



# Arthur Marble Aquifer and Te Waikoropupu Springs, New Zealand: flow contributions and nitrate sources

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## Abstract

This paper investigates sources of early signs of nitrate pollution in a karst aquifer and major karst spring. Te Waikoropupu Springs are the largest springs in New Zealand and amongst the largest in the Southern Hemisphere with a mean outflow of 13.4 m<sup>3</sup>/s. They discharge the clearest water ever recorded from a karst spring, but early warning of water quality deterioration is signalled by increasing nitrate concentrations. The springs discharge water from a partly artesian aquifer of 2.85 km<sup>3</sup> volume in karstified Ordovician marble with a transmission time averaging 7.9 years. Reservoir porosity is unusually high for a crystalline carbonate rock, estimated as 6%. Water emerging at Main Spring (9.89 m<sup>3</sup>/s) and adjacent Fish Creek Springs (3.48 m<sup>3</sup>/s) is a mix of recharge from four main sources. The paper investigates by means of mass balance modelling the proportions of water coming from each source, the origins of nitrate contamination and the contribution of nitrification. Water losses into the bed of the upper Takaka River, about 17 km from the Springs, provide 13% of the mean flow at Main Spring, 34% of the flow at Fish Creek Springs, and 83% of the volume discharged at submarine springs. Infiltration from rain falling over farmed land in the middle valley is estimated to supply 10% of total outflow at Main Spring and Fish Creek Springs but to contribute 82% of the NO<sub>3</sub>-N output discharged at the springs. Nitrification accounts for 16% of the nitrate discharged. A 34% reduction of leachate concentration in the middle valley would be required to bring NO<sub>3</sub>-N values at Main Spring back to levels recorded in the 1970s.

**Keywords** Karst aquifer · Mass balance modelling · Stable isotopes · Nitrate · Artesian · New Zealand

## Introduction

During millennia of human history, karst springs have had special importance. They have been used as dependable sources of water and have sometimes been revered as holy or tapu sites. Examples include Ras el Ain Spring in Lebanon (Burdon and Safadi 1963; Bakalowicz et al. 2002) and Jinan Spring in Shandong Province, China (Sweeting 1995). Many other examples are referred to in Drew and Hötzl (1999). But over the last century, land use changes in their catchments and increasing water abstraction have led to the diminution of spring discharge and deterioration of water quality. The problem of quality management of karst groundwater resources is both widespread and complex (Bakalowicz 2011) and elevated nitrate—which can be a surrogate for

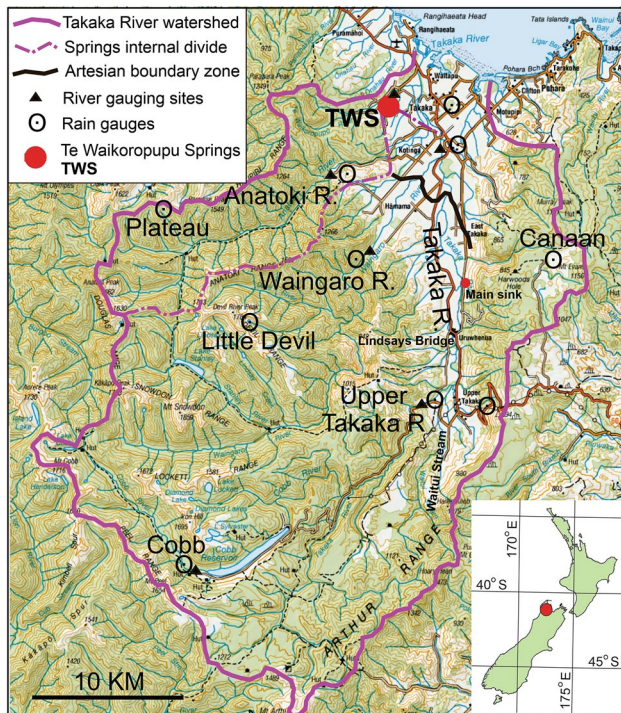
other forms of mainly agricultural contamination—is found in many karst groundwaters around the world (Coxon 2011). Consequently, the restoration of spring flow and improvement of water quality can present significant scientific and management challenges in karst terrains.

This paper discusses a large karst spring and aquifer in which early signs of water quality deterioration have been detected and where agriculture is intensifying and demand for water abstraction is increasing. The case concerned is that of Te Waikoropupu Springs (TWS) and its aquifer in New Zealand (Williams 1977, 2004a, b). The groundwater system is located in the Takaka River basin (Fig. 1) which covers an area of 940 km<sup>2</sup> and rises to 1650 m above sea level (Thomas and Harvey 2013).

The springs have special cultural significance (wahi tapu) for the indigenous Maori people who consider the maintenance of water quality and spring flow volume to be especially important. Since demand for water abstraction for agriculture is at a maximum during drought conditions, questions also arise concerning acceptable minimum flows

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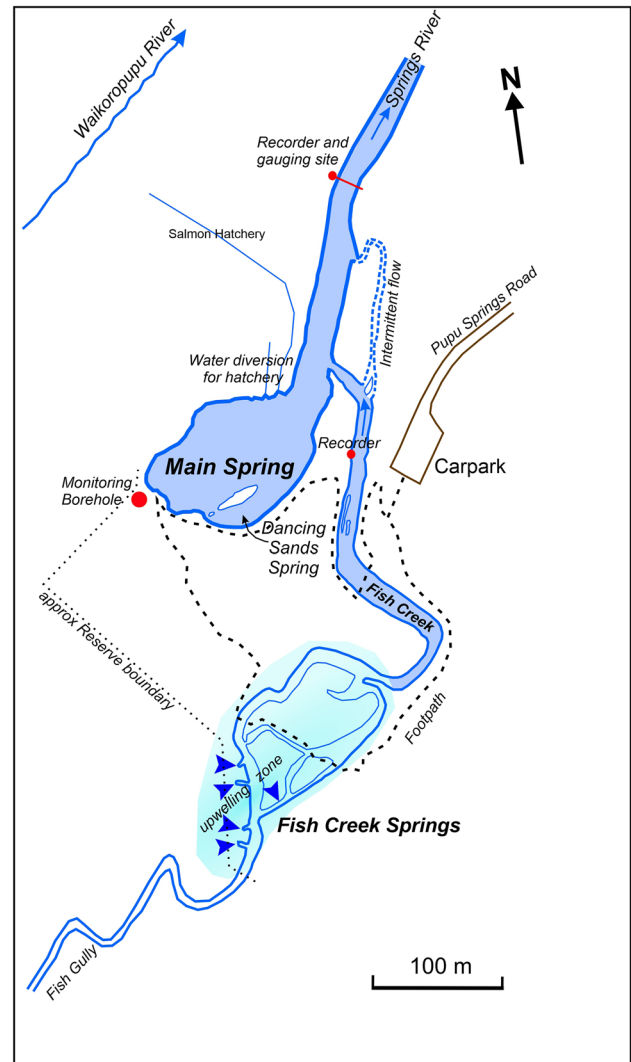
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**Fig. 1** Takaka River basin, showing the catchment watershed and gauging sites. Internal divides are shown that delimit areas draining towards Te Waikoropupu Springs (TWS). Inset shows the location of the basin in New Zealand. (Basemap from LINZ under Creative Commons Attribute 4 licence)

in the spring-fed river downstream, but these are considered elsewhere (Young and Hay 2017).

Te Waikoropupu Springs sometimes referred to informally as "Pupu" Springs (Henderson 1928), are one of New Zealand's natural treasures. They are the country's largest springs and feature on the World Karst Aquifer Map (BGR et al. 2017). The groundwater is contained within Arthur Marble of Ordovician age (Grindley 1971, 1980; Rattenbury et al. 1998) and has an estimated water volume of about 3 km<sup>3</sup> (Stewart and Thomas 2008). The springs are situated 4.6 km inland from the coast and emerge in a zone of about 150 m radius located 14.2–17.5 m above sea level (Fig. 2). The springs are tidal at a mm scale with a mix of about 0.4–0.6% sea water that increases with discharge (Williams 1977; Stewart and Thomas 2008). A monitoring borehole beside the Main Spring shows the marble to be capped with 9.1 m of coal measures above which is a veneer of floodplain gravels; so, the top of the marble aquifer is just 3 m above mean sea level. Water emerges with a boil under artesian pressure (Fig. 3). The aquifer is recharged by a mix of autogenic and allogenic sources, including influent water loss along two river channels. Discharge from the aquifer occurs at TWS and at submarine springs. Total outflow at TWS varies within the range of 5.7–14.3 m<sup>3</sup>/s



**Fig. 2** Plan of Te Waikoropupu Springs (modified from Rapiet 1975). The zone of upwelling water that supplies Fish Creek Springs varies in extent according to artesian pressure head conditions

with a mean annual discharge of 13.4 m<sup>3</sup>/s. The outflowing waters have an average transit time of 7.9 years but contain younger (1.2 years) and older (10.2 years) components (Stewart and Thomas 2008). They are cool with an almost unvarying temperature of around 11.7 °C (Michaelis 1976) and are so well mixed that annual water quality signals are not discernible (Stewart and Williams 1981). The combined outflow sustains the Springs River that is supplemented by the flow of Waikoropupu River prior to joining the Takaka River as a left-bank tributary near the head of a tide. It then discharges into Golden Bay, located at the northern end of South Island. Submarine springs 1–5 km offshore discharge up to a further 7.1 m<sup>3</sup>/s.

