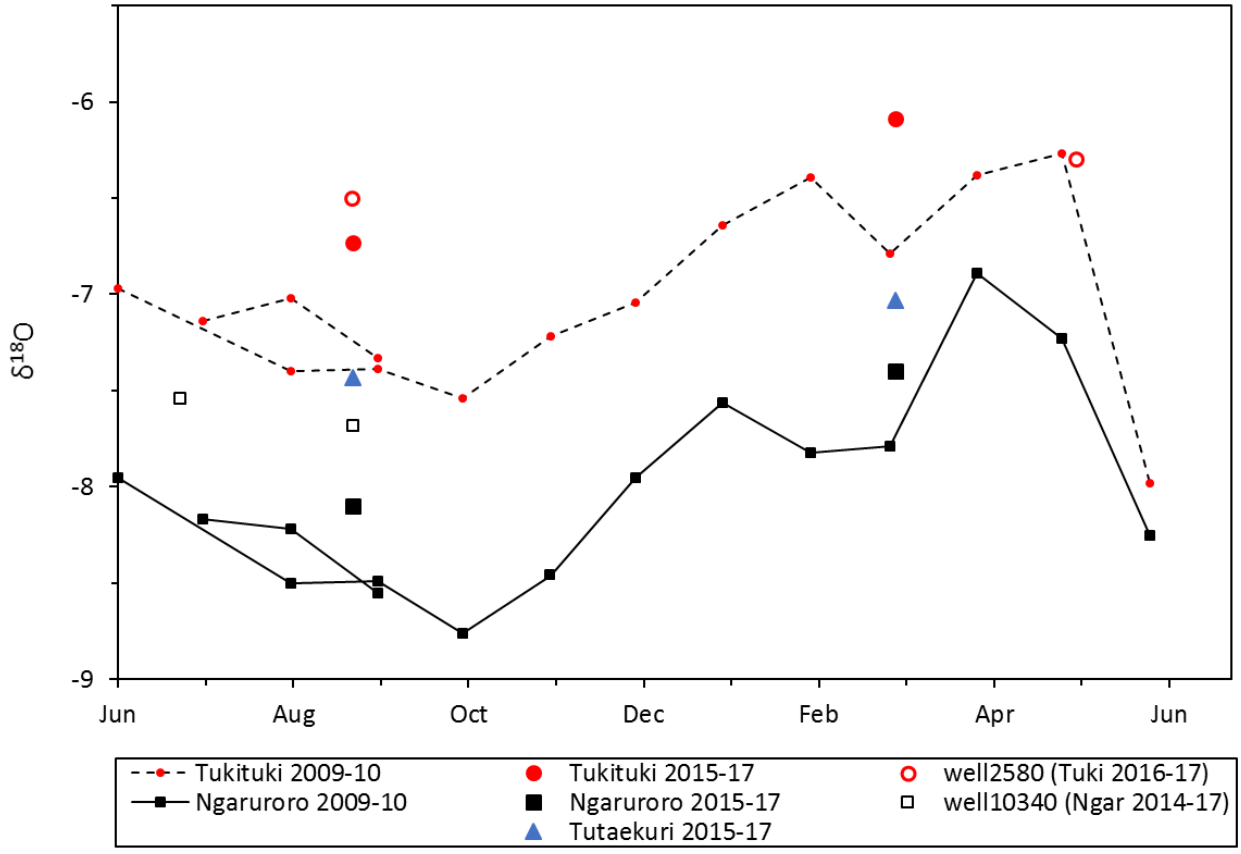


Figure 3-5: Oxygen stable isotopes in river water. The data for stable isotopes of oxygen ($\delta^{18}O$) from Morgenstern *et al.* (2018) for the period 2009-2010 (Ngaruroro at Chesterhope, Tukituki at Red Bridge) are plotted together with samples collected for this study in March 2015 and August 2017 (Ngaruroro, Tutaekuri and Tukituki). Over-plotting the datasets by calendar month demonstrates both the seasonal and the inter-annual variability in $\delta^{18}O$. This also demonstrates the consistent pairwise difference between the Ngaruroro and Tukituki rivers (paired t-test $P<0.0001$, $n=17$ pairs). Results for two wells are over-plotted, which were located close to the Tukituki River (well 2580) and the Ngaruroro River (well 10340).

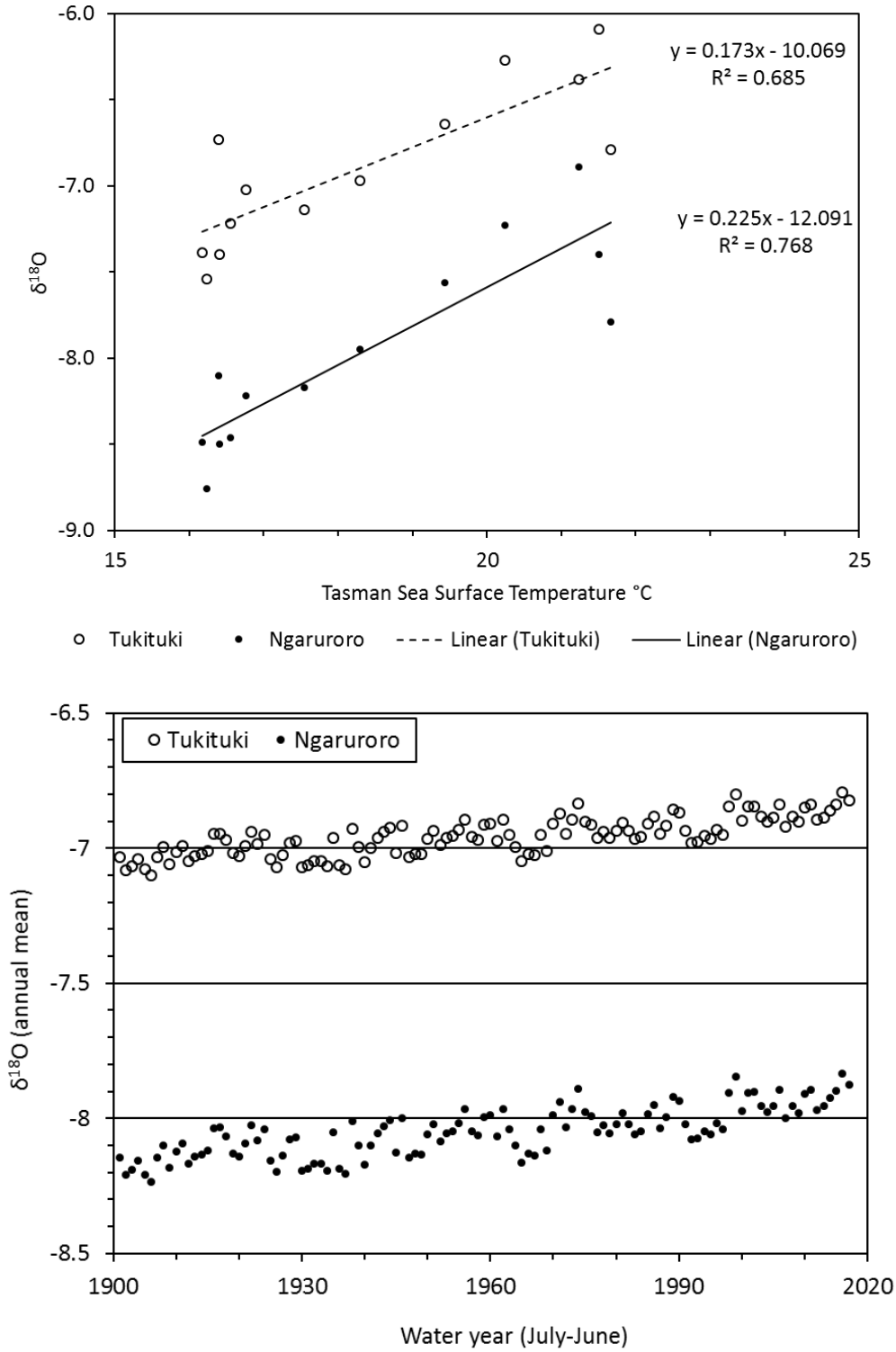


Figure 3-6: Predicting long-term $\delta^{18}O$ using sea surface temperature. The upper plot shows the relationship between river samples of $\delta^{18}O$ (Tukituki and Ngaruroro) and mean surface temperate of the Tasman Sea for the month prior to each sample. This excludes two sample dates when flow fluctuations were high (July 2010, October 2010). Using the equation from the upper plot, $\delta^{18}O$ was predicted for other years for which we have sea-surface temperature (annual mean for the water year).

Adopting a $\delta^{18}O$ signature of -7.9‰ for Ngaruroro River water still leaves a shortfall in explaining the many wells with $\delta^{18}O$ closer to -7.6‰. Morgenstern *et al.* (2018) concluded that these wells with a less negative $\delta^{18}O$ were influenced by local rainfall recharge, going so far as to suggest a $\delta^{18}O$ of -7.6‰ may represent the isotope signature of rainfall recharge. In the absence of lysimeter samples, or similar, to characterise rainfall recharge on the Heretaunga Plains, this report adopted an interim $\delta^{18}O$ of -6.0‰ (Section 2.1.5). This rainfall signature provides a strong contrast with the value of -7.6‰ seen in many wells, suggesting rainfall made a minor contribution to groundwater recharge. For example, if local rainwater recharge has a mean $\delta^{18}O$ of -6.0‰ (see Section 2.1.5) and Ngaruroro sourced groundwater had a mean $\delta^{18}O$ of -7.9‰, then a mix of river water with 15% rainfall recharge would be required to produce a mean $\delta^{18}O$ of -7.6‰, as observed in many wells. Seasonal bias in the transfer of river water to groundwater is minimised by the low variability for the major-loss reach (interquartile range 3900 to 4600 L/s loss).

The contrast between the Tukituki and Ngaruroro river water was sufficient to differentiate the two sources (Figure 3-5), with a mean difference of -1.1‰ between 17 paired samples (paired sample T-test P<0.0001, n=17 pairs). The isotopic signature of the Tukituki is described in Section 3.2. The few samples that were available from the Tutaekuri River indicate its isotopic signature lies between the Tukituki and Ngaruroro (see Section 3.3). The contrast in isotopic signatures for these potential water sources is summarised in Figure 3-7.

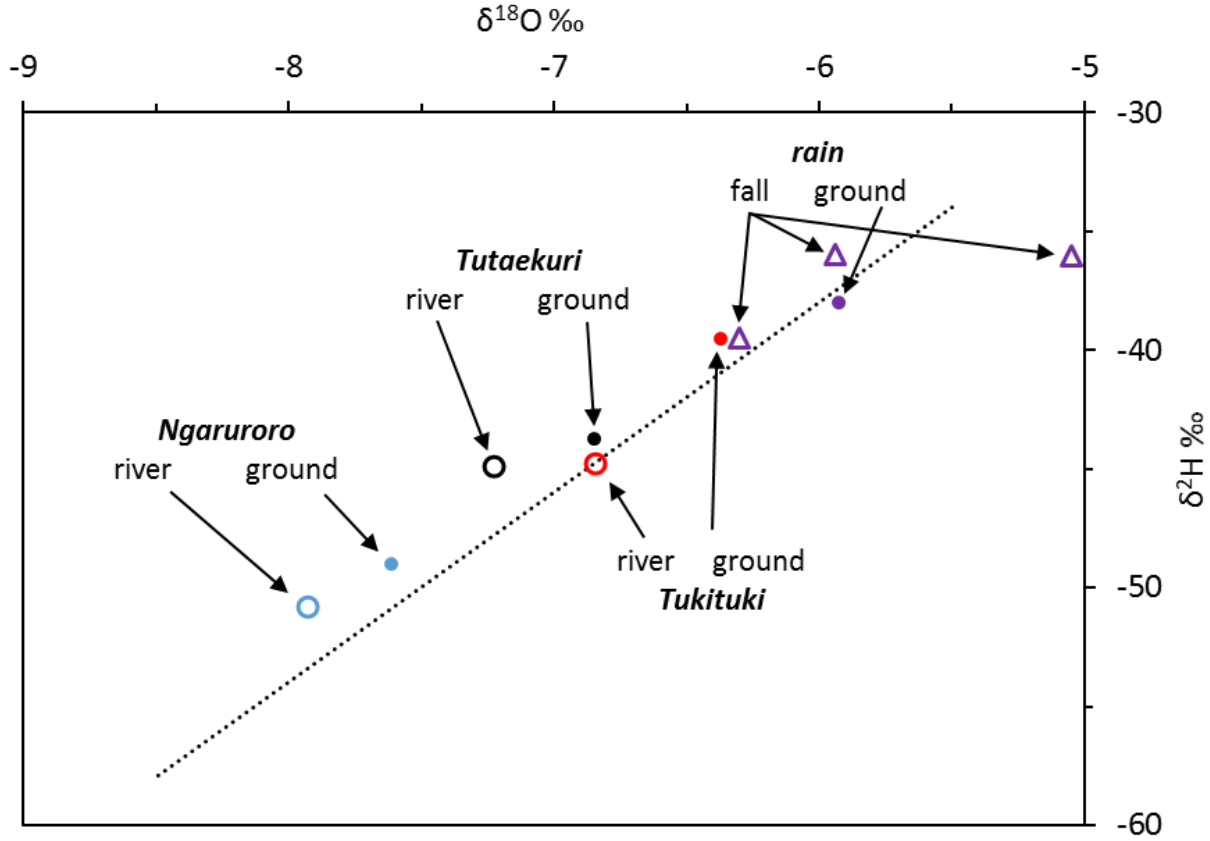


Figure 3-7: Stable isotope signatures. This plot of $\delta^{18}O$ versus $\delta^{2}H$ synthesizes the stable-isotope data for potential sources of spring water on the Heretaunga Plains. For example, employing the various datasets available, river water from the Ngaruroro River is expected to have a long-term mean $\delta^{18}O$ of -7.9‰ and $\delta^{2}H$ of -51‰, which translated to a groundwater signature of $\delta^{18}O$ of -7.6‰ and $\delta^{2}H$ of -49‰. This summarises the signature conclusions from this section, plus sections 3.2 (Tukituki), 3.3 (Tutaekuri) and 2.1.5 (rainwater). The meteoric water line is plotted as a dotted line (Stewart & Morgenstern, 2001).

History

Turning to the history of the Ngaruroro, this river has experienced notable channel changes. The Hawke's Bay Catchment Board diverted a 7 km section of the Ngaruroro River from what is now referred to as the Clive River, to instead flow downstream of Chesterhope (labelled "diversion" in Figure 3-4). The river was diverted to address siltation problems in an overflow channel, which was itself constructed in the 1930s to reduce flooding (MacGeorge, 1989). The subsequent 1969 diversion of the entire flow resulted in an 80% reduction in the MALF for the original channel, which now only conveys water from the Karamu Stream and Raupare Stream.

In addition to the 1969 artificial diversion, the Ngaruroro River has changed its own course in the past during major floods. Prior to 1867, the Ngaruroro flowed down what is now the Irongate Stream, into the Karamu Stream upstream of Havelock North (labelled "pre-1867 Ngaruroro" in Figure 3-4). A large flood in 1867 shifted the course of the Ngaruroro River to its present path north of Twyford (*NIWA events catalogue*; HBRC, 2014). Past river-avulsions may have also changed the location and magnitude of spring outflows on the Heretaunga Plains. The Heretaunga Plains is itself a sequence of gravels deposited by the Ngaruroro, Tutaekuri and Tukituki rivers. For example, the layer of alluvial pumice on the Heretaunga Plains (Figure 3-14) was deposited by the Ngaruroro River. The location of these pumice layers likely reveals the river's path (plus floodwater deposits) after Taupo's Hatepe eruption in 177 AD (Grant, 1996; Griffiths, 2001).

Monitoring sites

The flow losses from the Ngaruroro River raise a problem with the location of the monitoring site at Fernhill. The measured flow is unique to Fernhill, given that flow losses can occur both upstream and downstream of this point. This flow-loss problem adds to the technical challenges with maintaining a flow monitoring site in a mobile braided channel. The Fernhill site has a long data-record, with water level measured here since the SH50 Bridge was completed in 1952. The predecessor of NIWA (Ministry of Works) abandoned the Fernhill site in 1976, and relocated to Chesterhope Bridge (Figure 3-4). Motivations for abandoning the Fernhill site included the periodic movement of the flowing channel away from the monitoring point, as well as the need for a new stage-to-flow rating after each flood that changed the channel shape (Arnold & Horrell, 2013).

The predecessor of the Hawke's Bay Regional Council (Hawke's Bay Catchment Board) adopted the Fernhill monitoring site, after the Ministry of Works stopped monitoring here, because Fernhill was needed for minimum flow monitoring and flood monitoring. Since 2005, Hawke's Bay Regional Council has invested more heavily in flow gaugings at Fernhill to improve the accuracy of stage-to-flow ratings during low-flow periods (Arnold & Horrell, 2013). Diggers are sometimes used to redirect river flow past the monitoring point. The flow records from sites at Whanawhana and Chesterhope have also been used to provide a double check on flow estimates at Fernhill. The quality of flow data has improved since 2005 (Arnold & Horrell, 2013). However, the extra effort required to maintain the Fernhill site serves to highlight its problems.

Alternatives to the Fernhill site include Chesterhope bridge and the Hawke's Bay Expressway (SH50a) bridge. Monitoring low flows at the Chesterhope site would be complicated by the Tutaekuri-Waimate Stream inflow that nearly doubles the flow of the Ngaruroro River during low flow conditions (the confluence is located a short distance upstream of Chesterhope Bridge). The Chesterhope site therefore represents a short section of river (1.6 km of river between the Tutaekuri-Waimate confluence and the saltwater tidal reach). Further complicating matters, the Tutaekuri-Waimate has an overflow channel that bypasses the Ngaruroro at Chesterhope Bridge during floods.

The construction of the Hawke's Bay Expressway has provided a third option for a monitoring site that was not available when Ministry of Works relocated to Chesterhope Bridge (Figure 3-4). The river channel at the Expressway Bridge is mobile, but not as mobile as Fernhill (channel mobility was compared using a series of aerial photographs). The expressway site is located downstream of the variable-loss reach, so the flow here

represents an 8.4 km reach. However, the cost of constructing a new monitoring site on the Expressway Bridge would be significant, plus there is no footpath for site inspection and maintenance. Relocating flow monitoring from Fernhill to Chesterhope could therefore be the better option, provided an adequate correction for the Tutaekuri-Waimate inflows can be achieved.

3.2 Lower Tukituki

This report focusses on the section of the Tukituki River traversing the Heretaunga Plains. A separate report characterises the hydrology of the broader Tukituki catchment, including losses and gains across the Ruataniwha Plains (Wilding & Waldron, 2012).

Concurrent gaugings along the lower Tukituki River revealed a loss of river flow to groundwater (Figure 3-8). This flow loss is not as well quantified as the Ngaruroro loss, with ten concurrent gaugings completed to date (5 of those in the last 5 years). There are insufficient data to quantify how this loss varies over time. However, 14 data pairs were compiled by using stage-to-flow data from Red Bridge together with gaugings compiled from two downstream sites (Black Bridge and Tennant Rd, in order of preference). Using this expanded dataset, the median flow loss was 910 L/s, with an interquartile range of 640 to 1050 L/s. Only one concurrent gauging measured a significant flow gain (1683 L/s flow increase between Red Bridge and Black Bridge on 16/11/2005). This gauging set was excluded from the flow loss estimates (plotted as 'X' in Figure 3-8) because the downstream site was gauged when river flows were higher, resulting from a small freshet passing through the river.

Concurrent gaugings were also used to estimate where the flow loss occurs (Figure 3-9). Unfortunately, the variability between flow measurements is high relative to the flow loss between adjacent sites, which may reflect measurement error or local anomalies (e.g. hyporheic exchange flow). The variability makes interpretation difficult, in terms of where flow losses start and stop. Available information indicates that flow loss is mostly confined to the section of river between River Road and Tennant Rd. The valley is confined by hill-country upstream of River Road, limiting the space available for a gravel aquifer to develop. The gauging data supports this, with little change in flow upstream of River Road, either from the mapped gaugings (Figure 3-9) or earlier gauging data at nearby "Site E" (15/12/1966, 30/1/1967).

Additional concurrent-gaugings are recommended to confirm that flow losses are confined between River Road and Tennant Road. Of the four concurrent gaugings for the reach further downstream, between Tennant Road and Black Bridge, two measured a gain and two a loss. The emergence of perennial springs adjacent to the river at Tennant Road (feeding Grange Stream, Section 3.14) indicate that the water table is high enough downstream of Tennant Rd to limit further flow loss during summer. This information was used to inform mapping of the losing reach, as shown in Figure 3-24.

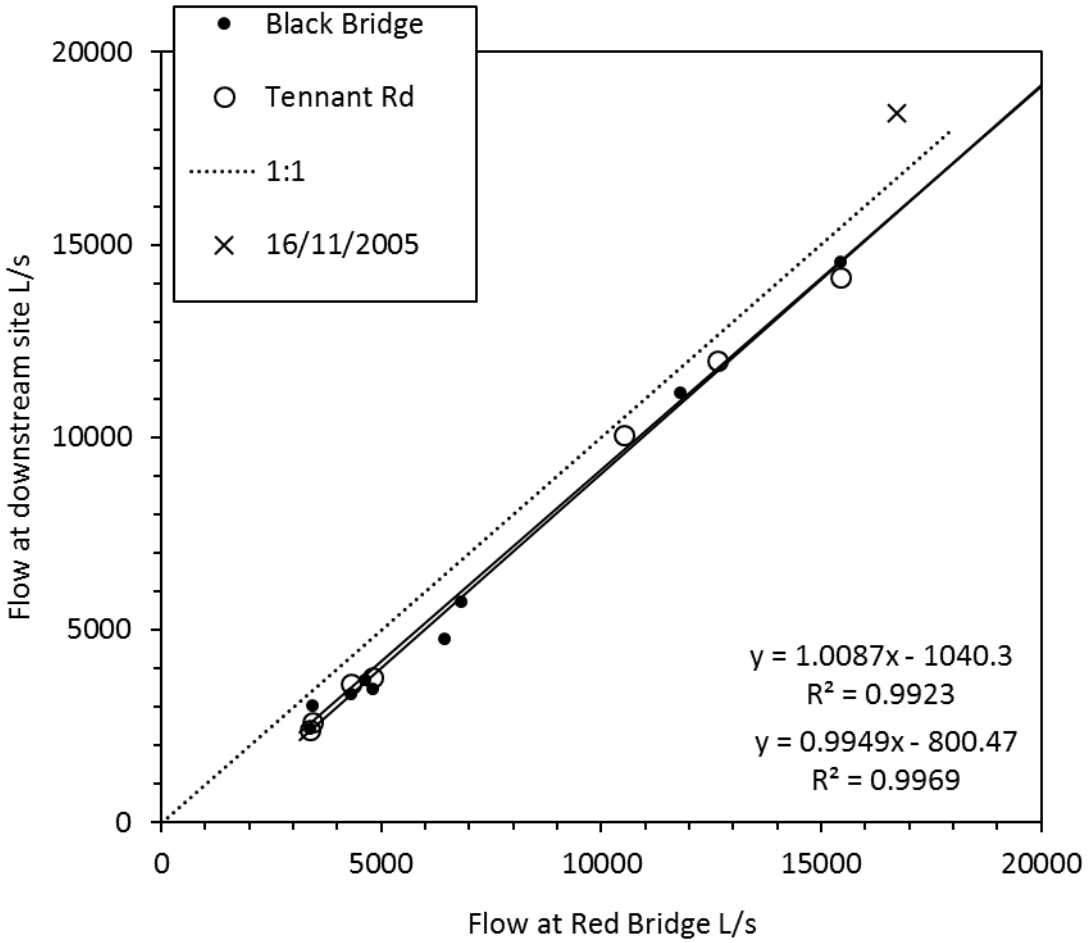


Figure 3-8: Tukituki River flow loss. Tukituki River flows near the river mouth (Black Bridge and Tennant Rd) are plotted against same day gaugings at the upstream monitoring site (Red Bridge). Points below the dotted line have lost flow, relative to Red Bridge. The median flow loss was 910 L/s. The gauging on 16 November 2005 (marked X) was not used for trend line estimation because of variable flows on the day. See Figure 3-9 for a map of gauging locations.

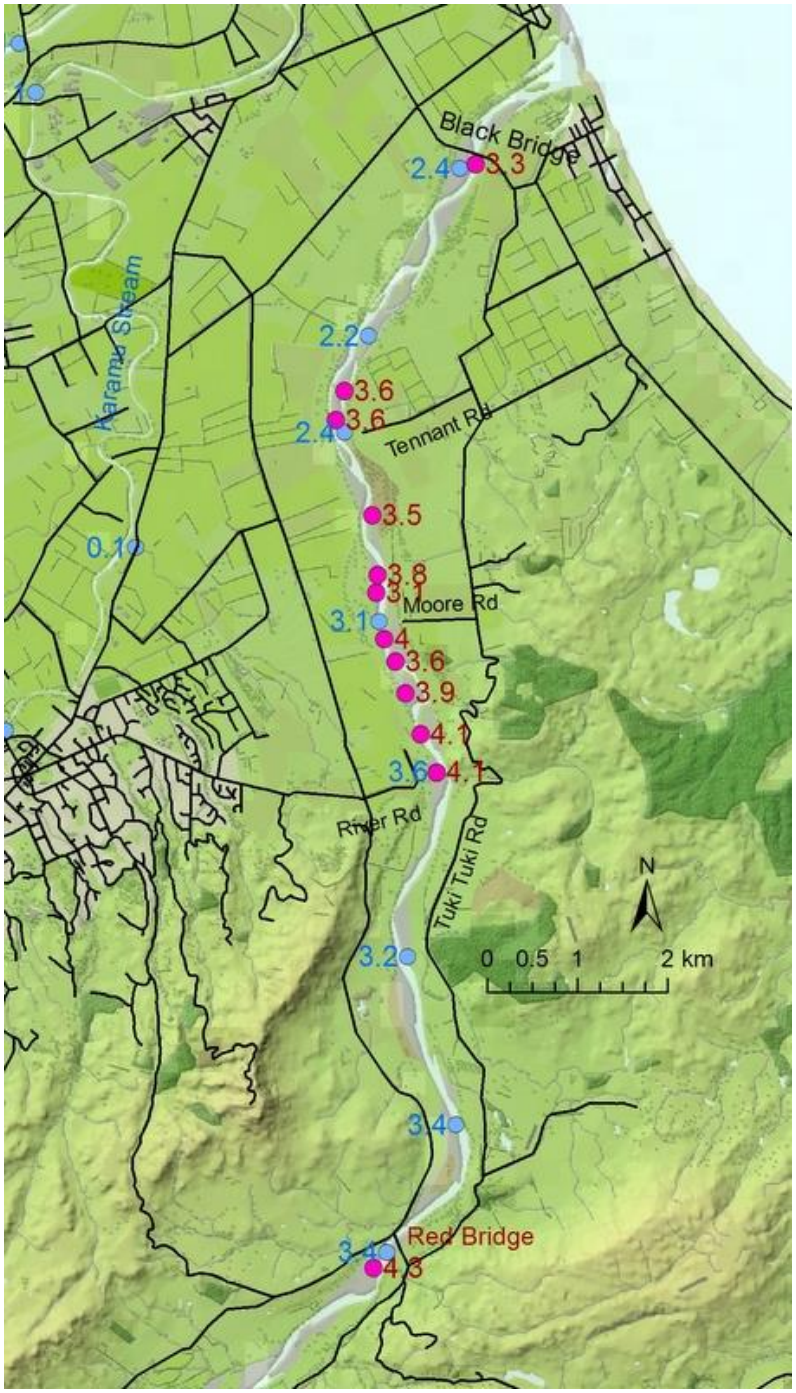


Figure 3-9: Lower Tukituki gaugings. Concurrent gaugings along the lower Tukituki River on 27 March 2013 (blue dots, left labels) and 5 February 2015 (red dots, right labels). Flows are labelled in m^3/s (multiply by 1000 to get L/s).

