Estimating Time Lags for Nitrate Response in Shallow Southland Groundwater

Technical Report

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Executive Summary

This report presents a region-wide study to estimate the length of time it will take for changes in land practices to reach the groundwater table. The aim of the study is to obtain preliminary answers to two questions. Firstly, if leaching limits are set, how long would it take to see an improvement in shallow groundwater quality? Secondly, have groundwater nitrate concentrations reached equilibrium from recent dairy expansion in the region, or could we expect future increases?

We have used a methodology that provides a balance between the detail and generalisation that is required for a regional-scale study. The first step involves the calculation of drainage through the soil profile by a daily soil moisture balance model. The travel of this water through the unsaturated zone was simulated using water retention curves. These curves enable an estimate of the average volumetric water content of the unsaturated zone. The percentage saturation can then be used to calculate the travel time through the unsaturated zone if effective porosity, depth to the water table and annual average soil drainage are known. A time for mixing in the uppermost part of the aquifer has also been calculated.

The calculations indicate that about 80% of the region is expected to have a transit time of less than five years, and 90% less than two years. By doing a large number of calculations we are able to present the results as maps of travel time through the unsaturated zone and shallow groundwater. The maps indicate that older transit times are associated with mid-Pleistocene outwash gravels. These deposits have lower permeability, and are also located at higher elevations above the rivers.

The results of our modelling indicate that shallow groundwater beneath properties in most of Southland will respond rapidly to a reduction in leaching rates. Most of Southland's shallow groundwater (close to the water table) is expected to show some response to a change in farming practices within five years. Large future increases in nitrate concentrations are only expected in discrete areas beneath older more elevated outwash gravel deposits.

We have good confidence in the relative values of the time lag estimates, which enable responsive areas to be managed differently to areas of longer travel times. We have less confidence in the absolute values of the travel times. Preliminary validation of the modelled values has been carried out by comparing transit times with tritium ages from shallow groundwater. The initial results are encouraging, but a proper comparison cannot be made without the availability of shallow groundwater samples from wells that are not used for production.

1. Background

1.1 Introduction

This study has been initiated to answer two main questions:

1. If land use practices are changed in Southland, how long will it take before there is a change in shallow groundwater quality?
2. Have groundwater nitrate concentrations in Southland reached an equilibrium, or can we expect a "time bomb" sometime in the future?

Answering these two questions involves estimating particle travel times through unsaturated media and into shallow groundwater. The approach taken in this report is to calculate the vertical travel time of a particle to make its way from the surface to the uppermost, rainfall-recharged portion of each groundwater zone. The approach is one-dimensional (vertical), and its simplicity allows the calculation of nitrate travel time at a large number of points. A large number of calculations enable us to interpolate the results and display them spatially as response time maps. These maps allow comparisons to be made between vertical time lags, land use, measured nitrate concentrations, and environmental tracer results.

Because of the regional nature of the project, the approach we have taken is quite generalised. The input data we have used is derived from regional-scale maps, and the calculations made assume a number of simplifying assumptions. This generalised approach allows us to generate regional-scale maps of time lags which may be used for informational purposes.

There are three main components to the project methodology:

1. calculating a daily soil moisture balance using the Rushton (2006) method. This characterises drainage through the soil, and its variability;
2. estimating the vadose zone transit time. This estimation is based on soil drainage, depth to the water table, and the average calculated vadose zone water content;
3. estimating a nitrate mixing time for the upper, dynamic portion of the aquifer.

The combination of these three components provides an average travel time from the surface into shallow groundwater. The resulting data can be interpolated to produce a contour map of estimated time lags for nitrate accrual in shallow groundwater.

This report is divided into five main sections:

1. Background
2. Methodology
3. Parameterisation
4. Results
5. Conclusion

1.2 Defining time lags

The time lags associated with diffuse discharges consist of three main components (after Meals et al. 2010):

1. time for a land-use change to have effect (social response);
2. time for the effect to be delivered to the water resource (in this case shallow groundwater);
3. time for the water body to respond to the effect (cumulative environmental impact).

The first of these components depends on changes in policy, and/or in land management. For example, the setting of limits for nitrate discharges could be written into a regional water plan, a process which would take some time. It may also take time for land managers to implement changes in their management practices. While these components cannot be easily estimated, they do need to be considered in the overall water quality response time.

This report deals specifically with the second of these time lag components. It assumes that a land-use change has occurred, and is starting to have an impact on soil water quality. This time lag can be considered as the *"response time for an individual change in land use to be delivered to shallow groundwater"*.

This definition distinguishes an individual effect from a cumulative water body response at a specific location (e.g. monitoring well, spring discharge, or base flow contribution to river).

In conceptual terms, the individual effect is characterised by a one-dimensional, gravity-driven vertical flow path from the soil into the top of the aquifer. By contrast, the cumulative water body response occurs in more of a horizontal direction along a groundwater flow path, and nitrate can be accrued with travel distance.

A distinction also needs to be made between groundwater pressure response and the actual flow of a particle of water through the saturated or unsaturated zone. When a recharge event occurs, such as a high flow event in a river, there is an associated rapid groundwater response in the form of a pressure wave that travels across the aquifer. However, the individual water particles infiltrated during a recharge event will take considerably longer to travel through the aquifer of unsaturated zone than the pressure wave. The reason for this is that particles need to negotiate the pore spaces of the sediment. Not all of the pore spaces will readily allow mobility, particularly around fine-grained sediments. For this reason, we use different analytical approaches to calculating pressure response times and contaminant transport response times.

The time for nitrate draining though the soil to be delivered to the water table depends on a number of factors including:

- rate of drainage through the soil profile;
- depth to the water table;
- water content of the unsaturated zone;
- physical properties of the unsaturated zone;
- any chemical attenuation processes that may occur.

This study also includes an estimation of time taken to penetrate or mix in the uppermost, rainfall recharged portion of the aquifer.

1.3 Conceptual Setting

The modelling of nitrate movement from the surface into shallow groundwater requires three main calculation steps:

1. calculation of rainfall recharge (drainage thought the soil profile);
2. modelling the rate of flow through the unsaturated zone;
3. a calculation to allow for time to mix in the shallow aquifer.

These calculations are placed in a conceptual hydrological setting in Figure 1, where the circled numbers refer to the calculation steps listed above. The rainfall recharge calculation provides the input flow rate for both the unsaturated and saturated calculations. The unsaturated and saturated time lag calculations are independent of each other, but when added together provide a total time lag estimate to the uppermost part of the aquifer.

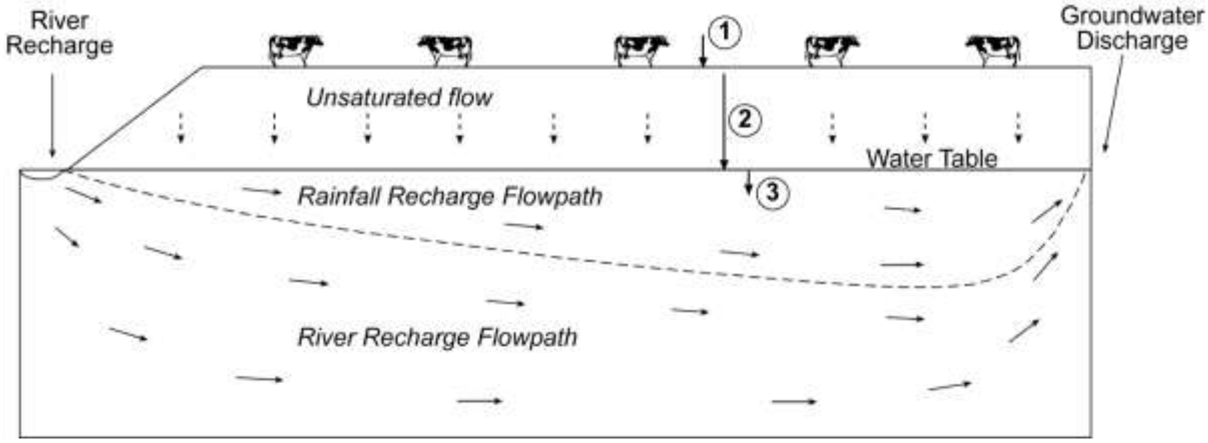

Figure 1: Conceptual hydrological setting for the time lag calculations

The calculations for the saturated zone are only carried out for a very thin part of the unconfined aquifer. The intention is to give an indication of the time it will take for leachate from the vadose zone to become mixed into shallow groundwater. It is hoped that a more comprehensive saturated flow assessment involving horizontal flow will be carried out in the near future.

2. Rainfall Recharge Methodology

The primary driver of travel time through the soil and unsaturated zone to the water table is vertical drainage of water through the soil profile. The estimation of long-term soil drainage rates through the soil is best approached with a daily soil moisture balance. The record from the soil moisture balance model can be used to derive annual statistics of soil drainage rates (or groundwater recharge). These soil drainage fluxes can then be used as an input dataset for estimating the drainage time through the unsaturated zone to the water table for a given pore water content.

The Rushton et al. (2006) soil moisture balance model was used to estimate soil drainage for this study. An outline of how this method works is provided in Appendix 1. The Rushton model performs well compared to other soil moisture balance methods commonly used in New Zealand (Wilson and Lu, 2011, Zemansky and White, 2012). It also has the advantage of being simple to use and modify. The Rushton model also predicts the daily soil moisture deficit, so it has useful applications for the management of irrigation and effluent discharge.

As a word of caution, any soil moisture balance model is strongly dependent on the soil hydraulic properties used. A comparison of cumulative model predictions with local lysimeter data is desirable if possible to verify the soil properties used.

The soil moisture balance model was the most time-intensive component of this study. The large amount of data involved for the regional scale simulation, and the care taken to parameterise the soil properties took up at least half of the total project time. The actual soil moisture balance calculations were automated using a VBA script. This meant that rainfall and soil drainage statistics for the 2,373 unique soil, rainfall and PET combinations could be carried out in about 15 minutes of computation time.

2.1 Rainfall

Spatial representation

The simulation of soil moisture balance with the Rushton model requires spatially distributed daily rainfall values for the whole modelled area. Unfortunately, the spatial extrapolation of daily rainfall values from the rain gauge measurements cannot be done through a simple spatial interpolation method. Precipitation varies greatly with several factors such as elevation, proximity to the coast or to mountains rather than with distance uniquely.

To accommodate this spatial variability we adopted the national precipitation grid developed by NIWA. The NIWA grid models average annual rainfall values at a 500 m scale and incorporates the major drivers of precipitation, such as orographic effect and coastal proximity. In order to limit the number of final hydrological response units (and therefore the complexity of the model) while conserving an acceptable spatial resolution, the 500 m grid was re-sampled to a 5 km scale. This was achieved with a bilinear interpolation technique, where each new pixel is assigned a value based on its four direct neighbours.

In order to define a zone of influence for each rain gauge, Thiessen polygons were created around each of them in ArcGIS[1]. The Thiessen polygons were then intersected with the re-sampled 5 km annual rainfall grid. This resulted in each groundwater zone being assigned one or parts of several Thiessen polygons containing daily rainfall values, and one or several parts of 5 km pixels containing average annual rainfall.

In order to extrapolate the daily rainfall measurements from a rain gauge to each 5 km pixel within its zone of influence (the Thiessen polygon), a ratio was calculated between the annual average rainfall value of each pixel and the annual average rainfall value calculated from daily measurements at the rain gauge. All calculations were performed with R - Statistics package version 9.3.1.

Daily rainfall values were finally assigned to each pixel, calculated as follows:

Daily rainfall in pixel = (Average annual rainfall ratio) x (Daily rainfall at rain gauge)

The output of this procedure is a 5 km grid of daily rainfall estimates for whole the region.

Irrigation

Irrigation can increase drainage volumes from spring to autumn, particularly if its application is inefficient. Irrigation is not widespread in Southland, but there are areas where irrigation is more common such as Edendale and Riversdale. Irrigation has not been included in this study because the mapping of land and irrigation use is outside of the budgetary scope of the project.