The clarity of the waters emerging at TWS is exceptional. Long residence time and slow water movement underground



**Fig. 3** View across Te Waikoropupu Springs showing the ‘boil’ from water released under artesian pressure. The view looks northeast across the Main Spring towards the outflowing Springs River

facilitates the settlement of sediment and, in addition, organic matter is filtered by stygofauna (Fenwick and Smith 2016; Bolton and Weaver 2021). Davies-Colley and Smith (1995) measured the optical properties of the springs, obtaining an average black-body visibility of 63 m, which is close to the theoretical maximum for optically pure water. Remeasurement by Gall (2018) using a continuously recording transmissometer confirmed exceptional clarity. The combination of exceptional quality on a world scale and other unusual features places Te Waikoropupu Springs amongst the world’s scientifically most important karst springs. The multifaceted challenges to water management presented by the Arthur Marble Aquifer (AMA) and Te Waikoropupu Springs provide an instructive case study that has significance for many other karst aquifers and springs around the world, especially because the long residence time of water in the aquifer restricts the effective application of commonly used karst water investigations techniques (for example as described by Benischke et al. 2007).

## Research objectives

The overriding objective is to provide information about the operation of the groundwater system that will assist water quality and quantity management. This paper first reviews what is known about the aquifer and its functioning, and then proceeds to address outstanding questions that need answering to facilitate appropriate aquifer management. These concern: (1) the main recharge areas contributing water to the springs; (2) the relative importance of each area as a source of recharge; and (3) the relative importance of each recharge area as a source of nitrate contamination.

The main input source of water that recharges the aquifer supplying TWS has long been considered to be the upper Takaka River because about 17–18 km inland from the springs it loses water into the gravels of its channel and often runs dry (Park 1890; Bell et al. 1907; Henderson 1928). Confirmation of this source was difficult to verify using conventional fluorometric water tracing techniques because tritium dating by Taylor in 1966 indicated the emerging water could be 3–4 years old (Stewart and Downes 1982). With such a long transmission time, introduced tracer dyes would be subject to considerable attenuation and mixing, leading to dilution below detection limits. The problem was overcome by setting up a pulse train experiment in which TWS were monitored to see if a timed pattern of water releases from a hydroelectric dam (Cobb Dam) in the upper catchment could be detected. The discharge pulses were identified 17 h later at the springs, transmission through the aquifer being by pressure pulse (Williams 1977). This demonstrated the hydraulic connection of the upper Takaka River to TWS but still left open questions concerning other possible sources of water and the relative contributions of each. Subsequent research by Stewart and Williams (1981), Stewart and Downes (1982), Mueller (1991), Williams (1992, 2004a, b), Stewart and Thomas (2008) and Weir and Fenemor (2017) has helped clarify our understanding of these issues, but uncertainties still remain especially in regard to sources of nitrate contamination, a matter addressed in this paper.

## Methods

The approach taken here is to:

1. Outline the background hydrogeological context;
2. Identify source areas where the AMA within the TWS catchment is recharged;
3. Quantify the mean annual recharge volume for each source area;
4. Identify mean annual tracer signatures ( $\delta^{18}\text{O}$ ;  $\text{NO}_3\text{-N}$ ) for each source of recharge;
5. Identify groundwater outflow sites and associated mean annual discharge volumes;
6. Model the average proportions of flow contributed by each recharge area; and
7. Calculate the average nitrate load introduced by each recharge area, the effect of nitrification during groundwater transit through the aquifer, and the mean annual  $\text{NO}_3\text{-N}$  budget for the karst aquifer.

Modelling in karst is discussed by Palmer et al. (1999), Kovács and Sauter (2007) and Bakalowicz (2011). However, this study is not concerned with modelling hydrological processes in the aquifer but with establishing the input–output

budget, and for that purpose a mass balance approach is preferred because it assumes conservation of mass (input = output + storage) and makes no assumptions about the nature of flow except that inputs are instantly mixed and immediately transferred to outputs. The model used here deals with average steady-state conditions and so does not address transient conditions.

For mean flow conditions, a mass balance model is used to estimate, firstly, the input flow proportions (and hence volumes) contributed by different aquifer recharge sources and, secondly, the nitrate load contributed by each recharge area to the outflowing spring waters. Implications for management are then considered because an undesirable trend of increasing nitrate concentrations has been observed over several decades at the springs (Moreau 2021).

The model proceeds in three steps:

1. From the  $\delta^{18}\text{O}$  mass balance, the proportion of outflow from the Main Spring and Fish Creek Springs that can be attributed to each recharge area is calculated.
2. These proportions of flow are converted to discharge (L/s) and together with their measured  $\text{NO}_3\text{-N}$  concentrations are used to determine the contributions (or loading) made by each recharge source to the total  $\text{NO}_3\text{-N}$  introduced into the AMA, then
3. The effect of nitrification during groundwater transmission through the aquifer is evaluated by identifying (through trial and error) the percentage increase that must be applied to nitrate input loadings to achieve the nitrate load discharged at the springs.

Modelling is constrained by the requirement to be consistent with measured values of discharge,  $\delta^{18}\text{O}$  and  $\text{NO}_3\text{-N}$  at both input and output sites.

## Hydrogeological background

### Geological setting

Takaka valley is a structural landform produced by compression and faulting during the emergence of the Southern Alps (Williams 2017), although there may have been an earlier phase of extension and normal faulting in the Eocene during which Brunner (formerly Motupipi) Coal Measures were deposited (Leask 1993). Reverse faulting and folding of early-middle Miocene Tarakohe Mudstone (to 518 m thickness) and underlying Brunner Coal Measures (to 350 m thickness) in Takaka valley implies compression and uplift from the mid-Miocene (Grindley 1971; Ravens 1990), and this ultimately led to the development of the Pikipiruna Fault as a north–south reverse fault expressed in the landscape as the 700 m Takaka valley escarpment. Compression fractured Paleozoic basement rocks and their covering sediments,

driving the western side down; a displacement that formed a wedged-shaped fault-angle depression about 9 km wide at the coast that tapers inland for 26 km, the Takaka valley. Downthrow on the western side pushed marble well below sea level and took overlying Brunner Coal Measures to a depth of around 340 m near Dry River (Fig. 4) and to 275 m in the middle valley near East Takaka (Ravens 1990; Leask 1993). Erosion has since removed most of the Cenozoic cover beds and, in the process, has exhumed an ancient erosion surface cut across the Paleozoic rocks, including the marble, which had been exposed to karstification in the Paleocene-Eocene.

Arthur Marble extends through northwest South Island in a discontinuous 90 km belt up to 12 km wide from Mt Owen (1867 m) through Mt Arthur to Takaka Hill and Golden Bay (Rattenbury et al. 1998) (Figs. 1 and 4). The marble varies in stratigraphic thickness from about 500–1500 m depending on location. It is often steeply dipping and is frequently faulted (Fig. 5). Near the headwaters of the Takaka River, marble is found to 1778 m on Mt Arthur, on Takaka Hill it underlies a plateau at around 760 m and in Takaka valley it descends well below sea level. Geological details are available in Grindley (1971, 1980), Ravens (1990), Mueller (1991), Leask (1993) and Rattenbury et al. (1998).

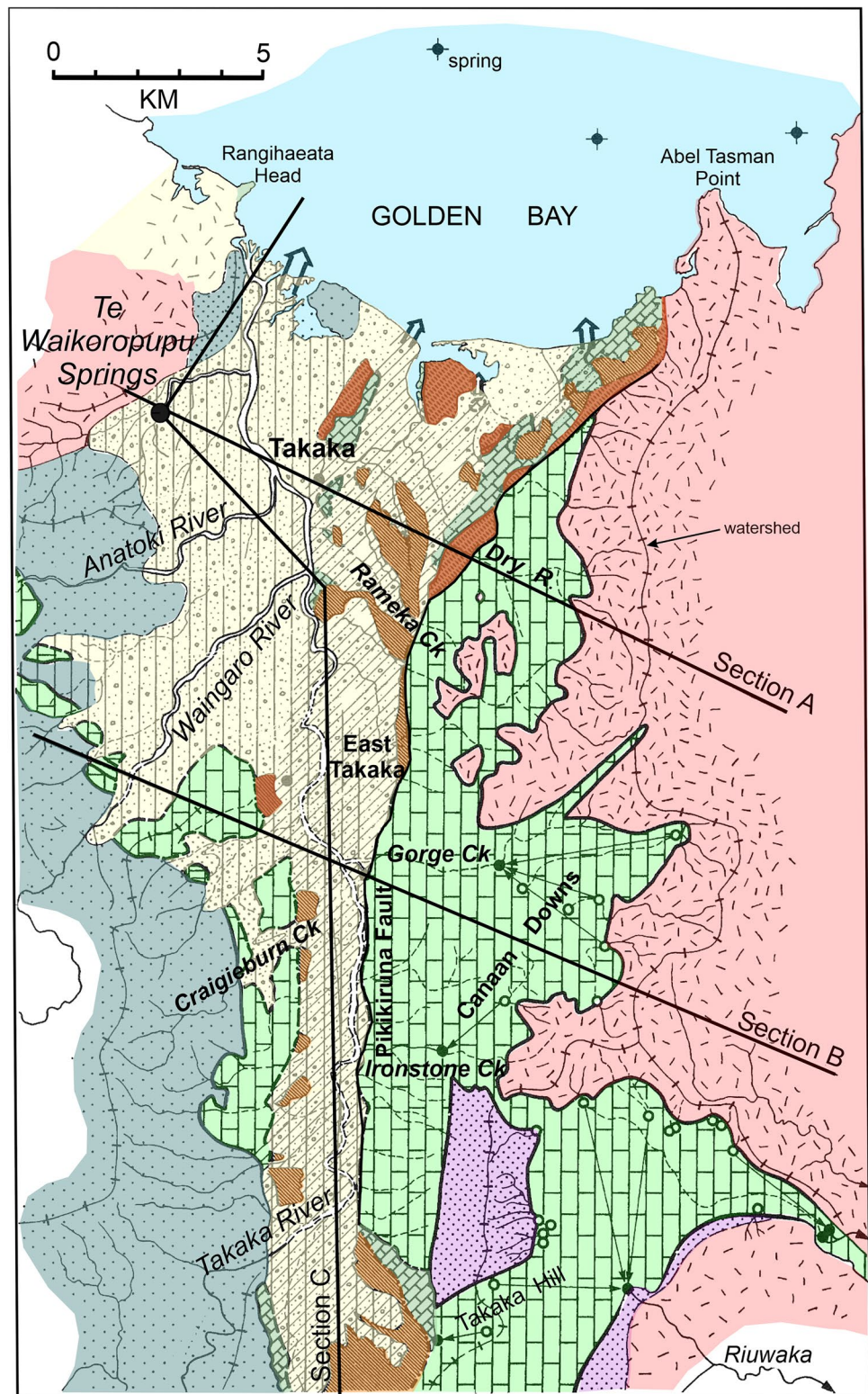
### The Arthur Marble aquifer

The catchment draining to Te Waikoropupu Springs has an area of about 714 km<sup>2</sup>. This includes about 141 km<sup>2</sup> of marble in the middle and lower parts of the Takaka River basin, there being another 18 km<sup>2</sup> of marble lying seawards of TWS.