The loss of flow from the lower Tukituki River makes it a potential source of the groundwater feeding springs in neighbouring catchments (e.g. Karamu, Section 3.8; Mangateretere, Section 3.9). Other potential sources of this spring water include the Ngaruroro River, adjacent hill-country and direct rainfall recharge. Distinguishing Tukituki water from the other sources was investigated using natural tracers, including electrical conductance and stable isotopes.

The lower electrical conductance in Tukituki water can be used to distinguish it from water originating from adjacent low-elevation hill country, where the predominance of limestone and other marine sedimentary rock increases the concentration of dissolved ions in water. For example, electrical conductance in the Tukituki River (188-219 µS/cm inter-quartile range, n = 94, Red Bridge) was lower than streams fed by hill-country (e.g. Awanui at flume interquartile range 670-771 µS/cm, n=15).

Electrical conductance also provides a contrast between the Tukituki River (188-219 µS/cm inter-quartile range) and the Ngaruroro River, which is less conductive (135-160 µS/cm inter-quartile range, n = 50, Fernhill). A low value for electrical conductance should be a reliable indicator of Ngaruroro sourced water. However, higher conductance values are not a reliable indicator of Tukituki water because small contributions from limestone areas can overwhelm the margin of difference (e.g. Ngaruroro water of 160 µS/cm mixed with 10% water at 600 µS/cm produces water at 200 µS/cm).

Stable isotopes were therefore used to distinguish Tukituki from Ngaruroro water. Using data reported by Morgenstern *et al.* (2018), the contrast in $\delta^{18}O$ between the Tukituki and Ngaruroro was sufficient to reliably differentiate the two sources, with a mean difference of -1.1‰ between 17 paired samples of river water (Section 3.1, Figure 3-5). To maximise the generality of isotope results from river samples collected 2009-2010, the relationship between $\delta^{18}O$ and ocean temperature was used to predict the mean $\delta^{18}O$ for a longer period (described in Section 3.1). This predicted a long-term mean $\delta^{18}O$ of -6.8‰ for the Tukituki River at Red Bridge (-6.83‰ for 2013 to 2017, -6.85 for 2008-2017, -6.85‰ for 1998 to 2017, using July to June water years).

Distinguishing Tukituki water from direct rainfall recharge on the plains is more difficult (Figure 3-7). In terms of stable isotopes, the long-term mean $\delta^{18}O$ of -6.8‰ for the Tukituki River at Red Bridge differs from the interim estimate for rainfall recharge on the Heretaunga Plains ($\delta^{18}O$ of -6.0‰, Section 2.1.5). However, the groundwater samples collected from a well adjacent to the Tukituki River were intermediate between Tukituki and rainfall recharge estimates. Groundwater sampled from a shallow well located just 0.7 km from the Tukituki losing reach (well 2580, screened at 9-12 m deep, Figure 3-24) provided a mean $\delta^{18}O$ of -6.4‰ (-6.3‰ on 6/5/2016 and -6.5‰ on 23/8/2017), (Figure 3-5).

There are several possible explanations for the less negative groundwater $\delta^{18}O$ of -6.4‰ (well 2580) compared to the long-term mean for river water of -6.8‰. Groundwater signatures vary over time and space, albeit less than surface water, and it is possible that more than two samples are needed to characterise the long-term average for Tukituki sourced groundwater.

It is also possible that river recharge is biased to warmer periods, such as when low groundwater levels lengthen the river's losing reach. Achieving a mean $\delta^{18}O$ of -6.4‰ through seasonal bias alone would require little or no flow losses from the Tukituki in winter or spring.

Another possibility is that local recharge from rainfall has reduced the $\delta^{18}O$ of groundwater, as proposed by Morgenstern *et al.* (2018). With estimates for rainfall recharge of approximately -6.0‰ (see Section 2.1.5), this would require dilution approaching 50% to bring river values of -6.8‰ up to the observed $\delta^{18}O$ for groundwater of -6.4‰. Considering the large volume of river water being lost (median 910 L/s), this magnitude of dilution is not supported by the low rainfall in this area. For example, 1400 to 2900 mm/year of rainfall recharge would be required to achieve 50% dilution (for a recharge area of 5 km^2 to 10 km^2 respectively), compared to a measured rainfall recharge of 260 mm/year (mean 2013 to 2017 for Bridge Pa, Maraekakaho and substation lysimeters). Using the same lysimeter data, annual rainfall recharge is estimated to contribute 4%-8% of inflow to this aquifer (assuming a recharge area of 5 km^2 to 10 km^2 respectively).

In summary, the Tukituki River water has a long-term mean $\delta^{18}O$ of -6.8‰. It is reasonable to assume that there is a change in the isotopic signature during the transfer of river water ($\delta^{18}O$ of -6.8‰) to groundwater ($\delta^{18}O$ of -6.4‰). A combination of groundwater sampling bias and seasonal bias in river recharge probably contributes to this transfer function, with a minor contribution from local rainfall recharge. Additional sampling is recommended to better characterise the Tukituki sourced groundwater. For further discussion on the potential sources of spring water in this area, see Section 3.8 (Karamu) and Section 3.9 (Mangateretere).

The loss of water from the Lower Tukituki described for this report is in addition to important gains and losses that occur further upstream. Most notably, as the Tukituki crosses the Ruataniwha Plains, the river water interacts with the Ruataniwha aquifer through losing and gaining sections (Johnson, 2011b; Wilding & Waldron, 2012). There are additional losses and gains from the Tukituki between the Ruataniwha Plains and the Heretaunga Plains. Wilding and Waldron (2012) suggested a loss of water from the Waipawa River (upstream of the Tukituki confluence) sustains springs in the Papanui Stream. The Papanui also gains 400 L/s to 500 L/s of flow downstream of Middle Road (e.g. gain of 443 L/s on 27/1/2015), as it crosses the alluvial valley upstream of the Tukituki confluence. This gain coincides with a loss from the Tukituki River of approximately 500 L/s over the reach within 2.5 km upstream of the Papanui confluence (from gaugings 27/1/2015). The Tukituki River is elevated higher than the Papanui through this reach (from 2003 LiDAR data), providing the necessary pressure-head for flow loss and gain.

No recent changes to the alignment of the lower Tukituki River were noted, with Wakefield *et al.* (2012) indicating the river mouth had changed little since the mid 1800's. The 1867 flood that caused the Ngaruroro to change course, also changed the course of the Waipawa just upstream of the Tukituki confluence (Engineering File 2991). This returned the Waipawa to its palaeochannel down what is now the Papanui Stream to re-join the Tukituki downstream of Middle Road. By 1887, local landowners had managed to divert the Waipawa back to the pre-1867 channel, enabling drainage works to continue in the Papanui catchment (i.e. draining the Roto-a-Tara lake/wetland complex).

3.3 Tutaekuri

This report focusses on the section of the Tutaekuri River downstream of the Mangaone confluence. The Tutaekuri River loses flow to groundwater, the majority of which appears to feed tributaries of the Tutaekuri-Waimate Stream. More recent calculations support a flow loss from the Tutaekuri of approximately 820 L/s (interquartile range 580 to 890 L/s, n=34 at median Puketapu flow of 3980 L/s), (Figure 3-10). Data from the Hakowai gauging site were supplemented using flows estimated by summing flows from the Mangaone River (Dartmoor site or the site upstream of the Tutaekuri confluence) with flows from the Tutaekuri River upstream of the Mangaone confluence (Figure 3-11).

The location of the loss was investigated using concurrent gaugings between the Mangaone confluence and Puketapu. These confirm the loss occurred between Hakowai and Silverford (Figure 3-11), with flow conserved between Silverford and Puketapu (from 8 concurrent gaugings at a median flow of 7560 L/s at Puketapu). This location conflicts with Dravid and Brown (1997), who indicated the loss occurred downstream of Silverford (their Table 6.4 contains flows labelled as Silverford, which do not match same day gaugings archived for Silverford in the Hilltop *allsites* database).

Dravid and Brown (1997) noted an absence of flow losses downstream of Puketapu, which was confirmed by more recent concurrent gaugings. Only 5 out of 30 concurrent gaugings measured a flow loss between Puketapu and the most downstream monitoring site (Brookfields Road bridge; Figure 3-4).

The loss of flow from the Tutaekuri River makes it a potential source of the groundwater feeding springs in neighbouring catchments (e.g. Tutaekuri-Waimate Stream, Section 3.13). Stable isotopes can be used to trace groundwater originating from the Tutaekuri, provided it has an isotope signature that is distinct from other sources. In an effort to characterise the isotope signature of the Tutaekuri River, stable isotopes were sampled from the Tutaekuri River (Puketapu) both in summer and winter, providing a mean $\delta^{18}O$ of -7.2‰ (-7.0‰ on 4/3/2015 and -7.4‰ on 23/8/2017). In addition to river water, isotopes were also sampled from a shallow well located less than 0.5 km from the river (Figure 3-11) in an effort to provide a temporally-integrated signature of groundwater originating from the Tutaekuri (well 15884, screened 9.4-10.4 m, Figure 3-11). Unfortunately, this well provided a poor match with the major ion chemistry of the Tutaekuri River. For example, electrical conductance in this well was 660 µS/cm (23/8/2017), compared to 300 µS/cm for the Tutaekuri River (median of 383 measurements at Puketapu and Hawke's Bay Expressway). Therefore, the less-negative isotope values from this well ($\delta^{18}O$ of -7.0‰ on 9/6/2016 and -6.7‰ on 23/8/2017) cannot be treated as representative of groundwater sourced from the Tutaekuri River. The mean $\delta^{18}O$ of -7.2‰ from river samples were therefore used as a preliminary signature for Tutaekuri sourced groundwater, placing it between the Ngaruroro and Tukituki sourced groundwater ($\delta^{18}O$ of -7.6‰ and -6.4‰ respectively). A similar $\delta^{18}O$ of -7.2‰ was obtained using the correlation between sea surface temperatures and the two river samples to predict the average $\delta^{18}O$ for the period July 2012 to June 2017 (mean slope of 0.194 from the Tukituki and Ngaruroro used, from Figure 3-6, then fitting an intercept of -10.85 to the Tutaekuri river-samples). Obtaining a more precise isotope signature for the Tutaekuri River will require additional sampling, such as samples from a more representative well, or time-series of river samples.

The Tutaekuri River has changed its course downstream of Puketapu. Prior to the 1931 earthquake, the Tutaekuri River flowed to the Ahuriri Estuary, dividing Napier from Taradale (Dravid & Brown, 1997). The Tutaekuri followed various flowpaths, including Riverbend Rd, Wellesley Rd and most recently following Willowbank Avenue and Georges Drive, with the historic flowpath drawn on Figure 3-4 (labelled "pre-1931 Tutaekuri"). Prior to the earthquake, various proposals for diverting the Tutaekuri River mouth from the Ahuriri Estuary to the Waitangi Estuary were rejected. However, the earthquake lifted the land sufficiently to compromise drainage of Tutaekuri flood waters to the Ahuriri Estuary, and the need for a diversion became "obvious to all" (Williams, 1987). The present diversion, starting approximately 1 kilometre downstream of the Hawke's Bay Expressway, was constructed between 1934 and 1940 (Williams, 1987). The Tutaekuri now shares a common river mouth to the ocean with the Ngaruroro River and Karamu Stream. Older paths of the Tutaekuri River are discussed in Section 3.13.

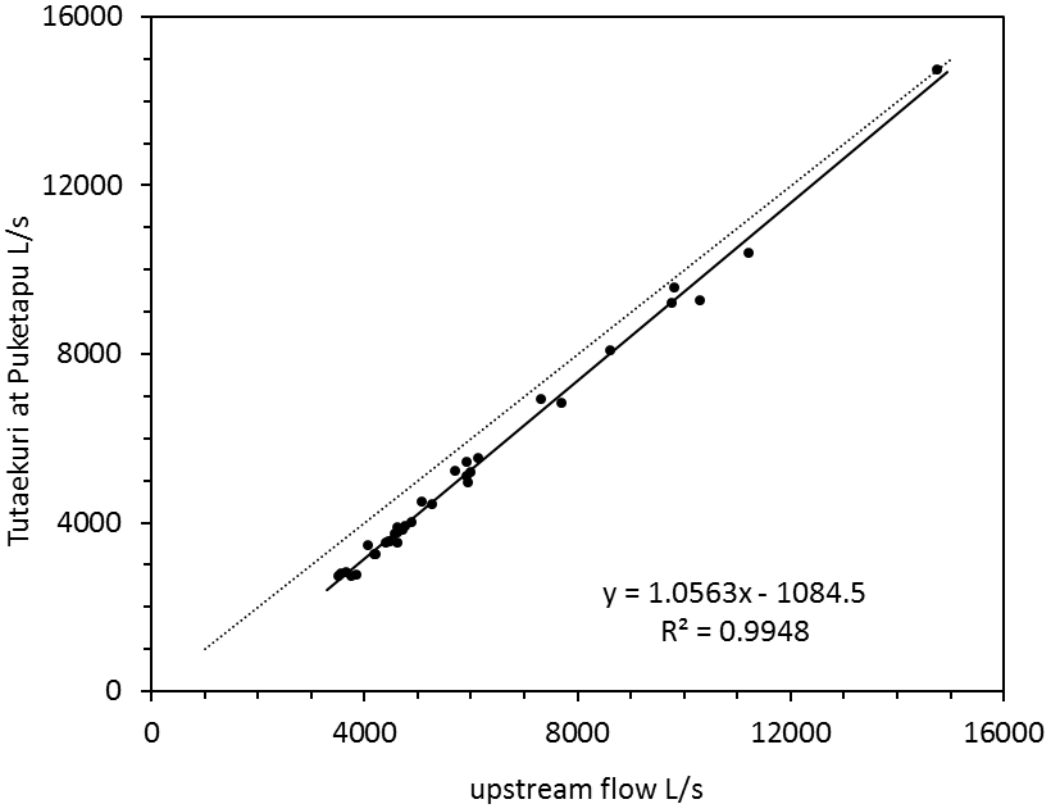


Figure 3-10: Tutaekuri River flow loss. Tutaekuri River flows gauged at Puketapu are plotted against same day gaugings at upstream sites (at Hakowai, or a sum of the upper Tutaekuri and Mangaone). The dotted line represents a condition of flow conservation between the sites. Points below the dotted line have lost flow, relative to the upstream sites. The median flow loss was 820 L/s. See Figure 3-11 for a map of named locations.

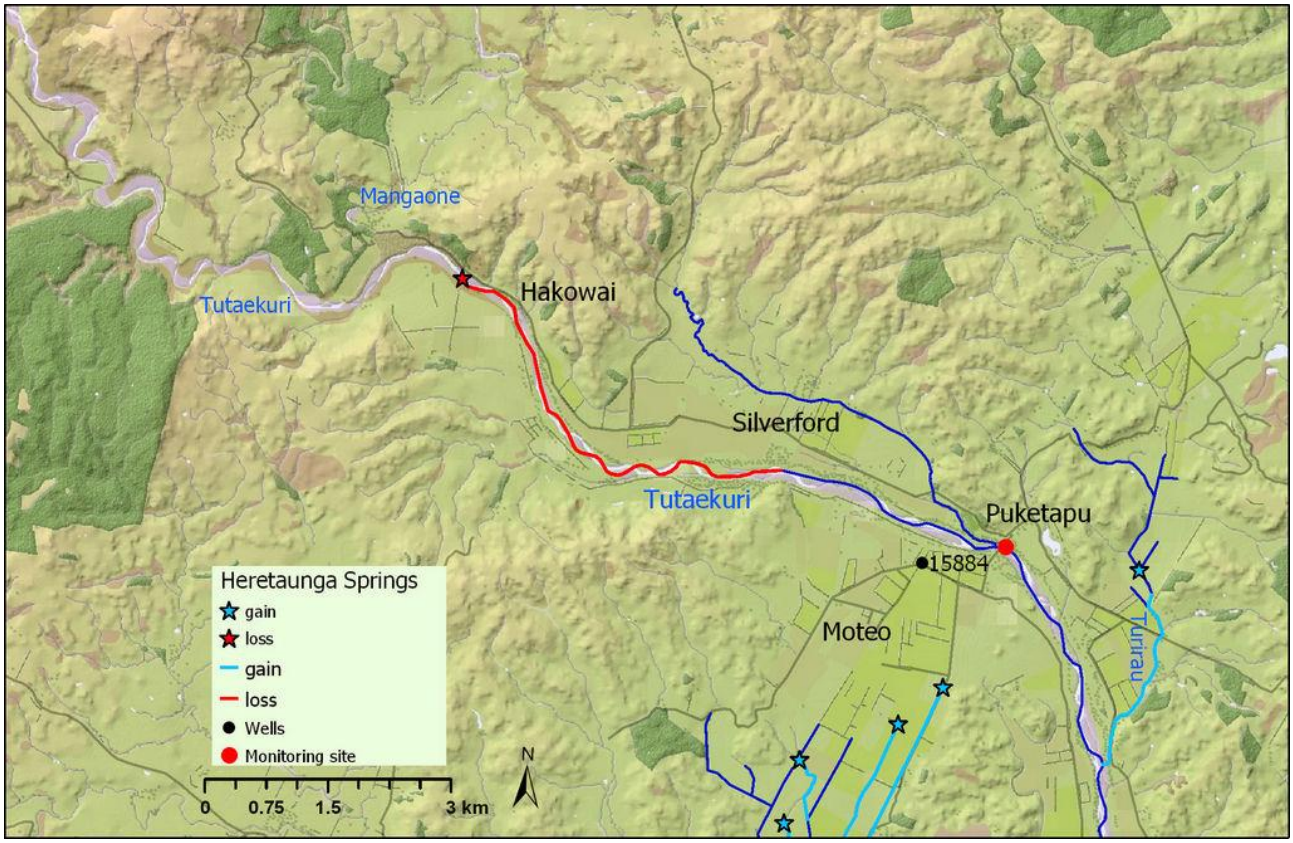


Figure 3-11: Tutaekuri River. The median flow loss from the Tutaekuri River was 820 L/s over the losing reach (approximated as a red line).

3.4 Paritua-Karewarewa

The Paritua Stream drains limestone hill-country (sandstone, limestone and mudstone of marine origin) before flowing across the Heretaunga Plains. Springs originating in the limestone hill-country were not surveyed for this report, beyond mapping the stream extent from aerial photographs.

Downstream of the Ongaru confluence (gauged at Washpool Station bridge), the stream loses flow to groundwater as it crosses the Heretaunga Plains (Waldron *et al.*, 2007). These losses occur where the stream flows across unconfined gravels that are perched several metres above the water table (Rabbitte, 2009). The start of this losing section coincides with the stream cutting through a layer of cemented pan that sits atop coarse gravels (Figure 3-12). At times, the stream runs dry before reaching Bridge Pa. Downstream of Bridge Pa, the stream name changes to Karewarewa, and the Karewarewa gains flow from diffuse springs (Waldron *et al.*, 2007).

The flow loss and spring gains were investigated further for this report. Additional concurrent gaugings confirmed the Paritua starts losing downstream of Washpool Station bridge and that this losing reach can extend more than 1 km downstream of Raukawa Road (7.5 km losing reach, Figure 3.4). There were insufficient gaugings between Raukawa Road and Rosser Road to characterise total flow losses over the losing reach. The site with the lowest flow (either Raukawa Road, or Rosser Road) was therefore used in calculations of total flow loss. This indicated that drying was most likely to occur when flows were less than 130 L/s at the start of the losing reach (Paritua at Washpool bridge, Washpool woolshed, or weir 1, in order of preference).

Hughes (2009a) considered the rate of loss to be slow for the losing reach, concluding the streambed sediments had low permeability. Given the stream is perched above the water table for much of the losing

reach, the rate of loss would be primarily determined by streambed permeability and stream depth (Hughes, 2009a; Rabbitte, 2009). A coating of tufa (Figure 3-12) weakly cements the cobbles and gravels together (revealed by fizzing when sulphuric acid was applied to a dry tufa surface). There was very little iron or phosphate in the tufa, which was composed mainly of calcite (from X-Ray Diffraction and X-Ray Fluorescence analyses by D. Hartland, University of Waikato), (G. Upchurch memo, January 2016). This tufa coating is the cause of the low permeability of the streambed.

Tufa forms when high concentrations of calcium carbonate precipitate out of solution (Ford & Pedley, 1996). The Paritua contains high concentrations of calcium (median 76 mg/L of calcium, 270 mg/L of bicarbonate at a pH of 8.2, Paritua at Raukawa Rd, n = 8). The limestone, and other marine sediments, that dominate the geology of this catchment provide ample sources of calcium (Griffiths, 2001), and the low rainfall reduces dilution (median 690 mm/year, Bridge Pa 1998-2016). Biofilms play an important role in tufa deposition (Pedley *et al.*, 2009), and the high abundance of aquatic plants in the Paritua would further promote the formation of tufa by consuming dissolved carbon for photosynthesis (de Montety *et al.*, 2011).

The tufa can reduce the permeability of the streambed. But this layer can be broken or bypassed, as demonstrated in 1998 during an attempt to divert the stream (Waldron *et al.*, 2007). The stream flow (in the order of 200 L/s) dried out completely within a short distance after diverting the stream into a newly excavated channel where the tufa layer had not yet had time to form (consent number WP980372D). Fine sediment was applied in an attempt to seal the new streambed.

Figure 3-12: Paritua losing section. The start of the flow loss coincides with stream level dropping below a cemented pan, visible in the upper photograph as a mossy layer, between the soil (grassed) and cobble layer (upper photograph taken at Paritua Vineyards, 13/8/2014). The Paritua Stream runs dry at times, with an excavated hole (presumably for stock water) revealing the shallow water table downstream of Raukawa Road (lower left, 5/5/2015). A layer of tufa cements the surface layer of gravel and cobble together, reducing the permeability of the streambed. Tufa was clearly visible on the upper surface of this partially exposed bottle (lower right).

The gaining section of the Paritua-Karewarewa was also investigated for this report. The start of this gain reveals the transition from being perched above the water table, to the streambed lying below groundwater level (Lough, 2014). Two additional concurrent gaugings (i.e. same day measurements) were undertaken, with closer site-spacing downstream of Bridge Pa. These gaugings indicated that the gaining section started downstream of Raukawa Road (Figure 3-13). The transition from dry channel to wet was visible from aerial photographs (e.g. Kiwi Image, Google Earth), which indicated a similar location for the start of the gaining reach during other years when the stream ran dry. An excavated hole revealed the shallow water table near the end of the losing reach (Figure 3-12), at a time when groundwater was at median levels (51%ile on 5/5/2015 at Bridge Pa well 1003). Flow gains would start further upstream when groundwater levels are higher. The magnitude of the flow increase over the gaining section was approximately 100 L/s on

12/11/2014, decreasing to 54 L/s by 5/5/2015. Given the estimated MALF was zero at Raukawa Road (i.e. dry), the estimated MALF of 25 L/s at Pakipaki could be seen as the typical magnitude of gain for the Karewarewa under low flow conditions. More concurrent gaugings are required to better characterise this gain and how it varies over time.

The start of the gaining section coincided with the edge of the confining layer, and with a change from unconfined gravels to a shallow layer of Taupo pumice sand and ash (Figure 3-14), as described by Grant (1996) and Griffiths (2001). From well records, the layer of the pumice sand is typically 4 m thick (e.g. well number 8512, 668, 10522), and this layer can be seen in eroded stream banks (Figure 3-15). Separating this layer of pumice sand from the deeper gravel-aquifer is a layer of confining marine clays that thickens southeast of Bridge Pa (Dravid & Brown, 1997). Downstream of Bridge Pa is therefore where this layer of pumice sand could provide a conduit for groundwater moving laterally to streams (e.g. Karewarewa, Awanui, Louisa), together with nutrients leached from overlying land use.

Potential sources of water recharge to the pumice sand layer include rainfall, flow loss from the Paritua-Karewarewa, and leakage/overflow from the deeper gravel aquifer. The shallow groundwater in this pumice layer has not been described, with no measurements of water level and water quality. The water is not used for irrigation, so it is not a productive aquifer. Water from the Heretaunga gravel aquifer could reach the Karewarewa via the pumice layer. However, the high electrical conductance of the Karewarewa is not consistent with a Ngaruroro source (Karewarewa inter-quartile range of 645-861 µS/cm, n=30; compared to the Ngaruroro 135-160 µS/cm, n=50). Additionally, stable isotopes from the Karewarewa, that were sampled while the Paritua ran dry (3/3/2015), did not match Ngaruroro sourced water ($\delta^{18}O$ -5.8 ‰ at Pakipaki, compared to -7.6 ‰ for Ngaruroro sourced groundwater, Section 3.1). In contrast, this sample from the Karewarewa had the least negative $\delta^{18}O$ level of any of the streams sampled on the Heretaunga Plains, and provided a closer match with streams sourced from warmer, lowland catchments (e.g. -6.0 ‰ for Paritua at waterwheel, -5.9 ‰ for Kaikora upstream of Papanui). These catchments are fed from rainfall recharge at low elevations, and so are expected to approximate the isotopic signature of local rainfall recharge (Section 2.1.5). Therefore the groundwater feeding the Karewarewa could be recharged by these streams or by direct rainfall recharge on the plains.

The Karewarewa has consistently higher nitrogen concentrations than the other Karamu tributaries (Haidekker, 2016). Given the more intense land use on the plains, compared to surrounding hill country, the elevated nutrients increase the likelihood that groundwater feeding the Karewarewa originated from local rainfall recharge. Further investigations would be required to understand the flow contributions from the pumice sand layer to the Louisa, upper Awanui and lower Awanui, given these are fed by perennial tributaries that arise within the same area of pumice sand (Figure 3-14).

Changes to the channel alignment of the Paritua-Karewarewa include the avulsion of Paritua flows into the Karewarewa. This avulsion resulted from the 1931 Earthquake, which raised the bed of the original channel that flowed toward the Irongate. Wetlands were drained and streams realigned through the lower Ongaru, the Paritua valley, Turamoe wetland and the Karewarewa Stream. One landowner was concerned that groundwater levels beneath their paddocks have decreased because of the channel straightening. This could have been exacerbated by incision of the channel into the soft pumice sands (HBRC, 2004). Flow from the Paritua was supplemented by releases from Te Tua Reservoir up until 2004 (Waldron *et al.*, 2007). Some of these flow-releases were at the request of the regional council, to improve instream conditions. After those releases stopped, drying of the Paritua became more frequent. People from local marae recalled that, historically, the springs arose upstream of Bridge Pa. This is a reasonable assertion, given that any lowering of the water table around Bridge Pa (e.g. by groundwater use or reduced recharge) would have moved the start point of the gaining section further downstream to its present location. The magnitude of spring flow could also have changed in response to the change in the Ngaruroro flowpath. In 1867, a flood on the

Ngaruroro River changed the channel location from flowing down what is now the Irongate and Karamu to its present path. As well as moving the source of river recharge further away from the Paritua, this may also have changed the magnitude of flow loss from the Ngaruroro (from changes to the zone of unconfined gravels traversed by the river). It is possible this changed the magnitude of flow loss and gain for the Paritua and Karewarewa.