2.2 Potential Evapotranspiration

The PET data used for this study are from NIWA's climate database. The values we have used are PET estimates derived from the Penman-Monteith equation (see Allen at al. 1998). A crop co-efficient has not been applied, since, for simplicity, the crop is assumed to be pasture.

[1] Thiessen polygons delineate a zone of influence around a point, so that any location within that polygon is closer to its central point than to any other.

3. Unsaturated Zone Methodology

This section describes the methodology used to calculate the pore water content and transit time for the percolation of water through the unsaturated (vadose) zone. The methodology has been developed with consideration that the investigation is of a regional nature. It calculates vertical transit times for an area of some 500,000 hectares in a region that has negligible data on vadose zone properties. Accordingly, the model resolution chosen for this study is quite coarse, and the results are expected to be reliable at a scale of around 25 hectares (500 m x 500 m).

There are many factors to consider for selecting the methodology and resolution of the modelling carried out. Ultimately, it is the availability of data which requires model and spatial simplification to be made. Many of the factors important to detailed studies are not applicable at a regional scale. For example, Dagan (1984) notes that a parameter that is critical at the site scale, such as dispersion coefficient, becomes meaningless at the kilometre scale because of the uncertainties involved in characterising variability. The model applied also needs to match the available information. For example, Sousa et al. (2013) found that sophisticated and simplified methods gave similar results when the same knowledge of hydraulic conductivity was assumed.

The approach to a regional-scale assessment therefore needs to be pragmatic, requiring generalisation, and some judgement of which methodologies and parameters are most suitable for the scale of the subject matter.

With these thoughts in mind, a number of assumptions have necessarily been made:

- drainage occurs in response to rainfall infiltration;
- flow is vertical (there are no flow barriers or artificial drainage);
- flow is steady state;
- hydraulic properties at each site are homogeneous and isotropic;
- the movement of nitrate is controlled by advective flow (seepage);
- nitrate behaves as a conservative solute (there is no sorption or denitrification);
- soil is "unmodified" (the soil parameters used represent undamaged, inherent properties).

Sorption (adherence) to particles can slow down the transport of solutes in the vadose zone. Close (2010) notes that nitrate does not tend to sorb within the vadose zone because nitrate and the medium it travels through are both negatively charged. Denitrification does occur within the soil horizon and can occur under certain conditions in the saturated zone. Studies overseas have shown that denitrification does not tend to occur in the unsaturated zone because of the abundance of oxygen (e.g. Cannavo et al 2002).

The assumption of artificial drainage does not apply to areas of Southland where the water table is shallow. This is the case for much of the coastal land. Unfortunately, maps for areas of artificial drainage are not available so we cannot readily remove them from our study. However, these areas are expected to have quite rapid vadose zone transit times even without drainage (less than a year) because of their association with a shallow water table. Thus, the occurrence of artificial drainage does not significantly impact the results of our calculations, but it does constrain the resolution of our results (years rather than months).

The assumption of advective flow is most valid for lower permeability sediments. If the sediments are coarse reworked gravels like on the Canterbury Plains, there is a greater possibility of non-linear (macropore or bypass) flow, which is difficult to quantify. Departures from advective flow can also occur as sediments approach saturation. The effect of saturation is to produce a large increase in the hydraulic conductivity of the sediments, which results in rapid flow through the larger pore spaces. This acceleration of the vadose zone travel time could be a common occurrence in the wetter areas of Southland, particularly during winter and spring.

Vadose zone travel times are expected to be more rapid in the wetter parts of Southland regardless of whether macropore flow occurs or not. For the purpose of this report, we consider it more useful to identify and characterise areas of longer time lags where the responses to land use change is expected to be quite slow. These longer time lags are expected to occur in the drier areas of Southland and are less likely to be subject to bypass flow caused by saturation.

It has been observed by Close (2010) that the centre of mass in tracer experiments does not vary as much as the rapid front or slower tail of the pulse. This allows us to derive a good first approximation of contaminant travel time using advective flow equations. The impact of bypass flow, if it does occur, would be that the effect of a land use change would be observed more rapidly than expected but the full effect would take longer as there would be a long tail related to the previous land use.

Dann et al. (2010) found that the total transporting water contents (matrix and macropore domains) were similar to the observed water contents in Canterbury vadose zone profiles. These observations led Close (2010) to use a measured 7.4% water content (Dann et al. 2009) for estimating vadose zone travel time in similar sandy gravels throughout in Canterbury.

The approach we have adopted for this study is similar to that of Close (2010) who made transit time estimates for select sites in Canterbury. Our method differs in that we have estimated the water content of different sediments using pedotransfer functions and soil water retention models to generate transit time values for the whole region. The reason for adopting pedotransfer functions is that the Quaternary sediments of Southland are relatively diverse, typically lower permeability than in Canterbury, and we have no data of their unsaturated characteristics.

Our vadose zone transit time calculations assume that the passage of water received from soil drainage moves to the aquifer by piston flow. This means that a volume of inflow from soil drainage displaces an equivalent volume from the vadose zone into the aquifer by a pressure response. The steady state transit time then becomes a function of the average volumetric water content of the vadose zone, effective porosity, and the depth to the water table:

$$t_{vz} = \frac{(\bar{\theta} \cdot \varphi \cdot l\,)}{R}$$

Where $\bar{\theta}$ is average volumetric water content (% or v/v), l is depth to water table below the soil profile (m), φ is effective porosity, and R is the soil drainage rate of land surface recharge (m/year).

Unfortunately there are very few values reported for volumetric water content of the vadose zone in New Zealand. The most comprehensive study reports water contents for a number of Canterbury gravels (Dann et al., 2009). These sediments are quite coarse, even by New Zealand standards, and are only likely to be equivalent to the more permeable gravels in Southland. The water contents ranged from 5 to 10% with a weighted mean of 7.4% for all samples. The range of water contents in Southland is expected to be higher than Canterbury because of the greater soil drainage rate and lower overall permeability of the sediments.

Because the hydraulic properties of the vadose zone are not well known for the Southland region, there is considerable uncertainty in any estimates of vadose zone water content. However, the range of expected water content values (around 10 to 40%) is small compared to the variability in depth to the water table and the annual soil drainage rate. This means that uncertainty in vadose zone properties will not contribute largely to the uncertainty of the transit time estimates. Accordingly, we can make reasonable estimates of vadose zone transit time by applying generalised estimates of vadose zone properties.

3.1 Theoretical background

The greatest challenge to estimating transit time in the unsaturated (vadose) zone is determining its average pore water content. This study makes use of reference water retention curves that have been developed for different sediment textures. These water retention curves describe the relationship between water content and capillary pressure for different sediments (Figure 2). Normally volumetric water contents are measured in the field using neutron probes, although such data is not available for the Southland region. In the absence of measurements, we can use models of water retention curves to estimate the average water content of the vadose zone.

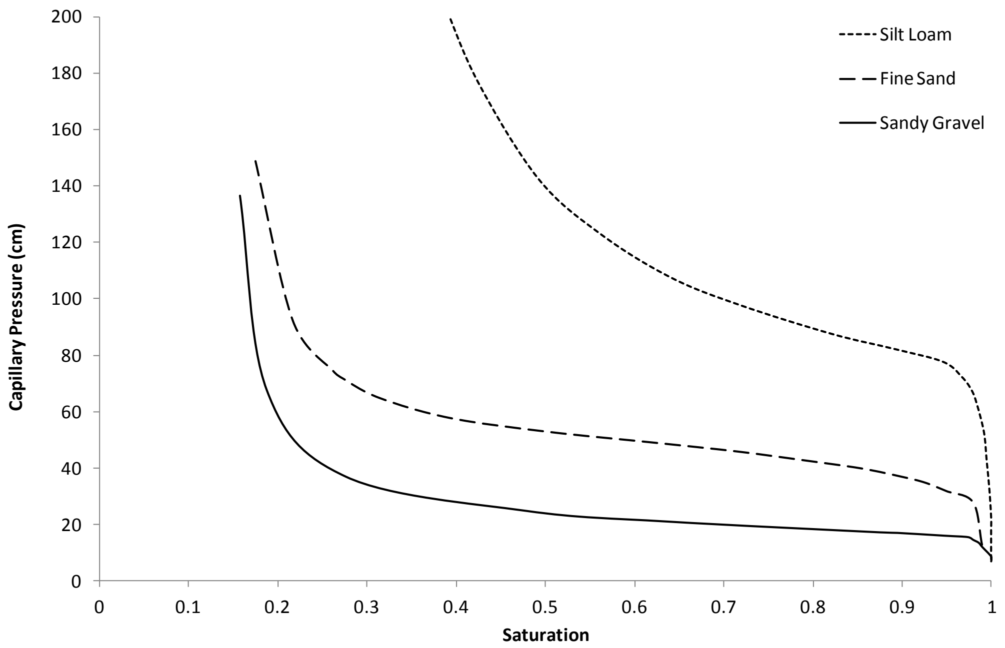

Figure 2: Examples of water retention curves

Water retention curves characterise how readily sediment will drain in response to land surface recharge. Because the vadose zone is semi-saturated, any inflow of water beneath the soil needs to overcome capillary tension in order for drainage to occur. It can be seen in Figure 2 that the amount of water held in the vadose zone is largely determined by soil texture. The reason for this is that the hydraulic conductivity of the medium increases rapidly as the degree of saturation increases.

Sand and gravel drain quickly as capillary tension decreases compared to a silt or clay. This is because coarse sediments have a lot of large pore spaces with water that is only weakly held by capillary tension. This characteristic allows coarse sediments to recharge and drain quickly under gravity. Conversely, silts and clays respond more slowly than sands and gravels, but their higher capillary tension enables them to stay saturated for longer. This means that the hydraulic conductivity of clay will be greater than that of gravel at low degrees of saturation, enabling the clay to continue to drain for a longer period of time.

The shape of water retention curves for different sediments has been modelled by a number of authors including Brooks and Corey (1964), Campbell (1974), Mualem (1976), and Van Genuchten (1980). For this report we have decided to adopt the two methods most commonly used overseas, the Brooks-Corey and Van Genuchten models. More complex models are available, such as the Richards equation. For this study it was decided that a more complex model would not add any value to the vadose zone transit time estimates because of uncertainties in parameter values.

In terms of applicability, the Brooks-Corey function is quite simple and easy to parameterise. This model performs well at high and medium water contents, as opposed to an arid environment. Predictions made with the Brooks-Corey model do tend to deviate from data observations in fine sediments as saturation is approached (Rossi and Nimmo, 1994). However, this function performs well in coarse textured sediments (Valiantzas, 2011).

By contrast, the Van Genuchten model is the approach most commonly used in the literature. One of its advantages is that the saturated hydraulic conductivity does not need to be estimated. However, two other parameters α and η do need to be estimated, which can easily be done if the sediment texture is known.

Rawls and Brakensiek (1985) used regression analysis to develop exponential relationships between sediment texture and water retention. These relationships allow some variables in both equations to be parameterised if the percentage of sand, clay, and porosity can be estimated. Accurate versions of the regression equations are provided in Chapter 5 of Rawls et al. (1992). The regressions are valid for sand contents from 5 to 70% and clay contents from 5 to 60%, a range which covers the majority of sediment textures in Southland.