The main Takaka river basin is delimited on the eastern side of Takaka valley by the principal watershed from Hailes Knob (1279 m) to Pikipiruna (1054 m), Mt Evans (1156 m) and Murray Peak (1101 m), the boundaries of the contributing area having been defined in places by tracing streams sinking into marble on Takaka Hill (Williams 1982). On the western side, the catchment includes Takaka River tributaries upstream of the Anatoki River confluence (Fig. 1) all of which contribute recharge to the marble aquifer. When the AMA is fully recharged, excess water overflows down the Takaka River and discharges into Golden Bay (Williams 1977; Mueller 1987; Edgar 1998; Stewart and Thomas 2008; Weir and Fenemor 2017).

The aquifer within the marble has characteristics described by Ford and Williams (2007) as being typical of karsts in pure, dense, crystalline carbonate rocks. Its porosity involves pores, fissures and conduits (cave passages), but most groundwater is held in dissolutionally widened interconnected fissures. Although individual conduits can be large (dm-m scale), total conduit porosity occupies only a few percent of the rock mass; yet almost all of the water is

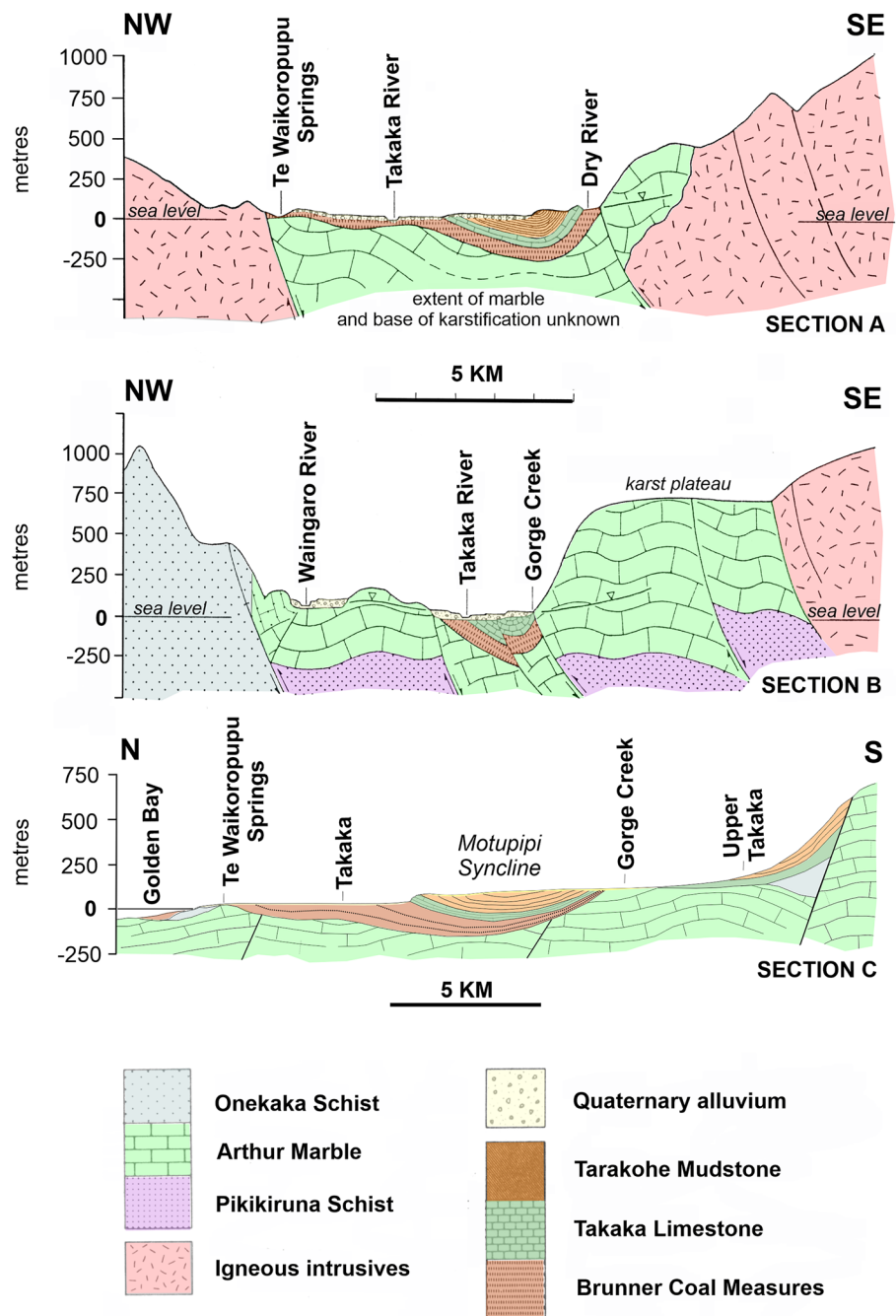
**Fig. 4** Simplified geological map of the Takaka valley showing the extent of Arthur Marble, and the location of Te Waikoropupu Springs and submarine springs. The basemap from Ford and Williams (2007, Fig. 11.7) shows proven water tracing connections in the marble area from stream-sinks to springs. Geology from Grindley (1971, 1980). Key as per Fig. 5



conveyed through them despite their containing relatively little water volume. Hence the aquifer can be envisaged to contain a few large but widely spaced conduits (functioning as main drains) that penetrate and traverse a water-saturated

fissure matrix comprising mm to sub-mm cracks and occasional interconnected pores. Thus, most groundwater storage is accommodated in a three-dimensional fissure network that extends vertically several hundred metres and laterally

**Fig. 5** Geological sections across Takaka valley with transect locations shown in Fig. 4. Sections A and B from Ford and Williams (2007, Fig. 11.7) and section C modified from Mueller (1991, Fig. 2)



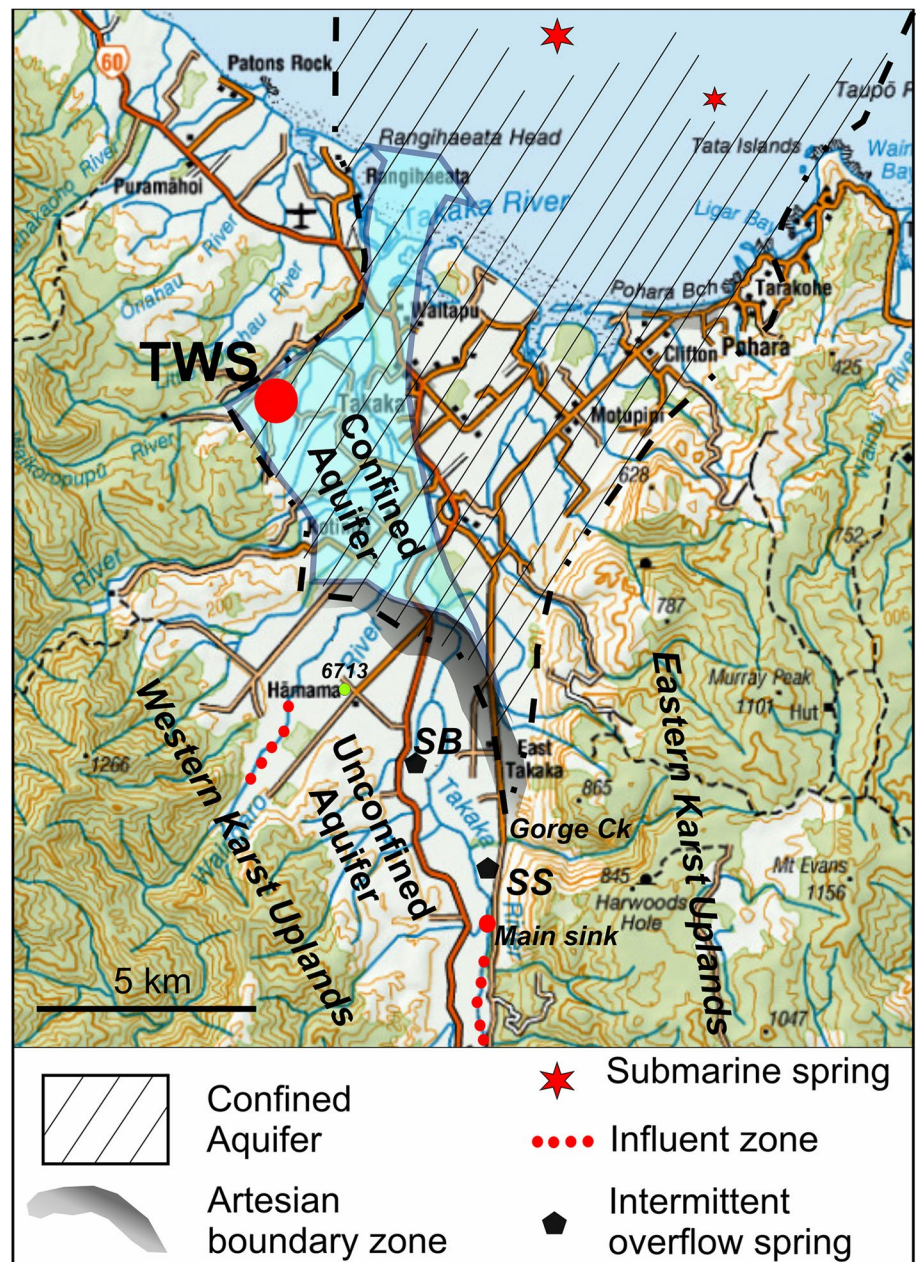
several kilometres. Water storage in some places can be remote from main drains in both time and distance (km away). However, while most are laterally remote, some water is held at depth, as signalled by chloride (Cl) in the spring water that comes from an underlying intrusion of sea water from Golden Bay. Outflow from the Main Spring discharges about 0.5% of sea water, and Fish Creek Springs a little less (Michaelis 1976; Williams 1977; Stewart and Thomas 2008). Halites are unknown in the catchment and Stewart and Downes (1982) determined SO<sub>4</sub> concentrations and δ<sup>34</sup>S values in Main Spring water to be consistent with a sea water