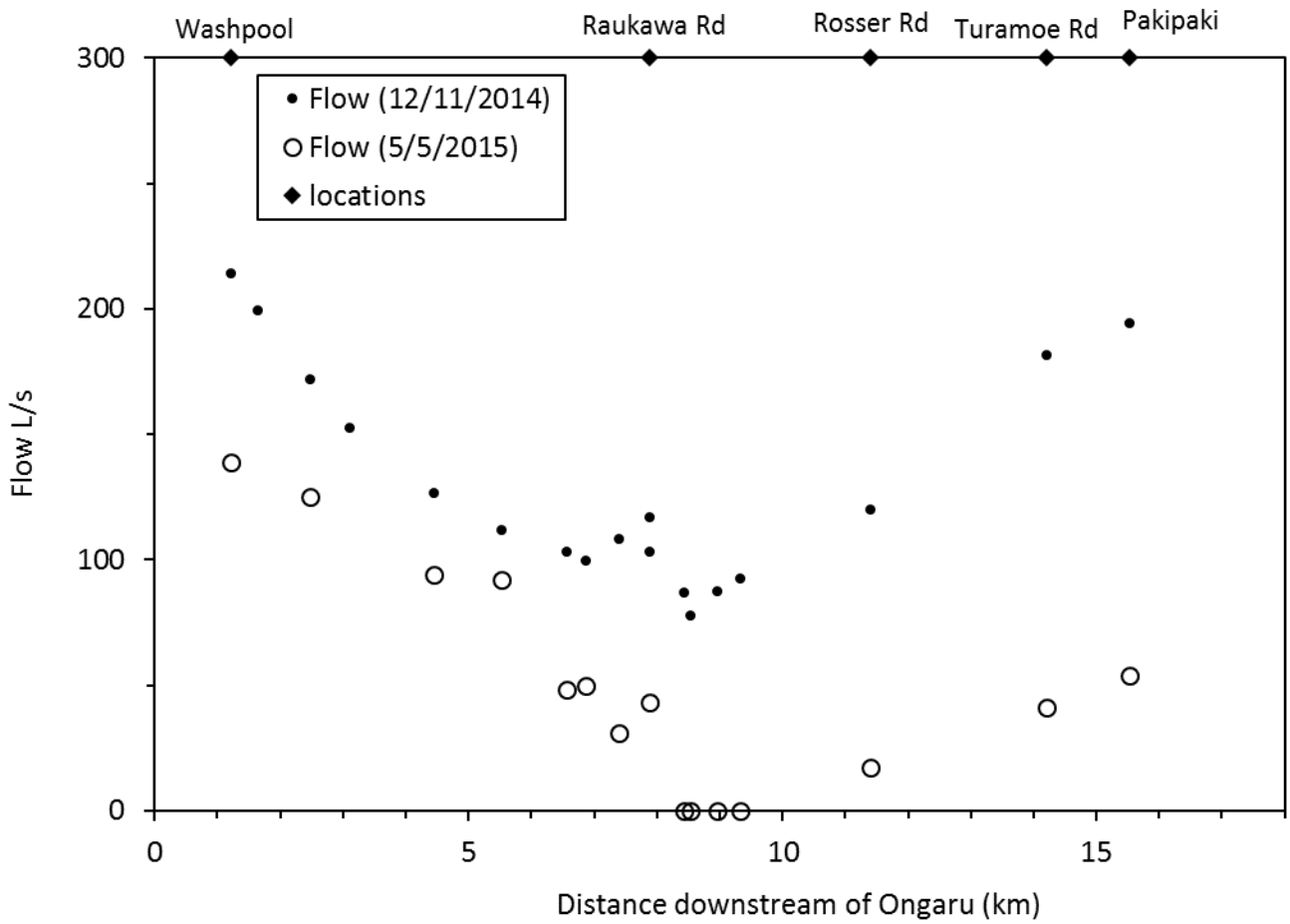


Figure 3-13: Paritua Karewarewa flow loss and gain. The Paritua loses flow as it flows across the Heretaunga Plains, before gaining flow downstream of Bridge Pa (Raukawa Road). Flow measurements were made along the length of the Paritua and Karewarewa stream on two occasions (12/11/2014 and 5/5/2015). Measurement locations are plotted as the distance downstream of the Ongaru confluence, with landmarks named at the top of the plot.

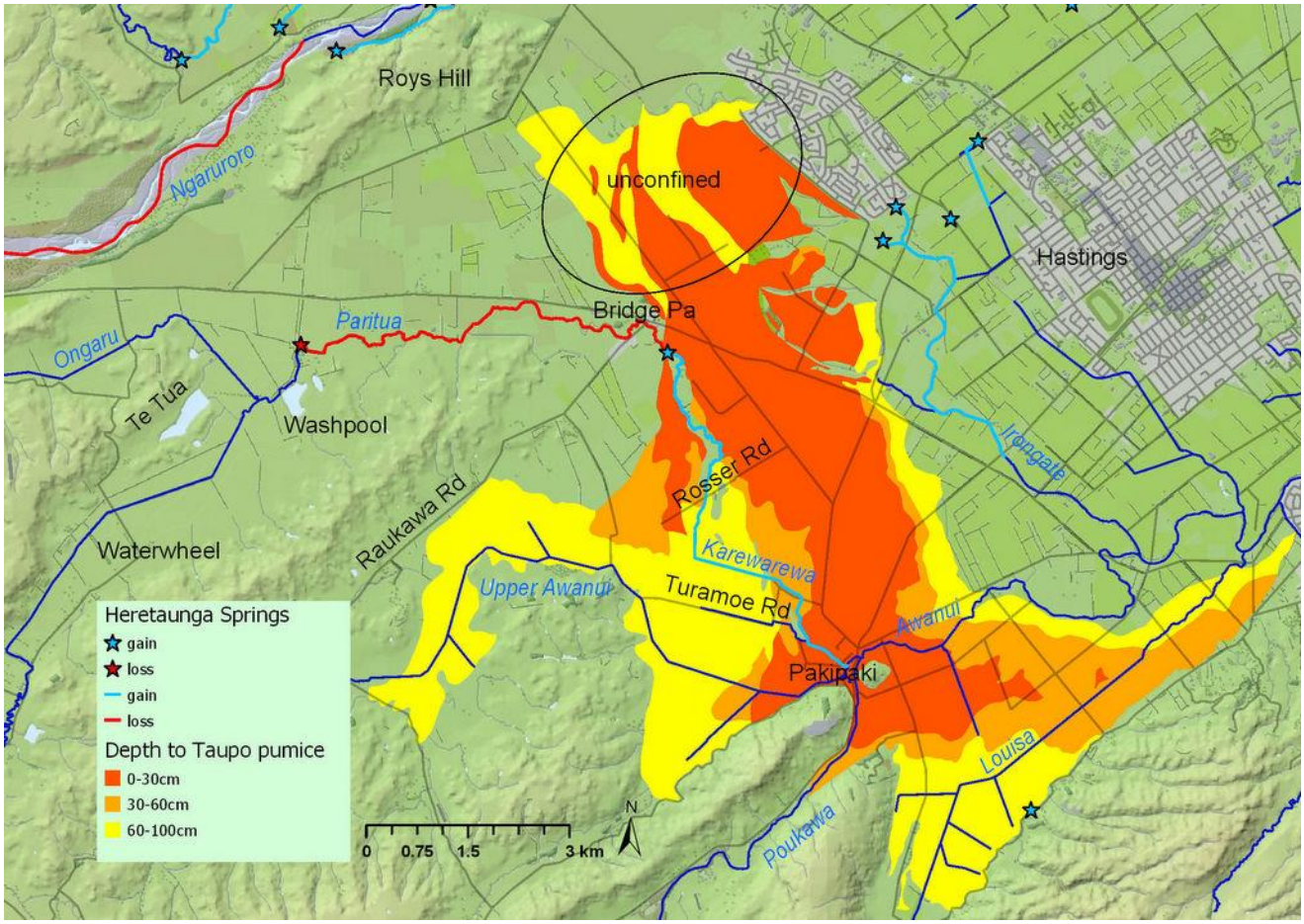


Figure 3-14: Taupo pumice sand layer. The soil depth to the top of a layer of Taupo pumice/ash sand is represented by coloured polygons (darker orange for shallower soil overlaying pumice). The polygons and soil-depths are from Griffiths (2001). Marine clays separate the pumice layer from deeper gravels, with the exception of the area north of Bridge Pa (labelled "unconfined").

Figure 3-15: Pumice layers by Karewarewa Stream. This report discusses the potential for this layer of shallow pumice sand to convey groundwater to the Karewarewa stream. Photo upstream of Turamoe Road (NZTM E1924177 N5600548, 6/9/2016). This 2.5 m high bank was exposed by slumping.

3.5 Awanui

For this report, the Awanui is divided between the Upper Awanui and the Lower Awanui (Figure 3-16). The Lower Awanui starts at the confluence with the Poukawa and the Karewarewa (Section 3.1). The Lower Awanui becomes the Karamu Stream (Section 3.8) at its confluence with the Irongate Stream (Section 3.7).

Combining the three sub-catchments, the Lower Awanui drains a large area of peat wetlands and limestone hill-country (limestone, mudstone and sandstone of marine origin). Concentrations of calcium bicarbonate in the Awanui are approximately 4 times the concentration in the Raupare, reflecting the Awanui's limestone geology. Electrical conductance was higher than streams fed by the Heretaunga aquifer (Awanui flume interquartile range 670-771 µS/cm, n=15; Raupare at Ormond Rd 179-226 µS/cm, n=40). Among the sub-catchments, the Poukawa includes two large peat-wetlands (Lake Poukawa and Pekapeka Swamp). Other wetlands in the Upper Awanui and Paritua catchments were drained, including Turamoe Wetland. The

influence of peat from Poukawa and Pekapeka wetlands increases the organic carbon concentrations to a median of 13 g/m^3 for the Awanui, compared to 0.7 g/m^3 for the Raupare (from 7 pairs of same-day measurements).

The surface catchment of the Lower Awanui is 17 times larger than the Raupare, producing larger flood flows. But the MALF is less than a third of the Raupare, reflecting less groundwater input from the Heretaunga Aquifer. Gains and losses of flow to groundwater were only investigated in detail for the Paritua-Karewarewa (see Section 3.1), which loses all its flow at times to groundwater, upstream of groundwater inflows to the Karewarewa. For the remainder of the catchment, the springs were only described by the extent of the stream network, which was delineated from aerial photographs.

A detailed account of springs in the Poukawa catchment is provided by Cameron *et al.* (2011). They described three types of springs. The first is equivalent to point springs, as described in this report, arising from limestone fractures, sometimes under pressure. Two types of diffuse springs were also identified, including those that start flowing where the water table intersects the streambed, plus those that occur where a buried limestone fracture seeps to the surface. The Poukawa Stream also experiences a significant loss of flow to evaporation from Lake Poukawa and Pekapeka wetland. Evaporation rates measured near the northern end of Lake Poukawa were reported by O'Shaugnessy (1988). For example, approximately 100 L/s could evaporate in January when evaporation rates are highest (194 mm/month), dropping to 70 L/s in March. This represents a large proportion of the summer median flow (monthly median flow for Awanui flume is 100-150 L/s for March to January respectively). The median flow at Awanui flume exceeds 1000 L/s in July, hence evaporation is a minor component of the winter water balance. Lower temperatures in July reduce evaporation to less than 50 L/s, despite the doubling in wetted area as the wetland reclaims grazed paddocks.

The extensive layer of pumice sands (Figure 3-14) coincides with the start of several tributaries in the Awanui catchment. Investigations are recommended to improve our understanding of this shallow groundwater and its interaction with flows of the Awanui and its tributaries (see Section 3.4).

The drainage pattern of the Awanui and its tributaries has been extensively modified. In addition to draining extensive wetlands (e.g. Paritua valley, Turamoe, Poukawa), the location of tributary confluences has changed. The 1931 earthquake changed the drainage pattern by raising ground levels, so that the Paritua no longer flowed toward the Irongate Stream (from Bridge Pa). The Paritua now flows down the Karewarewa into the Awanui. The Poukawa Stream has been diverted so that its confluence with the Awanui is located about a kilometre further upstream (near Pakipaki). Te Waikaha Stream was diverted out of the Awanui in the 1960s (original confluence upstream of Crystal Rd, see “Old Te Waikaha” in Figure 3-17), to reduce flooding in the Awanui (HBRC, 2004). Te Waikaha Stream now flows down the Louisa Stream (Section 3.6). The Lower Awanui has been eroding down into the pumice sands, with a drop in bed level of 0.5 m observed between 1982 and 1993 (HBRC, 2004). Active down-cutting was observed further upstream in the Upper Awanui, with head cuts also propagating into small tributaries through the soft pumice sands (at or around NZTM E1922344 N5599606, 31/8/2017).

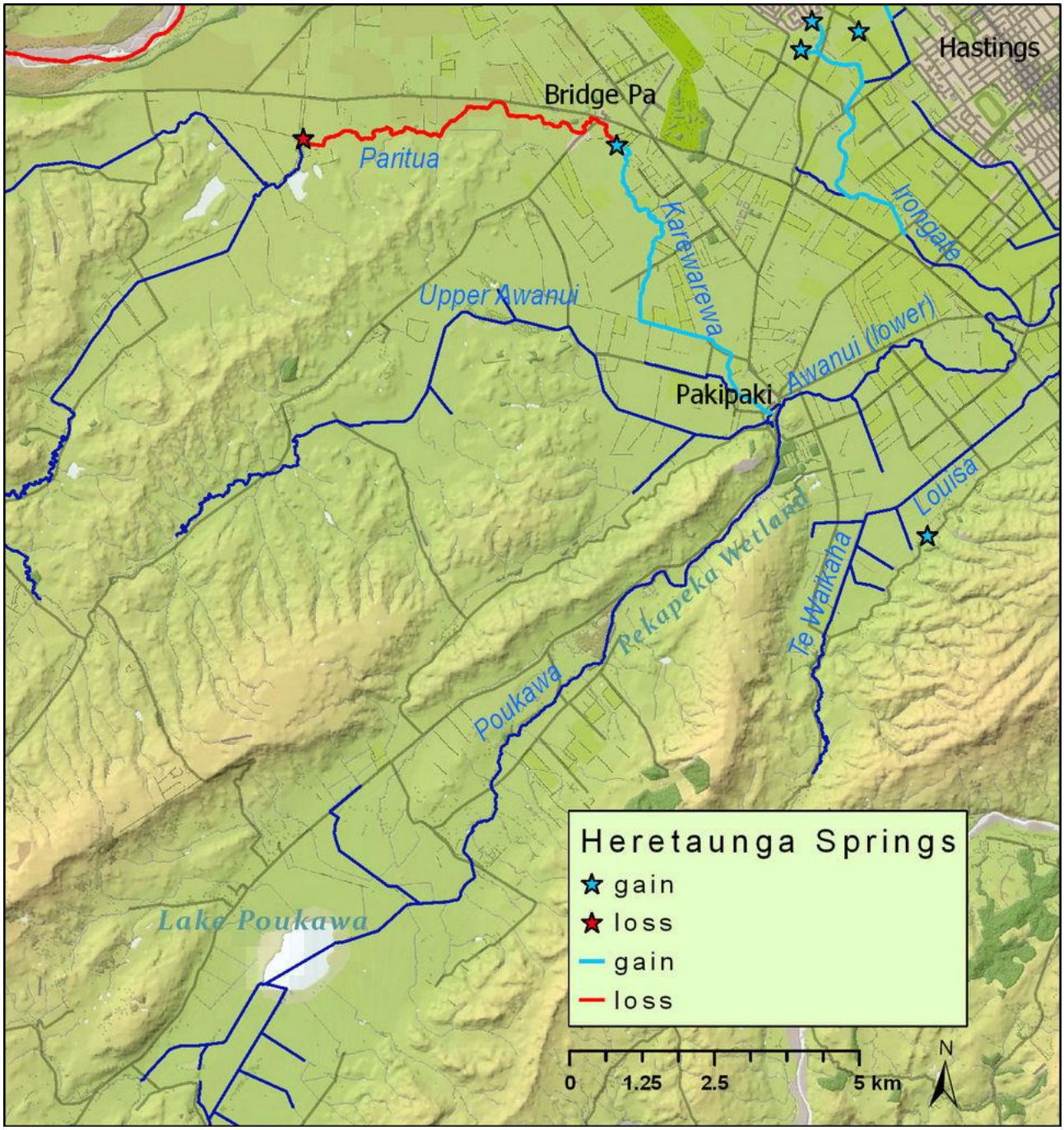


Figure 3-16: Awanui catchment. Map of the Awanui catchment, showing the Poukawa, Paritua-Karewarewa and the Upper Awanui as major tributaries of the Lower Awanui.

3.6 Louisa and Te Waikaha

The Te Waikaha Stream drains a narrow catchment of limestone hill country (limestone, mudstone and sandstone of marine origin), including the western slope of the highest peak in the Karamu catchment (Mt Erin, 490 m AMSL). Wedged between the Poukawa and Tukituki catchments (Figure 3-17), Te Waikaha Stream runs the length of an active fault line called the Tukituki Fault (Lee *et al.*, 2014). At Mutiny Road bridge, the valley opens out onto the Heretaunga Plains, draining peat soils and pumice sand (Griffiths, 2001). The Te

Waikaha was diverted into the Louisa Stream in the 1960s to alleviate flooding from the Awanui (HBRC, 2004). Hence, the stream name changes from Te Waikaha to Louisa at the diversion point. Peat soils and actively pumped drainage probably reveal the wetland extent prior to drainage for agriculture. The catchment's limestone geology is reflected in the high electrical conductance (Te Aute Rd site median 517 µS/cm, n=20; Mutiny Rd site median 548 µS/cm, n=5).

Springs were not surveyed in this catchment, beyond mapping the extent of wetted channels from aerial photographs (Figure 3-17). However, the gauging record is informative in terms of spring inflows. As flow recedes, a larger proportion of the Louisa flow originates upstream of Mutiny Road bridge (Figure 3-18). The inflows between Mutiny Road and Te Aute Road increase the MALF by only a third (24 L/s to 36 L/s), despite representing more than two-thirds of the stream's catchment area (cumulative area of 10 km^2 and 35 km^2, respectively). Morgenstern *et al.* (2018) estimated the water at Mutiny Road had a mean residence time of 120 years (sample 5/3/2015). The proportion of flow originating from the lower catchment increases at higher flows (Figure 3-18). This suggests a shorter residence time for groundwater originating from the lower catchment, which includes pumice sands (Figure 3-14) and limestone hill country. The start location of the perennial tributaries (draining to the Louisa and Te Waikaha) coincides with the extent of pumice-sands (Figure 3-17). Levy (2016a) noted a spring feeding the Louisa that was located along the margin between limestone hills and the plains (blue star in Figure 3-17, NZTM E1927877 N5597187, flow 2 L/s on 9/11/2016). Investigations into the contribution of groundwater from the pumice sands to these tributaries would improve our knowledge of surface-groundwater interactions in this area.

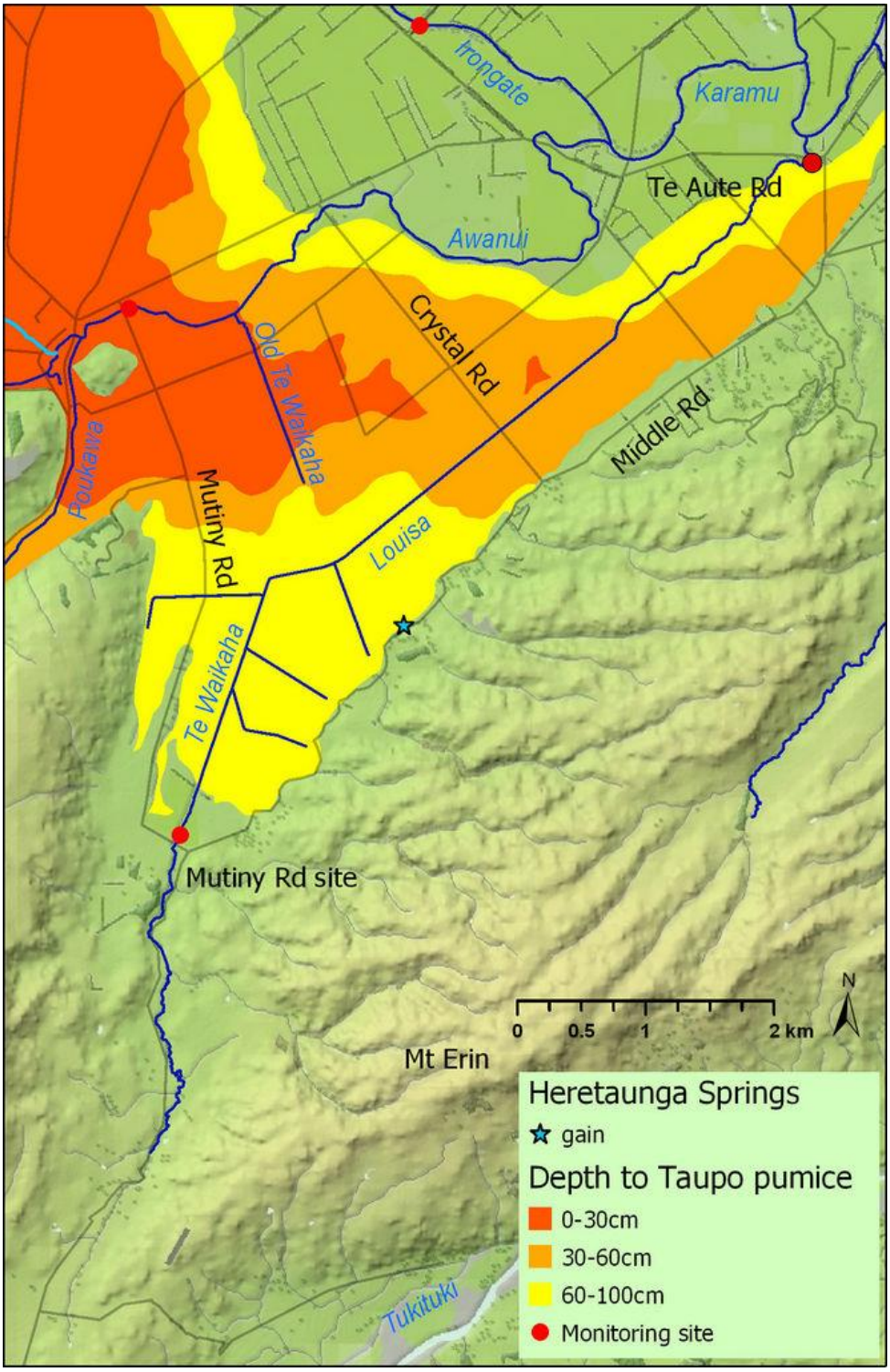


Figure 3-17: Louisa and Te Waikaha Stream. Springs were not mapped for this stream, beyond delineating stream extent from aerial photographs and Google Street View. A spring was reported by Levy (2016a) at the start of a smaller tributary that is not mapped. Flow gauging sites are indicated using red dots.

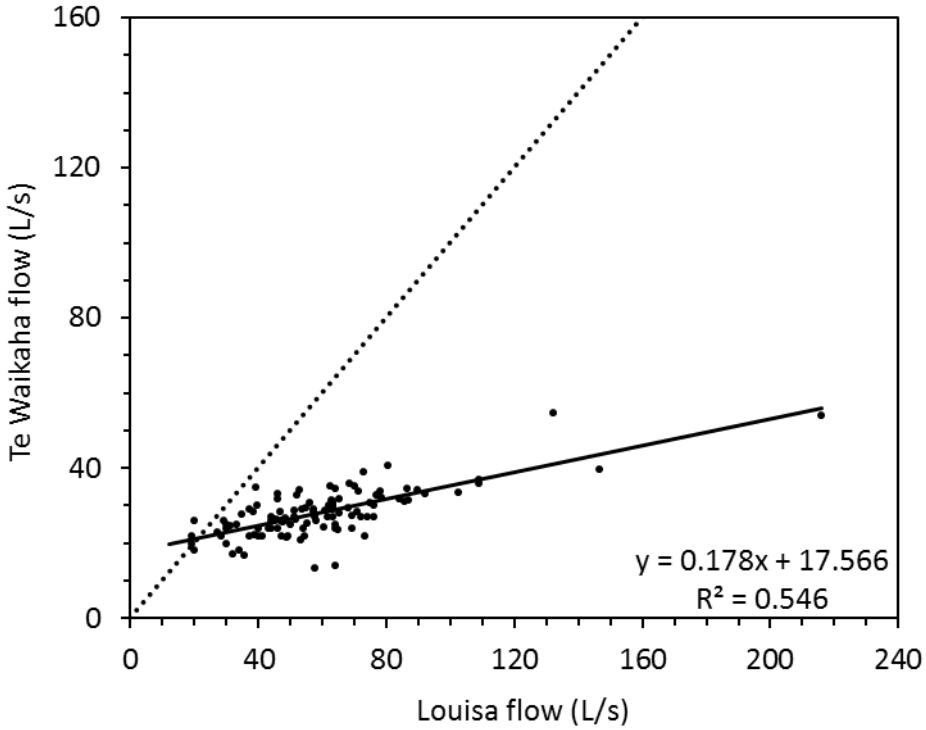


Figure 3-18: Louisa flow. The flow originating from the Te Waikaha Stream at Mutiny Road contributes a large proportion of the Louisa flow (measured at Te Aute Rd site, Figure 3-17). This proportion decreases at higher flows, dropping further below the dotted line (1:1). In total, 111 same-day gaugings were made at the two sites.

3.7 Irongate and Upper Southland

Flow in the Irongate Stream is probably dominated by springs. The magnitude of flow gauged at the monitoring site (Clarkes weir) was strongly correlated with groundwater levels at well 3737 (29.2 m deep), (Figure 3-19). The flow recession changed abruptly to a flatter response at groundwater levels less than 14.4 m, and flows less than 100 L/s. A similar abrupt change in recession was also apparent using a different well (Equestrian Park well 3698), confirming this pattern is not unique to the well 3737 (see Figure 3-21 for well locations). The stream flow was most often gauged during baseflow conditions, and gaugings rarely coincided with short-lived rainfall-runoff events. Peak flows are higher than other spring-dominated streams (e.g. Raupare Stream) because urban stormwater from Flaxmere is piped directly to the Irongate Stream and its tributaries.

No point-springs were observed in the Irongate catchment. Instead, groundwater seeps through the bed of the Irongate as diffuse springs. On the day of a concurrent gauging (2/12/2014), the stream water near the base of weed beds felt cooler than overlying stream water, which is consistent with seeping spring water. The concurrent gauging run demonstrated that the diffuse springs extended downstream to Railway Road (Figure 3-20). The flow measured at Clarkes weir was correlated with flow measured further downstream at Riverslea Road (R^2 = 0.95; n = 12). The Irongate has a mean annual low flow of 100 L/s at Clarkes weir (Waldron & Kozyniak, in prep 2014), which would equate to 170 L/s at Railway Rd (from concurrent gaugings).

Much of the gradual increase in flow occurs above a layer of clay that confines groundwater within the Heretaunga Aquifer (e.g. 12 m thickness of blue clay in well log 3322, 10 m in well log 10799). Above the confining layer sits a shallow layer of river gravel (Figure 3-21), which was left behind when the Ngaruroro River abandoned this flowpath in 1867 (HBRC, 2004). Well logs indicate this gravel layer is approximately 2.5

m to 5 m thick (e.g. well logs for 10799, 3322), transitioning to a similar thickness of pumice sand to the west of Hawke's Bay Expressway (e.g. well 8373). The gaining reaches of the Irongate, as mapped for this study (Figure 3-21), terminate at the downstream end of the shallow gravels that were included in soil maps by Griffiths (2001).

The shallow gravels would provide a longer groundwater pathway for Heretaunga groundwater spilling over the edge of the confining layer, compared to springs rising vertically to the Irongate through holes in the confining layer (Gyopari, 2011). Evidence for the longer pathway include the lack of response of flow in a tributary of the Irongate to 8 hours of pumping from nearby wells (30-60 m deep) (Rabbitte, 2012).

Stable isotopes collected from the Irongate at Clarkes weir ($\delta^{18}O$ -6.9 ‰, 4/3/2015) did not match Ngaruroro sourced groundwater ($\delta^{18}O$ -7.6 ‰, Section 3.1). For example, a mix of Ngaruroro sourced groundwater with 40% local rainfall recharge ($\delta^{18}O$ value of -6.0 ‰, Section 2.1.5) would produce the observed $\delta^{18}O$ for Irongate. Whatever the contribution from the Ngaruroro was at the time of sampling, this could change at different groundwater levels (e.g. more rainfall recharge in winter).

The length of the Irongate will increase when the widespread intermittent springs are flowing, compared to the length of perennial springs mapped in this report (see Section 1.2). For example, intermittent springs probably flow to the Irongate via Sisson Drain during winter. However, no springs were mapped for Sisson Drain because it was dry at the confluence with Irongate during the last concurrent gauging (2/12/2014). Both Gyopari (2011) and Rabbitte (2012) reported flows upstream of Portsmouth Rd for the Irongate during wetter months, and they anticipated that drying would extend below Portsmouth Rd as groundwater levels dropped in summer.