Brooks-Corey method

The methodology of applying the Brooks-Corey method to determine the average water content of the vadose zone is provided by Charbeneau and Daniel (1992). The permeability of the vadose zone changes with saturation, with higher permeability occurring at higher water contents. The Brooks-Corey power functions describe the drainage velocity as:

$$v = \frac{R}{\theta_r + (\varphi - \theta_r)(R/K_s)^{\lambda/(3\lambda+2)}}$$

Where v is pore water velocity (m/d), R is land surface recharge or soil drainage (m/d), φ is effective porosity, θ_r is the residual water content[2], K_s is saturated hydraulic conductivity (m/d),

[2] Four standard water contents are routinely measured and used in soil studies:

Name	Notation	Suction pressure (J/kg or kPa)	Typical water content (vol/vol)	Conditions
Saturated water content	θ_s	0	0.2-0.5	Fully saturated soil, equivalent to effective porosity
Field capacity	θ_{fc}	-33	0.1-0.35	Soil moisture 2-3 days after rain or irrigation
Permanent wilting point	θ_{pwp} or θ_{wp}	-1500	0.01-0.25	Minimum soil moisture at which a plant wilts
Residual water content	θ_r	$-\infty$	0.001-0.1	Remaining water at high tension

and λ is the pore size distribution index. Both the residual water content θ_r and the pore size distribution λ are based on the texture of the sediment. These values are calculated with the Rawls-Brakensiek regression equations and require an estimate of the percentage of clay and sand in the sediment, and also the effective porosity.

The drainage velocity can be used to determine the average volumetric water content of the vadose zone:

$$\bar{\theta} = \frac{R}{v}$$

This is expressed either as a depth (e.g. mm/m), or as a percentage. More permeable sediments tend to have lower volumetric water contents than fine grained sediments. Having estimated the average volumetric water content, the depth to the water table, and rate of soil drainage, we can then calculate the steady-state vadose zone transit time (t_{vz}).

Van Genuchten method

The approach for estimating transit time with the Van Genuchten equation is outlined well by Souza et al. (2013). Saturation in the vadose zone is calculated as a function of pressure head as follows:

$$\theta(\psi) = \theta_r + \frac{1 - \theta_r}{[1 + \alpha \cdot \psi^{\eta}]^{(1-1/\eta)}}$$

Where θ is water content at pressure head ψ (m), and θ_r is residual water content, and parameters a and η are empirical coefficients. Carsel and Parrish (1988) related the two empirical coefficients to the Rawls-Brakensiek (1985) regression equations as follows:

$$\alpha = {h_b}^{-1} \qquad \text{and} \qquad \eta = \lambda + 1$$

Where h_b is air entry or bubbling pressure, and λ is the pore size distribution index as used in the Brooks-Corey method. Both of these terms require estimates of effective porosity, and the percentage of sand and clay to populate the Rawls-Brakensiek regressions.

The advective travel time is then calculated as:

$$t = \int_0^l \frac{\varphi(z) \cdot \theta(z)}{R} dz$$

Where l is the thickness of the vadose zone (m), φ is effective porosity, θ is the average volumetric water content, and R is land surface recharge or soil drainage (m/d).

3.2 Monte Carlo Approach

The uncertainty of vadose zone properties can be accounted for by calculating pore water contents using a range of potential parameter values. This has been done for this study using a Monte Carlo simulation, which is a technique commonly applied to the vadose zone.

The Monte Carlo approach involves setting a probability distribution function for each parameter (e.g. normal or log-normal), including a mean and standard deviation of expected values to define its dimensions. The probability distributions for each parameter are then populated with random numbers, and a large number of time lag calculations (thousands) are carried out. This produces a new probability distribution for the time lag estimates, and provides a method to include parameter uncertainty in the vadose zone water content estimates.

4. Saturated Zone Methodology

This section of the report describes a method for calculating the time taken for nitrate from the vadose zone to mix with the upper most dynamic part of the shallow aquifer. Movement through groundwater is usually slower than movement though the vadose zone, particularly if the saturated flow is directed downwards.

To describe vertical movement of nitrate into shallow groundwater a mixing depth equivalent to a year of annual rainfall recharge has been selected. This mixing depth has been calculated by dividing the mean annual rainfall recharge by the effective porosity for each site. This depth is equivalent to 1 m ±0.5 m for the majority of sites used in this study. It is expected that groundwater in this interval will be entirely recharged from the land surface. Exceptions to this rule will apply in close proximity to streams.

Figure 3 shows the conceptual model for the saturated zone modelling undertaken for this study. For a homogenous unconfined aquifer with uniform infiltration and the boundary conditions presented in Figure 3, groundwater at a specific depth (t_1, t_2...) is all equal in age (Appelo & Postma, 2005). This simple conceptual model allows us to easily make analytical calculations of vertical groundwater travel time at a regional scale.

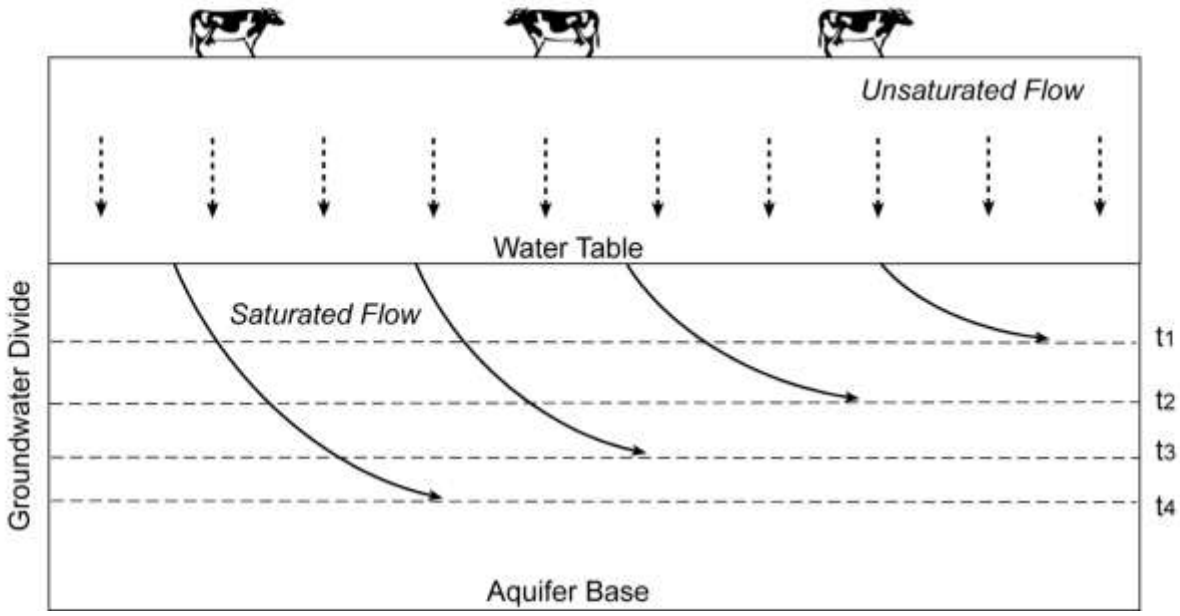

Figure 3: Conceptual model for saturated zone calculations

The boundary condition requirements for Figure 3 can be satisfied if the model is positioned within a regional setting at a shallow level (shown in Figure 1). The groundwater divide for the rainfall recharge pathway can be set at the river. The base of the aquifer can effectively be set where the flow lines are close to horizontal. These boundary configurations imply that a simple model will give a good estimate of vertical groundwater travel time for shallow depths. However, a simple model will not provide reliable results in the vicinity of river recharge and discharge areas. This is because groundwater flow paths in these areas diverge and converge respectively and deviate markedly from the horizontal.

A simple analytical model will now be presented that is based on the boundary conditions shown in Figure 3. Vertical flow through the uppermost portion of an unconfined aquifer that is recharged from the land surface can be approximated by the following equation:

$$t_{sat} = \ln\left(\frac{D}{D-d}\right)\frac{D\varphi}{R}$$

Where *D* is aquifer saturated thickness, *d* is depth below water table (m), *R* is land surface recharge (m/year), and φ is effective porosity (Appelo & Postma, 2005). For this estimation the aquifer is assumed to be homogeneous, and horizontal flow velocity is equal with depth. These assumptions can be considered to be reasonable for the purposes of providing regional estimates of saturated time lags. The reason for this is that at a regional scale, most aquifers have horizontal dimensions that are far greater than the vertical (kilometres compared to metres).

However, the horizontal flow assumption is not valid in the vicinity of surface water bodies. If we looked at a profile through an aquifer that is recharged by river losses, we would see groundwater flow lines diverging away from the river. Conversely, we would see flow lines converge as they approach an area where a groundwater contribution to river baseflow occurs. For this reason, our saturated mixing time calculations are not valid in the vicinity of surface water bodies. Calculations in these areas can be made with more complex 2D or 3D flow equations. This is outside the scope of this report, although we hope to calculate horizontal time lags in the near future.

5. Model Parameterisation

5.1 Rainfall Recharge

Record duration

Ideally, the minimum duration of rainfall and PET records for characterisation of long-term soil drainage is 28 years. The reason for this is that shorter term records are likely to be influenced by the southern oscillation in the Pacific Ocean. El Niño/la Niña events typically last up to seven years. If we assume that the southern oscillation is a good indicator of long-term climate variability then we would need at least 14 years of record to cover a full southern oscillation pair. This limits the number of sites which have records of suitable length for this study.

Rainfall records

Scrutiny of the rainfall dataset indicated that we could model with a period from 1 July 1991 to 1 March 2013 while retaining good spatial coverage. This dataset covers a period of over 21 years, and therefore includes at least three El Niño/la Niña cycles.

Rainfall records for each site were either downloaded from NIWA's Cliflo website, or extracted from the Environment from the Environment Southland rainfall database. Sites with short records or a large number of missing values were missing values were not used. The final list of sites used is provided in

Table 1.

Table 1: Rainfall sites used for each groundwater zone
Key: ***full record; good correlation; incomplete or poor correlation***

Groundwater Zone	Area (Ha)	Core Sites	Secondary Sites
Castlerock	6,800	Oreti at Lumsden	
Cattle Flat	2,638	Glenlapa	
Central Plains	26,257	CPA at Heddons, Inver Aero, Otahuti	
Edendale	7,529	Mataura at Tuturau	
Five Rivers	13,795	Five Rivers Stn, Oreti at Lumsden	
Knapdale	8,185	Gore, Waimea, Otama	Mandeville
Longridge	4,393	Riversdale (Liverpool)	
Lower Aparima	34,327	Otahuti	
Lower Mataura	40,084	Mataura at Tuturau, Tiwai, Gore, Woodlands	Waimea, Inver Aero
Lower Oreti	41,155	Winton, Winton Dam, Inver Aero, Dipton	Otahuti
Lower Waiau	35,218	Monowai, Orepuki, Lillburn	Waiau at Clifden, Round Hill
Makarewa	78,924	Woodlands, Waihopai, Winton, Gore	Winton Dam, Otahuti, Waimea
Orepuki	13,699	Orepuki	Round Hill
Oreti	6,128	Lumsden	Oreti at Lumsden
Riversdale	10,342	Riversdale (Liverpool)	
Te Anau	56,478	Oreti 3 Kings, Plains Stn, Manapouri, Whitestone	
Tiwai	2,489	Tiwai	Inver Aero
Upper Aparima	49,290	Hamilton Burn, CPA at Heddons	Otahuti
Waihopai	73,965	Inver Aero, Tiwai, Waihopai, Woodlands	
Waimatuku	26,889	CPA at Heddons, Otahuti, Inver Aero	
Waimea Plain	25,200	Lumsden, Kaweku, Waimea	Oreti at Lumsden
Waipounamu	3,213	Riversdale (Liverpool)	

Groundwater Zone	Area (Ha)	Core Sites	Secondary Sites
Wendon	4,066	Otama	Waimea, Mandeville
Wendonside	8,769	Otama	Waimea, Mandeville
Whitestone	2,277	Whitestone	Plains Stn, Manapouri

Most datasets had some missing values, and these required estimation to develop a complete record for the site. This was achieved by correlating the record with a nearby site located within the same orographic rainfall zone. The nearby site(s) with the best correlation was chosen and corrected values, derived using linear equations, were used to fill gaps. In many cases more than one nearby site was used to fill gaps to complete a single site record. Due to the lack of complete data and poor correlations with nearby sites some sites were discarded.