origin. Mixing at the fresh/salt interface is enhanced during times of high flow when Cl values are found to increase in the Springs' discharge (to about 0.7%).

**Artesian boundary**

The floor of Takaka valley covers about 109 km<sup>2</sup>. Much is underlain by marble, but at its northern (seawards) end marble is unconformably overlain by Cenozoic sediments [coal measures, limestones and mudstones (Grindley 1971; Leask 1993)] that provide an impermeable caprock (section

**Fig. 6** The northern end of the Takaka valley shows the location of the artesian boundary zone and the confined aquifer. Groundwater in the confined aquifer is schematically indicated in blue, the flow being along the western edge of the Motupipi Syncline. Eroded caprock permits water under pressure to escape at TWS with the rest continuing north to Golden Bay and offshore springs. Excess water from the unconfined aquifer emerges at intermittent springs: SB Spring Brook Spring and some at SS Spittal Spring most of which is water from the eastern karst uplands. The site of groundwater well 6713 is shown near Hamama. (Basemap from LINZ under Creative Commons Attribute 4 licence)



C in Fig. 5). This confines the aquifer as far inland as East Takaka in the middle valley (Fig. 4) and so gives rise to artesian conditions. Over the 8 km distance from the artesian boundary to TWS, the difference in the hydraulic head is of the order of 25–30 m, varying with recharge conditions.

The position of the artesian boundary is coincident with the edge of the confining cover of Brunner Coal Measures. But its position cannot be determined exactly, being buried beneath a thick veneer of Quaternary gravels, so is inferred from outcrops, borehole and geophysical data. A seismic profile across the valley from East Takaka to Hamama is interpreted to show a maximum thickness of Coal Measures of 275 m around East Takaka thinning to the northwest

near Hamama (Fig. 6), where thick Quaternary gravels rest directly on marble (Ravens 1990; Leask 1993). Although the artesian boundary extends generally northwest across the valley from East Takaka, its trace in plan is irregular, partly because a buried course of the Takaka River (formed during glacial low-stands of the sea) cut a downstream extending re-entrant into and across the Coal Measures. This is consistent with drilling reports of gravels up to 59 m thick (Stewart and Thomas 2008; Thomas and Harvey 2013) that probably infill a valley graded to glacial sea level. In addition, the tapering edge of the artesian caprock is sometimes punctured by collapse sinkholes, where surface materials have slumped into underlying karst voids. Consequently, the artesian boundary

is irregular in plan and perforated in places; so is best considered a zone of 1 km width rather than a simple sharp edge. Seawards of the boundary confined artesian conditions extend across about 60 km<sup>2</sup>. Marble is unlikely to underlie the entire width of the lower valley because schist outcrops and the eastern side is occupied by coal measures in a deep N–S synclinal structure, but to the west of the synclinal axis groundwater in about 30 km<sup>2</sup> of marble flows mainly to TWS but with excess water draining directly to the sea.

The Takaka River channel near East Takaka is about 40 m above sea level and this provides the threshold elevation for groundwater overflow in the marble aquifer. Borehole data in this part of the valley shows groundwater levels can vary by up to about 30 m vertically, which indicates that at the artesian boundary the head difference to Te Waikoropupu Springs can be as little as 7 m (ca. 21–14.3 m in Main Spring) in drought conditions yet reach around 32 m (ca. 49–17.5 m in Fish Ck Springs) when the aquifer is full. High water overflow occurs at Spring Brook (southeast of Hamama) and at Spittal Spring (at 41.5 m near the exit of Gorge Creek, Fig. 6). When the inland head is at a maximum, the discharge of offshore springs will peak, but when

head inland is at a minimum the submarine springs may hardly flow, because of the more effective resistance to outflow provided by the hydrostatic pressure of the denser sea water column.

The 40 m groundwater head at the inland artesian boundary is in itself insufficient to drive deeply circulating flow paths, but when moving water is trapped beneath descending impervious caprock then it will be forced downwards before it can escape upwards. This is the situation in the middle valley below the Motupipi Syncline, where Coal Measures descend about 275 m below sea level (Ravens 1990; Leask 1993). Most groundwater will be directed along the western edge of the impervious rocks that constitute the syncline, which accounts for the northwest offset of the Springs in an otherwise north–south valley. Coastal intrusion of salt water along the eastern side of the valley is limited by impervious sediments of the Motupipi Syncline to – 340 m near Dry River (Fig. 5 section A) and by a possible diorite intrusion (Jones 1939).

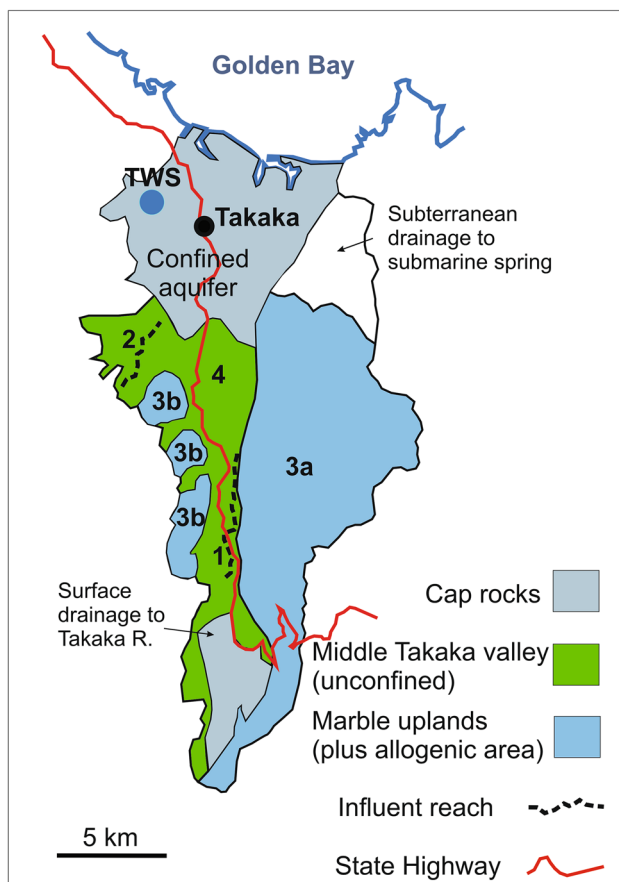
### Aquifer recharge

Inland of the artesian boundary zone recharge of marble occurs in four areas (Fig. 7):

1. The influent zone of Takaka River, a reach of about 4.5 km from just above Lindsay's Bridge to just downstream of the Main Sink (Fig. 6), where water is lost into the riverbed and at a high stage also into the foot of the marble escarpment.
2. The influent zone of the lower Waingaro River (Fig. 1), a reach of about 4 km downstream of Hanging Rock river gauging station towards Hamama (Fig. 6), where water is lost into thick channel gravels overlying marble.
3. The marble (karst) uplands on the western and eastern sides of Takaka valley (Figs. 4, 6). The eastern side comprises 66.3 km<sup>2</sup> of marble plus 22.8 km<sup>2</sup> of adjacent impervious uplands (mainly granodiorite) that shed allo-genic runoff onto and into the marble. The western side comprises 11.2 km<sup>2</sup> of marble plus 0.6 km<sup>2</sup> of allo-genic catchment.

In addition, about 5.2 km<sup>2</sup> of marble with runoff from 12.4 km<sup>2</sup> of adjacent granodiorite on the north-eastern seawards side of the valley (Dry R and Kitty Ck area) contribute recharge to the lower aquifer adjacent to the sea without influencing TWS.

4. The middle Takaka valley immediately upstream of the artesian boundary that extends across about 48.9 km<sup>2</sup> of floodplains and terraces from around Hamama to beyond Upper Takaka into the Waitui valley (which contributes an unconfined area of 6.1 km<sup>2</sup>) (Fig. 1).