The Irongate Stream has occupied an old river bed of the Ngaruroro River (Figure 3-4) since a major flood moved the Ngaruroro to its current path in 1867 (NIWA events catalogue; HBRC, 2004). A section of the Southland Stream was diverted into the Irongate in 1964 to alleviate flooding of the lower Southland Stream (HBRC, 2004). This additional water enters the Irongate immediately below Clarkes Weir via a 1.2 km long pipe. Other changes to the channel included straightening and enlargement of the Irongate in the late 1960s (page 62 in HBRC, 2004). Agricultural land in the Portsmouth Rd area is tile drained, and the Flaxmere stormwater network appears to perform a similar drainage function, with flows persisting after rain has passed (Rabbitte, 2012).

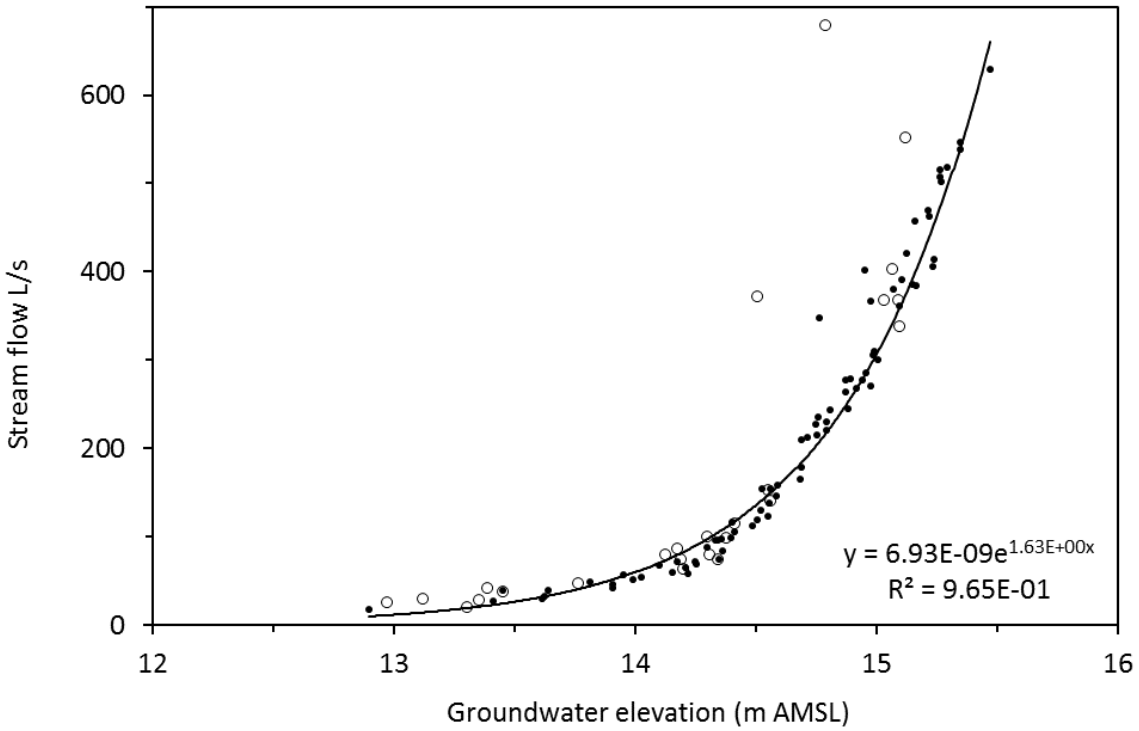


Figure 3-19: Irongate flow versus groundwater elevation. Flow for the Irongate at Clarke's weir is plotted against groundwater level in Flaxmere well (Well 3737, depth 29.2 m). Flows are plotted as open circles if there was rapid variation in flow (from rated stage to flow record) on the day of the manual flow measurement. These were isolated in case the flow measurements included rainfall runoff, which is unrelated to groundwater level.

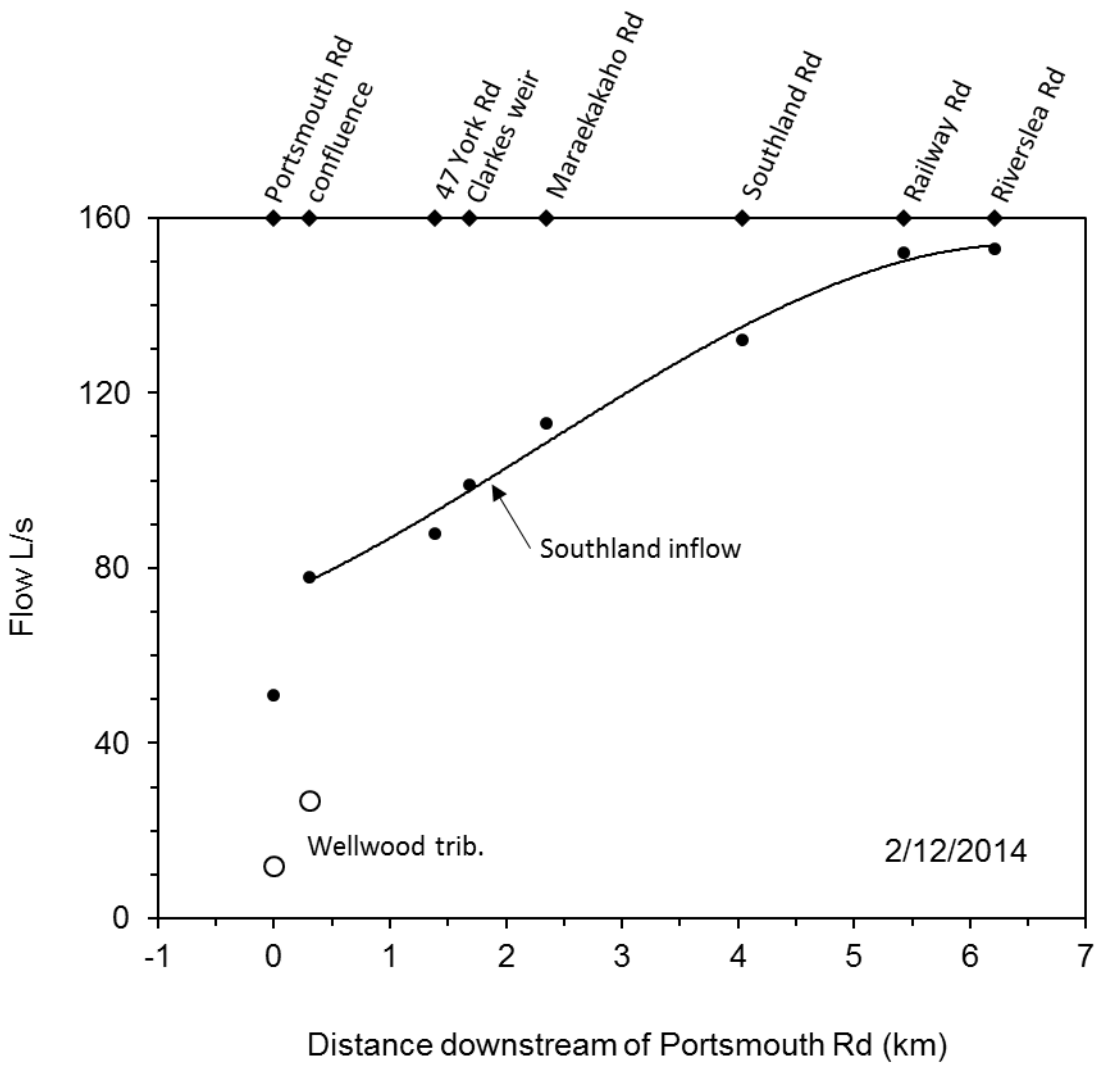


Figure 3-20: Irongate Stream flow increase. Flow measurements on 2/12/2014 are plotted against the distance downstream of Portsmouth Rd (Stock Rd intersection). The flow in the Irongate increased downstream of Portsmouth Rd. This gain levelled off at Railway Road (5.4 km). Measurements of a tributary inflow (Wellwood Stream) are plotted as open circles. The flow at the Wellwood confluence (0.3 km) was estimated by summing inflows from Wellwood plus Irongate. The location of the small inflow from Upper Southland Stream is arrowed (not measured).

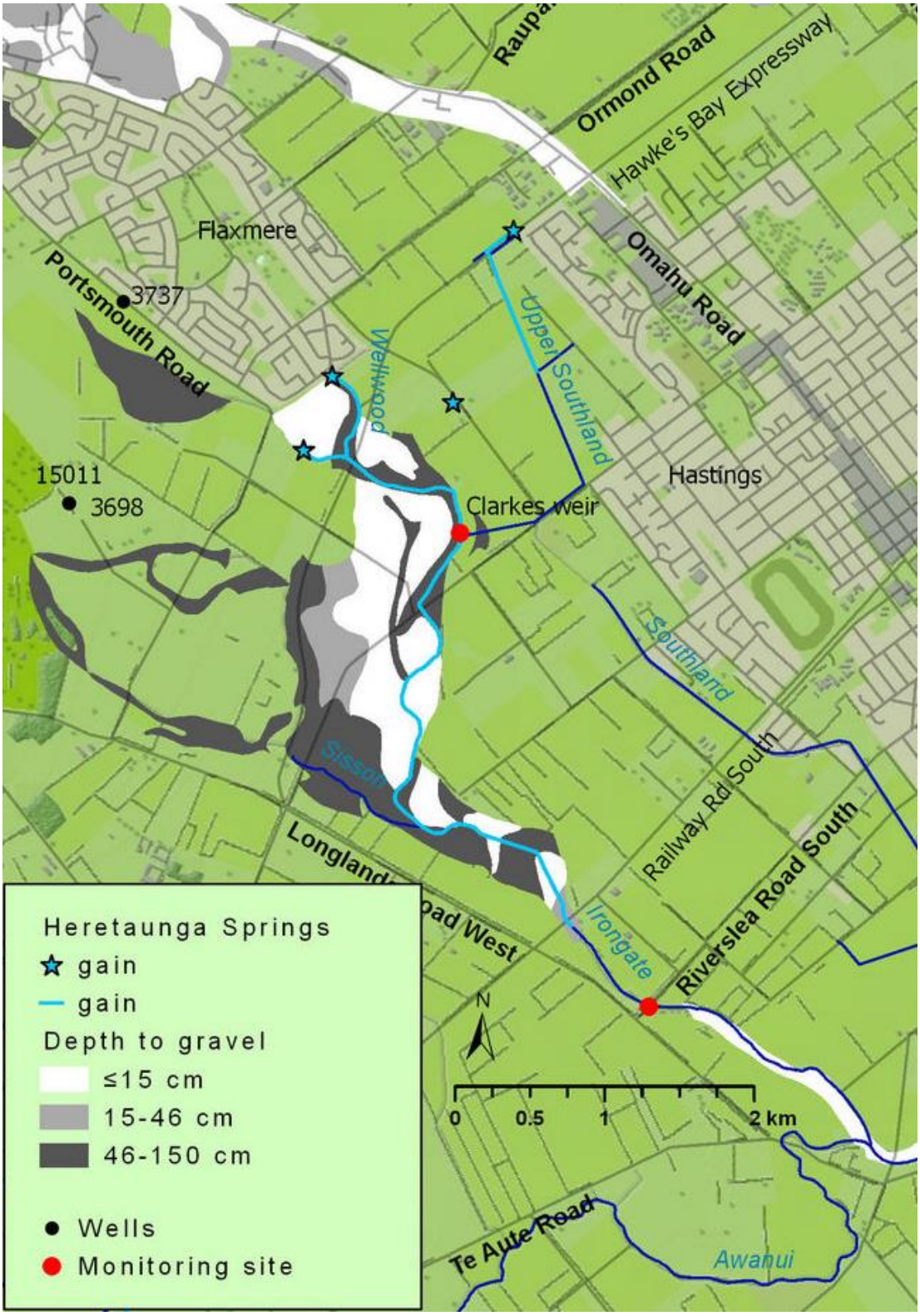


Figure 3-21: Irongate Stream. Perennial streams are shown as blue lines, with the start of perennial springs shown as blue stars. For Irongate, the springs are diffuse and extend all the way down to Railway Road (light blue line). Shallow gravels from soil maps are displayed as shaded polygons (e.g. white polygons represent gravel within 15 cm of the soil surface), (Griffiths, 2001). Note, soil mapping did not extend into urban areas like Flaxmere.

3.8 Karamu

A large component of the flow in the Karamu Stream has an unidentified source. Adding together the known tributary inflows accounts for less than half of the Karamu outflow, as measured at the floodgates monitoring site (Table 3-1). Estimates of the shortfall varied from 570 L/s, at mean annual low flow, to 920 L/s for the 75th percentile of concurrent gaugings (Table 3-2). Dravid and Brown (1997, page 167) were aware of the shortfall, and they suggested the additional inflow originated from irrigation return flows. They offered no explanation for the return flows concentrating here, but lacking elsewhere on the Heretaunga Plains.

Table 3-1: Unidentified inflows to the Karamu Stream. Tributary inflows to the Karamu Stream, upstream of the Raupare confluence, are substantially less than the measured flows at the floodgates monitoring site. The unknown inflow was investigated for this report. Existing mean annual low flows were estimated by Wilding (2016).

Stream	Mean Annual Low Flow (L/s)
Awanui at flume (Te Aute Rd)	87
Irongate at Riverslea Rd	168
Southland at St Georges Rd	10
Louisa at Te Aute Rd	36
Herehere at Te Aute Rd	7
Mangarau at Te Aute Rd	4
Karituwhenua at Napier Rd	10
Mangateretere at Napier Rd	48
Awahou at St Georges Rd	15
Ruahapia at Ruahapia Rd	15
Sum of tributaries	**400**
Karamu at Floodgates	**970**
shortfall	*570*

Table 3-2: Flow statistics for the unknown inflow to the Karamu Stream. Concurrent gaugings of major tributaries (Karamu at Havelock Rd; Mangateretere at Napier Rd) were used to calculate the shortfall needed to balance with the outflow measured at the floodgates site, together with approximate flows for minor tributaries (Karituwhenua, Awahou, Ruahapia).

Flow statistic	Flow L/s	method
Mean annual low flow	570	Subtract balance of tributary MALF (Table 3-1)
25%ile	640	Balance of concurrent gaugings
Median (50%ile)	710	Balance of concurrent gaugings
75%ile	920	Balance of concurrent gaugings

The location of the flow gains was estimated from longitudinal kayak-surveys of electrical conductance (Section 2.1). This took advantage of the natural contrast between the low-conductance of spring inputs (160 µS/cm on 5/2/2015 to 280 µS/cm on 13/3/2014 from concurrent gauging mass balance) and consistently higher conductance of surface inflows (310-500 µS/cm for Karamu upstream of Awahou

confluence). This kayak survey was repeated three times (13/3/2014, 1/4/2014 and 5/2/2015), and each survey indicated a similar pattern of declining conductance that started 0.5 km downstream of the Awahou confluence (Figure 3-22). There was a smaller contrast between spring inflows and river conductance at points further downstream where spring inputs started to dominate the river flow.

After converting the declining conductance to an estimate of flow (Equation 2-1), all three surveys produced a similar pattern of flow gains that were concentrated within three sections (Figure 3-23). The start of each section is mapped as a blue star in Figure 3-24. The largest input was from the section crossed by State Highway 2.

The gaining sections coincide with riparian willow wetlands, with waterlogged soils, puddles, seepages and small upwellings of less than 1 L/s (Figure 3-25). There was some evidence of the inflows originating primarily from the true right bank of the channel (eastern bank), with the willow wetlands located on the true right, and lower electrical conductance observed closer to the true-right bank through gaining sections.

In addition to the diffuse gaining sections through the willow wetlands, there is one point spring located by Golflands (1.2 km downstream of SH2). Several features distinguish this point spring from the diffuse springs further upstream. Firstly it is a larger (visual estimate 10 L/s) point-spring with boiling sands (Figure 3-25). Secondly, the chemistry of this point spring is different to the diffuse springs, including low electrical conductance (156 µS/cm).

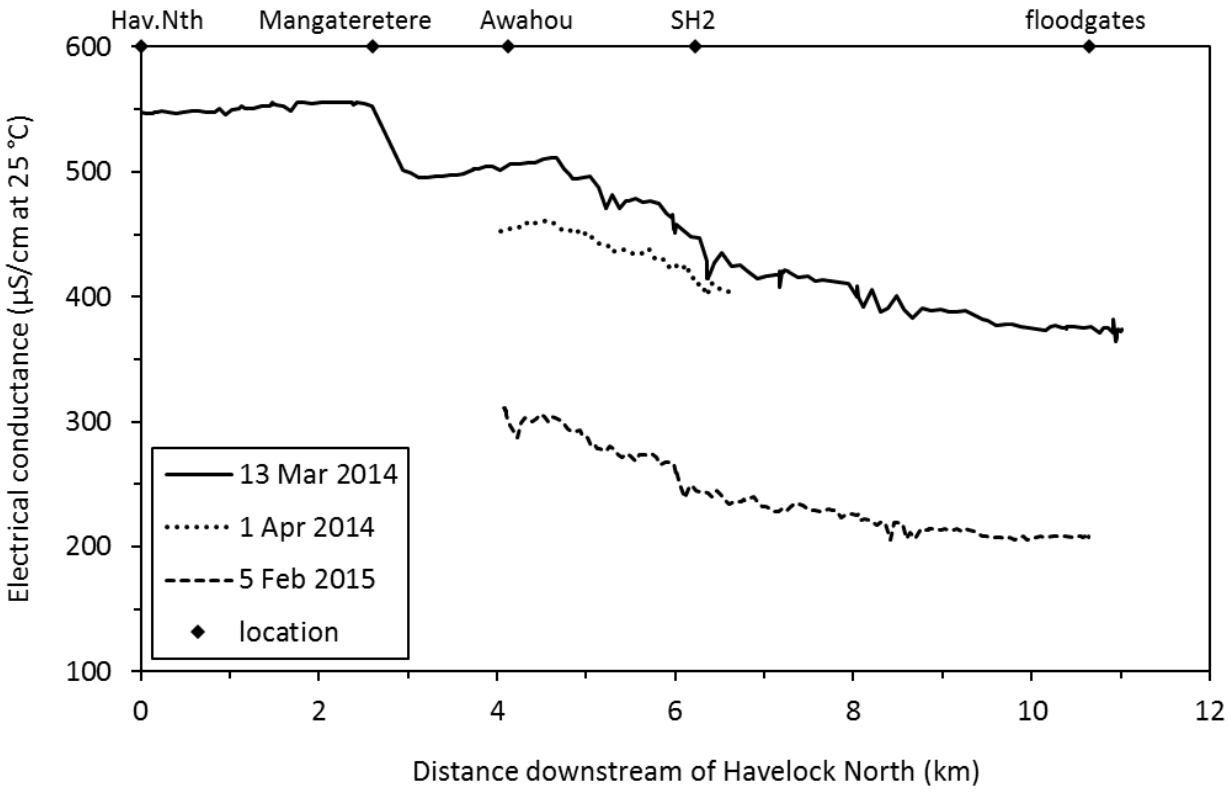


Figure 3-22: Conductance profile for Karamu Stream. Drop in electrical conductance in the Karamu Stream between Havelock Rd (0 km) and the Raupare confluence at Whakatu (11 km). Reference locations are labelled above the plot (black diamonds), including confluences with the Mangateretere and Awahou. The kayak survey was repeated on three occasions, plotted as separate lines.

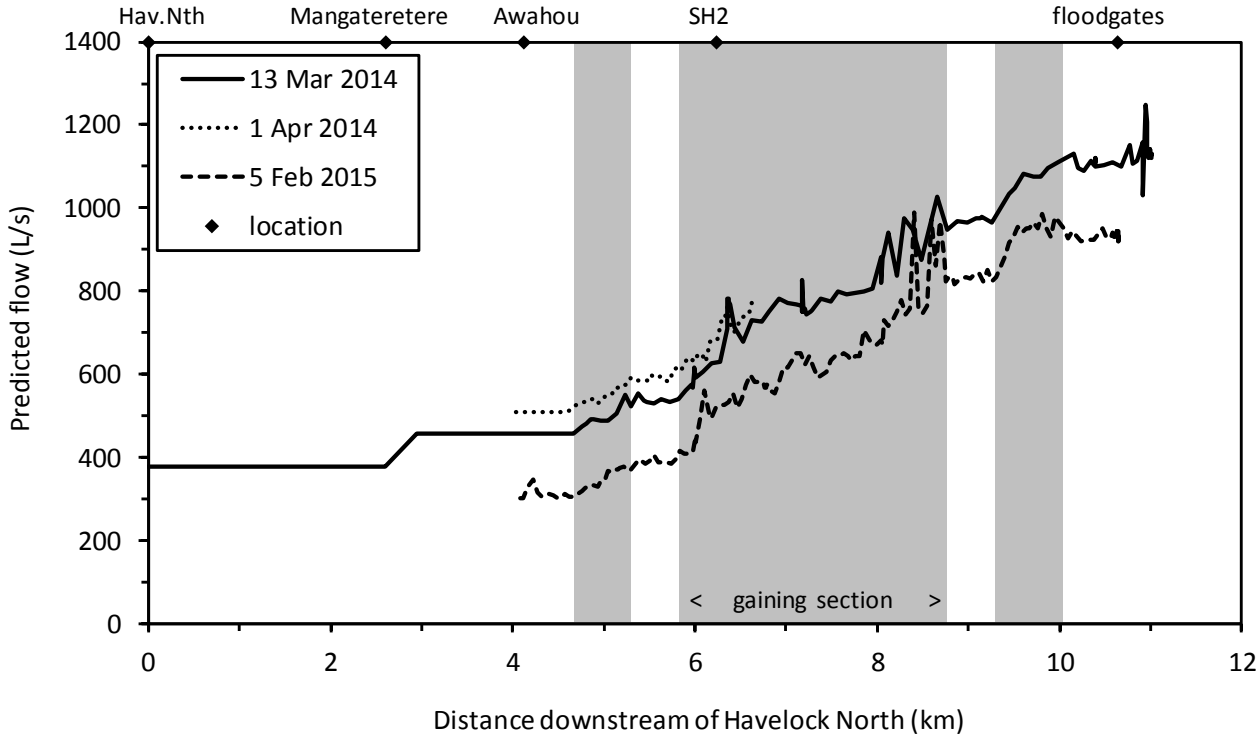


Figure 3-23: Predicted flow increase for Karamu Stream. The location of flow increases for the Karamu Stream was estimated from the drop in electrical conductance. Flow gains were more concentrated over three sections of river, highlighted by grey columns.

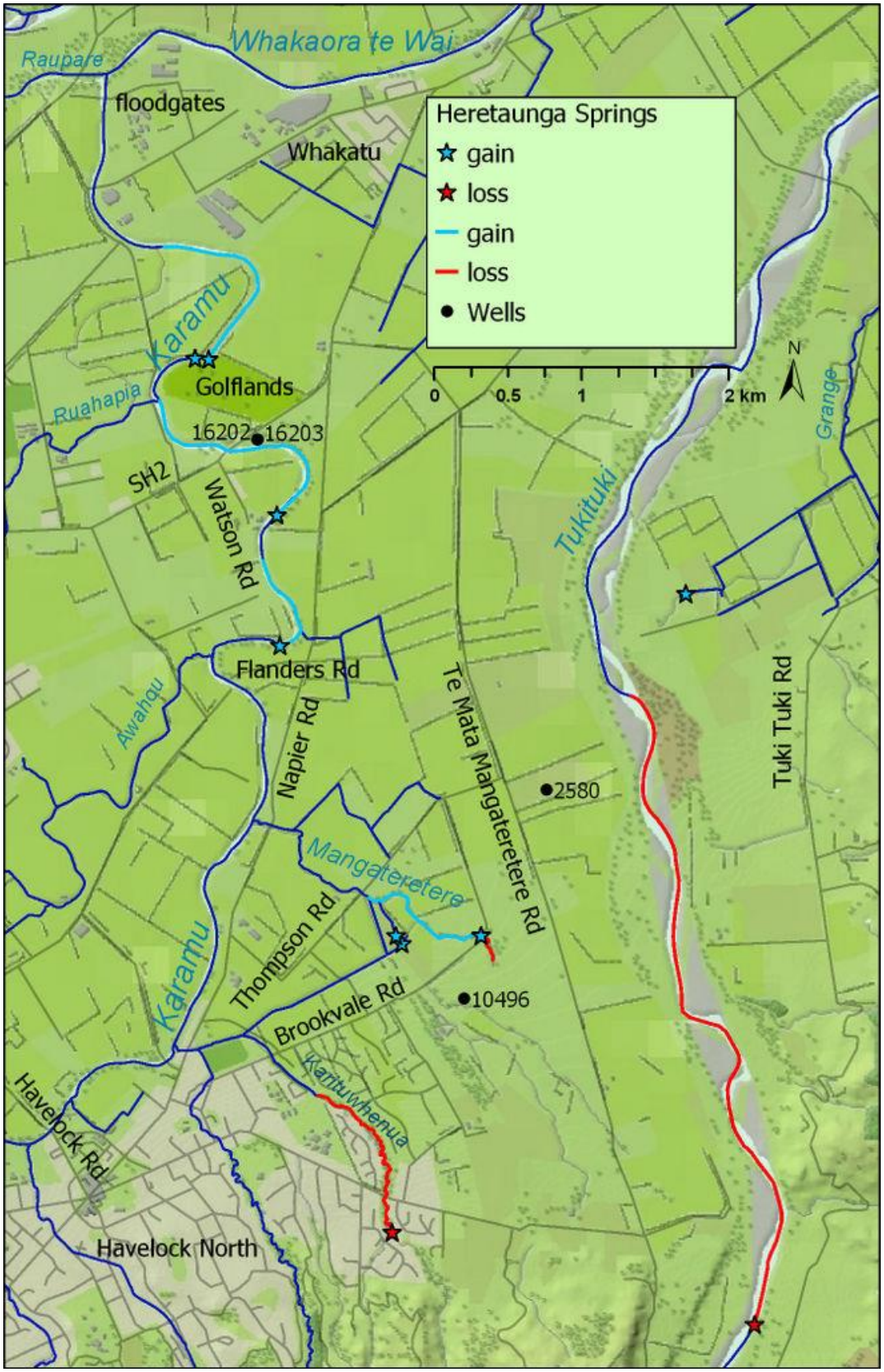


Figure 3-24: Karamu Stream flow gains and losses. Springs were located using the change in electrical conductance, with blue stars indicating the start of a gaining section. The losing section of the Tukituki River (red star at start) was located using concurrent gaugings (see Section 3.2). Selected wells are mapped.

Figure 3-25: Karamu Stream spring photos. The Karamu Stream gains flow from this section (top photo) that is bordered by willow wetlands on the true right bank (near Flanders Rd). Within the willow wetlands (lower left photo), the soils are damp and muddy, with some pools and small upwellings (photographs Kelvin Fergusson 1/4/2014). The point spring by Golflands (lower right photo) differs from the more diffuse springs associated with willow wetlands. Boiling sands reveal the stronger outflow (photograph 13/3/2014), although the total contribution to flow was <2%.

Local rainfall recharge was largely discounted as the dominant source of the unknown inflow. Using lysimeter recharge data from the Heretaunga Plains indicates a catchment area of 65 km^2 to 137 km^2 would be required to achieve the observed median inflow of 710 L/s (substation and Bridge Pa lysimeters, respectively, using 2013-2014 period). This area is based on an annual average recharge, ignoring the effect of reduced recharge in summer on spring inflows (e.g. <1 mm of total recharge depth for the 2013/14 summer, averaged across Bridge Pa, Maraekakaho and Substation lysimeters). A large aquifer would be required to sustain the magnitude of unidentified inflows through recent droughts. To achieve the inflow of 700 L/s, as measured in March 2013, would require an area larger than the Heretaunga Plains. For example, a 400 km^2 catchment area would be required if using the specific discharge of the Louisa Stream during March 2013, or 1000 km^2 if using the Paritua specific discharge at this time. The Louisa and Paritua were chosen for this example because they have negligible trans-basin gains, and many flow gaugings.