Some key sites were not used because of poor quality data. In particular, the Kingston record was very poor, which meant that an assessment of the Upper Mataura was abandoned. The Lower Aparima rainfall record was also very poor, but in this case there were other nearby sites that could be used.

Potential evapotranspiration records

Five sites were used for this study to create continuous PET records from 1991 onwards. Table 2 lists the sites used to generate a complete PET record for this study. The records for Gore and Invercargill consisted of two sites quite close to each other. These records were combined to create composite records for these two areas.

Table 2: List of PET sites used for this study

PET Site	Site Number	Easting	Northing	Start	End	Duration (years)
Gore Aws	5778	1282189	4885125	14-Jul-86	2-Mar-13	26.7
Gore Grasslands DSIR	5780	1282576	4885145	2-Jan-72	1-Nov-86	14.8
Invercargill Aero	5814	1241176	4849147	2-Jan-60	2-Mar-13	53.2
Invercargill Aero Aws	11104	1240108	4849575	27-May-95	2-Mar-13	17.8
Manapouri Aero Aws	5430	1181660	4943814	29-Oct-91	2-Mar-13	21.4
Tiwai Point Ews	5823	1245776	4830511	5-Jul-91	2-Mar-13	21.7
Winton 2	5768	1239273	4878067	2-Jan-72	2-Mar-13	41.2

Most of the sites had incomplete records, which required a process of filling in the missing values. Fortunately PET is cyclic on a seasonal basis and is therefore fairly predictable, being primarily driven by energy received from the sun. This seasonal predictability enables a reasonable estimate of the missing values to be made.

To do this we made correlations between pairs of PET sites by plotting median PET values for each day of the year (day 1 to 365) for the full available record at the two sites. The correlation could then be used to create a synthetic record that we could use fill missing values for the site in question. To avoid negative PET values, the intercept for the trend was set to zero. This is a realistic assumption to make, because in winter, PET values of 0 mm/day typically occur.

The correlation between most sites is very good with an r^2 of 0.98 to 0.99. The poorest correlation used was between Manapouri and Winton (r^2 of 0.97).

5.2 Soil hydraulic properties

There are four soil hydraulic parameters that need to be determined for the soil moisture balance. The S-map only covers the mid-Mataura area, and does not have accurate values of the properties required for soil moisture balance modelling. Additional work was required to determine soil hydraulic properties for the whole region. The properties required to be parameterised are as follows:

1. ***PAW***[3]: profile available water, sometimes called total available water (TAW);
2. ***PRAW***: plant readily available water (crop-dependant, typically pasture);
3. ***SCS Curve number*** which characterises runoff: There are four hydrological soil classifications according to their infiltration rates when saturated;
4. ***Fracstor***: near-surface soil retention (a pre-wetting coefficient).

Estimation of PAW and PRAW values

In most cases the PAW and PRAW values were derived from the SoilPro dataset within the Topoclimate South database. PRAW is calculated by summing the water capacity volume (mm) for each horizon to the potential rooting depth. The water capacity volume is estimated as the difference in volumetric water content between –10 kPa and –1500 kPa in the 0-0.4 m layer, and between –10 kPa and –100 kPa in deeper layers. PRAW values in the Soilpro dataset are based on laboratory-measured volumetric water content where possible. However, most values are estimated by using the pedotransfer functions of Giltrap (2002), which were derived from the National Soils Database.

PAW in the root zone is defined as the difference between the water content at field capacity and wilting point. PAW is calculated as the difference in volumetric water content between -10 kPa (the pressure level at field capacity) and –1500 kPa (the pressure level at wilting point). PAW values in the Soilpro dataset were developed using the same approach by Giltrap (2002) as the PRAW values.

Where PAW and PRAW values were missing, surrogate values were used from sibling soils in the Toloclimate dataset with similar profile features. Organic soils had no PAW or PRAW values attributed to them in the SoilPro dataset, however raw data were available to make calculations on. Raw data were evaluated and PAW and PRAW values were calculated and compared with values from similar organic soils found in the S-MAP dataset. In general, good correlations were found which gave us confidence that the values generated were suitable for use.

Hydrological Soil Group development

Permeability classes were attributed to each soil in the Topoclimate database based on their drainage characteristics and depth to the lowest permeability horizon. This gave us a profile permeability class for each soil in the region, as shown in Table 3.

[3] PAW = (Field Capacity – Wilting Point) x Rooting depth

Table 3: Classes of soil profile permeability

Drainage characteristic of slowest horizon within 1.2 m	Depth to horizon with lowest permeability (m)	Profile permeability Class
Slow	0 - 0.45	S1
Slow	0.45 - 0.6	S2
Slow	0.6 - 0.9	S3
Slow	0.9 - 1.2	S4
Moderate	0 - 0.3	M1
Moderate	0.3 - 0.6	M2
Moderate	0.6 - 0.9	M3
Moderate	0.9 - 1.2	M4
Rapid	> 1.2	R

We then attributed hydrological soil groups and SCS curve numbers to each profile permeability class (Table 4). SCS curve numbers were then allocated to each corresponding permeability profile class in the entire Topoclimate dataset. This provides us with the means to estimate runoff response for each soil based on the depth and drainage characteristics of the slowest permeable horizon.

Table 4: Relationship between SCS curve number, Hydrological Soil Group and Topoclimate Permeability Profile Class

Permeability Profile Classes	Hydrological Soil Group	SCS Curve Number
R	A	40
M2 & M3	B	50 - 60
S3	C	65
S1 & S2	D	75 - 80

Fracstor

Fracstor has a small influence on drainage calculations, so estimated values are sufficient. Typical values are 0 for a coarse sandy soil, 0.4 for a sandy loam, 0.75 for a clay loam (Rushton et al. 2006). If the soil dries quickly, Fracstor will be less than 0.3. If the soil surface remains wet after heavy rain and can't be worked for a few days, Fracstor is likely to be 0.6-0.8 (de Silva & Rushton, 2007).

5.3 Vadose Zone

Average water content

A number of sediment hydraulic properties have needed to be estimated for modelling the average water content of the vadose one. The calculation of the residual water content and pore size distribution both require estimates for the percentage of clay and sand as well as the porosity of the sediment.

The regional-scale geological maps (Turnbull 2000; Turnbull and Allibone 2003, Turnbull et al. 2010) were used as baseline database for information on the region's geology. The GIS database for these maps was used to group mapped units by their 'unique code' (e.g. Q1Q2.alv.gvl). This provided a summary list of 230 geological units to populate with characteristic textural values.

Textural values were assigned to each "unique code" based on the "main rock" and "sub rock" descriptions. This was done by positioning the unit on the Folk classification and soil texture triangles. This gave a mean sand and clay percentage for each unit, thereby allowing the Rawls-Brakensiek regression equations to be populated with appropriate values. A margin of error of clay and sand content was expressed as a standard deviation for the Monte Carlo simulation. Standard deviation values typically range from 1 to 20% depending on confidence of placement in the textural triangle.

Relevant effective porosity values were assigned from representative values reported in the literature (e.g. McWhorter & Sunada, 1977). A standard deviation of 0.025 was also attributed to most sediment codes.

The Brooks-Corey calculation of the average volumetric water content of the vadose zone requires an estimate of saturated hydraulic conductivity. This was more difficult than estimating textural parameters. To do this, each "unique code" in the geological database was attributed a representative mean value hydraulic conductivity value and a standard deviation.

Hydraulic conductivity values representative of the mean were derived from the results of aquifer tests from throughout the Southland region. Higher conductivity values from the aquifer test database were attributed to the more recent reworked alluvial gravels, whereas the mean value for each groundwater zone was considered to reflect the dominant geological unit. Older deposits, and those with lower depositional energy, were attributed indicative values based on those reported in the literature (e.g. Domenico & Schwartz, 1998).

It is well known that horizontal hydraulic conductivity values tend to be higher than vertical values. This anisotropy is partly formed by the processes responsible for the deposit's deposition, and partly a result of subsequent compaction. It was considered that there is insufficient information in Southland to enable an anisotropy adjustment to be made. Furthermore it is debatable that such an adjustment would be justified given the uncertainties involved in estimating hydraulic conductivity values.

For the Van Genuchten model, the empirical coefficients a and η were calculated using the Rawls-Brakensiek functions. The estimation of a was modified to account for higher values expected for coarser gravels. This was done by assigning appropriate values for gravel dominated sediments as quoted in Sousa et al. (2013) for values of a greater than 3.5.

Transit time calculations

The calculation of transit time requires an estimate for the depth to the water table. An extensive database of water depths has been generated for the entire region by Brydon Hughes of Liquid Earth Limited. A detailed methodology for determining water table depths is provided in Appendix 2.

Standard deviations of the water table fluctuation were not used in the Monte Carlo simulation. While the seasonal variability of water level may be significant, the variation in annual average water levels is not considered to be significant in an unconfined aquifer for the purposes of this study.

The positions of the water depth data points were used for transit time calculations for the vadose zone and uppermost part of the aquifer. Figure 4 shows the locations of the 6,450 water level sites that coincided with the input data sets. The interpolated results of our modelling are more representative in areas with a high density of points.

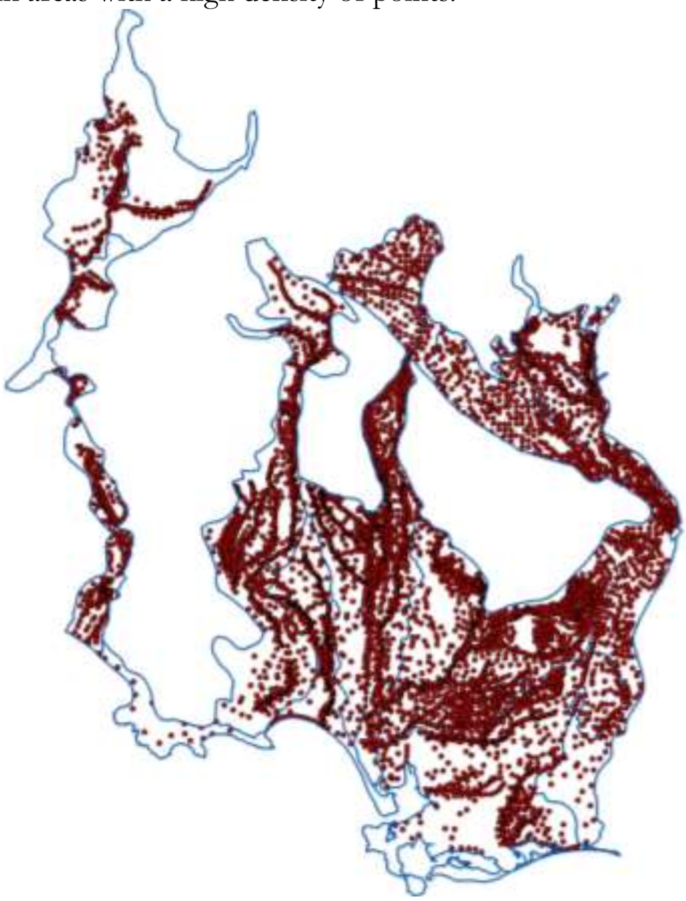

Figure 4: Location of points where static water level data were used for time lag modelling

6. Model results

6.1 Soil drainage

Summary statistics from the Rushton soil moisture balance model are shown in Table 5, and a box and whisker plot is provided in Figure 5. To generate these statistics, the weighted average of each soil type and rainfall combination was calculated for each groundwater management zone. The expression of soil drainage as mm/year allows comparisons to be made between groundwater management zones and also with annual rainfall. The soil drainage rate can easily be multiplied by the area of the groundwater zone to derive annual rainfall recharge volumes.