**Fig. 7** Source areas of recharge input to the Arthur Marble Aquifer in the lower and middle Takaka valley and adjacent uplands



For the purpose of mass balance modelling, estimates are required of the mean annual recharge contributed to the AMA by each of the above input source areas. Rainfall and discharge data are available from the network of monitoring sites maintained in the Takaka basin (Fig. 1).

The principal input of allogenic water comes from the Takaka River (recharge source 1 above) that flows from the Tasman Mountains and averages about 14.3 m<sup>3</sup>/s at the Upper Takaka gauging site (Harwoods) (Thomas and Harvey 2013). The river loses water from about Lindsay's Bridge downstream (Fig. 1), where the channel is formed in bouldery sands and gravels. The river runs along the eastern foot of the fault-line escarpment (Fig. 4), and under high-water from Ironstone Creek confluence to beyond Craigieburn confluence water sinks directly into marble at several sites at around 44.3 m above sea level. Some of these sinks reverse to become springs when the head of groundwater escaping from the adjacent karst uplands rises sufficiently (Castle 2020, 2021) (Main Sink, Figs. 1, 6). Subterranean convergence of groundwaters from Takaka Hill and floodwaters from Takaka River leads to overflow at Spittal Spring (45.1 m) a few hundred metres further downstream. So, some recharge input is lost to overflow near the artesian boundary.

Water loss along the Takaka River channel downstream of Lindsay's Bridge (recharge source 1, Fig. 7) varies in

volume, being at a maximum when low groundwater levels coincide with high river discharge. Under such circumstances up to about 15 m<sup>3</sup>/s can be absorbed underground. When river discharge falls to 11 m<sup>3</sup>/s or less, which can occur for several months per year (Weir and Fenemor 2017), the riverbed becomes dry in the middle valley and the water table can fall to around 19 m above sea level. Under such conditions, flowing water is not seen again in the channel until it receives water at the confluence of the Waingaro River some 9 km further downstream. Average loss from the river in this reach to the AMA is estimated as 8–8.35 m<sup>3</sup>/s (Stewart and Thomas 2008; Thomas and Harvey 2013).

Influent flow also occurs along a stretch of the Waingaro River (recharge source 2, Fig. 7) southwest of Hamama (Fig. 6). In this reach the river crosses gravel that veneers marble for about 4 km (Fig. 5 section B). Mean discharge is about 17.8 m<sup>3</sup>/s and losses of 0.5–0.8 m<sup>3</sup>/s have been measured at low flow (Thomas and Harvey 2013). The estimate of recharge input from this source is not well constrained by measured data at higher flows but is of the order of 1.5–2 m<sup>3</sup>/s (Table 1).

The neighbouring Anatoki River (Figs. 1, 4) can also lose about 0.5 m<sup>3</sup>/s during low flow below the Happy Sams gauging site, but this appears to just infiltrate gravels without entering marble (Thomas and Harvey 2013).

**Table 1** Input sources of recharge to the Arthur Marble Aquifer: their locations, contributing areas, average input flows, and water quality signatures

Recharge input source	Location	Area (km <sup>2</sup> )	Water surplus (mm)	Recharge input to marble (L/s)	δ <sup>18</sup> O, ‰	NO <sub>3</sub> -N mg/L
1a	<b>Upper Takaka R</b> influent flow			8350	– 8.67	0.0535
1b	Waitui valley runoff from confined lowland	14.2				
2	<b>Lower Waingaro R</b> influent flow			1500	– 7.2	0.01
3a	<b>Karst uplands E</b> autogenic marble	89.1	2750	7764	– 7.4	0.1
	allogenic runoff from granitic rocks	<b>66.3</b> 22.8				
3b	<b>Karst uplands W</b> autogenic marble	11.8	2350	834	– 7.0	0.1
	allogenic runoff from schistose rocks	<b>11.2</b> 0.6				
4a	<b>Middle Takaka valley</b> (unconfined)	<b>42.8</b>	1500	2324	– 6.0	3.04
4b	Waitui valley (unconfined)	<b>6.1</b>				
	<b>Lower Takaka valley</b> (confined)	60.36				
	Marble draining to TWS	~15				
	Marble draining to sea	~15				
	Non-marble rocks	~30				
	<b>Marble draining to TWS</b>	<b>141.4</b>				
	Takaka valley lowlands	109.3				
	Dry R and Kitty Ck area draining to sea	17.6				

Data sources: Stewart and Williams (1981), Stewart and Thomas (2008), Thomas and Harvey (2013), Moreau (2021), Weir and Fenemor (2017), FoGB  
Bold text in 'Location' column identifies principal sites and source areas and in 'Area' column identifies areas of marble that drain to TWS

Along the eastern watershed, the marble uplands (recharge source 3, Fig. 7) receive streams from the adjacent hills of granodiorite and schist (Fig. 4) that sink in swallow holes close to the lithological boundary, their flow directions having been traced (Williams 1982). The marble plateau around 760 m also receives recharge directly from rainfall. In the adjacent Riuwaka basin (Fig. 4), rainfall across the marble averages 2184 mm and, with mean annual actual evapotranspiration of 525 mm, yields an annual specific discharge of 51.75 L/s/km<sup>2</sup> (Williams and Dowling 1979). Thus rain quickly infiltrates into the fissured marble and water surplus can be about 75% of precipitation. On the karst plateau near Canaan (Figs. 1, 4) that drains west towards the Takaka valley, specific discharge may be greater because short-term rainfall records show an annual mean of 3484 mm (Thomas and Harvey 2013) and rain on the adjacent granitic mountains is likely greater. Consequently, rainfall over this part of the uplands (recharge source 3a) is estimated to be about 3400 mm. The marble surface is covered in places by granitic 'grus' soil which impedes rapid runoff into open joints, so actual evapotranspiration is probably the order of 650 mm. Therefore a water surplus of about 2750 mm is generated across 89 km<sup>2</sup>, giving an average input of 7764 L/s (Table 1).

Along the western side of Takaka valley, smaller and lower karst uplands to 300 m altitude (recharge source 3b, Fig. 7) receive less precipitation, up to about 3000 mm, noting average annual rain gauge measurements in Upper Takaka (Harwoods) of 2215 mm and Waingaro (Hanging Rock) of 3128 mm. Evapotranspiration losses will be similar to those on the eastern karst plateau, so the annual water surplus yielded from 11.8 km<sup>2</sup> of the western karst uplands is about 834 L/s (Table 1).

In the lower Takaka valley, mean annual rainfall at Kotinga near Takaka township is 1978 mm, somewhat less than the 2215 mm at Upper Takaka, so the average across the intervening middle valley (recharge source 4) is about 2100 mm with actual evapotranspiration losses of around 600 mm, given the summer soil moisture deficit in the area. Where floodplain and terrace river sediments directly overlie marble, the recharge from rain infiltrates the gravelly veneer and percolates vertically into the marble with the excess draining laterally towards the Takaka River. Across the terraces, sinkholes with ponds mark areas where gravels are water-saturated for most of the year. Numerous sinkholes are the expression of subsidence of gravels into karstic voids in the marble, so localised point recharge from farmland into the marble aquifer via sinkholes is widespread.

The four source areas identified above (and shown in Fig. 7) therefore account for recharge to the marble. In Table 1 the estimated recharge volumes from each are expressed in litres per second (L/s). Although most rain gauges and river gauging sites have been monitored for

decades as part of a national hydro-meteorological network, uncertainties associated with these estimates exceed  $\pm 10\%$ . In addition, about 0.5% of spring flow has a sea water origin (Williams 1977; Stewart and Thomas 2008) but is not included as a water source in the modelling.

### Tracer signatures ( $\delta^{18}\text{O}$ ; $\text{NO}_3\text{-N}$ ) for each water input source

The mass balance model used here estimates (1) the proportions of flow contributing to TWS that are derived from the different recharge sources discussed above and (2) the proportions of nitrate discharged at TWS that can be attributed to each source of recharge. Water quality attributes  $\delta^{18}\text{O}$  and  $\text{NO}_3\text{-N}$  are used to characterise each input. Being part of the water molecule,  $\delta^{18}\text{O}$  is an excellent conservative tracer, i.e., does not change with passage through the aquifer. However, nitrate may be generated in groundwater by nitrification during its passage through an aquifer, so  $\text{NO}_3\text{-N}$  is only acceptable as a tracer if this process can be accounted for. This is the case here and is discussed later.

### Oxygen isotopes

New Zealand participated in the global network for isotopes in precipitation established by IAEA/WMO in the 1960s (Stewart and Taylor 1981) and subsequent sampling in the Takaka basin contributed to a national survey of stable isotopes in rivers and lakes (Stewart et al. 1983, 1987). Following initial reconnaissance in the Takaka basin, regular four-weekly sampling for oxygen isotopes was carried out at eight sites between 1976 and 1979 (Stewart and Williams 1981). This sampling included the groundwaters that emerged at Te Waikoropupu Springs. Mean annual values have been published by Stewart and Thomas (2008). Thus with respect to the input areas discussed in "Aquifer recharge", the basin-wide sampling of stable isotopes provides a reasonable basis for estimating mean annual isotopic values of recharge water despite an uneven geographical coverage and frequency of measured samples.

The upper Takaka River drains a mountainous area subject to westerly wind flow from the Tasman Sea and so its precipitation, which can fall as snow, is characterised by relatively negative values ( $-8.67\text{‰}$   $\delta^{18}\text{O}$ ), whereas rainfall in the sheltered lowland valley is of lower volume and much less negative ( $-6.00\text{‰}$   $\delta^{18}\text{O}$ ). Estimated mean annual  $\delta$  values for water inputs in each source area and for groundwaters emerging at TWS are entered in Table 1. Three spring-fed streams emerging from the eastern karst uplands have  $\delta^{18}\text{O}$  values in the range  $-6.5$  to  $-8.0$ , Ironstone Ck averaging  $-7.8\text{‰}$ . From available data Stewart and Thomas (2008) assigned a mean of  $-7.4\text{‰}$  to the east karst tributaries and

– 7.0‰ to the western tributaries. These values are adopted here.