In reality, most of the Heretaunga Plains drains to other tributaries that are already accounted for, and the surface catchment draining directly to the Karamu is less than 7 km^2. Because the surface-catchment inflows can only explain a small fraction of the unknown inflow, groundwater inflows from neighbouring catchments are likely to form the bulk of the unknown inflow. This is further supported by the low electrical conductance of the inflow, compared to tributaries lacking trans-basin groundwater inputs (e.g. median 714 µS/cm for Awanui at Flume, n=15).

The two most likely sources of groundwater inflows are losses from the Tukituki River, or losses from the Ngaruroro River (via the Heretaunga Aquifer). These two sources were proposed by Dravid and Brown (1997) for nearby springs feeding the Mangateretere Stream. The authors suggested the spring source changed seasonally from predominantly Tukituki water in summer, to include Ngaruroro water in winter when groundwater levels in the Heretaunga aquifer are elevated (page 167 in Dravid & Brown, 1997). The hydraulic connection between the Heretaunga Aquifer and the shallow aquifer feeding the Mangateretere was thought to occur via overlapping gravel layers in the Thompson Road area (near the Mangateretere springs in Figure 3-24).

This same hydraulic connection near Thompson Road could connect the Heretaunga aquifer to the groundwater feeding the Karamu Stream via a shallow aquifer. Otherwise, a direct connection to the Heretaunga Aquifer would require flowpaths through a thick confining layer of marine clays (e.g. 8.3 m thick in well log 16203) in order to reach the gaining sections of the Karamu Stream. Such flowpaths could be produced by earthquake faults, for example, through inactive faults mapped in the Whakatu area by Lee *et al.* (2014), in addition to aquifer heterogeneities.

A potential connection also exists between the Tukituki River and the Karamu springs. The gaining section of the Karamu Stream passes within 2 km of the Tukituki River. As described in Section 3.2, the Tukituki River loses water from this adjacent reach, and the flow losses (median 910 L/s, Section 3.2) are similar in magnitude to the Karamu gains (median 710 L/s Karamu direct gain, plus 180 L/s via Mangateretere). A layer of gravel traverses the area between the Tukituki River and the Karamu Stream, which provides a potential conduit between the two. This layer is often recorded as brown or blue gravel in well logs, and the bottom of this gravel layer is typically 9 m to 13 m below ground. The top of the gravel layer was more variable in depth, within 1 m to 3 m below the ground surface in many bore logs (e.g. wells 2580, 1073, 5444, 16202), compared to more than 9 m deep in some (e.g. wells 5432, 830). This gravel layer also extends to the springs feeding the Mangateretere Stream. From soil maps, the material overlaying the shallow gravels is predominantly in the "Mangateretere" class (Griffiths, 2001). As discussed in Section 3.9, drainage through this soil type is limited by a clay pan and its susceptibility to soil compaction. Drainage barriers between the soil and shallow gravels are lacking from the lower-terraces located closer to the Tukituki River (Omarunui and Esk soil-classes).

The spatial extent of the shallow gravels was mapped by John Begg and others at GNS, in a revision Heretaunga plains geological model (Lee *et al.*, in prep 2018). This 3-dimensional geological model of the gravel fan deposited by the Tukituki River shows two boundaries (Figure 3-26). Firstly, the maximum extent boundary includes deeper gravels that lie beneath the confining clays (light green area in plan-view). The second boundary outlines the smaller area of gravels that are shallow enough to intersect the stream bed ("Tukituki fan shallow gravels" dark-green line in Figure 3-26 plan view). The boundary of the shallow gravels provides a close match with the losing reach of the Tukituki River. These shallow gravels also extend across to the Karamu Stream and the Mangateretere Stream, confirming the potential for a groundwater connection.

The elevation of the Karamu Stream is lower than the Tukituki River, providing the necessary gradient for groundwater to flow from the Tukituki to the Karamu, via the shallow gravel layer. The Karamu gains flow at elevations less than 2.2 m above sea level, compared to losses from the Tukituki that were measured between 7.6 and 15.3 m above sea level (using LiDAR data for water surface elevations, Section 3.2).

The 3 dimensional geological model also shows the potential connection between the deeper gravels of the Heretaunga aquifer and the shallow gravels (Figure 3-26). This could provide a conduit between Karamu Stream and Ngaruroro sourced water. The layering of gravel, sand, silts and clays is spatially complex within this alluvial fan. Representing this complexity within a geological model is limited by a variety of factors, such as the number of bore logs, consistent description of sediments across decades of drilling, and the interpretation of those bore logs within the software (Rakowski, in prep 2018). Because of these limitations, the geological map cannot resolve the exact nature of these groundwater pathways.

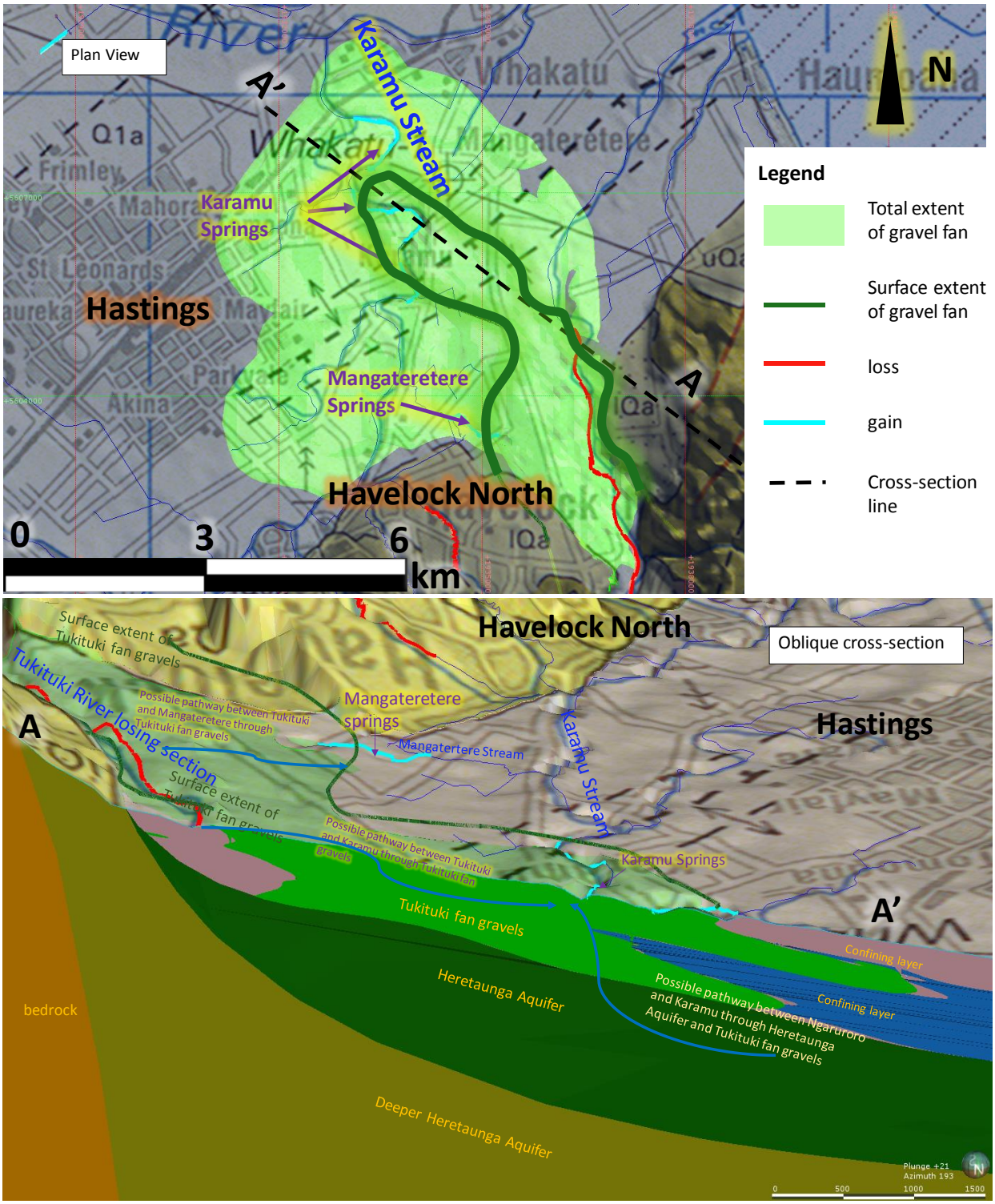


Figure 3-26: Shallow gravels between the Tukituki and Karamu. A 3-dimensional geological model was constructed from well logs, which reveals the shallow layer of gravel ("Tukituki fan shallow gravels") connecting the losing reach of the Tukituki River to the gaining reach of the Karamu Stream. These maps were prepared by Rakowski (in prep 2018)based on John Beck's September 2017 revision of the Heretaunga geological model (Lee *et al.*, in prep 2018). The dashed cross-section line shown in the plan-view shows the orientation of the cross-section in the lower oblique image.

Having demonstrated the hydrogeological potential for water to originate from the Tukituki, or the Ngaruroro, the next step is to determine which path the water follows at any point in time. This was investigated by comparing the chemistry of the Karamu springs to that of the Ngaruroro and Tukituki. Electrical conductance provides a contrast between the Tukituki River (188-219 µS/cm inter-quartile range, n = 94, Red Bridge) and the Ngaruroro River (135-160 µS/cm inter-quartile range, n = 50, Fernhill). Estimates of electrical conductance for the spring-inputs to the Karamu range from 160 µS/cm (5/2/2015) to 280 µS/cm (13/3/2014), calculated using mass balance of the missing inflow from kayak surveys. This range of values is consistent with the contribution changing over time from the Tukituki to the Ngaruroro. A low value for electrical conductance should be a reliable indicator of Ngaruroro sourced water. However, higher conductance values are not a reliable indicator of Tukituki water because small contributions from limestone areas can overwhelm the margin of difference (e.g. Ngaruroro water of 160 µS/cm mixed with 10% water at 600 µS/cm produces a 200 µS/cm mixture).

The relative contribution of the two rivers could not be resolved from the composition of major ions, because the Ngaruroro and Tukituki contain similar proportions of calcium, bicarbonate, sodium and chloride (e.g. Piper plots). Therefore, isotope composition has better potential for resolving the flow source (Taylor *et al.*, 1989). As explained in Section 2.1, the stable isotopes $\delta^{18}O$ and $\delta^{2}H$ are often reliable tracers of water source.

The two likely groundwater sources of the Karamu springs are the Ngaruroro and Tukituki rivers. As detailed in Section 3.1, there was a consistent difference between river water from the Ngaruroro and Tukituki, with Ngaruroro water consistently more negative than the Tukituki ($\delta^{18}O$ mean difference -1.1‰). Some uncertainty remains in the specific values associated with groundwater from each source (see Section 3.1 and 3.2). However, the $\delta^{18}O$ of groundwater from beside the losing reach of Ngaruroro was consistently more negative than groundwater beside the Tukituki losing reach (-6.4‰ and -7.6‰ respectively). For comparison with these river-recharge signatures, the isotope ratios for the Karamu spring inflows were calculated using a mass-balance equation (i.e. subtracting concentration multiplied by flow from all inflows). This indicated that the Tukituki River was the dominant source of Karamu inflows on 3/3/2015 when the Karamu was experiencing low flows (808 L/s at floodgates). The $\delta^{18}O$ ratio for the combined spring inflow was -6.2‰, which is a closer match to Tukituki groundwater (-6.4‰) than Ngaruroro groundwater (-7.6‰).

The point spring by Golflands had a $\delta^{18}O$ value of -7.6‰, reinforcing the contrast between this spring and the dominant Karamu inflow. Both the Golflands spring and a small spring closer to SH2 bridge (spring 22, Appendix B) had isotope ratios more like the Ngaruroro water. Therefore, the gaining reach of the Karamu did receive Ngaruroro sourced water, albeit a minor component on 3/3/2015.

To investigate whether the dominant spring source changes over time, the sampling was repeated at higher flows in winter (Karamu 3757 L/s at floodgates on 23/8/2017). This time, the mass balance calculation revealed a mean $\delta^{18}O$ value of -7.0‰ for the spring inflows. This stable isotope signature can be achieved by mixing 40-50% Ngaruroro sourced groundwater with the remainder Tukituki sourced groundwater ($\delta^{18}O$ of -7.6‰ and -6.4‰ respectively).

A shift from the Tukituki being the dominant source during summer low flows, to increasing inputs from Ngaruroro sourced water during winter is consistent with the seasonal variability in contribution proposed by Dravid and Brown (1997) for the Mangateretere Stream. The Mangateretere Stream is probably fed by the same groundwater system as the Karamu (see Section 3.9). Adding support to this, the relationship between groundwater levels at Brookvale and Mangateretere Stream flow is similar in form to the relationship with Karamu spring inflows (Figure 3-28). The shallow aquifer feeding the Mangateretere Stream has been referred to as the "Te Mata Aquifer" (DIA, 2017), or the "Tukituki Aquifer" (Dravid & Brown, 1997). The implication of the isotope results is that the Te Mata/Tukituki Aquifer is bigger than previously realised, both in terms of the area and the flow contribution to Karamu Stream.

The alignment of the Karamu Stream has seen little change in recent history. But the catchment area feeding the Karamu has changed drastically. Prior to 1867, the Ngaruroro followed the entire length of what is now the Irongate and Karamu (see Figure 3-4). After this river avulsion, the Ngaruroro entered the Karamu at what is now the Raupare confluence. In 1969, the Ngaruroro was diverted by the Hawke's Bay Catchment Board down a channel that was originally constructed as an overflow channel in 1940. The overflow channel remains as the flowpath for the Ngaruroro to this day. The most recently abandoned channel (downstream of the Raupare confluence) is often referred to as the Clive River, or Whakaora te Wai. Flow in the Clive River now only carries the Karamu and Raupare waters, which represents an 80% reduction in mean annual low flow, compared to when the Ngaruroro followed this path.

3.9 Mangateretere

The Mangateretere Stream is a major tributary of the Karamu Stream. The length of the stream was walked for this investigation to identify the start of the stream and the location of visible springs (21/5/2014, 11/7/2014). The first springs were observed immediately downstream of Brookvale Road, with a conspicuous increase in flow within the first 500 m (Figure 3-24). Some inputs were visible as discrete point-springs, including upwelling around old tree roots (Figure 3-27). However, the observed springs only account for a small proportion of total flow. The remainder of the flow may arise as diffuse springs, or the sum of many inconspicuous point-springs (e.g. submerged, overgrown). The combined outflow from these springs generates a mean annual low flow of 48 L/s, and a median flow of 180 L/s, at the Napier Rd monitoring site located 130 m upstream of the Karamu confluence (rated stage for period 1999 to 2015).

Flow has been monitored continuously at Napier Road since November 1999. However, there have only been three concurrent gaugings at upstream locations to date. The three concurrent gaugings indicate that most of the groundwater inflows are located upstream of Thompson Rd (flow at Thompson Rd was approximately 90% of the flow at Napier Rd). The flow in the Mangateretere Stream was correlated with groundwater level (Figure 3-28).

Local rainfall recharge cannot account for the magnitude of flow originating from the Mangateretere springs. Compared to the Mangateretere's surface catchment area of 6.2 km^2, a catchment area of 17 km^2 to 35 km^2 would be required to achieve the observed median flow of 180 L/s from the rainfall-recharge rates measured on the Heretaunga Plains (lysimeters at substation and Bridge Pa, respectively, using the 2013-2014 period). This area is based on an annual average recharge, ignoring the effect of reduced recharge in summer on spring inflows (e.g. <1 mm of total recharge depth for the 2013/14 summer, averaged across Bridge Pa, Maraekakaho and Substation lysimeters). A larger area would be required to sustain the magnitude of spring flow through recent droughts. For example, the specific discharge of neighbouring catchments, which lack trans-basin groundwater inputs, was approximately a third of the specific discharge of the Mangateretere (26% for Herehere, 35% for Mangarau, from gaugings on the 27/2/2013). Added to this shortfall is the large portion of flow taken from the Mangateretere by stream-depleting groundwater takes (e.g. average 140 L/s metered use from the Brookvale municipal wells on 27/2/2013, compared to a gauged stream flow of 18 L/s).

Because the surface-catchment inflows can only explain a small proportion of the spring flow, groundwater inflows from neighbouring catchments are likely to form the bulk of the flow. This is further supported by the low electrical conductance of the Mangateretere, compared to tributaries lacking trans-basin groundwater inputs (median 244 µS/cm at Napier Rd, compared to 714 µS/cm for Awanui at Flume).

Dravid and Brown (1997) proposed that the shallow groundwater feeding the Mangateretere Stream originated from the Tukituki River, with increasing inputs of Ngaruroro sourced water during winter when water levels in the Heretaunga aquifer were elevated (page 167). Neighbouring landowners also considered the Tukituki to be the source of the spring water (*pers. comm*. Jim Frogley, May 2016). The stable isotope ratios in the Mangateretere Stream support this, with $\delta^{18}O$ of -6.4‰ measured in the Mangateretere (Napier

Road) on 3/3/2015 (flow 38 L/s at Napier Rd), compared to -6.4‰ for Tukituki sourced groundwater (Section 3.2), and -7.6‰ for Ngaruroro groundwater (Section 3.1). A spring immediately downstream of Brookvale Rd (spring 13) also produced isotope signatures closer to the Tukituki water (-6.5‰ 6/5/2016).

A second set of isotope samples were collected at higher flows in winter (258 L/s at Napier Rd on 23/8/2017), revealing a more negative $\delta^{18}O$ value of -6.9‰. This stable isotope signature can be achieved by mixing 40-50% Ngaruroro sourced groundwater with the remainder Tukituki sourced groundwater ($\delta^{18}O$ of -7.6‰ and -6.4‰ respectively). The increased contribution of Ngaruroro-sourced groundwater during winter is consistent with the seasonal variability proposed by Dravid and Brown (1997) for the Mangateretere Stream.

The shallow aquifer feeding the Mangateretere extends from the Tukituki River, which presumably deposited this layer of gravel. The Mangateretere springs start at an elevation that is equivalent to the end of the Tukituki losing reach (7.7 m versus 7.6-15.3 m respectively), creating the gradient necessary for Tukituki sourced groundwater to flow towards the Mangateretere Stream (using LiDAR data for water surface elevations, Section 3.2).

The gravels lie on the edge of a limestone basement material (limestone, mudstone and sandstone of marine origin), which increases the electrical conductance of groundwater and springs closer to Te Mata Mushrooms and Arataki Road. For example, an electrical conductance of 417 µS/cm was measured in the true left tributary (closer to the hills) compared to 225 µS/cm in the true right tributary (both sampled 6/5/2016). Likewise, the conductance was higher in a shallow well closer to the hills (581 µS/cm in well 10496, 8/3/2016), compared to a shallow well closer to the Tukituki River (200 µS/cm in well 2580, 6/5/2016). The median conductance of the combined Mangateretere outflow (measured at Napier Road) was 244 µS/cm (n=159 days of data logging). This indicates that water originating from the limestone hill country represents a minor component of the total outflow. For example, 12% limestone water at 600 µS/cm plus 88% of Tukituki water at 200 µS/cm would produce a solution equivalent to the observed outflow at Napier Road.

A capping of clay pan is noted from the classification of “Mangateretere” soils, which overly the shallow gravels (Griffiths, 2001). If this pan were extensive, it would limit the rainfall recharge of this shallow aquifer. Augering near Brookvale Road in 2016 revealed about 1.5 m of alluvial soils over dense loess (Grant Upchurch, memo 26/9/2016). Poor soil structure in some paddocks further reduced water infiltration to 1 mm per hour. Peat soils were observed in stream cuttings downstream of Brookvale Road, revealing historical wetlands that were likely sustained by springs in this area. Some springs emerge via the decaying roots of tree stumps (lower left, Figure 3-27). One landowner commented that ancient logs within the peat seem to have fallen in the same direction (*pers. comm*. Jim Frogley, May 2016), which is consistent with wind disturbance patterns proposed for Hawke’s Bay by Grant (1996).

Surface-groundwater interactions between the Mangateretere Stream and the shallow aquifer were the subject of detailed investigations, following the *Campylobacter* contamination of Havelock North’s water supply in August 2016 (DIA, 2017). Dye tracer tests revealed that water could be drawn into the aquifer through the bed of the Mangateretere Stream, bypassing the clay pan that was observed in surrounding paddocks. A dye solution was applied to the drying reach upstream of the perennial springs (i.e. upstream of Brookvale Rd), and was subsequently detected in nearby groundwater wells, including municipal supply wells (DIA, 2017). The findings of the government enquiry were wide ranging (DIA, 2017). One outcome that is relevant to this report is the potential for intermittent springs (i.e. stream reaches that seasonally change from gaining to losing) to become a conduit for particulate contaminants (e.g. animal faeces) from floodwater to enter groundwater.

Some historic changes were noted for the Mangateretere Stream with the most prominent being wetland drainage. The LiDAR elevation imagery indicates the main channel may have been cut through a mound 50 m downstream of Thompson Road, perhaps to facilitate wetland drainage. The wetland drainage perhaps

was directed toward the Karituwhenua Stream for some time, given that a constructed barrier on Crombie Drain now prevents flow to the Karituwhenua. In the 1800s, the name Wahaparata Stream was used for part or all of the Mangateretere, and this name was also given to a flour mill located near the stream (Fowler, 2017). Springs arise downstream of Brookvale Road during summer. From discussion with local landowners, water was normally present upstream of Brookvale Rd, prior to the municipal supply wells increasing supply to Havelock North.

Figure 3-27: Mangateretere Stream. Many small point-springs were observed flowing into Mangateretere Stream, which gains much of its flow over a short reach downstream of Brookvale Road. Springs arise around ancient tree roots (bottom left) and through clay (bottom right, pictured with isotope sampling pump in place underwater).

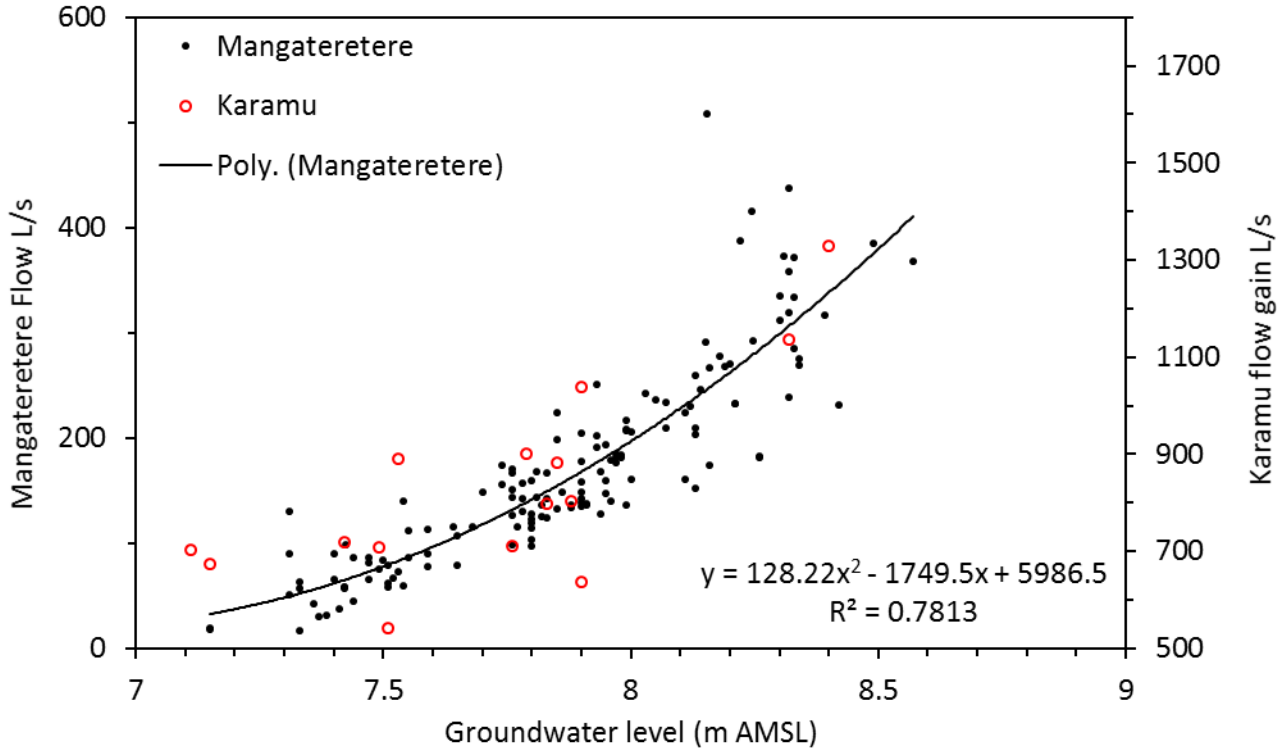


Figure 3-28: Mangateretere flow versus groundwater elevation. Gauged flow for the Mangateretere Stream at Napier Road is plotted against groundwater level (metres above mean sea level) at Brookvale (well no. 10496) measured within seven days of the gauging (from dipped groundwater level). Also over-plotted are spring inflows to the Karamu Stream (estimated from concurrent gaugings) plotted against Brookvale groundwater levels.

3.10 Raupare

The Raupare is a spring-dominated stream that arises close to the variable-loss reach of the Ngaruroro River (Figure 3-29). The springs in this catchment were mapped using aerial photographs, together with conversations with landowners, plus walking lengths of the stream. Concurrent gaugings provided information on the magnitude of inflows from the many springs in this catchment. Point-springs with boiling sands were found between Raupare Road and Twyford Road (Figure 3-30), presumably revealing holes (i.e. gravel/sand discontinuities) in the clay confining layers (Figure 1-1, Figure 3-30). Discussion with landowners reveals that these springs sometimes change (i.e. start or stop flowing), in response to earthquakes or removing tree stumps.

Springs arising where the groundwater level intersects the surface are typically more diffuse, as found west of Twyford Road and Trotter Road (Figure 3-30). Harper and Hughes (2009) described this part of the aquifer as semi-confined, because sand and gravel is interspersed through the thin edge of confining clays, allowing groundwater to overflow as diffuse springs (Figure 1-1). Land adjoining the diffuse springs has been drained by networks of buried tile-drainage pipes to enable deeper rooting of orchard trees. Reportedly, some tile-drains require pumping where the drain outlet is lower than the receiving surface drain.

Most of the flow in the Raupare can be accounted for from the tributaries (Burns, Flowers, Twyford, Gudsell, plus Raupare at Carrick Road, Figure 3-29). These leave a shortfall of approximately 10% of the flow measured at Ormond Road, which could be attributable to gauging error or springs arising in the bed of the main stem. . The gaugings to date suggest that the diffuse springs closer to the Ngaruroro River have more variable flow than the point springs closer to Raupare Road (coefficient of variation 87% for Raupare at Twyford Rd; 34%

for Burns at Nicholl Rd; n=9 concurrent gaugings). Morgenstern *et al.* (2018) estimated a younger age for the springs with more variable flow, however, both were relatively young (mean residence time 0.5 years for Raupare at Twyford Rd; 2 years for Raupare spring that feeds Burns tributary). There was also a difference in the stable isotope ratios between the two areas ($\delta^{18}O$ -7.5‰ at Twyford Rd, -8.0‰ for Raupare spring that feeds Burns tributary on 3/3/2015).

The combined outflow of these springs, measured at Ormond Road (2014-2016), produced a mean annual low flow of 400 L/s and a median flow of 650 L/s (synthetic flow record for 1977-2015 from concurrent gauging with Irongate at Clarkes weir). The flow at Ormond Road was correlated with groundwater level at the nearest monitoring well (well no. 15006, 30 m deep, Figure 3-31), using the more recent velocity meter record (from January 2014). An exponential equation provided a better fit than a linear function, especially at groundwater levels higher than 14 m (Figure 3-31).

Results from stable isotope sampling are consistent with groundwater recharged by the Ngaruroro River being the primary source of Raupare water. The $\delta^{18}O$ was -7.7‰ for Raupare at Ormond Rd (4/3/2015), which is a close match to values of -7.6‰ for groundwater adjacent to the losing reach of the Ngaruroro (Section 3.1).