Table 5: Drainage (mm/yr) and rainfall (mm/yr) statistics for the groundwater management zones

Groundwater	Area	Recharge					Rainfall			Drainage
Zone	(Ha)	Max	UQ	Median	LQ	Min	Median	Mean	SD	%
Castlerock	6,698	405	291	229	168	120	897	892	137	25
Cattle Flat	714	315	215	148	96	16	800	809	137	18
Central Plains	26,316	415	332	270	229	151	913	924	96	29
Edendale	7,541	493	369	321	267	156	1,053	1,072	109	30
Five Rivers	13,333	394	289	239	195	129	901	919	113	26
Knapdale	7,844	308	177	133	102	48	772	807	111	17
Longridge	4,402	355	200	158	107	67	802	826	135	20
L Aparima	31,817	632	485	415	363	282	1,080	1,091	102	38
L Mataura	36,020	481	354	298	248	145	1,030	1,050	109	29
L Oreti	39,820	502	349	293	251	178	937	951	94	31
L Waiau	25,333	763	567	470	435	178	1,160	1,149	157	40
Makarewa	73,537	514	372	316	263	179	1,030	1,043	101	31
Orepuki	11,851	791	576	510	438	372	1,334	1,354	138	38
Oreti	5,950	412	289	234	174	94	855	878	149	27
Riversdale	10,329	374	216	166	112	75	787	820	139	21
Te Anau	26,433	635	463	361	315	209	1,003	1,010	113	35
U Aparima	39,920	484	396	314	265	183	938	960	113	32
Waihopai	61,705	512	373	293	229	157	1,046	1,066	114	28
Waimatuku	22,324	479	375	312	265	193	974	989	94	32
Waimea Plain	25,256	338	212	165	126	45	810	838	139	20
Waipounamu	3,220	421	275	221	145	103	851	868	142	26
Wendon	3,212	429	295	247	139	90	927	926	143	26
Wendonside	8,488	454	328	260	172	113	922	924	145	28
Whitestone	2,280	546	388	284	236	154	915	919	105	31

Figure 5 shows that soil drainage rates vary considerably across the Southland region. The highest median annual soil drainage rates occur in the Orepuki, Lower Waiau and Lower Aparima areas. These areas also receive the highest annual rainfall. Conversely, the least soil drainage occurs in the Knapdale, Cattle Flat, Longridge, Waimea Plain, and Riversdale management zones. These areas also receive the least annual rainfall.

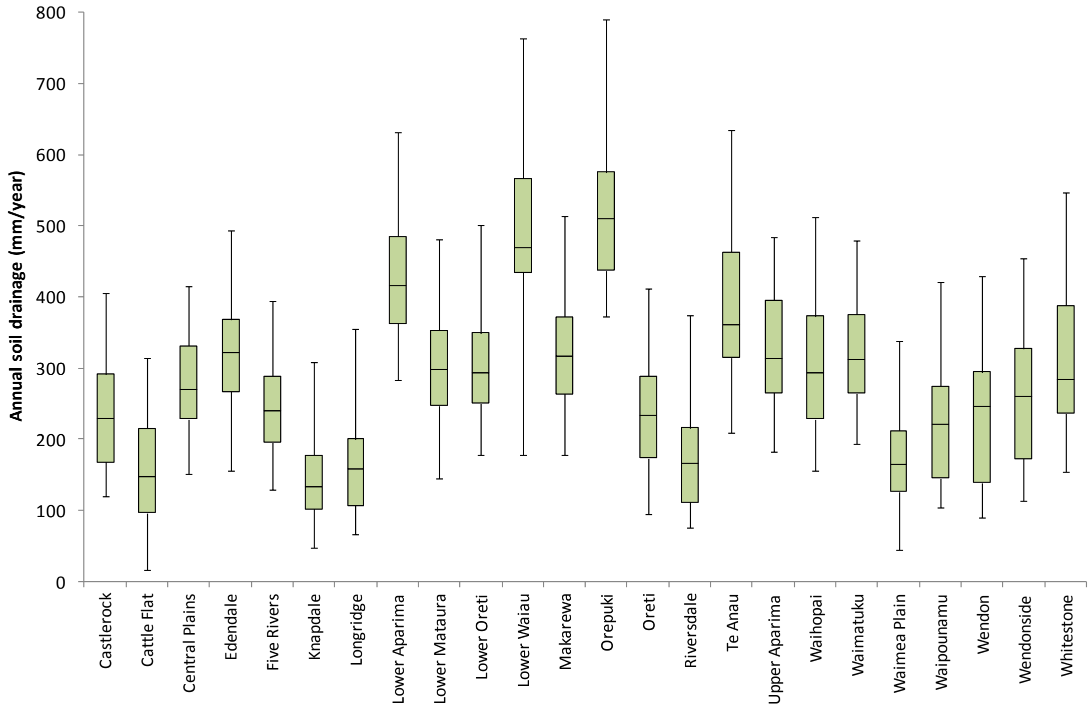

Figure 5: Box and whisker plot of annual soil drainage recharge statistics

The annual soil drainage variability does not match the pattern of median annual soil drainage particularly well. Normally, we would expect areas with higher drainage rates to have a greater variability, and this is what we see for extreme events across the region (longer whiskers on Figure 5). However, the quartile data suggests there is greater variability of annual drainage in the northern regions. This is manifest in a larger interquartile range for the Wendon, Wendonside, Te Anau and Whitestone zones.

Figure 6 shows the median annual rainfall recharge rate to the groundwater as a percentage of median annual rainfall. The pattern is similar to that shown by the soil drainage statistics, with higher percentages of recharge occurring in wetter areas, and lower percentages in drier areas. The Rushton model estimates that recharge varies from 17% to 40% of median annual rainfall. This is much higher than has been estimated for Otago (Wilson & Lu, 2011), which is most likely due to the considerably higher rainfall observed in Southland.

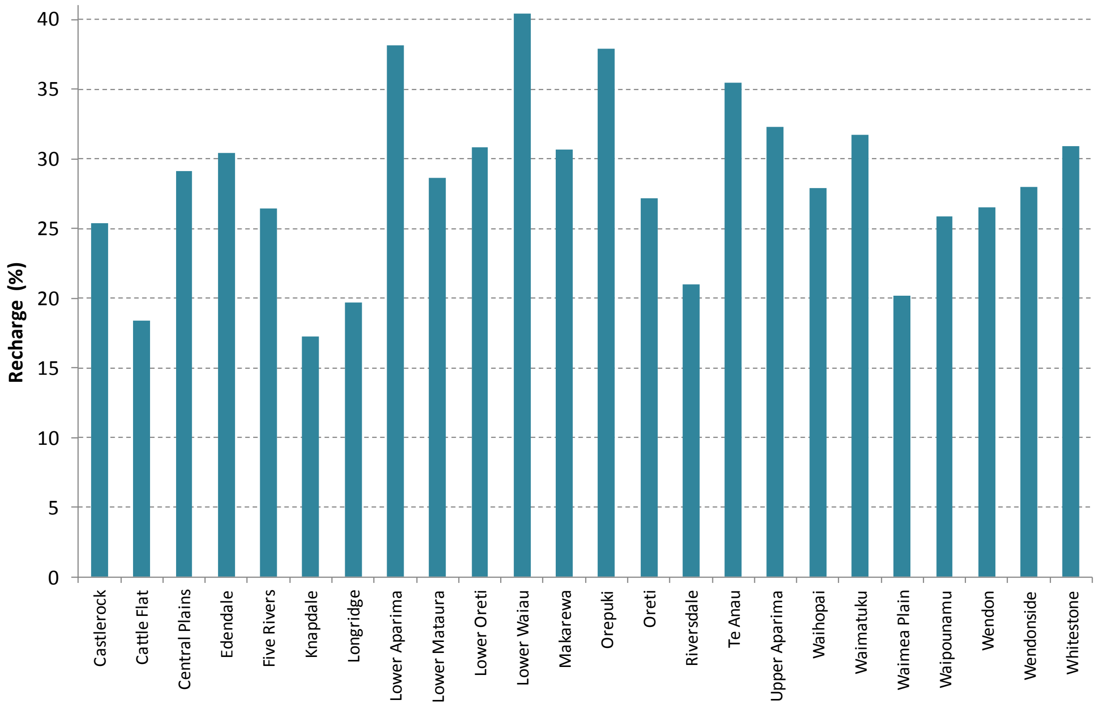

Figure 6: Rainfall recharge as a percentage of mean annual rainfall

The statistics derived from the Rushton model suggests that recharge rate at a groundwater zone scale is largely a function of mean annual rainfall. If rainfall and recharge are correlated, a power relationship can be seen with a very good coefficient of determination (R^2 0.92). Despite this relationship, we can still see considerable variation in recharge with different soil types.

Figure 7 shows a map of the mean annual recharge, which is the dataset used for modelling time lags. Considerable variation can be seen within individual groundwater zones. The pixelation used for distributing rainfall across the region is also evident in this map.

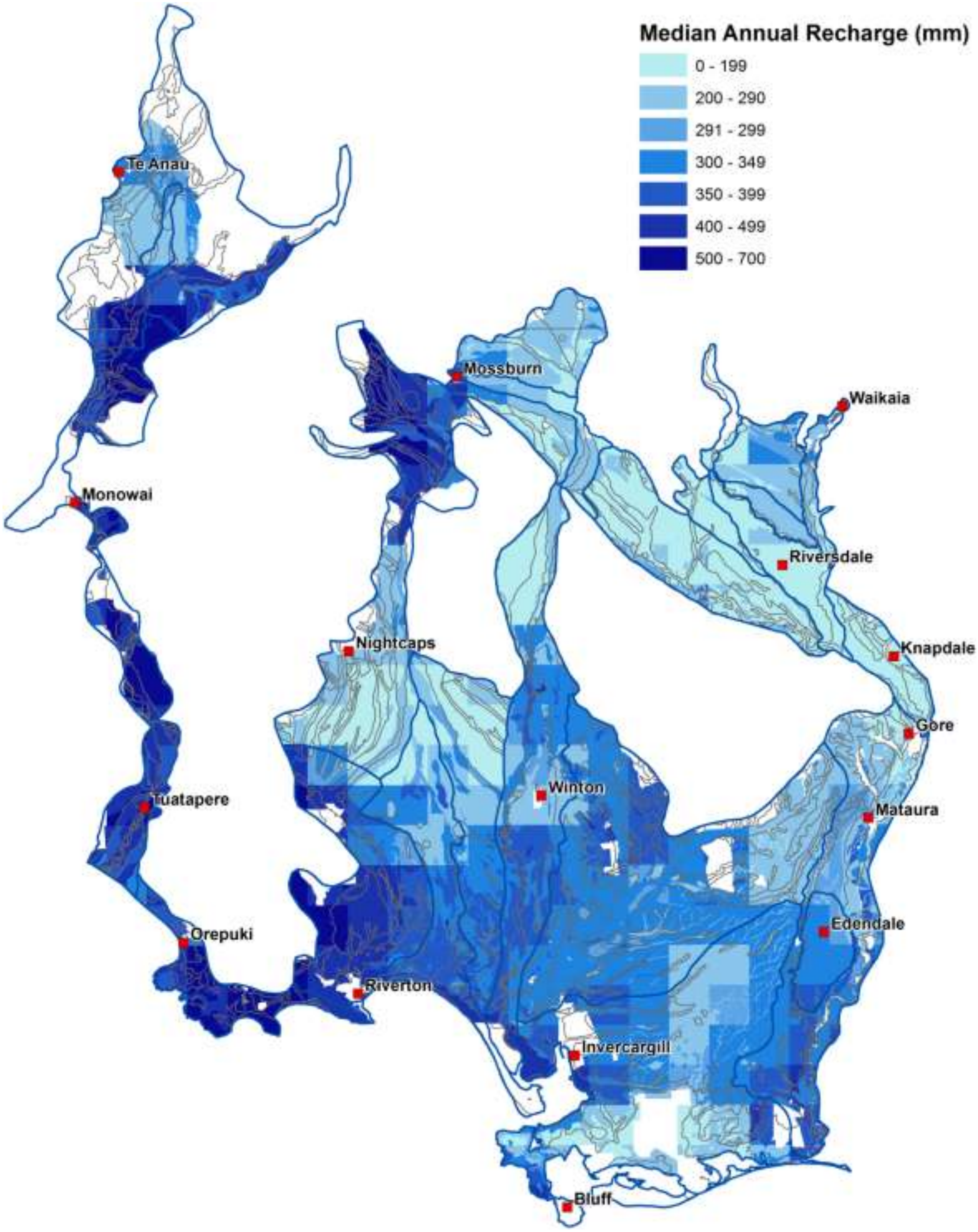

Figure 7: Map of mean annual recharge

6.2 Vadose zone transit time

Calculations for the average volumetric water content of the vadose zone for the 6,450 sites modelled ranged from 8.5% to 40.7% for the Brooks-Corey model and 12.9 % to 36.3% for the Van Genuchten model. The mean results for the two models were 16.3 and 17.1% respectively.