### Nitrates

Of particular concern for water resources management in the catchment is the 50% increase in nitrate ( $\text{NO}_3\text{-N}$ ) concentration recorded at TWS between 1970 and 2020. During this time interval there was a significant increase from around 0.30 to 0.45 mg/L confirmed by robust statistical analysis (Moreau 2021). Consequently,  $\text{NO}_3\text{-N}$  is used here as a tracer to try to identify its main input sources. Over the years, samples have been measured using different techniques in different laboratories, so to ensure consistency and comparability, Moreau (2021) combined, cleaned and screened for duplicates all known samples collected for the Main Spring. Prior to 1994 analysis was by cadmium reduction colorimetry but since then has been by ion chromatography. Samples include those collected by the regional government (Tasman District Council; reviewed in Weir and Fenemor 2017) and a citizen science group (Friends of Golden Bay, FoGB). The national Geological and Nuclear Sciences laboratory responsible for analysing and quality control of the records used here (and the laboratory used by FoGB) cites a lower detection limit for  $\text{NO}_3\text{-N}$  of 0.003 mg/L. Uncertainty of measurements in the range of 0.1–0.5 mg/L is  $\pm 1.5\%$  and for 0.5–5 mg/L is  $\pm 0.3\%$ . Ammonia–nitrogen ( $\text{NH}_3\text{-N}$ ) in the springs has consistently measured below the detection limit.

Nitrate-N input values used represent the average concentration of  $\text{NO}_3\text{-N}$  in recharge waters entering the aquifer at the water table. Values for recharge sources were obtained mainly from streams flowing from largely unmodified forested catchments of national parks. Samples from the upper Takaka River where it leaves the forested uplands at the Harwoods gauging station have low values (median of 0.004 mg/L from 39 monthly samples) but 2.7 km downstream after the Waitui confluence and at the start of the influent reach (recharge source 1) where input into the marble aquifer occurs  $\text{NO}_3\text{-N}$  concentration in the river has increased to 0.0535 mg/L (median of 68 samples) (Table 1).

Monthly samples from the Waingaro River as it emerges from the forest-covered schists of Kahurangi National Park (recharge source 2) also show low  $\text{NO}_3\text{-N}$  concentrations, 39 samples having a median of 0.008 mg/L, here rounded up to 0.01 mg/L.

Nine samples from each of the karst springs fed Gorge Creek and Ironstone Creek (Figs. 1, 4) are available (FoGB data) to characterise  $\text{NO}_3\text{-N}$  concentrations from the relatively undisturbed eastern karst uplands (recharge source 3). These show a range of 0.007–0.18, with a median of 0.03 for Ironstone Ck and 0.06 for Gorge Ck. Nearby Rameka Ck has greater human disturbance with nitrate values ranging from 0.005 to 1.0 mg/L depending on the sampling site, but this

catchment may not contribute to the recharge that reaches TWS. Since most of the eastern karst drains forested national park, a rounded average input value of 0.1 mg/L  $\text{NO}_3\text{-N}$  was adopted for the area.

Water quality data available to characterise nitrate values in infiltrating recharge from the middle Takaka valley (recharge source 4) are difficult to interpret objectively. Weir and Fenemor (2017, Fig. 25) compiled groundwater  $\text{NO}_3\text{-N}$  data from farm wells collected by the Tasman District Council. In the middle Takaka valley upstream of the artesian boundary information is presented for 16 wells. Those drilled into floodplain gravels closest to rivers (12 wells) have groundwater nitrate values  $< 2$  mg/L, whereas wells in terrace gravels most remote from rivers have concentrations in the range of 2–8 mg/L. Therefore, it is evident that mixing with river-derived recharge dilutes nitrate concentrations in some wells. Groundwater nitrate values most likely to represent recharge from farmland are to be found in wells located farthest from adjacent rivers in local groundwater divide positions. For example, a well (WWD6713) in farmland on a gravel terrace at 57 m above sea level near Hamama (Fig. 6) penetrates gravels for 50 m before entering marble. Samples of its waters returned nitrate concentrations of 2.5–5.0 mg/L (mean 3.74, median 3.75 mg/L). This does not represent leachate from the farm soil but leachate mixed with receiving groundwater near the top of the saturated zone. The Hamama site is in an area of relatively intensive irrigated dairy farming. In an agricultural case study from the valley, Litherland et al (2018) calculated leachate concentration of 7.38 mg/L from farm drainage water, which emphasises the point that a high yield of N is to be expected from intensive agriculture (Dymond et al. 2013; Snelder et al. 2018).

The concentration of  $\text{NO}_3\text{-N}$  in waters representing recharge source 4 can also be deduced from the mass balance model, because it is the only recharge value that is unknown, and so can be determined from the concentration required to enable the mass balance model to balance. The value determined is 4.05 mg/L which represents the average leachate input across a farmed area of 49.8 km<sup>2</sup> of floodplains and river terraces plus the subsequent effects of nitrification during the passage of this water through the aquifer. Comparison of the value deduced with that measured from the Hamama well (4.05 versus 3.75 mg/L) suggests that although it overestimates initial input concentrations, the effect of nitrification is not great, probably less than 20 percent.

It is assumed that some nitrification in water from the karst uplands (recharge source 3) is accomplished in the epikarst, where litterfall is captured, detained and decomposed, and in the long percolation through the vadose zone beneath it (about 500 m deep). This component from diffuse recharge is joined by the contribution of small sinking

allogenic streams flowing from granodiorite that provide about 23% of the water input from the forested karst uplands. However, these streams are unlikely to carry much nitrate, because in undisturbed beech/podocarp native forest typical of the region most of the  $\text{NH}_4$  produced is utilised by plants rather than being converted to  $\text{NO}_3$ . McGroddy et al. (2008) commented that most estimates of total or dissolved N loss for New Zealand lands not influenced by prior agricultural use are  $\leq 2$  kg N/ha. In four monitored native forest catchments at Maimai (located c.150 km to the southwest) total nitrogen yields varied over 6 years from 0.01 to 3.67, averaging 1.16 kg/ha/year, and a catchment in Big Bush Forest (c. 65 km to the south-southwest) ranged from 0.18–0.44, averaging 0.29 kg/ha/year (equivalent to 0.02 mg/L) (Fahey and Jackson 1997). Although these cases are not in the Takaka catchment, McGroddy et al. (2008) found that dissolved N concentrations are not strongly correlated with climate or geology. These examples show that organic N loads carried by streams from the native forest are low because most is recycled within the forest ecosystem.

As noted earlier, nitrate is only reliable as a water tracer if denitrification or nitrification are not major issues. Significant denitrification is unlikely in this aquifer because water issuing from Main Spring is oxygenated with a median value of 6.2 mg/L (Moreau 2021), whereas denitrification requires dissolved oxygen concentrations in groundwater to be less than 1–2 mg/L (Rivetta et al. 2008). However, some nitrification is indicated because dissolved oxygen levels at Main Spring are below saturation (54–66% saturation from 19 samples at Main Spring (Moreau 2021)), probably as a result of the biologically moderated decomposition of organic matter. This releases N thereby making it available for nitrification. A contribution from this process to  $\text{NO}_3$ -N concentrations measured in water samples from TWS is therefore to be expected.

By a repeated process of trial and error, a nitrification factor of 16%, when applied to all recharge source input values, was found to permit the model to balance. This yielded an average input value from across all farmlands in the middle valley of 3.04 mg/L  $\text{NO}_3$ -N (Table 1) which is well within the range of concentrations measured in farm wells (Weir and Fenemor (2017)).

## Storage

The average residence time (or transit time) of water in the aquifer is 7.9 years (Stewart and Thomas 2008). The Main Spring (including adjacent Dancing Sands) has an average discharge of 9.89  $\text{m}^3/\text{s}$  (Thomas and Harvey 2013); so, over a period of 7.9 years a total water volume of 2.47  $\text{km}^3$  flows from the aquifer. In addition, Fish Creek Springs yield an average of 3.48  $\text{m}^3/\text{s}$  with a transit time of 3.5 years (Stewart and Thomas 2008); so, in that interval these springs

discharge 0.38  $\text{km}^3$ . Consequently, their combined outflow yields an aquifer volume of about 2.85  $\text{km}^3$ . This is stored beneath the valley floor and under marble hills on each side of Takaka valley, including in complex cave systems (Stephenson and Davies 2018; Castle 2020, 2021). Although most storage is below the water table some waters are also held for months in the epikarst, which is the intensely weathered zone that extends for about 10 m beneath the marble surface (Williams 2004b). Storage in this subcutaneous zone is held in dissolutionally widened fissures that with an estimated secondary porosity of up to 10% could account for 3% of karst water storage in the Takaka region, and storage within alluvial gravel floodplains and terraces over the unconfined aquifer could accommodate a further 5%.

The thickness (or depth) of the aquifer is difficult to estimate because the base of karstification is unknown. The stratigraphic thickness of the marble varies up to 1500 m and down-faulted tilted blocks of marble in the valley probably descend at least several hundred metres below sea level (Fig. 5). Karstification will extend to the depth to which karst water has circulated and this will be at least to the level of low glacial sea levels, although Golden Bay was dry during glacials (Williams 2017) and stood well above the lowest seas (– 125 m). However, in places artesian circulation has forced deep water movement because Cenozoic caprocks descend to 275 m or more below sea level (Leask 1993). Further, Paleogene caves may also have been pushed down by reverse faulting but, even if water-filled, may not necessarily participate in active circulation. Taking the above factors into account, water storage in marble is conceivable in places up to 300 m below sea level and active circulation may have penetrated at least that far.