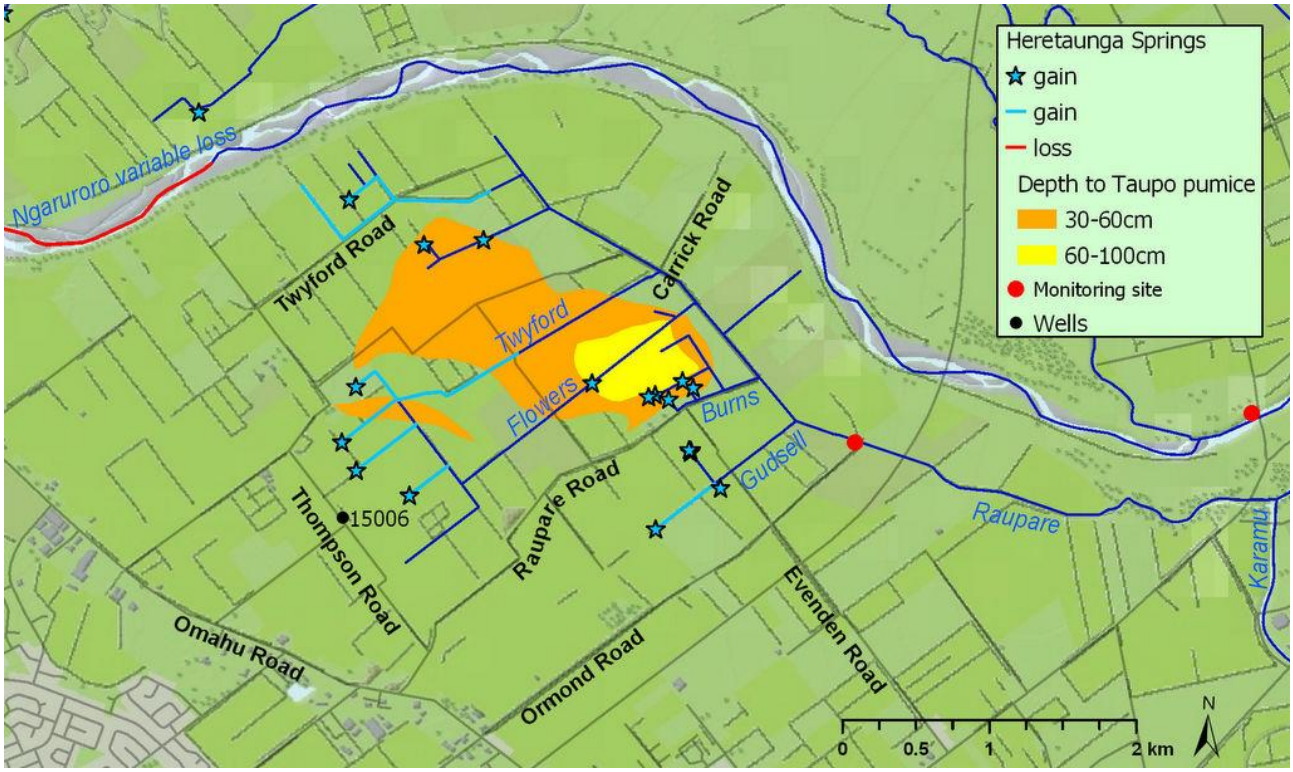


Figure 3-29: Raupare Stream map. Springs were mapped in the Raupare catchment, with point springs shown only as blue stars, and more diffuse springs combined with a light blue line to give an indication of the length of gaining stream. The soil depth (cm) to the top of a layer of Taupo pumice/ash sand is represented by coloured polygons from Griffiths (2001). The flow monitoring site at Ormond Road is marked with a red dot, and selected groundwater monitoring wells with a black dot.

Figure 3-30: Raupare Stream and springs. At the monitoring site (upper photo, Ormond Road, 20/7/2012), the Raupare is fed by both large point-springs (lower left, Raupare Rd, 3/2/2016) and small diffuse-springs (lower right, Thompson Rd, 27/3/2013).

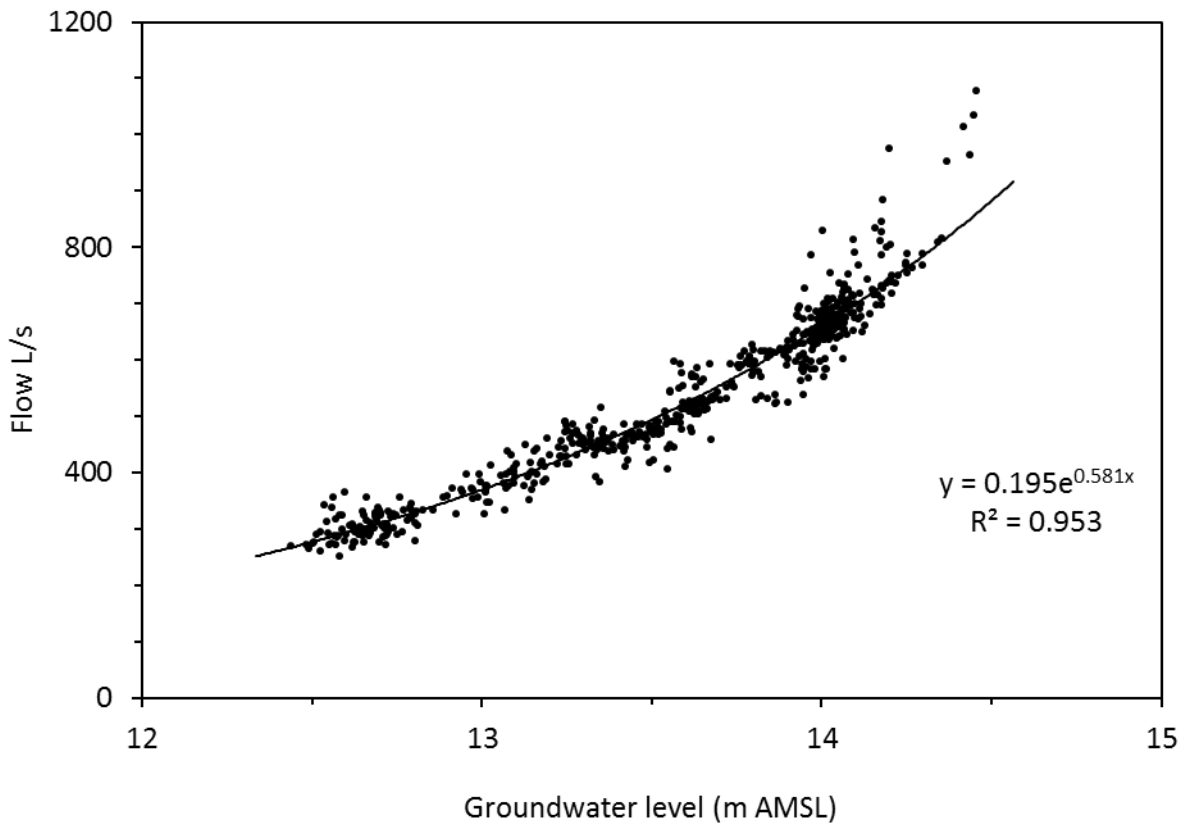


Figure 3-31: Raupare flow correlation with groundwater level. The daily mean flow, as measured using a Sontek IQ velocity meter (1/1/2014 to 20/9/2016), is plotted against daily mean groundwater level, as measured at well 15006. This was the period of overlap between groundwater level data and stream velocity data. Flow estimates from the velocity meter better account for the effect of aquatic plant growth on stream depth than the longer-term record of rated water-level.

The quality of the emerging groundwater is generally excellent (HBRC, 2014), with 95% of nitrate samples less than 0.5 mg/L (from 71 samples collected between 1995 and 2014 from Monitoring Well 1674). The water temperature is also stable, with spring water emerging from the ground at 14.5 °C.

In 1867 the course of the Ngaruroro River shifted (Figure 3-4), so that the Raupare catchment was no longer on the north bank of the Ngaruroro, but instead was situated on the south bank (*NIWA events catalogue*; HBRC, 2014). The river avulsions may have changed the location and magnitude of spring outflows. This river avulsion also cut off flows from the Waitio and Ohiwia catchment, which previously traversed this area (page 34 in Dravid & Brown, 1997). The natural springs and wetlands were modified to their present channel alignment at least 75 years ago. The stream network as delineated in a soil map from 1938 (Bell *et al.*, 1938), and in 1950 aerial photographs, was substantially similar to today's network. Boyd (1984), page 102) mentions drainage of Frimley Estate by the early 1890s. Frimley Estate extended downstream of Ormond Rd at a time when dairy farming was a significant industry. This preceded several decades of rapid orchard expansion in the Hastings area (Boyd, 1984; Wright, 2001) that may have accelerated efforts to drain the Raupare catchment.

Some older bores have failed in a way that contributes flow to Raupare Stream in an uncontrolled (and uncontrollable) manner. For example, a bore on Evenden Road developed a problem in December 1966 when water started escaping around the well casing while it was being drilled. Initially, water was free flowing at 50 L/s, eventually reducing to 18 L/s (Dunlop, 1992). More recent measurements (14 April 2014) of outflow

from this failed bore installation on Evenden Road (NZTM E1928711 N5609610) were less than 3 L/s. It was therefore contributing less than 1% of flow in Raupare Stream, and it enters the Raupare via Gudsell Drain (Figure 3-29).

3.11 Maraekakaho

The Maraekakaho Stream drains limestone hill-country (sandstone, limestone and mudstone of marine origin) together with some old river terraces, before flowing across more recent alluvial terraces as it approaches the Ngaruroro. This transition to the more recent alluvial terraces occurs downstream of Tait Road (Figure 3-32), coinciding with changes in the surface-groundwater interactions for the Maraekakaho. The Maraekakaho Stream loses flow to groundwater over the reach downstream of Tait Rd (red star in Figure 3-32). Complete drying can occur over a section starting about 2 km upstream of Kereru Rd (Christie, 2010). When this section is dry, the stream starts flowing again about 0.5 km further downstream (blue star in Figure 3-32), and regains much of its flow before reaching the Ngaruroro River (from concurrent gaugings 31/1/1994 and 1/19/1967).

More concurrent gaugings along the Maraekakaho Stream (Tait Rd to Ngaruroro confluence) would improve our understanding of the flow losses and gains. The transition from losing to gaining reach is likely determined by groundwater levels, and those groundwater levels could be influenced by water levels in the Ngaruroro River (Rosen, 1996).

The recorded water levels at the Tait Rd site have a pronounced diurnal variation, which was the topic of a master's thesis (Lee, 2011). That research considered changes in water use, evapotranspiration, and groundwater pressure as potential drivers of the diurnal pattern. But work since then has revealed that thermal expansion of the recorder tower was inflating that diurnal pattern. For example, an 11 mm error in water level was created by air temperature increasing from 5 to 15 °C (float encoder compared to submerged hobo logger, correlated with air temperature inside the tower). Since then, the shaft encoder was replaced with an OTT bubbler (21/10/2015). A diurnal pattern still appears in the record at times. However, the timing is no longer an artefact of tower expansion. A similar problem was observed in the Raupare Stream, where the diurnal pattern was in part a consequence of thermal expansion of the tower. For the Raupare, that expansion was confirmed from three independent sources (manual measurements at dawn, water level from Sontek IQ acoustic sensor mounted on bed, plus a submerged Hobo pressure logger). Dravid and Brown (1997) provides a detailed examination of diurnal fluctuations in groundwater level (their Section 5.3.4), which probably contributes to the diurnal fluctuations in stream flow.

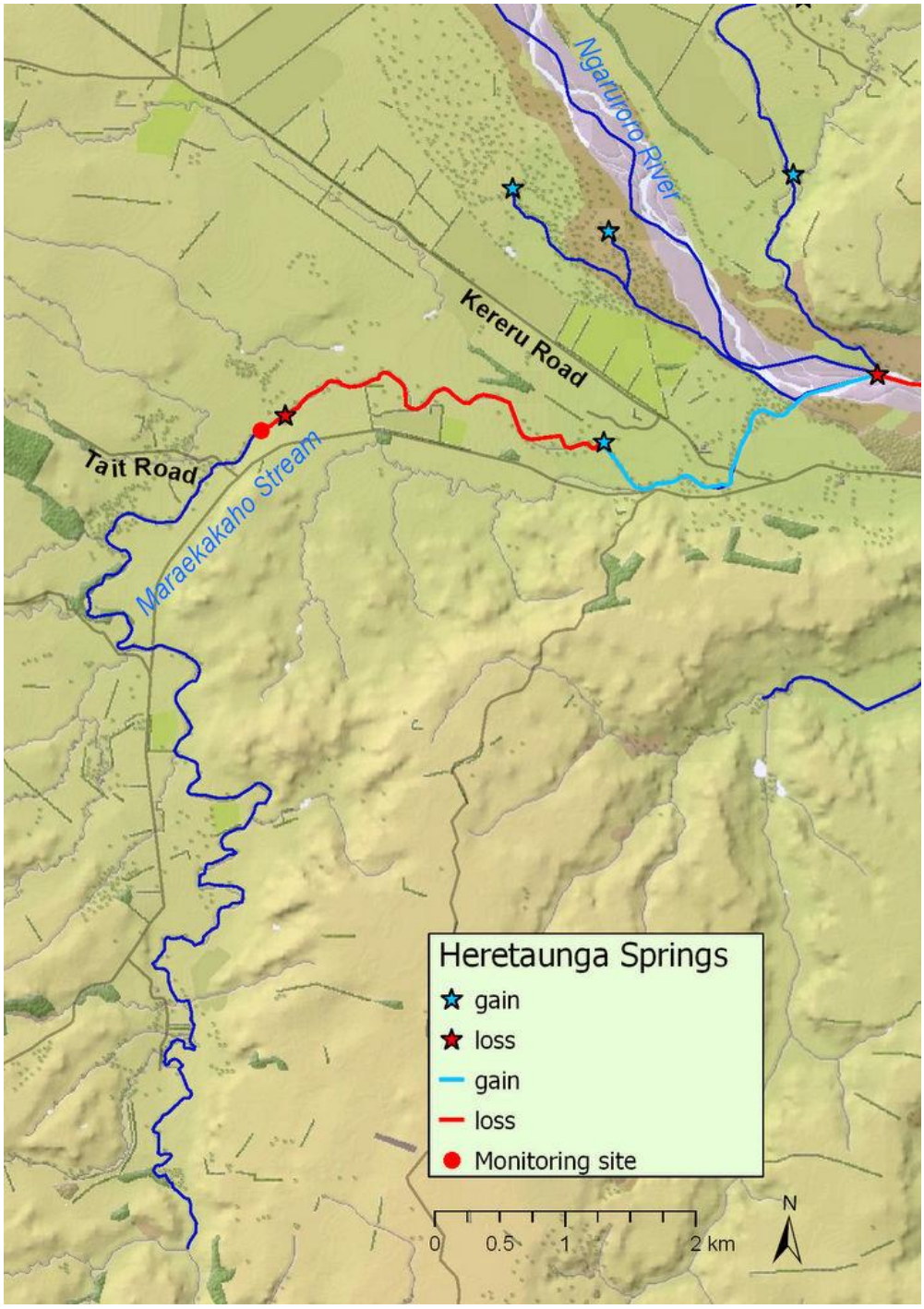


Figure 3-32: Maraekakaho Stream. The existing flow monitoring site is located downstream of Tait Rd. The section of stream that loses flow to groundwater is displayed as a red line, with the gaining reach is a light-blue line.

3.12 Waitio

This stream drains the Ohiti area on the true left of the Ngaruroro opposite Roys Hill, and gains flow from springs as it crosses the Heretaunga Plains (Figure 3-33). The MALF increases from <10 L/s as the Waitio leaves the hill country, to 570 L/s before its confluence with the Ngaruroro River. Grant (1965) estimated 480 L/s of the Waitio flow originated from the Ngaruroro. Evidence for the additional water originating from the Ngaruroro River includes a physical elevation gradient between the two (Waitio gaining section starts at 45 m elevation compared to 59 m in the Ngaruroro), separated by alluvial gravels. Further evidence comes from a decrease in electrical conductance of the Waitio, from 500 µS/cm, before reaching the plains, to 148 µS/cm before reaching the Ngaruroro confluence. The Ngaruroro has equivalent electrical conductance (median 150 µS/cm, n=50). Stable isotopes sampled from the Waitio had a $\delta^{18}O$ of -7.6‰ (Ohiti Rd, 5/3/2015), which matched the Ngaruroro sourced groundwater ($\delta^{18}O$ -7.6‰, Section 3.1).There was some correlation between Waitio flow at Ohiti Rd and groundwater levels at well 10371 (R^2 = 0.48, n = 56).

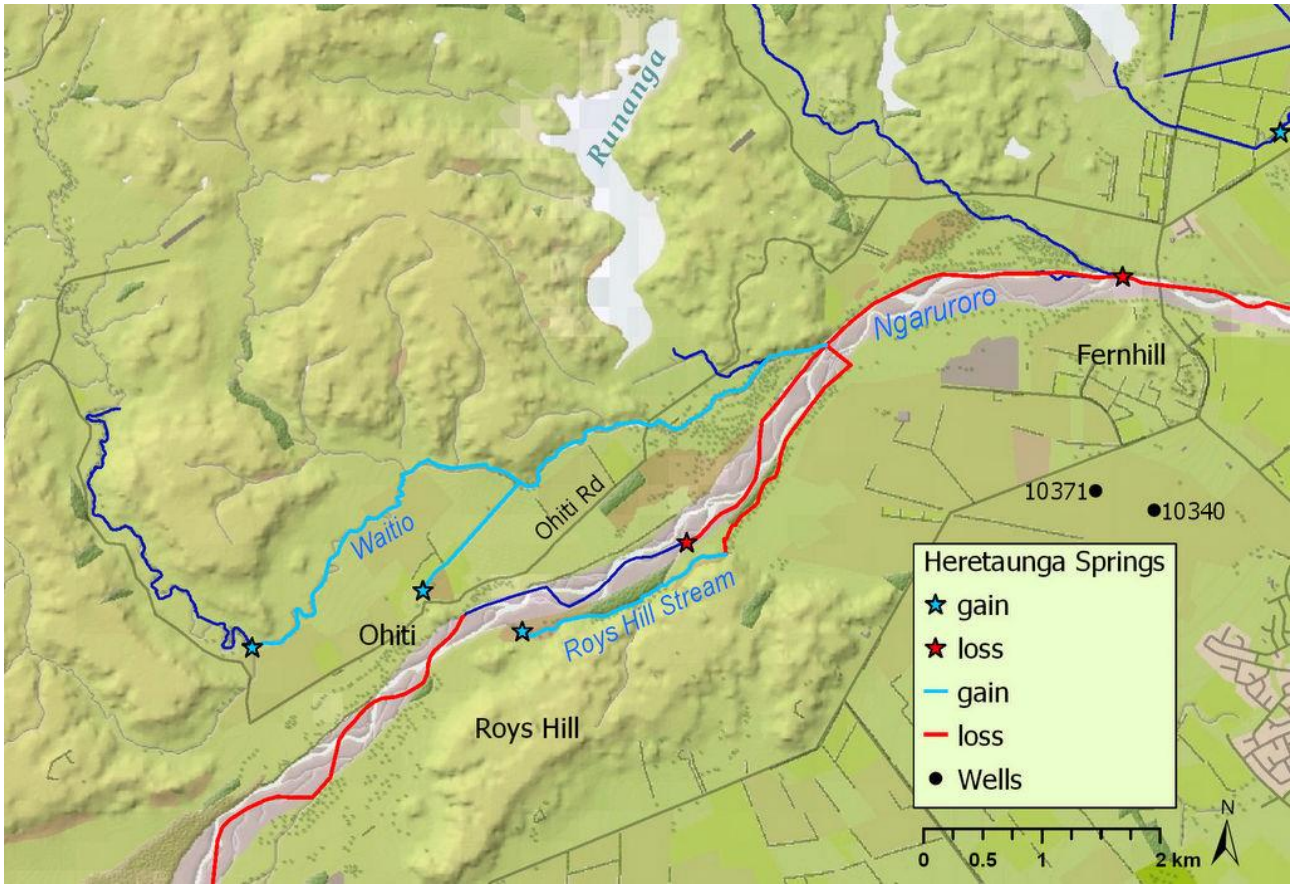


Figure 3-33: Waitio Stream. Spring inflows were detected where the stream flows across the Heretaunga Plains.

3.13 Tutaekuri-Waimate

The Tutaekuri-Waimate appears to be fed by two gravel aquifers – the Moteo Valley aquifer and the Heretaunga aquifer. Evidence for the two sources includes distinct chemistry and differing water levels at which the springs arise. Springs arise in the Moteo Valley, with electrical conductance ranging from 348 µS/cm to 411 µS/cm at 25 °C (Figure 3-34), compared to springs on the Heretaunga Plains, with conductance of 160 µS/cm at 25 °C (Korokipo Stream, 5/5/2016). The Moteo conductance data (July 2016 measurements) were mostly obtained from Levy (2016b). The Moteo springs probably originate from the Tutaekuri River upstream of Puketapu, which has an electrical conductance of approximately 300 µS/cm at 25 °C (median of 383 measurements at Puketapu and Hawke's Bay Expressway). The increase in conductance from 300 µS/cm to more than 400 µS/cm in some tributaries is consistent with addition of ions from the surrounding limestone hill-country (sandstone, limestone and mudstone of marine origin). For example, electrical conductance of 620 µS/cm at 25 °C was measured in one hill-country tributary where it reaches the Moteo Valley (Kelly Stream, 28 July 2016). This contribution could enter as tributary inflow, interaction with basement rock, or interaction with eroded limestone that is deposited in the valley alluvium (Levy, 2016b).

The Tutaekuri is connected to the Tutaekuri-Waimate via a gravel aquifer that feeds springs in the Moteo Valley (Dravid & Brown, 1997; Hughes, 2009b; Levy, 2016b). Median flow losses from the Tutaekuri were estimated at 820 L/s (see Section 3.3), which recharges the Moteo gravels. Groundwater is confined to deeper gravels by clays in the downstream (southern) part of the valley. Springs are expected to arise near the edge of that confining layer where artesian pressure becomes positive and the confining clays are thin and imperfect. The combined flow of those springs was estimated (MALF 640 L/s, median flow 750 L/s) using the correlation with the Tutaekuri-Waimate at SH50 (10 pairs of concurrent gaugings). This median flow of 750 L/s is slightly less than the median loss from the Tutaekuri River (820 L/s).

Our limited surveys to date indicate that most springs originate 2-4 km from the Tutaekuri River (Figure 3-34), (Johnson, 2010). Exceptions that arise lower in the valley were reported by Levy (2016b), including springs arising near the confluence of the Repokai te Rotoroa and Windy streams ("Nuprop springs" are labelled in Figure 3-34). Two possible sources of these springs, that are located lower in the valley, include holes through the thick layer of confining clays (Levy, 2016b), or lateral seepage through shallow Taupo pumice sands that overlay the confining clays. The pumice sands, as mapped by Griffiths (2001), terminate about where the Nuprop springs arise (Figure 3-34). The Nuprop Springs also had lower electrical conductance (348 µS/cm at 25 °C; Levy (2016b), indicating less interaction of the Tutaekuri-sourced water with limestone.

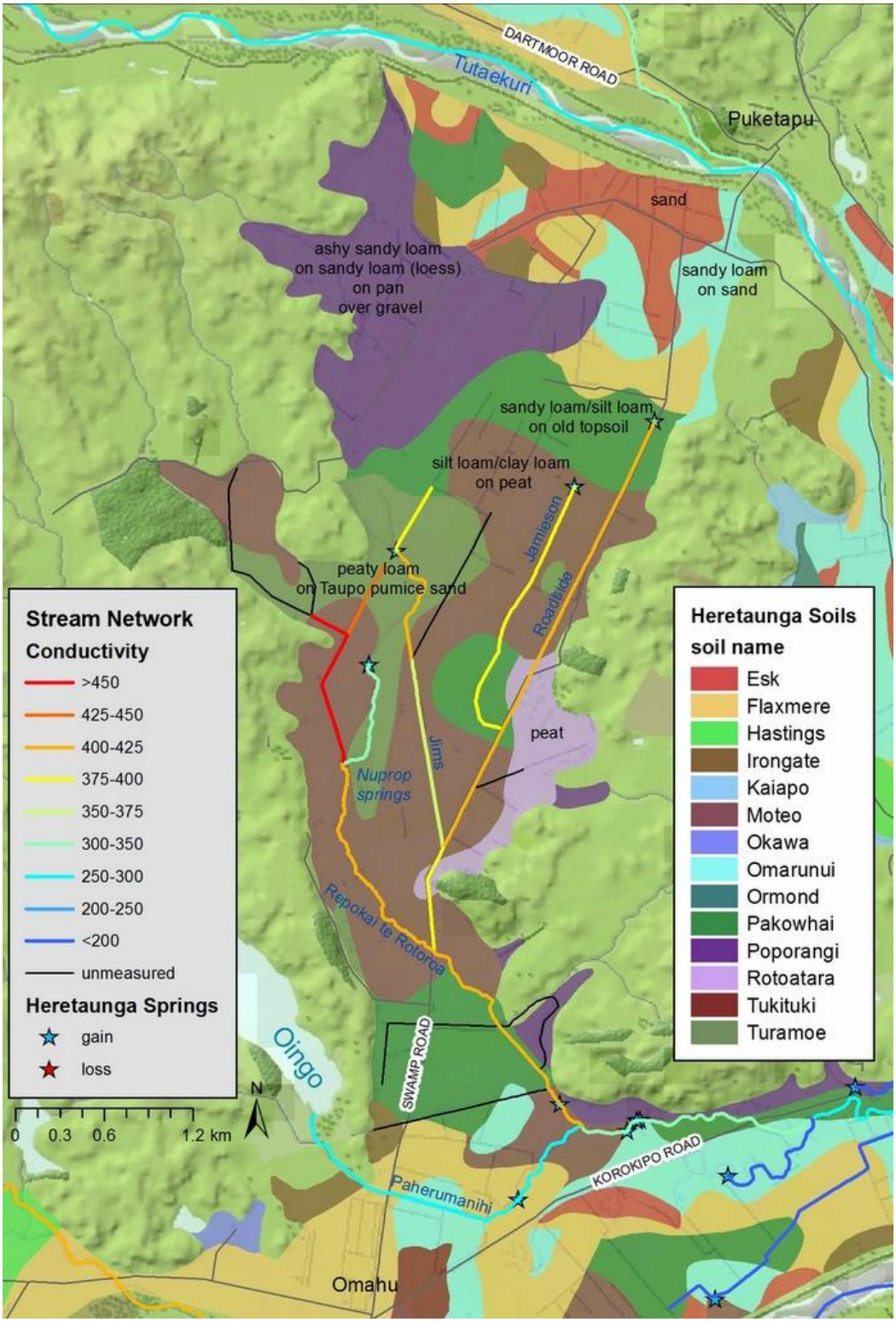


Figure 3-34: Moteo Valley springs and soils. Streams are mapped with lines colour coded using electrical conductance (µS/cm at 25 °C). Soils were mapped by Griffiths (2001), with polygons mapped and relevant properties of soil classes described in overlaid text. The stars represent the start of gains from groundwater, which are mostly diffuse springs in this area.

Stable isotopes were sampled from the Tutaekuri-Waimate Stream. However, the isotope signature of the Tutaekuri River is not clear (see Section 3.3), so its potential contribution to the Tutaekuri-Waimate via Moteo Valley springs is also unclear. Preliminary data indicate the Tutaekuri $\delta^{18}O$ is approximately -7.2‰, which places it somewhere between the likely alternative sources ($\delta^{18}O$ -7.6‰ for Ngaruroro sourced groundwater, Section 3.1; -6.0‰ for rainfall recharge, Section 2.1.5).

Moteo tributaries of the Tutaekuri-Waimate were sampled for stable isotopes in winter 2017 ($\delta^{18}O$ of -7.1‰ for Roadside Drain upstream of Repokai Te Rotoroa, Jim's Drain at Swamp Road and Jamieson Stream at Swamp Rd; $\delta^{18}O$ of -7.0‰ for Repokai Te Rotoroa at Swamp Rd, 23/8/2017). These results are close to Tutaekuri River water (-7.2‰). However, one sample collected during summer (3/3/2015) from Repokai Te Rotoroa at Swamp Rd had a $\delta^{18}O$ of -7.4‰, which is at the high end of the range for Tutaekuri, and approaching that of Ngaruroro sourced water.