The results of the vadose zone models accord well with reasonable estimates in the literature (uniform sand 13%, mixed grain sand 16%, silt 42%; Peck et al. 1974). In Southland we would not expect values as low as those reported by Dann et al. (2009) for Canterbury sandy gravels, although the more permeable Southland gravels do approach their observations of 3.5 to 13.9%.

The vadose zone transit times predicted by the Brooks-Corey and Van Genuchten models are very similar. Figure 8 shows the similarity of predictions made by the two models. The Brooks-Corey method gives slightly longer transit times overall. The results from two sites at Knapdale depart quite significantly from the trend shown by the rest of the dataset. These two sites are the only two sites included in the study that have a loess substrate. Their departure from the trend of the rest of the dataset reflects the uncertainty of estimating the hydraulic properties of loess. If these two loess sites are omitted from the dataset the slope of the trend line increases to 0.94 with a correlation coefficient of 0.99.

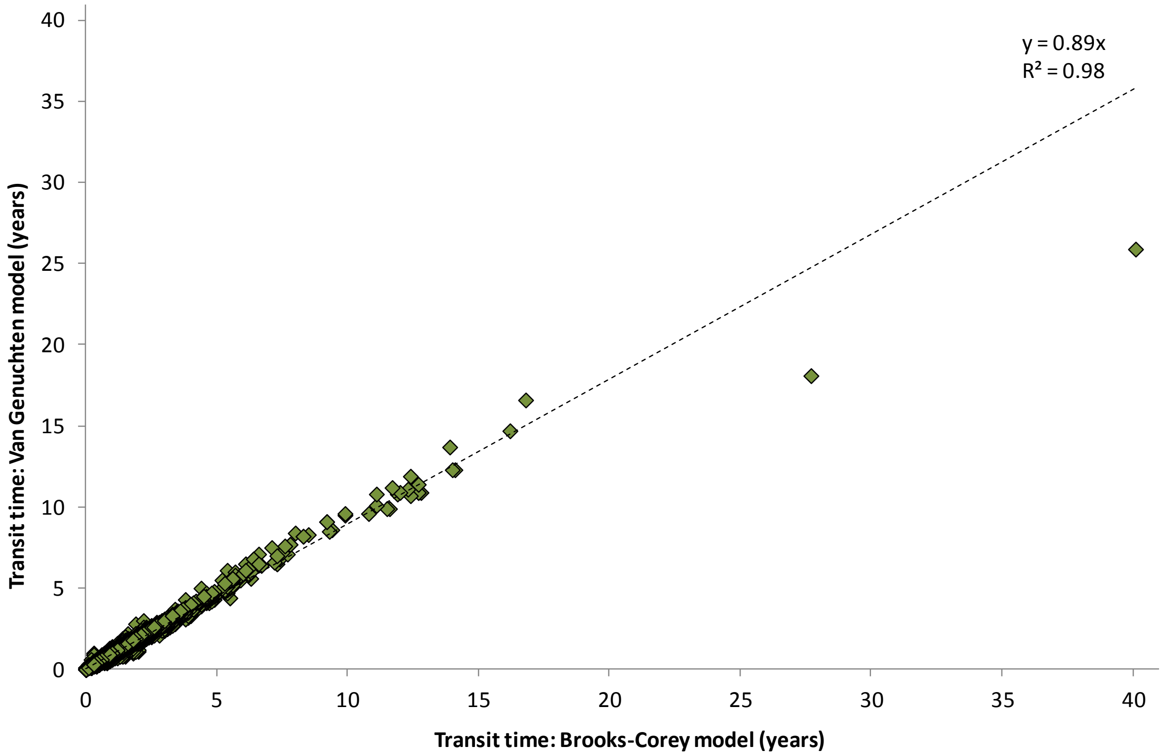

Figure 8: Comparison of vadose zone transit time estimates for the two models used

The mean transit time at the 6,450 water level sites was less than a year for both models. When the dataset is spatially interpolated and analysed for area, about 80% of the region has a vadose zone transit time of less than a year, and 90% has a transit time of less than two years.

Figure 9 shows a map of the results of the vadose zone modelling as contours of transit time. This map was created by interpolating the results of the Van Genuchten model using the natural neighbour method. The boundaries of the main geological units have been traced in grey to illustrate how they relate to our modelling results. We refer the reader to the relevant regional-scale geological maps (Turnbull 2000; Turnbull and Allibone 2003, Turnbull et al. 2010) for details of these units.

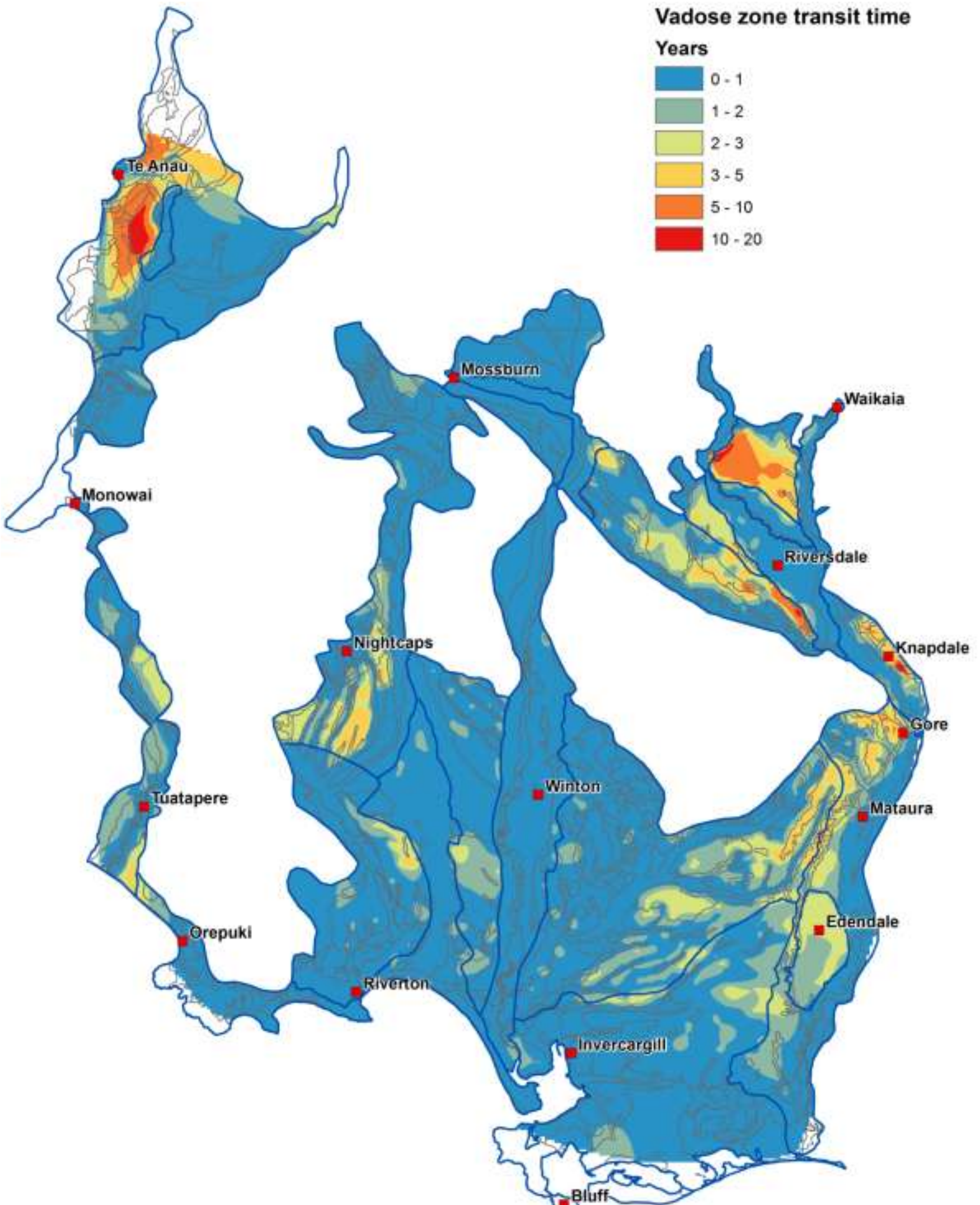

Figure 9: Mean vadose zone transit time estimates for the Van Genuchten model
Groundwater Zones are shown in blue, and the outlines of the main geological units are shown in grey

A general pattern can be seen in
Figure 9 that relates well to the distribution of different Quaternary units. Very rapid transit times (<1 year) are seen in the Holocene Q1 gravels that occur adjacent to the main rivers in the region. Even in the more widespread Q2 gravels deposited during the last glacial period, the transit times are estimated to be two years or less.

Longer transit times (>five years) are found beneath older terrace surfaces (Q6-Q10). These terraces are located at higher elevations in the catchments, and are also situated high above the active river terraces. This results in a greater depth to the water table, often over 20 m. The sediments beneath these terraces also have a lower permeability because they have had less alluvial reworking than the younger, lower level terraces.

The distinction between residence times in younger and older sediments is most clear in the Wendonside and Longridge areas. The interpolation has not followed the geological boundaries particularly well, but some broad distinctions can be seen. Here the more weathered Q6 to Q10 gravels have estimated vadose zone transit times of up to 10 years. Extremely long transit times in the Knapdale groundwater zone are associated with loess deposits to the southwest of Knapdale.

Longer transit times are also predicted in the northwest Te Anau basin, particularly beneath the Q6 moraine deposits southeast of Te Anau. The water table beneath these terraces is over 20 m deep. The static water level coverage is not as detailed as elsewhere in the region, so the interpolated contours can be misleading in this area if the distribution of source data is not considered.

Figure 10 shows the standard error resulting from the Monte Carlo simulations performed on the Van Genuchten model. This error accounts for the uncertainty arising from lack of specific knowledge of vadose zone hydraulic properties. At least 90% of the region has a standard error of less than a year, which is less than the variation caused by annual fluctuations in rainfall recharge.

Standard errors of over half a year tend to occur in areas of lower permeability sediments such as older clay-bound gravels and loess deposits. Larger errors are also associated with deeper water tables. This is because uncertainties in calculating the average vadose zone water content become compounded when multiplied by the travel distance.