Karst groundwater is held in interconnected, dissolutionally opened cavities. Porosity data for crystalline carbonate rocks of Silurian to Carboniferous age in North America show (primary) porosities ranging from 2.4 to 6.6%, fracture (or fissure) porosities of 0.02–0.03% and conduit porosities of 0.003–0.06% (Ford and Williams 2007). In the Arthur Marble, primary porosity is negligible and most storage is in the secondary porosity of interconnected fissures (bedding planes, joints and faults), but most groundwater transmission is through the tertiary porosity of conduits (water-filled caves). An estimate of total porosity in the Arthur Marble can be obtained by comparing the volume of water storage in the aquifer with the volume of the host rock.

The width of exposed and unexposed Arthur Marble in the Takaka catchment varies from 4 to 12 km, and it extends inland for about 26 km. The boundaries of marble beneath caprock and gravel are uncertain, but the area of marble that drains to TWS is of the order of 141  $\text{km}^2$  (Table 1). Of this about 78  $\text{km}^2$  occurs across karstified marble uplands to the east and west of the valley, approximately 50  $\text{km}^2$  lies below the middle valley floor, and the rest is buried beneath

confining caprock, mainly in the lower valley, amounting to about 30 km<sup>2</sup>. Within the confined part of the aquifer, a groundwater divergence directs most water to TWS with the remainder moving offshore to Golden Bay through about 15 km<sup>2</sup> of marble (that also continues offshore but is not considered here) (Fig. 6).

If the depth of karstification is assumed to be around 300 m, then the host rock volume is ~42 km<sup>3</sup> (or if 400 m depth then ~57 km<sup>3</sup>). Since this has the capacity to store 2.85 km<sup>3</sup> of water, the required interconnected porosity approaches 7% (or 5% with 400 m depth). Even allowing for some water storage being in saturated gravel over unconfined marble (0.2 km<sup>3</sup>) and some in the epikarst (0.1 km<sup>3</sup>), the estimated porosity value of about 6% is still high for a dense crystalline carbonate rock. This can be attributed to high precipitation and alpine runoff that has focussed considerable dissolution potential onto a relatively small wedge of marble in the fault-angle valley.

## Discharge

The principal outflow site for the AMA is Te Waikoropupu Springs. Their combined yield averages about 13.4 m<sup>3</sup>/s but ranges from 14.3 to 5.7 m<sup>3</sup>/s. Almost 10 m<sup>3</sup>/s of mean flow comes from Main Spring and adjacent Dancing Sands Spring (the latter contributing about 20%) with the balance from Fish Creek Springs, although these run dry when Main Spring output falls below about 7.8 m<sup>3</sup>/s (Weir and Fenemor 2017). The 7-day mean annual low flow for the Springs is 7.661 m<sup>3</sup>/s, that being derived from the combined output of the Main and Dancing Sands Springs (McGlinchey 2016).

Submarine springs are known to emerge through the sandy floor of Golden Bay at 1–5 km offshore and at depths of 12–14 m (hydrographic survey in 1957 by the *Lachlan* shown on Oceanographic Chart NZ61, plus photo in Mueller 1987, Fig. 14). Although the source of their waters has not been determined, some submarine springs (as near Rangiahaeta Head) are likely to discharge water from the marble aquifer and would have been exposed when sea level lowered in the Pleistocene. During low sea level, the potentiometric surface of water confined in the marble would also have

fallen, and so Te Waikoropupu Springs would not have operated under such circumstances.

The volume of water discharged by the submarine springs has not been measured, but it can be estimated from the water balance, from the excess of recharge over measured discharge, the surplus providing a guide to the combined submarine outflow plus the overflow of water (occurring during floods) that had previously recharged the marble. Thus, Stewart and Thomas (2008) concluded that total submarine discharge could be about 6.45 m<sup>3</sup>/s, which is equivalent to roughly twice the flow of Fish Creek Springs.

Rain and floodwater generate recharge pulses that pass through the groundwater system and, like a piston, displace water already in storage. Hence new water pushes out and partly mixes with old water but does not itself finally resurge at the springs until some years later. These floodwater pulses are transmitted through conduits under pressure and in the process some water is forced into lateral fissures as the flood peak passes, but later returns to the conduit as the pressure declines. This explains the observation by Stewart and Downes (1982) that under high discharge at the Springs the proportion of old water decreases. This natural pumping encourages the exchange of water in underground stores, especially in the fissure matrix. However, the picture is not simple because pulses transmitted through conduits from different source areas can be expected to arrive at TWS at different times, and so the age of the water sampled may vary with the rising and falling stage. Extended periods of gradually reducing pressure head permits water in distant fissures to drain towards conduits and so be expelled from the system. This results in old water from distant parts mixing with the flow that discharges at the springs.

## Modelling water sources and their contributions to Te Waikoropupu Springs

The purpose of the mass balance model is to establish the relative importance of each input source area, first, as a source of recharge for the AMA and Te Waikoropupu Springs and, secondly, as a source of nitrate load discharged at the springs. Locations of aquifer recharge areas

**Table 2** Recharge source contributions to the outflow at Te Waikoropupu Springs

Source of recharge	Proportion of flow from source contributed to:		Volume of flow (L/s) from source contributed to	
	Main spring	Fish Ck springs	Main spring	Fish Ck springs
Upper Takaka River	0.13	0.34	1253	1190
Waingaro River	0.08	0.26	750	891
Karst Uplands (E&W)	0.70	0.30	6887	1056
Middle Takaka valley	0.10	0.10	1007	361
Total	1	1	9890	3499

**Table 3** Balance of nitrate input and output in the Arthur Marble Aquifer. Arrows schematically represent the input and passage of water and nitrate load through the aquifer to the springs

INPUTS NO <sub>3</sub> -N tonnes/year		Mixing & Nitrification	OUTPUTS NO <sub>3</sub> -N tonnes/year	
Upper Takaka River	4.1	<p style="text-align: center;"><b>ARTHUR MARBLE AQUIFER</b></p>	140.4	Main Spring
Mid Valley Infiltration	131.3			
Karst Uplands E	22.4		46.1	Fish Creek Springs
Karst Uplands W	2.6			
Waingaro River	0.6		88.1	Submarine Springs
Excess by-passing TWS	75.9			
TWS SUM	161		+16%	187
AMA SUM	237	+16%	275	Arthur Marble Aquifer

**Table 4** Comparison of modelled and measured values at Main and Fish Springs

Variable	Discharge, m <sup>3</sup> /s		δ <sup>18</sup> O ‰		Output NO <sub>3</sub> -N concentration, mg/L		Output NO <sub>3</sub> -N load g/s	
	Main	Fish	Main	Fish	Main	Fish	Main	Fish
Measured	9890	3467	- 7.38	- 7.64	0.45	0.42	4.45	1.46
Modelled	9896	3494	- 7.38	- 7.64	0.445 <sup>a</sup>	0.426 <sup>a</sup>	4.44 <sup>a</sup>	1.48 <sup>a</sup>
Modelled/measured	1.001	1.008	1	1	0.989	1.014	0.998	1.014

<sup>a</sup>Assumes 16% nitrification during aquifer transit

are identified in (Fig. 7). Tracer signatures (δ<sup>18</sup>O, NO<sub>3</sub>-N) for each input source were identified earlier with statistical information in Table 1. For the purposes of determining input contributions, some source areas were divided into constituent parts (a and b) to assemble data (Table 1), but

results were later combined to represent the input source area as a whole. For example, the karst uplands (recharge source 3) were subdivided into eastern and western sectors (Figs. 4, 7), because Takaka Hill-Canaan Downs is higher

and wetter than the smaller area of karst uplands on the western side of Takaka valley.

Important constraints on the model are the measured values at Te Waikoropupu Springs of discharge,  $\delta^{18}\text{O}$ , and  $\text{NO}_3\text{-N}$ . The present state of nitrate concentration is defined as the 5-year median to 2021 of  $\text{NO}_3\text{-N}$  at the Main Spring and at Fish Creek Springs, which are 0.45 mg/L and 0.42 mg/L, respectively.

The model is unconstrained in the sense that some of the input areas and values could be estimated differently according to the criteria used, but the model inputs and outputs are still required to balance, measured values are used, and some results are able to be verified against independent evidence.

The model proceeds in three steps, as explained in “**Methods**”, and mass balance calculations were made using an Excel spreadsheet (Appendix). Thus, for example, in step 1 influent flow from the upper Takaka River is shown to contribute 0.13 of the discharge of Main Spring and 0.34 of the flow at Fish Creek Springs (Table 2); and then in step 2, for example, 13 percent of Main Spring outflow is shown to come from upper Takaka River water that has an average  $\text{NO}_3\text{-N}$  concentration in the influent zone of 0.0535 mg/L (Table 1). This accounts for a loading contribution of 4.1 t/year to the 161 t/year of nitrate–N introduced into the AMA (Table 3). Step 3 shows that nitrification as discussed earlier accounts for about 16% of  $\text{NO}_3\text{-N}$  discharged at TWS.

## Results

All the aims outlined in Sect. “**Methods**” have been met. Results of mass balance modelling are shown in Tables 2 and 3 and the performance of the model can be assessed in Table 4 where modelled results are compared with measured values. The ratio of modelled to measured values across all variables shows modelled results always to be within 1% of measured values at both Main and Fish Creek Springs.