Spring inputs downstream of Moteo Valley are more likely to originate from the Heretaunga Aquifer (Figure 3-35), as indicated by the lower conductance of spring-fed tributaries (e.g. Korokipo Stream conductance 160 µS/cm at 25 °C, 5/5/2016) that are a closer match to the Ngaruroro River (150 µS/cm at 25 °C, median of 50 lab samples from Fernhill). The elevation of these Heretaunga springs is also lower than the Moteo aquifer springs (9.7-13.6 m and 15.2-19.8 m respectively).

Downstream of Swamp Road, the Tutaekuri-Waimate gains more water than can be accounted for from the tributaries, and these additional gains probably arise from streambed springs within the mainstem. The springs were the focus of a GNS study using fibre optic distributed temperature sensing to quantify flow gains upstream of State Highway 50 (Moridnejad, 2015). This located several large inputs downstream of the Paherumanihi confluence. Hawke's Bay Regional Council expanded on this study, using a longitudinal survey of electrical conductance between Swamp Road and Goods Bridge. This kayak conductance survey was completed on 4 December 2015, with the methods described in Section 2.1. Spring inflows were detected, including a large gain (435 L/s) over a short distance downstream of the Paherumanihi confluence (between 2 km and 2.6 km downstream of Swamp Rd in Figure 3-37), that coincided with the springs located by Moridnejad (2015). Additionally, the kayak survey detected a gradual gain that was located downstream of State Highway 50 (325 L/s between 3 km and 4.3 km, Figure 3-37); and a rapid gain between the Waipiropiro and Waima confluences (169 L/s between 4.8 km and 5.2 km, Figure 3-37).

Hughes (2011) proposed the Heretaunga springs feeding the Tutaekuri-Waimate had originated along an unconfined/semi-confined transition area. Rabbitte (2011) expanded on this, proposing some of the springs (e.g. Waipiropiro) were fed from shallow groundwater flowing across the top of the confining layer directly from the adjacent Ngaruroro River, rather than via the Heretaunga aquifer. The mean annual low flow of the Tutaekuri-Waimate at Goods bridge was 1900 L/s, and median flow 2300 L/s (synthetic flow record 1977 to 2014 based on concurrent gaugings with the Irongate at Clarkes weir). Of this, 1200 L/s is estimated to arise downstream of Moteo Valley from the Heretaunga springs (1857 L/s minus 643 L/s).

There appears to be negligible flow-gain downstream of the Waima confluence. Concurrent gaugings between Goods Bridge and the Ngaruroro confluence (Chesterhope site) measured small differences in flow (interquartile range -14% to +3%, n=13), with an average net loss of approximately 100 L/s. Flow at Goods bridge, was correlated with the groundwater levels in well 15006 (Figure 3-36). This well is located closer to the Raupare springs (Figure 3-29), however, provided better overlap with flow records than those wells located closer to the Tutaekuri-Waimate (e.g. well 1695).

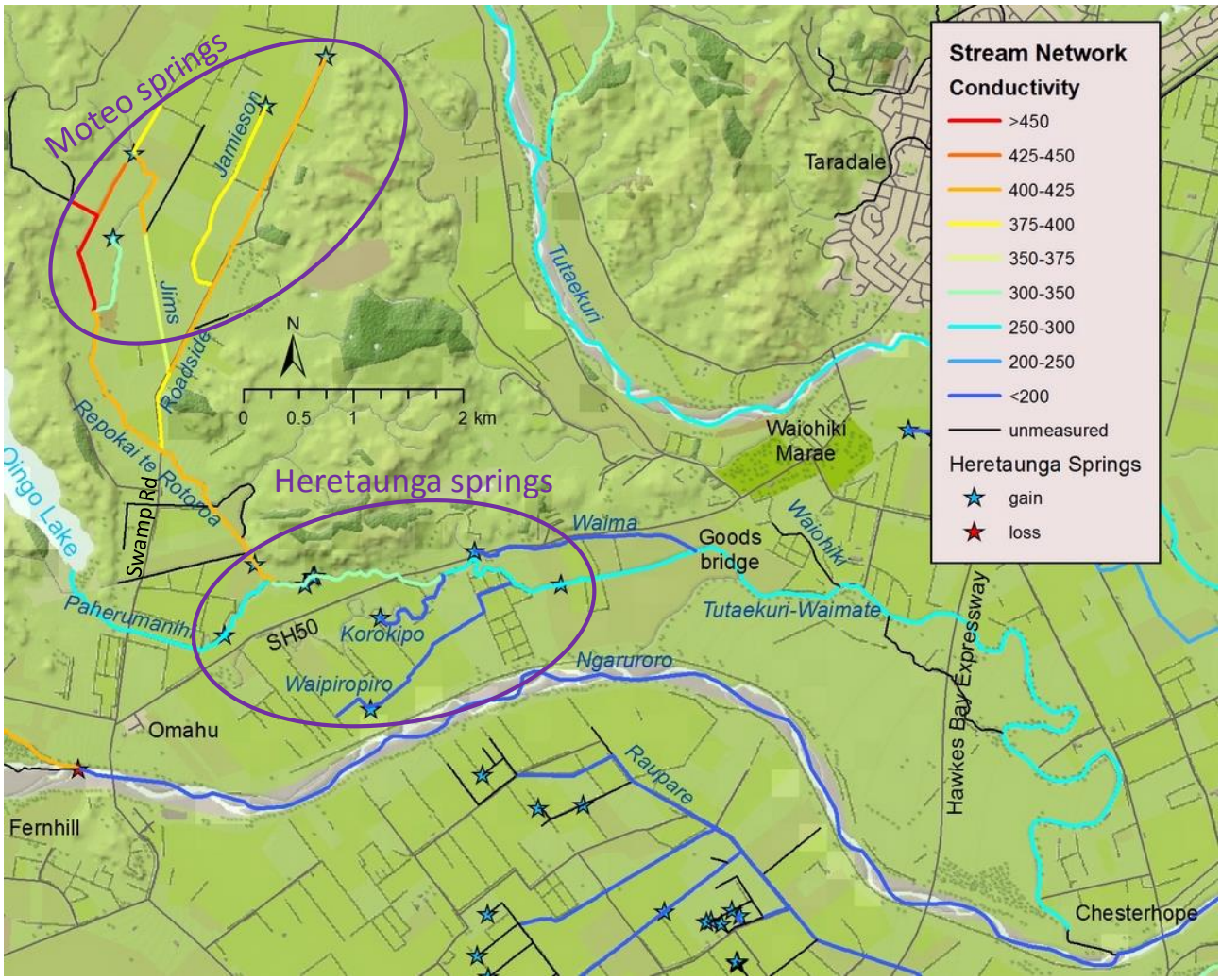


Figure 3-35: Tutaekuri-Waimate springs. Streams are mapped with lines colour-coded using electrical conductance (µS/cm at 25 °C). Lower conductance springs originate on the Heretaunga Plains (blue colours), compared to the higher conductance water from Moteo Valley (Repokai te Rotoroa exits Moteo). The stars represent spring locations that take the form of point springs or gaining sections.

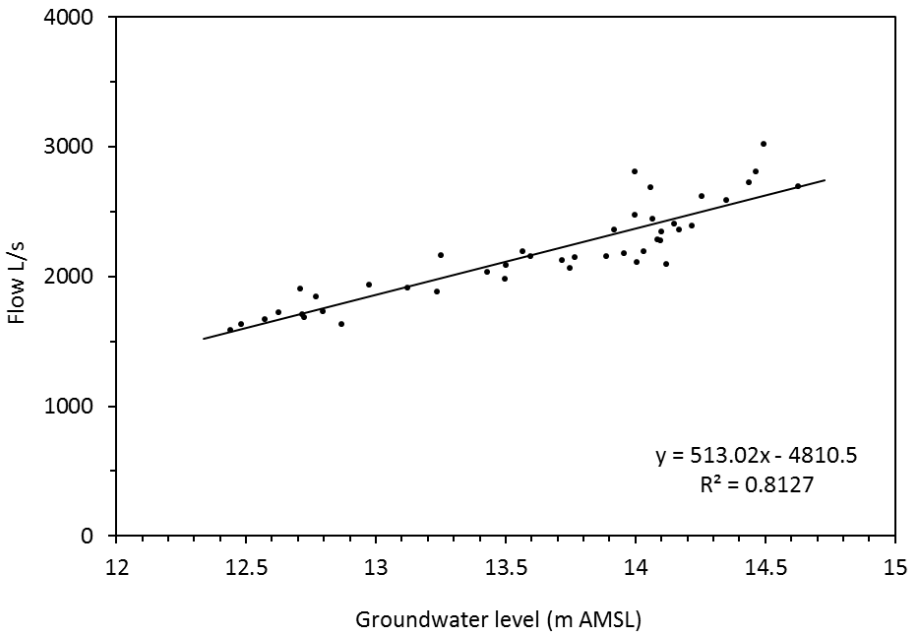


Figure 3-36: Tutaekuri-Waimate flow correlation with groundwater level. Flow measurements at Goods bridge are plotted against daily mean groundwater level, as measured at well 15006 on the day of the gauging (n=44, period 1991-1994 and 2014-2016).

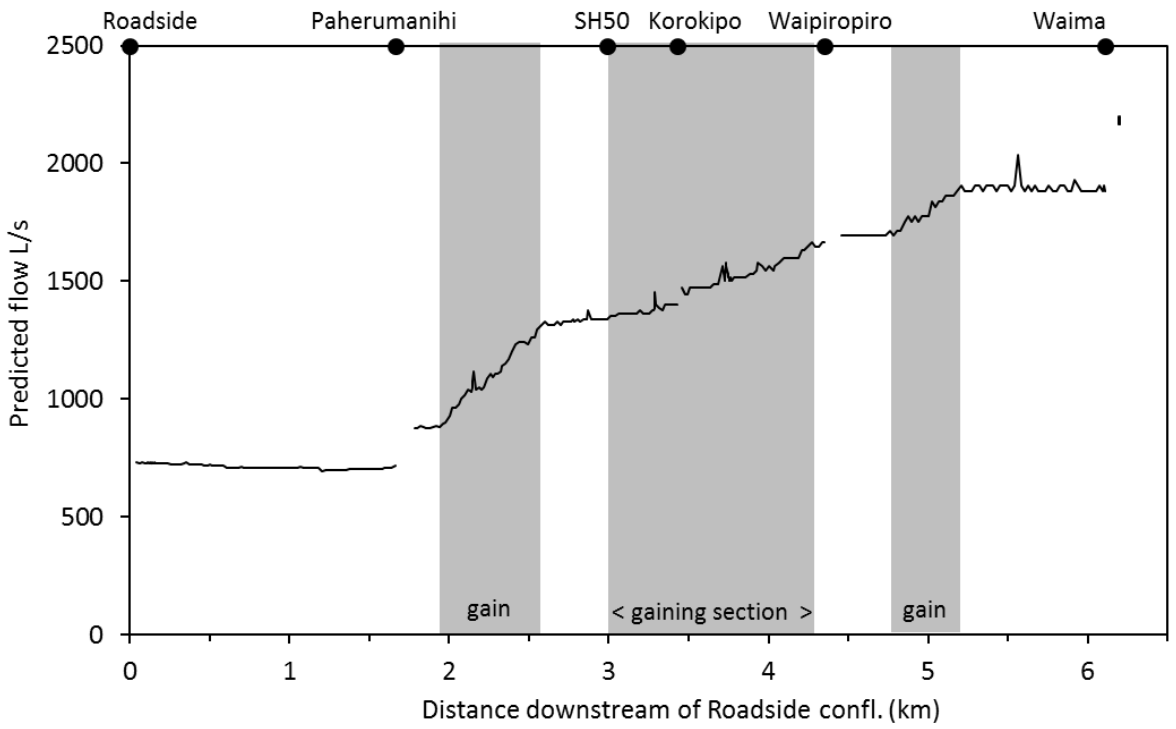


Figure 3-37: Predicted spring inflows to the lower Tutaekuri-Waimate. Change in flow along the Tutaekuri-Waimate Stream (4/12/2015), with distance measured downstream of the starting point (Roadside drain confluence) to Goods bridge (6.2 km). Spring inflows were detected using the decrease in electrical conductance by the spring water which has lower conductance than water originating upstream of Swamp Road. Flow gains were more concentrated over three sections of river, highlighted by grey columns. The location of tributary inflows and State Highway 50 Bridge is indicated by black dots. Mixing zones for the tributary inflows were omitted (gaps in black line).

The name Tutaekuri-Waimate is a clue to this river's past. Loosely translating to the dead-waters of the Tutaekuri, the Tutaekuri-Waimate occupies abandoned channels of the Tutaekuri River (except for realigned sections). There are probably two old flow-paths of the Tutaekuri occupied by the Tutaekuri-Waimate, including the Moteo Valley and the Waiohiki Stream (passing Waiohiki Marae and Links Road). A map from 1875 indicates that the Tutaekuri River occupied the Waiohiki Stream channel and carried on down the Tutaekuri-Waimate (Williams, 1987). The name Tutaekuri-Waimate is given to the mainstem, starting at the confluence of Repokai te Rotoroa Stream and Paherumanihi Stream (outlet from Oingo Lake). This nomenclature is from the LINZ gazeteer, which contradicts the Topomap (NZMS260) that extends the name Paherumanihi further downstream toward Waiohiki.

The Moteo Valley springs that now arise through the drainage networks, have historically sustained wetlands (www.rotokare.co.nz). Evidence of those wetlands is still found in the place names, including "Swamp Road" that crosses "Repokai te Rotoroa" Stream (literal translation - swamp food of the long lake). The physical setting for wetlands remain, with the valley-floor sloping backwards away from the Ngaruroro River toward Repokai te Rotoroa Stream (from LiDAR elevation maps). The obstructed valley creates a depositional environment where peat soils and fine sediments dominate (Figure 3-34). This geomorphic setting, which enables wetland development, is repeated around the border of the Heretaunga Plains where confined valleys are obstructed at their outlet by Ngaruroro sediments (e.g. Moteo, Oingo, Turamoe, Runanga, Louisa, Paritua, Rotokare).

The conversion of wetlands to agriculture required drainage schemes (www.rotokare.co.nz). Floods overwhelmed the Moteo drainage scheme for a short period in 1974, recreating a Moteo wetland. Several of these drainage systems now have resource consents to use pumped subsurface drainage (DP020244W, DP030219W, DP040216W, DP000305W).

3.14 Other small streams

There are many other small streams on the Heretaunga Plains that are not detailed in this report. These streams were mapped to capture their spatial extent, using aerial photographs and Google Street View. This section provides limited additional information for eight of those small streams.

Kikowhero

The headwaters of the Kikowhero Stream drain hill country that receives more rainfall than the Heretaunga Plains (>1000 mm annual total, compared to about 800 mm for the plains). Much of the stream flow originates from that wetter hill country. After leaving the hill country, the stream path then follows the border between the Ngaruroro river terraces and the hill country. The Kikowhero flows through a small gully with the stream-bed cutting down into white sedimentary rock in places, interspersed by some gravel substrates. No increase in flow was detected over the long section of stream flowing alongside the river terraces (11 concurrent gaugings 1992 to 2015; median 21 L/s at Crownthorpe Rd and 18 L/s at Omapere Rd).

A loss of flow likely occurs downstream of Omapere Rd, where the Kikowhero turns and crosses the river terraces to reach the Ngaruroro River. This 2 km of losing-reach can dry completely, as evidenced by a dry braided channel visible from aerial photos. This losing reach is also elevated above the Ngaruroro River (from LiDAR elevation data). Springs emerge at the end of the drying reach, marking the start of the gaining section (visited 5 December 2014). The start of the gaining reach coincides with the channel elevation dropping to the same level as the Ngaruroro River. The magnitude of that flow gain prior to the Ngaruroro confluence has not been quantified. Channel substrate appears to be gravel through the losing and gaining reaches, transitioning from white sedimentary rock somewhere upstream of Matapiro Road.

The only channel modifications noted were in the losing reach, where there are signs of gravel extraction from the dry braided bed, plus a pond constructed for jet boat racing.

Ohiwia

No springs were detected that originate from the Heretaunga aquifer. Evidence for limited interaction includes negligible drop in electrical conductance (416 µS/cm to 409 µS/cm between Kawera bridge and Broughton's bridge, 5/12/2014). There was also little change in flow detected as the stream crossed the Heretaunga Plains (MALF increased from 135 L/s to 169 L/s between Kawera bridge and Broughton's bridge). However, this needs confirmation, given that Grant (1965) concluded that flow losses from the Ngaruroro Rives contribute approximately 85 L/s to the Ohiwia. Surface water interactions with groundwater were not investigated within the smaller Okawa basin.

Roys Hill

This stream runs adjacent to the Ngaruroro River, with flow an artefact of channel excavation that dropped the bed level below the groundwater level. The location where spring inflows start was mapped from aerial photographs (Figure 3-33). The flow was originally diverted into an artificial groundwater recharge scheme. However, that scheme no longer operates (Brooks, 1999; Gordon, 2009). The channel now runs back toward the Ngaruroro River. It is possible that flow returns to the ground before reaching the Ngaruroro, given that it traverses gravels adjacent to the major loss reach of the Ngaruroro River.

Grange

Little is presently known of the hydrology of this stream. It arises from springs adjacent to the Tukituki River within 5 km of the coast, and flows to its confluence with the Tukituki estuary at Haumoana (Figure 3-38). The springs could be sourced from shallow groundwater fed by flow losses from the Tukituki River (see Section 3.2). Evidence for this link with the Tukituki includes the proximity to the end of the losing reach of the Tukituki, with alluvial gravels enabling groundwater movement from the river to the Grange springs. The Tukituki River is only 0.5 km from the nearest Grange springs. Also, the electrical conductance of the springs feeding Grange Stream is lower than streams originating from limestone hill country (180 µS/cm on 12/12/2014, compared to Awanui flume interquartile range of 670-771 µS/cm). A single flow gauging was completed for Grange Stream (at its most downstream crossing of Tuki Tuki Road), measuring 4 L/s on a day when the Tukituki River was flowing at less than 60% of MALF (3433 L/s gauged at Red Bridge, 20/1/2017).

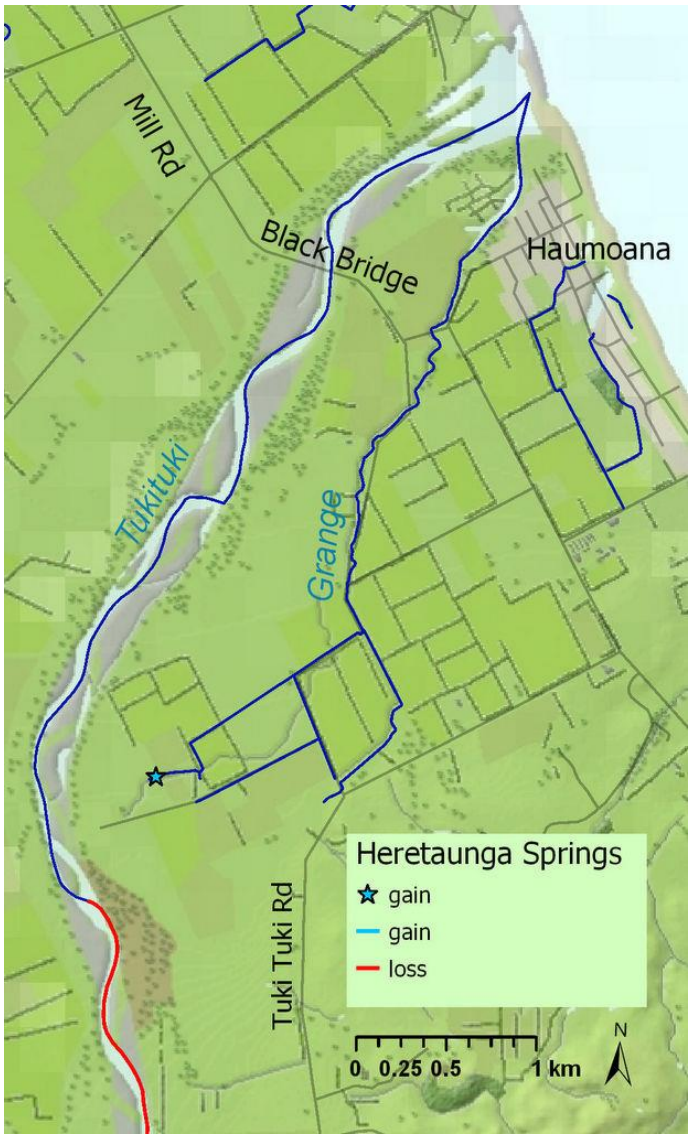


Figure 3-38: Grange Stream. Grange Stream arises close to the end of the losing reach of the Tukituki River, which may be the source of water feeding springs on Grange Stream.

Karituwhenua

The Karituwhenua Stream was walked on several occasions. The stream was dry for the entire length through Havelock North on at least one occasion. However, more often the stream was flowing upstream of Te Mata Road and gradually lost flow before drying completely at some point downstream of Te Mata Road. The channel is incised within old river terraces, and intersects a layer of gravel about where it appears to lose flow. A losing reach was mapped based on these observations (Figure 3-24).

Ruahapia

The source of flow for this stream was not identified. The gauging record for the Ruahapia Stream indicates a reduction in flow from a median of 102 L/s pre 1986 to 21 L/s post 2009 (n = 17 and 8, respectively). This reduction could have occurred any time during a gap in the gauging record (1987 to 2008).

Heinz Watties Limited hold a consent to discharge 100 L/s of cooling water to the Ruahapia Stream (Consent No. DP110144Wa). Improved efficiency of cool store operations could explain the reduced outflow. More importantly, it suggests the bulk of the stream flow was, and perhaps still is, groundwater used for cooling. The source of this stream may not be spring flow. But, given the fish kills (e.g. 11/6/2015), sewage fungus and objectionable odour as it passes through Hawke's Bay Show Grounds, the source of discharges to the Ruahapia flow deserves further investigation. In this respect, it is possible that the larger discharges of cooling water are beneficial in diluting the contaminants from smaller, more hazardous sources (*pers. comm*., Armann Einarsson, Hastings District Council, 31/7/2015).

Turirau and Wharerangi

The Turirau is a tributary of the Tutaekuri River (Figure 3-11). The headwaters of the Turirau were diverted out of the Tutaekuri catchment into the Ahuriri Estuary via a channel that was dynamited out of a hill. The headwaters are called the Wharerangi Stream, which flows through a small valley before reaching the diversion into the Ahuriri. Spring inputs to the Turirau and Wharerangi have not been investigated.

Georges

This small urban stream is sustained by saline groundwater. Electrical conductance around 10,000 µS/cm (compared with >50,000 µS/cm for seawater and <200 µS/cm for Heretaunga aquifer springs) was measured upstream of the floodgates that block tidal penetration (Kennedy Rd). Georges Stream is separated from the Heretaunga gravel aquifer by a thick confining layer. A geological model (Lee *et al.*, 2014) reveals a lens of beach gravel and sand on the true-right bank that could connect Georges Stream with the ocean (Figure 3-39). Local rainfall recharge from the central Napier area would dilute any seawater influx. From 1897 to 1931, this stream channel was occupied by the Tutaekuri River making its way to Ahuriri Estuary (page 45 in Dravid & Brown, 1997). The 1931 earthquake lifted the land sufficiently to compromise drainage of Tutaekuri flood waters to the Ahuriri Estuary, and the river was diverted to the Waitangi Estuary by 1940 (Williams, 1987).

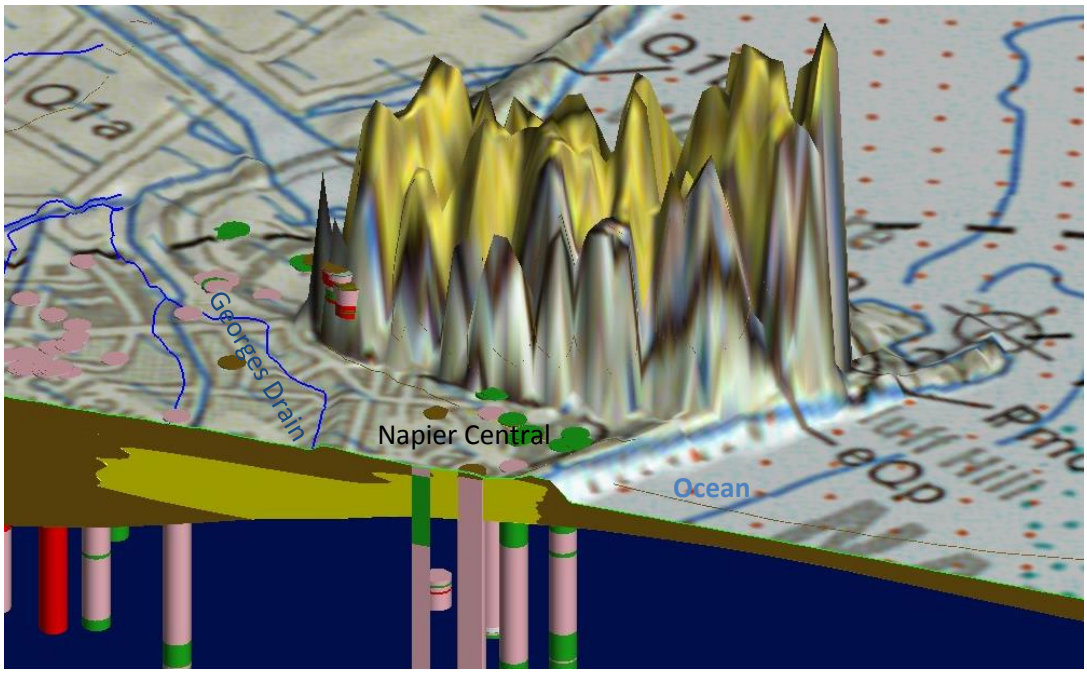


Figure 3-39: Napier Geology. Oblique view of Napier, with exaggerated relief. Beach gravels underlying Napier are shown in yellow, which may offer a groundwater connection between Georges Stream (blue line) and the ocean (orange dots).

4 Synthesis of results - springs of the Heretaunga Plains

This section brings together the results from the individual sub-catchments to provide a broader synthesis of the surface water – groundwater interactions on the Heretaunga Plains. Many simplifications are required to distil the complex and dynamic surface water – groundwater interactions into a more digestible form. This section focuses on the low-flow season when groundwater levels are low and soil moisture is low, and use of water resources is highest. Therefore, Figure 4-1 portrays the losses and gains as a static estimate at mean annual low flow. These patterns should not be taken as representative of cool wet periods when flow inputs will extend over a larger area and the source of groundwater can change. Starting with inputs to the Heretaunga system, the greatest loss of river flow was from the Ngaruroro River (Figure 4-1, Section 3.1). Compared to other sources, the Ngaruroro sourced groundwater was characterised by lower dissolved ions and stable isotopes that are more negative than the Tukituki and Tutaekuri. The consistency of the flow loss from the Ngaruroro River keeps the aquifer recharged and the springs flowing through the dry Heretaunga summers when there is little rainfall recharge. The Ngaruroro contribution is vital to ecosystems and water users of the Heretaunga Plains. In particular, the Raupare (Section 3.10), Tutaekuri-Waimate (Section 3.13), Waitio (Section 3.12) and Irongate (Section 3.7) streams are probably fed by groundwater originating from the Ngaruroro.

The Tukituki (Section 3.2) and Tutaekuri rivers (Section 3.3) also lose large volumes of water to groundwater aquifers that sustain spring flows (Figure 4-1). For example, results from this study indicate that Tukituki-sourced groundwater contributed more to the mean annual low flow of the Karamu Stream than all other sources combined (Section 3.8). The Karamu becomes the Clive (or Whakaora te Wai) at the confluence with Raupare Stream and, at this point, the amount of groundwater originating from the Ngaruroro approaches that of the Tukituki. Understanding the Tukituki sourced groundwater is therefore of equal importance for understanding the effects of groundwater use on the Karamu and Clive streams.