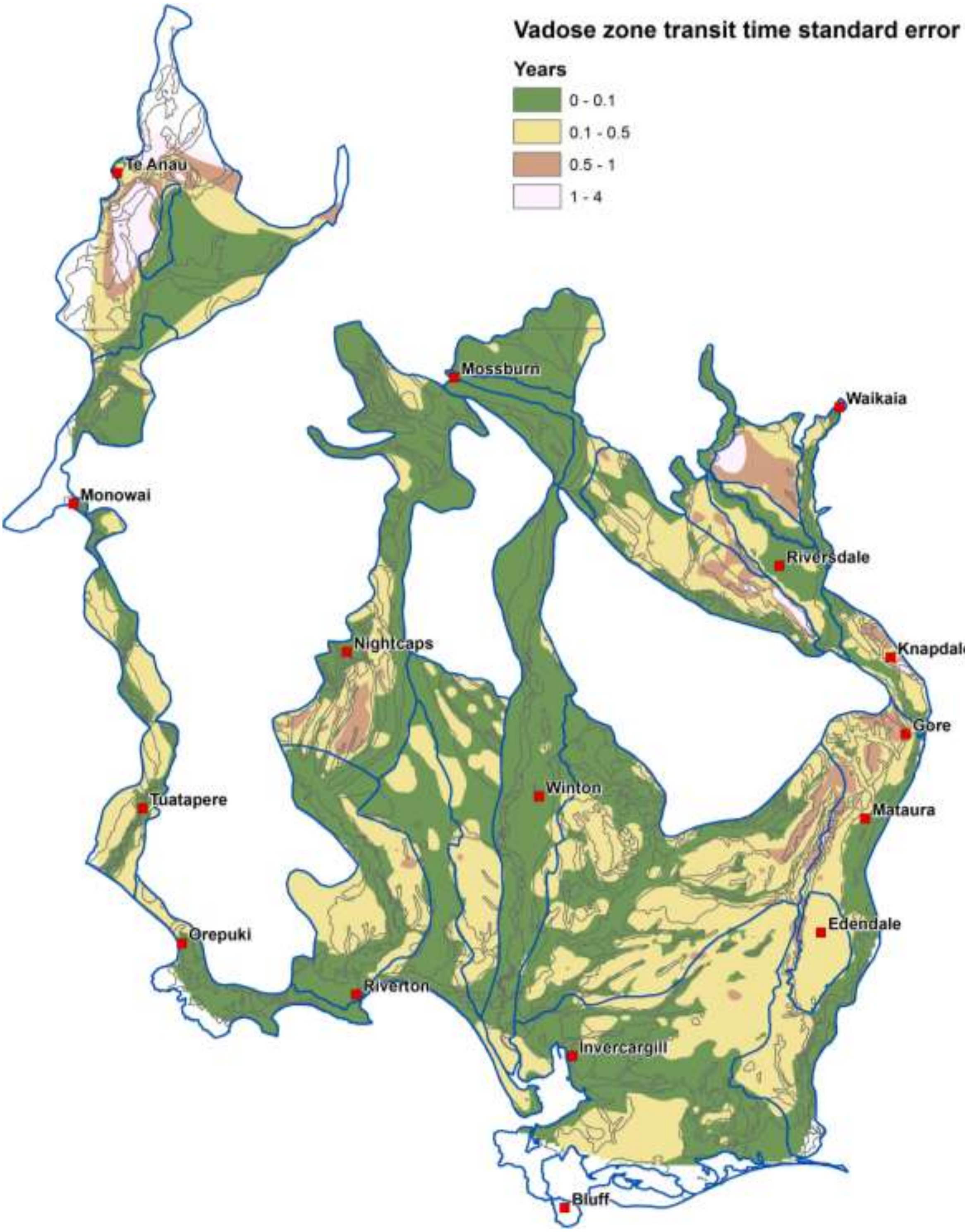

Figure 10: Standard error for vadose zone transit time estimates expressed as a percentage deviation from the mean

6.3 Saturated mixing time

To assess the time lag for a change in land use to impact on groundwater quality we need to allow for nitrate to penetrate the shallow aquifer. Figure 11 shows the result for the calculation of saturated flow time to a depth that is equivalent to a year of rainfall recharge. The vertical percolation time in the saturated environment ranges from around one to five years. The longest groundwater mixing times tend to occur in the thicker aquifer deposits. Care should be taken when interpreting this map because proximity to streams will greatly speed up the travel time in shallow groundwater, which tends to be the case with thinner aquifers. Figure 11 should therefore be considered as a map of the maximum groundwater mixing time.

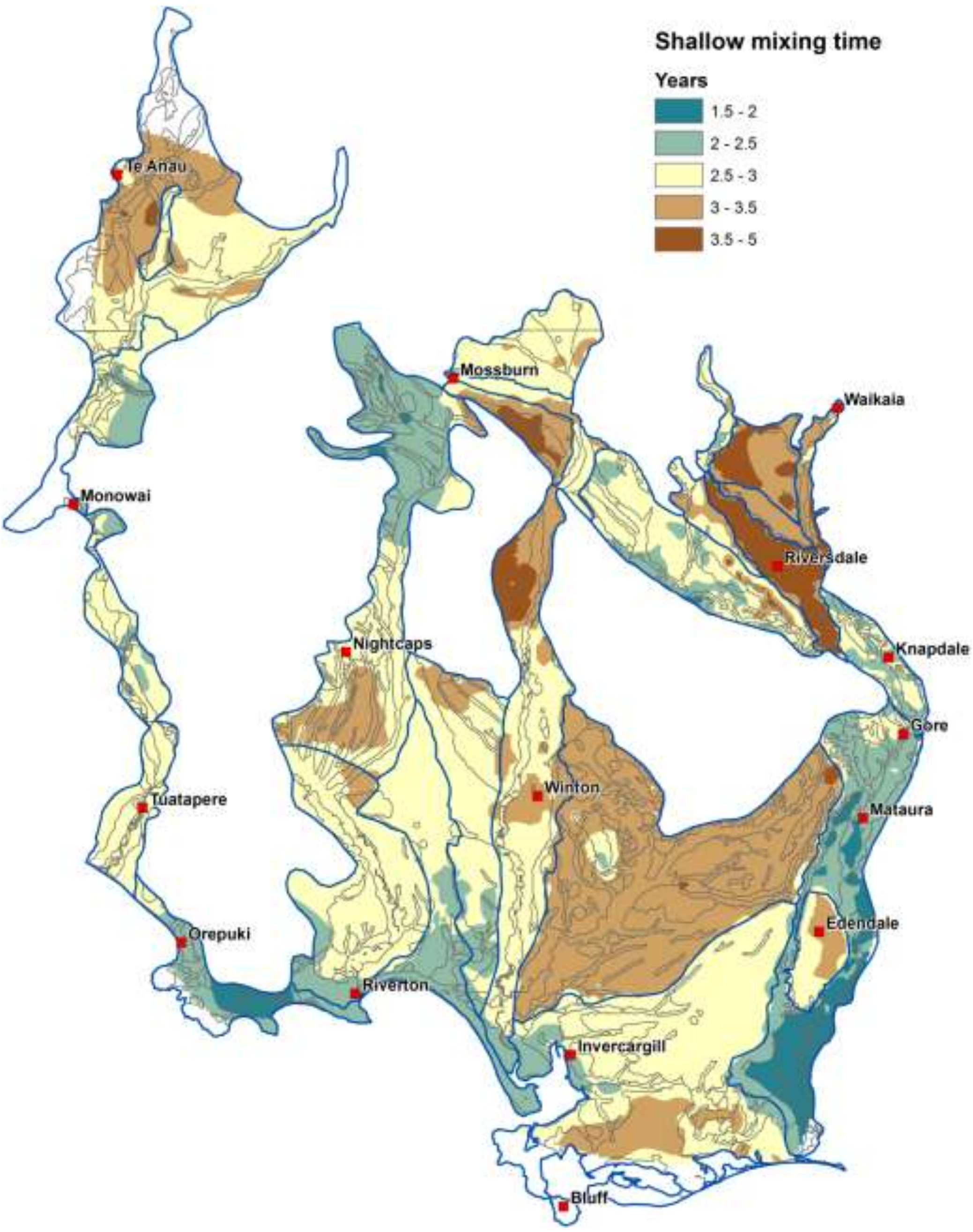

Figure 11: Time taken for a year of rainfall recharge to mix into shallow groundwater

6.4 Total time lag

The vadose zone transit time and shallow aquifer mixing time can be combined to estimate the total time lag beneath a property. By combining the two travel-time estimates we can determine the time it will take for nitrate leached at the surface to impact on groundwater quality beneath properties. Figure 12 shows a map of the combined vadose zone and saturated time lags.

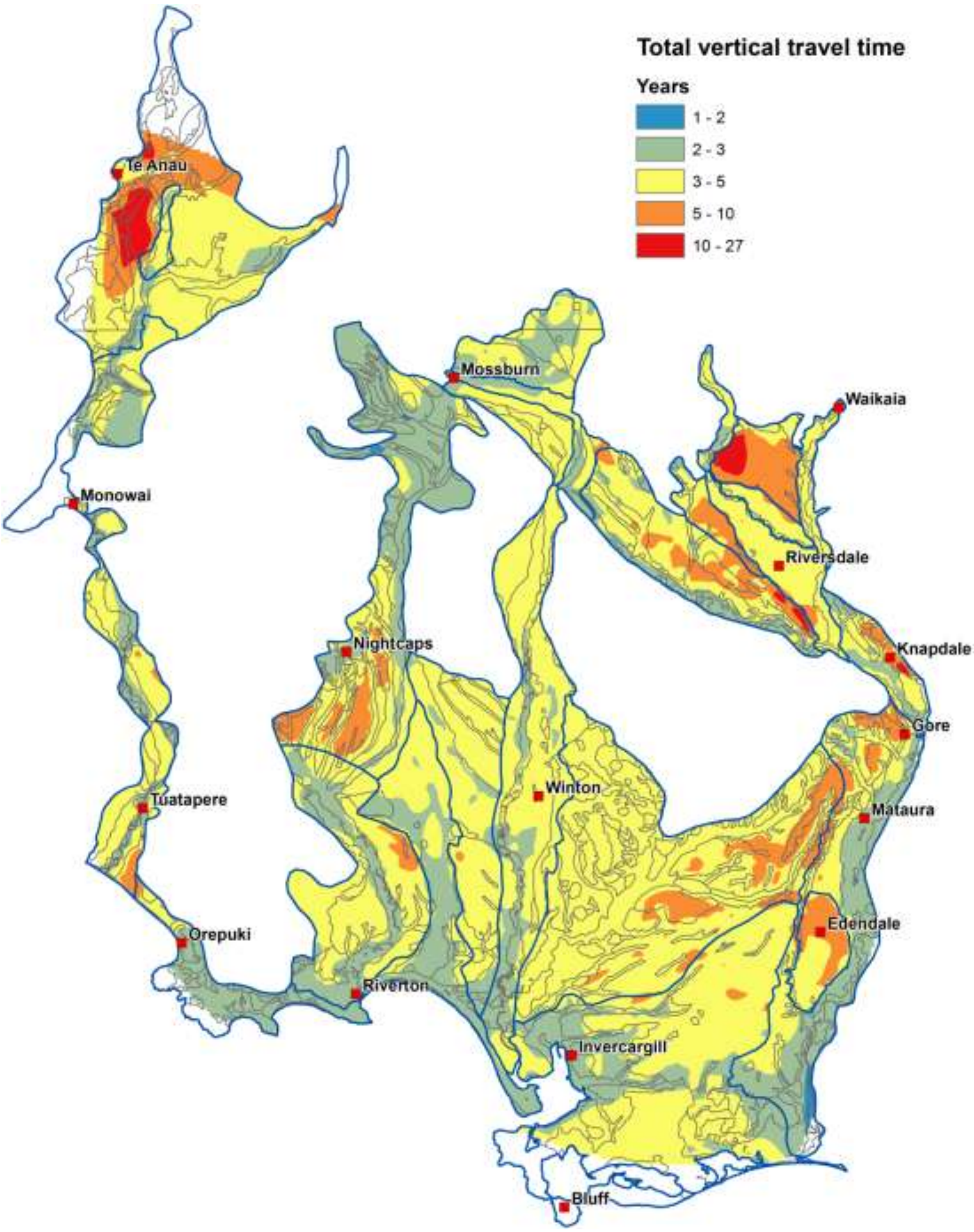

Figure 12: Map of total vertical travel time through the unsaturated zone and into the uppermost saturated zone

The results of combining the two estimates indicate that about 40% of nitrate leached in Southland would have mixed with shallow groundwater within three years. About 95% of nitrate leached in Southland would have mixed within five years.