The first part of the model uses  $\delta^{18}\text{O}$  as a tracer to determine the proportions of flow derived from the different source areas. This indicates that the Main Spring derives 70% of its outflow (6.86  $\text{m}^3/\text{s}$ ) from the Karst Uplands (E and W), 13% from the upper Takaka River (1.25  $\text{m}^3/\text{s}$ ), about 10% from rainwater infiltration in the middle Takaka valley, and 8% from water losses in the Waingaro River. These discharge component values for the Main Spring broadly confirm earlier estimates by Stewart and Thomas (2008). However, the model shows the proportion of water contributed to flow in Fish Creek Springs to be rather different from those estimated previously. This is partly because water losses from the Waingaro River have been explicitly included as a recharge source. The model indicates that about 34% of the outflow (1.19  $\text{m}^3/\text{s}$ ) of Fish Creek Springs comes from the upper Takaka River and has similar proportions from the

Waingaro River (26%) and Karst Uplands (30%). Infiltrating water from the unconfined part of the middle Takaka valley contributes about 10% of the flow of Fish Creek Springs, the same proportion as found in Main Spring (Table 2).

The excess of recharge over discharge at Te Waikoropupu Springs is 7.09  $\text{m}^3/\text{s}$ . This includes the outflow of submarine springs, plus the overflow of re-emerging recharge water near the artesian boundary during times of flood. The excess is larger than an earlier estimate (6.45  $\text{m}^3/\text{s}$ ) by Stewart and Thomas (2008), but one should note that both incorporate uncertainties associated with determining recharge input values and cannot be independently verified. The modelled isotopic composition of this excess water is  $-8.33\text{‰}$   $\delta^{18}\text{O}$  which is more negative than at TWS because it contains a higher proportion of river water. About 83% of the submarine outflow is derived from the upper Takaka River, which is to be expected because water is probably transmitted along a gravel-filled buried channel and through karst conduits developed during glacial low sea levels. Other contributions to submarine discharge include 9% from the eastern Karst Uplands and 9% from infiltrated middle valley rainfall. In addition, water sourced from about 17.6  $\text{km}^2$  of marble and allogenic granodiorite runoff at the seaward north-eastern end of the catchment (Fig. 7) yields 0.99  $\text{m}^3/\text{s}$  to the groundwater system beyond the TWS catchment. This probably contributes to submarine springs off Abel Tasman Point in Golden Bay.

The second and third parts of the model are concerned with the mass flux of nitrate (Table 3). Comparison of the total nitrate input to the aquifer with the output of nitrate discharged at TWS (186.85 t/year, Table 3) indicates a nitrification increment of 16 percent during groundwater transit. This confirms the estimate made in Sect. “**Nitrates**”. The sum of input flow loadings of nitrate from each recharge source when enhanced by 16% nitrification closely matches the outflow of  $\text{NO}_3\text{-N}$  measured at Main Spring (4.44 g/s estimated compared to 4.45 g/s measured, Table 4) and Fish Creek Springs (1.48 g/s estimated compared to 1.46 g/s measured, Table 4). The results clearly reveal the dominant source of nitrate load discharged at Te Waikoropupu Springs to be the middle Takaka valley (81.5%) and the second largest source to be the karst uplands (15.5%).

Main Spring discharges 140 tonnes per year (t/year) of nitrate–N and Fish Creek Springs 46 t/year, a total output from TWS approaching 187 t/year most being sourced from farmlands of the middle Takaka valley (131.3 t/year input), but the karst uplands also contribute 25 t/year. About 3 t/year of nitrate discharge at TWS can be attributed to the dissolution of marble rock in the aquifer (Young and Hay 2017). Given that another 76 t/year of input bypasses TWS to be discharged via submarine springs, total throughput of nitrate via the marble aquifer is around 275 t/year (Table 3).

Consideration has been given thus far to groundwater flow through the AMA, but the Takaka River also discharges overflow water that is not transmitted via groundwater. Weir and Fenemor (2017) estimated that at Kotinga, near Takaka, the river flow averages 33.4 m<sup>3</sup>/s and carries an annual NO<sub>3</sub>-N load of 132 t/year, and its tributary the Anatoki River discharges 11.9 m<sup>3</sup>/s, conveying 10 t/year NO<sub>3</sub>-N. They concluded that the combined surface water and groundwater components carry a total nitrate–N load from the Takaka catchment of around 699–714 t/year. Results presented here show that about 40% of this is conveyed through the marble aquifer system with 27% of the nitrate load carried from the Takaka catchment being discharged at Te Waikoropupu Springs.

## Conclusions

Prior to human habitation, the Takaka region was forested, and nitrification of dissolved and particulate organic matter would have accounted for most of the base load of nitrate discharged by Te Waikoropupu Springs. Natural values were probably close to those measured in rivers and springs issuing from the forested landscapes of surrounding national parks (<0.2 mg/L NO<sub>3</sub>-N). European settlements, forest clearance and farming had continued for about a century before the first samples of Main Spring waters were measured in the early 1970s when nitrate concentrations were around 0.3 mg/L (Michaelis 1976). They now exceed 0.45 mg/L NO<sub>3</sub>-N (Moreau 2021).

This study investigates by means of mass balance modelling both the proportions of flow provided by input sources sustaining the springs and the amounts of nitrate load contributed to the aquifer by each source. The proportions of flow are shown in Table 2 which indicates 13% of the flow to come from the upper Takaka River and 70% from the surrounding karst uplands. Taking account of these proportions and comparing measured and modelled nitrate concentrations in source waters and output flows, it is concluded that nitrification during transmission of groundwater contributes about 16% of the nitrate concentration measured at the springs. The modelling indicates that the main input sources of nitrate appearing at the springs are 82% from farmed lands of the middle Takaka valley, 16% from surrounding Karst Uplands and less than 3% from the upper Takaka river.

Since the average flow-through time of the aquifer is 7.9 years (Stewart and Thomas 2008), but the model assumes instant mixing, it is evident that the above contributions reflect past inputs. So if, for example, there have been changes in farming practices in the interim, any water quality effects will emerge over the next few years.

## Implications for aquifer management

The variable but relatively long residence times of water in different parts of the AMA have considerable implications for water quality management. The relatively unvarying values of oxygen-stable isotopes in the waters emerging at TWS show that the groundwaters from different input sources are relatively well mixed (Stewart and Williams 1981). Thus, if unwanted contaminants penetrate the aquifer, their residence time will be long and they will take even longer to dispel, because of the exponential process of freshwater recharge, mixing and flushing. The implication is that the mismanagement of water quality by this generation will take at least another generation to rectify.

If the water quality at Te Waikoropupu Springs is to be managed in a condition that is as close to natural as possible (given prior land use changes), then the recharge zones (Fig. 7) must have land use practices and vegetation cover that permits both diffuse and point source infiltration to be of high quality. The 49 km<sup>2</sup> of agricultural land across floodplains and terraces in the middle Takaka valley upstream of the artesian boundary (Table 1) is the most critical area from the point-of-view of water quality management because it is part of the unconfined aquifer and an area of relatively intensive farming. Although only 10% of discharge at TWS is derived from that source, it introduces most of the NO<sub>3</sub>-N in the groundwater (and could be associated with other agricultural chemicals). Modelling indicates that it would require a 38% reduction of leachate concentration in this part of the Takaka valley to bring NO<sub>3</sub>-N values back to levels recorded by Michaelis (1976) at Main Spring. If inorganic fertilizers are the dominant source of nitrogen introduced in this area, then this could be investigated further by analysing water samples for δ<sup>15</sup>N, a method used successfully for determining the source of nitrate contamination in karst springs (Panno et al 2001; Katz 2004).

In addition to water quality considerations, the volume of water resurging at Te Waikoropupu Springs is one of its outstanding characteristics and requires careful management, especially at low flows when the demand for abstraction is highest. The amount of spring flow depends on the state of aquifer storage. When water loss in the channel leaves the middle reaches of Takaka River dry, the aquifer is not full and discharge from TWS is sustained by draw-down of storage supplemented by ongoing recharge. Under such conditions, the rate of storage depletion is increased by simultaneous water abstractions or water capture upstream. When the river's middle reaches are dry (> 100 days per year), then water taken from the recharge zone or from bores into the marble aquifer depletes the flow in the groundwater system by a similar amount. The downstream impact of that is distributed roughly equally between the springs and offshore outflows, and impacts on spring flow are maximised under



such conditions. By contrast, when the middle Takaka River is flowing, the marble aquifer is full or close to capacity, and any water abstractions upstream or from bores into the confined aquifer will have no discernible effect on the outflow of the springs. Thus exploitation of water resources in the catchment needs careful consideration of location, amount and timing to achieve an acceptable balance that takes full account of competing agricultural, ecological and cultural needs and demands.

Compared to many other parts of the world there would seem to be abundant water resources for exploitation in the Te Waikoropupu Springs' aquifer with an estimated groundwater volume of 2.85 km<sup>3</sup>, an output averaging 13.4 m<sup>3</sup>/s and abstractions of less than 0.7 m<sup>3</sup>/s. But TWS are in a national conservation reserve and are recognised for their spiritual value and considered wahi tapu by the indigenous Maori people. Their natural heritage value is also highly appreciated by many others, especially for the abundance and clarity of the 'boiling' spring waters. Consequently, the springs are seen by many as a national treasure that should not be degraded, requiring effective protection and sustainable use. But to achieve this requires careful management of the entire marble aquifer and its recharge sources, including an understanding of the consequences for the springs of water exploitation upstream. The results presented in this paper are a contribution to those requirements.

## Appendix

Spreadsheet showing recharge contributions to Te Waikoropupu Springs when adjusted to meet measured  $\delta^{18}\text{O}$  and  $\text{NO}_3\text{-N}$  constraints at input and output sites, and assuming a 16 percent nitrification effect during groundwater transmission through the marble aquifer.

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