The Karamu and Tutaekuri-Waimate are the two largest spring-dominated streams on the Heretaunga Plains. Springs feeding the Tutaekuri-Waimate appear to be dominated by Ngaruroro sourced groundwater during low flows (Section 3.13). Losses from the Tutaekuri also make a significant contribution to the Tutaekuri-Waimate, particularly for springs arising in Moteo Valley. However, the stable isotope results were inconclusive, and further investigation of the source of springs feeding the Tutaekuri-Waimate is recommended.

The contribution of direct rainfall recharge to groundwater diminishes in summer on the Heretaunga plains. Lysimeters were used to measure how much rainfall makes it through the soil (after evaporation and runoff), to recharge groundwater. These have demonstrated that rainfall recharge is small to negligible in hot dry summers on the Heretaunga plains. However, rainfall is an important part of the water balance. For example, springs feeding the Karewarewa Stream (Figure 4-1) may be recharged by local rainfall (Section 3.4).

An extensive area of shallow Taupo pumice sands contributes groundwater to several streams, including the Karewarewa, Louisa (Section 3.6) and Awanui (Section 3.5). Little is known of the groundwater in this pumice sand layer because it is not used for irrigation or domestic water supply. However, it may be an important source of flow and nutrients for these streams, and hence deserves further investigation.

These investigations also revealed a tufa coating (calcite sourced from limestone) across the bed of the Paritua Stream (Section 3.4). The Paritua Stream has run dry in summers past. This stream loses flow to groundwater where it crosses unconfined alluvial gravels upstream of Bridge Pa. The tufa coating is therefore important in extending the length of flowing stream because it probably reduces the rate of flow loss to groundwater.

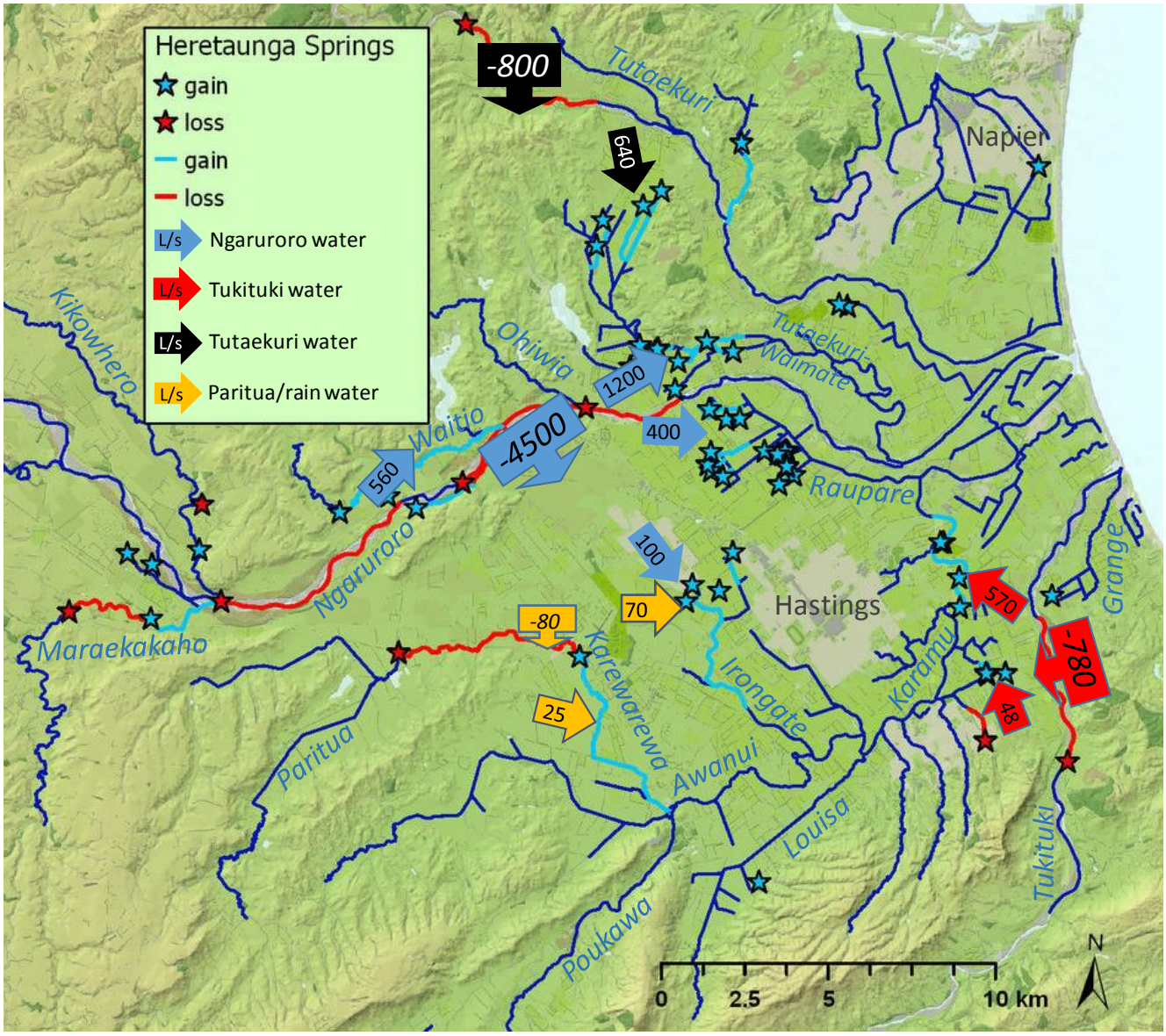


Figure 4-1: Flow gains and losses on the Heretaunga Plains. The various sub-catchments reported in the results are mapped together in this overview. Stream reaches that lose flow to groundwater are represented as red lines, with gaining reaches mapped light blue. Many simplifications are required to distil the complex and dynamic surface water / groundwater interactions into one map. For example, the dynamic losses and gains are presented as a static estimate at mean annual low flow. Arrows are labelled with flow gain in L/s, with flow losses presented as a negative value. Note that the Ngaruroro flow loss (-4500 L/s) combines both the major-loss and variable-loss reaches. Colour coding of the arrows is indicative of dominant source of water during the low-flow season.

The Irongate Stream (Section 3.7) probably receives groundwater originating from the Ngaruroro in addition to large contributions from rainfall recharge. A shallow layer of gravel, above the clay confining layer, appears to be an important pathway for the diffuse springs that feed the Irongate Stream.

This report describes the known springs and flow losses on the Heretaunga plains. Investigations for this report focussed on the major gains and losses for the defined sub-catchments of the Heretaunga plains (Section 1.1). Many springs will have been missed, and some of these may be of local significance. However, the undescribed springs are expected to contribute less to the sub-catchment flows during the low-flow season. Springs discharging to the marine area could be significant for the catchment water balance, but were outside the scope of this investigation.

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Thanks to Stephen Swabey for identifying and explaining the tufa deposits in Paritua Stream, and helping with kayak surveys of the Karamu and Tutaekuri-Waimate. Thanks to Grant Upchurch for further investigating the tufa deposits and relevant soil properties of the Paritua and Mangateretere.

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Appendix A Spring locations and attributes

This table lists the grid reference and recorded attributes of springs mapped to date across the Heretaunga Plains. Included are locations where losses of flow to groundwater were determined. All grid references are the start location, with the length attribute indicating how far downstream the loss or gain extends. This list is not exhaustive, instead capturing the major inputs that were encountered during this investigation.

SpringID	NZTM Easting	NZTM Northing	Stream	MALF L/s	Length m	Width m	Elevation m	Permanence	Morphology	LossGain	FlowQual	Comments
1	1928734	5610040	Burns	10	5	5	5.1	Permanent	Point source	gain	visual	Field surveyed
2	1928567	5609961	Burns	10	7	7	5.2	Permanent	Point source	gain	visual	Field surveyed. Litres per second visual only.
3	1928475	5609991	Burns	40	5	5	5.3	Permanent	Point source	gain	visual	Field surveyed. Spring plus large pipe inflow (tile drains?). Sum E
4	1928661	5610086	Burns	40	4	4	5.3	Permanent	Point source	gain	measured	Field surveyed
5	1926900	5611015	Centre	17	0.5	0.5	10.4	Permanent	Point source	gain	measured	Field surveyed, flow balance Carrick Rd minus Twyford Rd
6	1926391	5611321	Raupare	107	250	1	11.9	Permanent	Channel	gain	visual	Field surveyed
7	1927305	5611047	Centre	8	2	2	9.6	Permanent	Point source	gain	visual	Field surveyed
8	1926439	5609477	Flowers	35	300	0.5	12.1	Permanent	Channel	gain	visual	Field surveyed
9	1926341	5609669	Twyford	22	1000	0.5	12.7	Permanent	Channel	gain	visual	Field surveyed
10	1926436	5610047	Twyford	14	200	0.5	11.7	Permanent	Channel	gain	visual	Sited at downstream confluence. Estimate source from aerial ph
11	1926801	5609307	Flowers	8	300	0.5	11.7	Permanent	Channel	gain	visual	Field surveyed
12	1934713	5603376	Crombie	10	0.5	0.5	7.1	Permanent	Point source	gain	visual	Field surveyed 21052014 and 11072014.
13	1935249	5603431	Mangateretere	34	600	3	7.7	Permanent	Channel	gain	visual	Field surveyed. Sum flow at Thompson Rd 180Ls.
14	1933401	5607344	Karamu	20	15	10	1.9	Permanent	Point source	gain	measured	Field surveyed. Spoke to Kit Halford: "Spring always there and n
15	1933884	5605400	Karamu	80	700	12	2.2	Permanent	Channel	gain	measured	Derived from kayak conductivity study (fine spatial resolution).
16	1925693	5605583	Irongate	136	5156	0.5	11.2	Permanent	Channel	gain	visual	Field survey start of water.
17	1927097	5607053	Southland	5	1100	0.5	10.9	Permanent	Channel	gain	measured	Field survey start of water.
19	1924414	5617418	Jamieson	107	1000	0.5	19.8	Permanent	Channel	gain	measured	Field survey start of water.
20	1924961	5617864	Roadside	75	2300	0.5	22.9	Permanent	Channel	gain	measured	Field survey plus aerial photo.
21	1923212	5616984	Jims	283	2000	2	16.5	unknown	Unknown	gain	measured	Estimate from aerial photo
22	1933865	5606286	Karamu	225	1000	18	2.1	Permanent	Channel	gain	visual	Derived from kayak conductivity study (fine spatial resolution).
23	1925458	5612753	Korokipo	64	1000	5	13.9	unknown	Channel	gain	measured	Dougall G spoke to B Kilmister - confirmed pond as a start point.
24	1924860	5613130	Tutaekuri Waimate	410	2000	5	12.3	Permanent	Channel	gain	measured	DTS results from GNS (Stew Cameron). Kayak conductivity 0412
25	1928045	5610072	Flowers	71	700	1.5	7.2	Permanent	Unknown	gain	measured	12 Ls Trotter Rd to 51Ls 12Feb. Point (aerial photo) where weed
26	1933309	5607350	Karamu	250	1500	20	1.9	Permanent	Channel	gain	visual	Derived from kayak conductivity study (fine spatial resolution).
27	1934671	5603429	Mangateretere trib	2	0.3	0.3	7.0	Unknown	Point source	gain	measured	Field surveyed 11 July 2014. Red colour water (iron flocc). Small
28	1922514	5603933	Karewarewa	25	4000	4.5	15.0	Unknown	Channel	gain	measured	Where Karewarewa transitions from losing to gaining, from con
29	1924315	5613239	Tutaekuri Waimate	27	160	6	13.6	Permanent	Channel	gain	measured	DTS results from GNS (Stew Cameron).
30	1924770	5613056	Tutaekuri Waimate	99	6	6	12.7	Permanent	Point source	gain	measured	DTS results from GNS (Stew Cameron).
31	1924840	5613128	Tutaekuri Waimate	63	6	6	12.4	Permanent	Point source	gain	measured	DTS results from GNS (Stew Cameron).
32	1936241	5618593	Harakeke (Plantation)	10	0.2	0.2	-0.1	Permanent	Artificial point	gain	measured	HBRC well to supplement stream flow (well 15918)
33	1919015	5609132	Ngaruroro major loss	-4200	4834	100	21.8-37.2	Permanent	Channel	loss	measured	Boundary interpolated from gaugings and topography

SpringID	NZTM Easting	NZTM Northing	Stream	MALF L/s	Length m	Width m	Elevation m	Permanence	Morphology	LossGain	FlowQual	Comments
34	1922699	5611375	Ngaruroro variable loss	-50	3000	70	15.5-21.8	Permanent	Channel	loss	measured	Varies between + or - 500L/s (losing and gaining). Boundary inte
35	1911787	5605586	Ngaruroro minor loss	-150	6500	100	68.5	Permanent	Channel	loss	measured	Boundary interpolated from gaugings, uncertain length
36	1917099	5604045	Paritua losing	-100	7500	5.2	43.8	Permanent	Channel	loss	measured	From geomorphology. Calcite bed seal possibly variable over tim
37	1937105	5600789	Tukituki losing	-780	4800	70	7.6-15.3	Permanent	Channel	loss	measured	Boundary interpolated from gaugings
38	1919103	5622844	Tutaekuri losing	-800	5500	53	31.7-45.0	Permanent	Channel	loss	measured	From 34 concurrent gaugings up to 27032013. Reach length fror
39	1928714	5609601	Gudsell	1	15	15	5.2	Permanent	Point source	gain	measured	Evenden rogue well 1 (5.7 L/s total for all 3 heads)
40	1928919	5609357	Gudsell	1	4	5	5.2	Permanent	Point source	gain	visual	Evenden spring (appeared when removing apple tree stumps)
41	1928430	5609979	Burns	15	7	7	5.5	Permanent	Point source	gain	measured	Possible spriing - not surveyed. Flow is balance of MALF.
42	1928718	5609614	Gudsell	1	7	7	5.2	Permanent	Point source	gain	measured	Evenden rogue well 2
43	1928706	5609617	Gudsell	1	4	4	5.2	Permanent	Point source	gain	measured	Evenden rogue well 3
44	1928478	5609077	Gudsell	1	100	1	7.0	Permanent	Channel	gain	visual	Gudsell upstream spring
45	1911142	5607123	Kikowhero				76.0	Permanent	Channel	gain	measured	From conductivity survey 05122014 first water downstream of c
46	1916784	5608726	Ohiti	68	1230		42.8	Permanent	Channel	gain	measured	Inspected at confluence with Waitio, but source from aerial pho
47	1915341	5608245	Waitio	498	5000		45.5	Permanent	Channel	gain	visual	Inspected 500m downstream at which point conductivity had dr
48	1930523	5614408	Tattersal	5	7	7	5.8	Permanent	Point source	gain	visual	Pipe to stream surveyed 19122014. Source not checked but assu
49	1930276	5614463	Tattersal	5			6.1	Permanent		gain	visual	Pipe flowing to stream 19122014. Source unknown. Suspect spri
50	1927353	5619294	Turirau		500		18.4	Unknown	Channel	gain	visual	springs inferred from conductivity and elevation (uncertain). L K
51	1936639	5605752	Grange		500	1.9	6.8	Permanent	Channel	gain	measured	Surveyed 12122014 by SF. Water backs up 40m, but conductivit
52	1923023	5616214	Repokai te Rotoroa	145	1000	2	15.2	Permanent	Channel	gain	visual	From aerial photo.
53	1934645	5601415	Karituwhenua	-5	1000	2.5	39.0	Intermittent	Channel	loss	measured	Visual transition from flowing to dry while out walking. Length u
54	1925884	5606079	Wellwood	27	650	3	13.2	Permanent	Channel	gain	visual	Inflows start upstream. Point of outflow from Flaxmere stormwa
55	1917624	5608382	Roys Hill	50	1200	3	41.1	Permanent	Artificial channel	gain	visual	Excavated to harvest water for recharge scheme then frost prot
56	1909726	5606687	Kereru		500	7.4	77.5	Permanent	Channel	gain	visual	From aerial photo and elevation vs Ngaruroro.
57	1908990	5607017	KereruTrib		1300	6.75	80.1	Permanent	Channel	gain	measured	From aerial photo and elevation vs Ngaruroro.
58	1909690	5605068	Maraekakaho	120	2000	20	80.6	Permanent	Channel	gain	measured	Christie 2010 reported start of flow here (below dry section). Co
59	1907246	5605271	Maraekakaho	-120	3200	20	98.0	Intermittent	Channel	loss	visual	Christie 2010 reported end of loss. I estimated start of loss at en
60	1911218	5608481	Kikowhero	-20	1360	20	87.6	Permanent	Channel	loss	measured	Drying reach visible in aerial photo. Spring start surveyed. Loss s
61	1925367	5611916	Waipiropiro		60	2	14.3	Permanent	Channel	gain	visual	Compliance officer noted this one on Waipiropiro when investig
62	1927104	5613054	Tutaekuri Waimate	148	350	5	9.8	Permanent	Channel	gain	measured	Kayak conductivity survey 04122015
63	1926318	5613356	Waima	245	2000	3	11.5	Permanent	Channel	gain	measured	MALF flow estimated from concurrent gaugings
64	1924187	5612509	Paherumanihi	140	600	2.6	16.5	Permanent	Channel	gain	measured	Location from aerial photos. Assume Oingo outlet dry in summe
65	1927877	5597187	Louisa			1.5	10.6	Permanent	Channel	gain	measured	A Levy recorded for WP170001T

Appendix B Stable Isotope data

This table lists the individual results from stable isotope samples collected from springs, streams and groundwater wells. Included are locations (New Zealand Transverse Mercator easting and northing), the date the sample was collected, as well as electrical conductance and stream flow. For groundwater wells, the depth range over which the well is screened is provided (or total depth of well in parentheses if screen interval was not defined). Methods are described in Section 2.1.5.

Site Name	NZTM East	NZTM North	Date	$\delta^{18}O$	$\delta^{2}H$	Electrical conductance (µS/cm 25°C)	Stream flow L/s	Well screen-interval m
Tukituki at Red Bridge	1936570	5596260	4/3/2015	-6.09	-37.5	175	4138	
Papanui at Walker Rd	1911310	5569834	17/4/2015	-6.73	-41.0	215	57	
Kaikora at College Rd	1906682	5578966	17/4/2015	-6.03	-37.8	774		
Te Aute School at Spring	1908730	5585731	17/4/2015	-7.01	-43.2	624		
Kaikora U/S Papanui	1915456	5579780	17/4/2015	-5.85	-36.9	726	83	
Papanui at Newmans ford	1915530	5579735	17/4/2015	-6.36	-38.3	301	50	
Papanui at Middle Rd	1917832	5581536	17/4/2015	-6.00	-37.6	488	126	
Papanui at Camp David	1919167	5584050	17/4/2015	-6.35	-38.3	317	523	
Karamu at Havelock Nth bridge	1932190	5602020	3/3/2015	-6.19	-41.2	460	219	
Karamu at Floodgates	1932710	5609270	3/3/2015	-6.21	-39.9	276	808	
Poukawa at Stock Rd	1925313	5598912	3/3/2015	-6.35	-40.8	497	15	
Paritua at Water wheel	1914020	5601310	4/3/2015	-5.98	-38.8	931	35	
Karewarewa at Pakipaki	1925160	5599310	3/3/2015	-5.83	-38.6	769	12	
Te Waikaha at Mutiny Rd	1926137	5595551	5/3/2015	-6.80	-41.4	499	33	
Awanui at Flume	1925735	5599654	5/3/2015	-6.02	-38.6	651	42	
Spring 22 D/S SH2	1933499	5606783	12/5/2016	-7.52	-47.2	267		
Spring 14 at Golflands	1933401	5607331	12/5/2016	-7.56	-46.1	165		
Mangateretere at Napier Rd	1933819	5604172	3/3/2015	-6.38	-40.9	270	38	
Spring 13 at Brookvale Rd	1935190	5603427	6/5/2016	-6.51	-39.0	225		
Spring 12 Mangateretere trib.	1934715	5603375	6/5/2016	-6.32	-38.8	417		
Irongate at Clarkes weir	1926737	5605025	4/3/2015	-6.94	-44.6	210	36	
Raupare at Ormond Rd	1929836	5609665	4/3/2015	-7.71	-47.6	165	356	
Raupare at Twyford Rd	1926730	5611316	3/3/2015	-7.54	-46.8	155	50	
Raupare Rd Spring	1928734	5610040	3/3/2015	-8.03	-50.7	169		
Ngaruroro at Whanawhana	1891976	5615740	4/3/2015	-7.26	-46.8	163	5388	
Ngaruroro D/S Maraekakaho	1911787	5605586	3/3/2015	-7.35	-45.9	171	5743	
Ngaruroro at Chesterhope bridge	1932547	5610113	4/3/2015	-7.40	-45.2	207	3728	
Maraekakaho at Kereru Rd	1910677	5604844	5/5/2016	-6.54	-40.6	732		
Maraekakaho at Tait Rd	1906785	5604891	5/5/2016	-6.74	-41.6	712		
Waitio at Ohiti Rd	1918690	5610114	5/3/2015	-7.60	-48.9	225		
Repokai te Rotoroa at Swamp Rd	1923529	5614267	3/3/2015	-7.44	-45.6		622	
Tutaekuri-Waimate at Goods bridge	1928340	5613360	5/3/2015	-7.47	-46.1	249	1649	
Tutaekuri-Waimate U/S of Ngaruroro	1931723	5609827	5/3/2015	-7.68	-47.3		1902	
Tutaekuri at Puketapu	1925620	5619570	4/3/2015	-6.96	-42.5		3285	
Tutaekuri at Brookfields bridge	1933670	5612730	3/3/2015	-7.10	-43.2		3965	

Site Name	NZTM East	NZTM North	Date	$\delta^{18}O$	$\delta^{2}H$	Electrical conductance (µS/cm 25°C)	Stream flow L/s	Well screen-interval m
Tukituki at Red Bridge	1936570	5596260	23/8/2017	-6.73	-43.6	252	27489	
Tutaekuri at Puketapu	1925620	5619570	23/8/2017	-7.43	-46.9	314	12183	
Karamu at Havelock Nth bridge	1932370	5602146	23/8/2017	-6.23	-41.9	714	2527	
Karamu at floodgates	1932710	5609270	23/8/2017	-6.47	-43.2	594	3757	
Mangateretere at Napier Rd	1933819	5604172	23/8/2017	-6.92	-44.8	272	258	
Ngaruroro D/S Maraekakaho	1911182	5605498	23/8/2017	-8.17	-53.4	114		
Ngaruroro at Fernhill	1922943	5611292	23/8/2017	-8.15	-52.8	125	48281	
Ngaruroro at Expressway Bridge	1930419	5610100	23/8/2017	-8.15	-52.2	131		
Ngaruroro U/S Tutaekuri-Waimate	1932072	5609661	23/8/2017	-8.10	-52.5	133		
Roadside U/S RepokaiTeRotoroa	1923452	5614357	23/8/2017	-7.15	-45.8	428		
Jims at Swamp Road	1923516	5615014	23/8/2017	-7.13	-45.6	408		
Jamieson at Swamp Rd	1923896	5615791	23/8/2017	-7.14	-45.9	438		
Repokai te Rotoroa at Swamp Rd	1923434	5614294	23/8/2017	-7.00	-45.1	536		
Tutaekuri-Waimate U/S Ngaruroro	1931723	5609827	23/8/2017	-7.14	-46.4	327	2692	
Well 16383	1934970	5603163	30/6/2016	-6.34	-38.9	765		(147)
Well 5915	1929066	5610864	20/6/2016	-7.95	-49.2	157		34.9-37.8
Well 2502	1929836	5611166	23/6/2016	-7.94	-48.7	160		45.4-46.9
Well 4362	1904776	5613228	20/6/2016	-7.23	-46.0	241		11.5-15.5
Well 5023	1904773	5613220	23/6/2016	-7.50	-47.1	264		53-54
Well 15006	1926346	5609157	23/6/2016	-7.66	-47.4	159		-30
Well 3525	1927032	5598451	8/6/2016	-7.40	-48.9	371		(32)
Well 844	1925640	5612874	10/6/2016	-7.65	-47.8	151		11.5
Well 15465	1924136	5602754	9/6/2016	-7.05	-47.0	458		12.0-14.9
Well 5988	1924136	5602754	8/6/2016	-6.67	-43.4	581		106-109
Well 5690	1923774	5615445	9/6/2016	-7.67	-48.3	366		13.5-14.7
Well 15012	1928379	5603871	8/6/2016	-8.18	-52.5	287		113.5-114
Well 15884	1924708	5619381	9/6/2016	-7.02	-43.7	411		9.4-10.4
Well 2580	1935702	5604421	6/5/2016	-6.25	-38.3	200		9.0-12.0
Well 16202	1933732	5606791	26/6/2014	-7.75	-49.6	160		(11)
Well 16203	1933736	5606797	12/5/2016	-7.60	-47.4	159		19.4-21.6
Well 5006	1913126	5578864	2/4/2015	-6.92	-42.8	1087		28.8-29.8
Well 16208	1911247	5569726	2/4/2015	-7.44	-44.7	185		36.8-38.7
Well 16256	1912999	5578868	1/4/2015	-6.96	-43.2	734		7.2-9.3
Well 16212	1913289	5574865	1/4/2015	-6.39	-39.6	572		8.36-10.5
Well 16211	1912255	5572240	1/4/2015	-6.79	-39.0	148		6.7-8.8
Well 16209	1911246	5569722	1/4/2015	-7.05	-42.9	479		5.3-7.4
Well 16300-J-2	1924843	5610616	20/1/2015	-5.30	-29.6			-126.5
Well 16300-J-1	1924843	5610616	20/11/2014	-5.61	-33.2			-126.5
Well 16360	1924830	5610618	10/6/2016	-7.81	-51.3	149		64.3-65.3
Well 16361	1924830	5610619	10/6/2016	-7.68	-49.6	157		22.9-23.9
Well 10496	1935135	5603000	17/6/2014	-6.72	-43.6	509		(8)

Site Name	NZTM East	NZTM North	Date	$\delta^{18}O$	$\delta^{2}H$	Electrical conductance (µS/cm 25°C)	Stream flow L/s	Well screen-interval m
Well 3336	1928379	5603871	17/6/2014	-7.36	-47.5	335		47.0-47.5
Well 1459	1932239	5603942	17/6/2014	-7.59	-48.3	236		44.0-53.0
Well 413	1936994	5620244	17/6/2014	-8.08	-53.0	220		76.5-82.6
Well 705	1930388	5607410	19/6/2014	-7.75	-49.1	161		39.0-40.5
Well 15022	1935268	5612897	19/6/2014	-7.79	-50.5	175		29.5-30.5
Well 15003	1935268	5612897	19/6/2014	-7.64	-49.5	176		54.5-55.5
Well 15002	1935268	5612897	19/6/2014	-8.23	-50.0	164		(90)
Well 1191	1919583	5606896	23/6/2014	-7.53	-49.8	436		18.9-21.9
Well 8521	1923104	5605948	23/6/2014	-7.72	-50.6	308		(30)
Well 10340	1922964	5609399	23/6/2014	-7.54	-48.5	188		(17)
Well 1799	1935927	5608503	25/6/2014	-7.83	-49.7	162		28.6-32.6
Well 222	1936827	5615561	25/6/2014	-7.84	-50.1	179		57.3-59.1
Well 15795	1923397	5613906	26/6/2014	-7.24	-45.8	2693		37.2-37.7
Well 16078	1916485	5605277	26/6/2014	-7.00	-44.7	1689		37.2-40.0
Well 1674	1928677	5608418	23/6/2014	-8.23	-53.2	166		32-38
Well 15011	1924123	5605229	25/6/2014	-7.82	-49.2	233		112-114
Well 2580	1935702	5604421	23/8/2017	-6.49	-40.9	355		9.0-12.0
Well 10340	1922964	5609399	23/8/2017	-7.68	-49.5	205		(17)
Well 15884	1924708	5619381	23/8/2017	-6.67	-43.9	660		9.4-10.4