The distribution of longer time lags closely follows the distribution of vadose zone transit time. Longer time lags are expected in the older, more elevated Q6-Q10 terrace deposits, particularly in the Longridge, Wendonside, and Te Anau areas.

7. Validation

In order to assess the performance of the regional vadose zone time lag model, a validation process was required. The validation of a model consists in comparing the modelled results with observed data. The performance of each of the two components of the model was assessed separately - the soil drainage model was validated against estimated aquifer recharge values, and the modelled travel times through the vadose zone and upper part of the aquifer was compared to age tracer data.

7.1 Methodology

Rushton Model

The assessment of the performance of the Rushton model was based on two criteria - its ability to predict annual recharge, and its performance at predicting individual recharge events.

The Rushton model has previously been compared to other soil moisture balance models and validated against lysimeter data in Canterbury (Wilson and Lu, 2011), where it was found to perform best in predicting drainage below the root zone. In the context of this study, no lysimeter data were available for validation purposes, and the comparison with shallow soil moisture tapes was not possible owing to the difference in water holding capacity between the shallow and deeper soil horizons: The Rushton model calculates soil moisture deficit for an entire composite soil profile, whereas the soil moisture tapes are buried at a maximum depth of 20 cm.

The Rushton model was therefore validated against an estimation of drainage based on water table fluctuations. This method uses the long-term hydrograph of water table level to calculate annual aquifer recharge and the volume of each recharge event.

The major assumptions and requirements when using this method to calculate land surface recharge are:

- rainfall recharge is the dominant mode of aquifer recharge, river influence is minimal;
- any recharge event results in a peak on the hydrograph (baseline recharge is minimal);
- the water table fluctuation has to exhibit a typical peak-recession type curve, as opposed to a sinusoid curve;
- pumping should be minimal for the same reasons.

Only three adequate sites were identified and used in this validation procedure.

A practical way to assess graphically the performance of the model is to plot the observed cumulative aquifer recharge and the cumulative modelled recharge on the same chart. The observed recharge was calculated with the water table fluctuation (WTF) method. Therefore, the comparison between the modelled and observed recharge was made using the specific yield as an indicator. This is because:

$$\textit{Aquifer Recharge} = \textit{Observed Fluctuation} \times \textit{Specific Yield}$$

For unconfined aquifers, the specific yield should range between 0.05 and 0.3.

A linear model was fitted between the observed fluctuation and the modelled recharge, where the regression coefficient is the specific yield. In all cases, the multiple R squares of the regression model were in excess of 0.9.

For each of the selected three sites, aquifer recharge was estimated with the WTF method which involves determining the recharge depth from a long-term hydrograph (several years at least). With the WTF method, a recharge depth is calculated for each recharge event (peak in the hydrograph) and the annual recharge is obtained by adding these.

The multi-year hydrograph was separated into individual events (recharge peak + recession), and each recession curve was isolated and incorporated into a mathematical model which calculates a global recession equation for the hydrograph based on each individual recession curve (Posavec et al. 2006).

The obtained recession equation was applied to each recession curve in order to project the level the aquifer would have reached had the recharge event not occurred. Finally, the recharge depth for each individual event was calculated as the difference between the highest value of the recharge peak and the corresponding lowest point on the projected recession curve, multiplied by the specific yield obtained from the linear model described earlier.

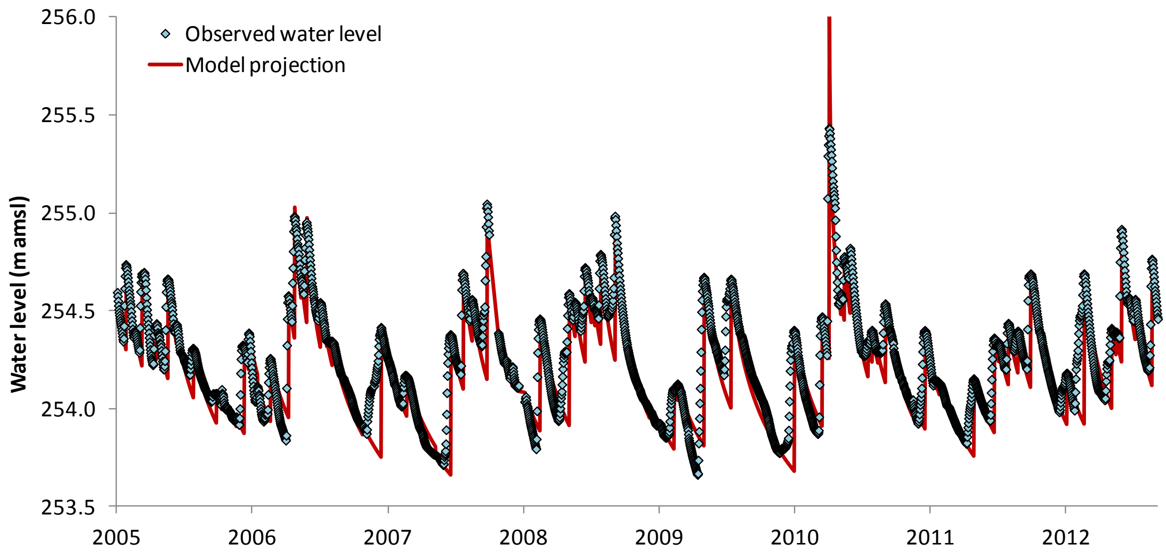

Figure 13: Hydrograph of groundwater level and modelled recession curves

Time lag model

The major assumption we have made for the validation of the time lag model is that the vertical travel time equals "groundwater age", or mean residence time. The conditions under which this assumption is valid are developed below. When using tritium for age dating water from streams or aquifers, the age is determined with reference to the time when the molecules of water precipitated (separation from the atmosphere).

Tritium samples were collected in several wells across the region and the availability of time-series measurements allowed the age dating of these waters with a reasonable accuracy (plus or minus a year). However, several sites required additional tritium samples to be collected in order to assign a mean residence time in a more robust manner. The requirement for long-term time series of tritium measurements (at least two samples at one year intervals) means that reactionary sampling to determine groundwater age is not possible in short timeframes such as this study.

The mean age of the water sample is determined by relating the tritium concentration measured in groundwater to the input function of atmospheric tritium with a lumped parameter model. The most commonly used models in New Zealand are the Piston Flow Model (PFM) which models water flowing through a constrained structure (pipe-like), or any flow without mixing or dispersion. For any situation where several flow paths are mixed together, an Exponential Piston Flow Model (EPM) is used. The EPM simulates the flow of water through the vadose zone (piston flow component) and the exponential mixing of flow paths in the saturated zone. In our study, the EPM was used exclusively, as none of the tritium originates directly from the bottom of the vadose zone or from a confined structure.

The mean age of a groundwater sample collected below the water table is the result of the mixing of several flow paths, and therefore waters with different ages. As described in the conceptual model in this report, it is assumed that a nearly vertical downward flux occurs in the very upper part of the aquifer, whereas the flow becomes more horizontal with depth. The assumption that the mean residence time of the groundwater sample reflects the vertical travel time is therefore only valid in the uppermost part of the aquifer. Because of this major assumption, all wells screened deeper than 5 metres below the water table were excluded from our validation dataset.

Furthermore, the tritium validation method cannot be applied in areas where a significant amount of interaction between aquifer and streams occurs. If there is significant surface water influence, the mean residence time of groundwater will no longer reflect vertical travel time exclusively.

7.2 Results

Rushton model

Graphs relating the modelled and observed cumulative recharge values indicate a reasonably good model performance. The model seems to generally predict recharge events with a good accuracy. For the Central Plains aquifer, the linear model fitted on observed recharge resulted in a specific yield (regression coefficient) of 0.062. The modelled recharge seems to match the observed recharge reasonably well, although the curves suggest that the actual recharge is more spread in the year than the model suggests (recharge primarily in winter).

For the Castlerock aquifer, the best fit is obtained with a specific yield of 0.11, which is within the expected range of variation. Aquifer tests carried in this aquifer provided values ranging between 0.06 and 0.2 depending on the method and solution used.

For the Five Rivers aquifer, for the best correlation was obtained with a specific yield value of 0.08, and the plot suggests a reasonably good fit between observed and modelled aquifer recharge. A step-drawdown pump test in this aquifer gave a value of 0.1, which is consistent with our findings.

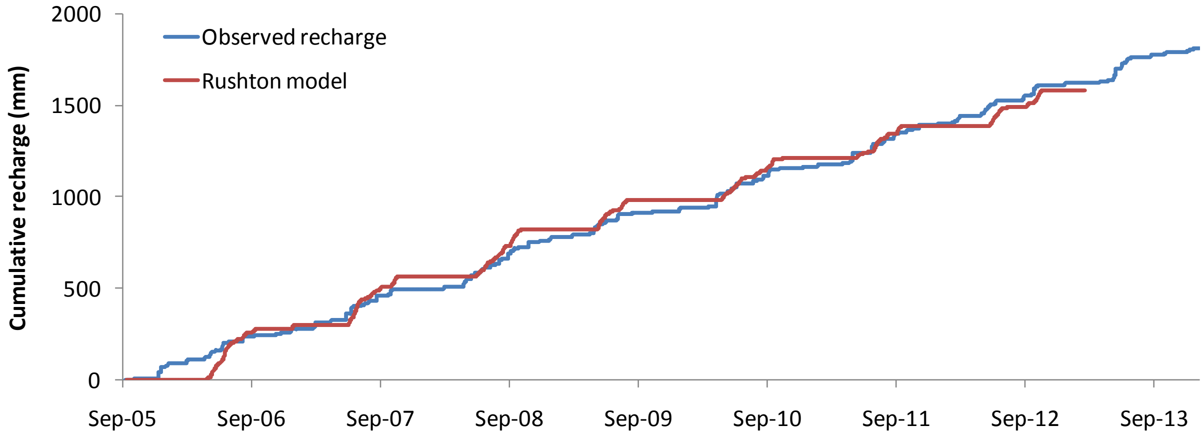

Figure 14: Comparison of observed (blue) and modelled (red) recharge, Central Plains aquifer

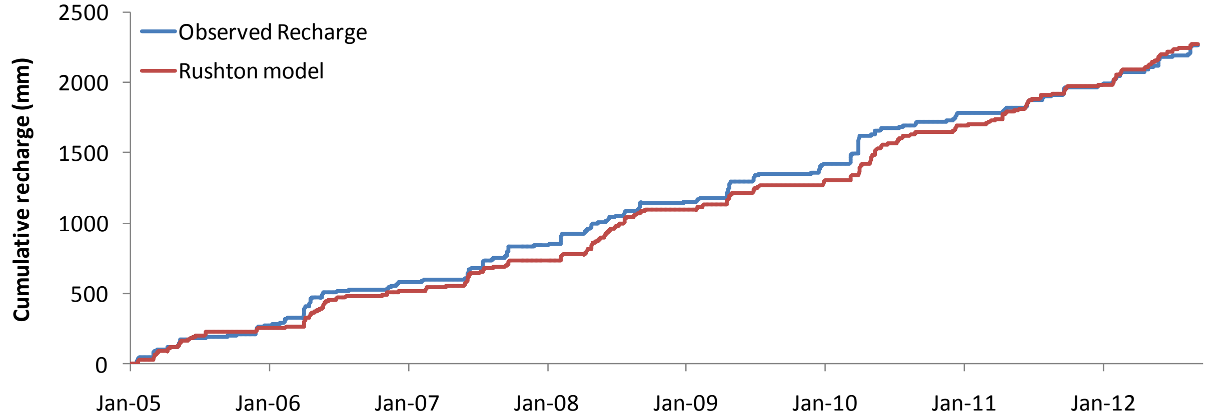

Figure 15: Comparison of observed (blue) and modelled (red) recharge, Castlerock aquifer

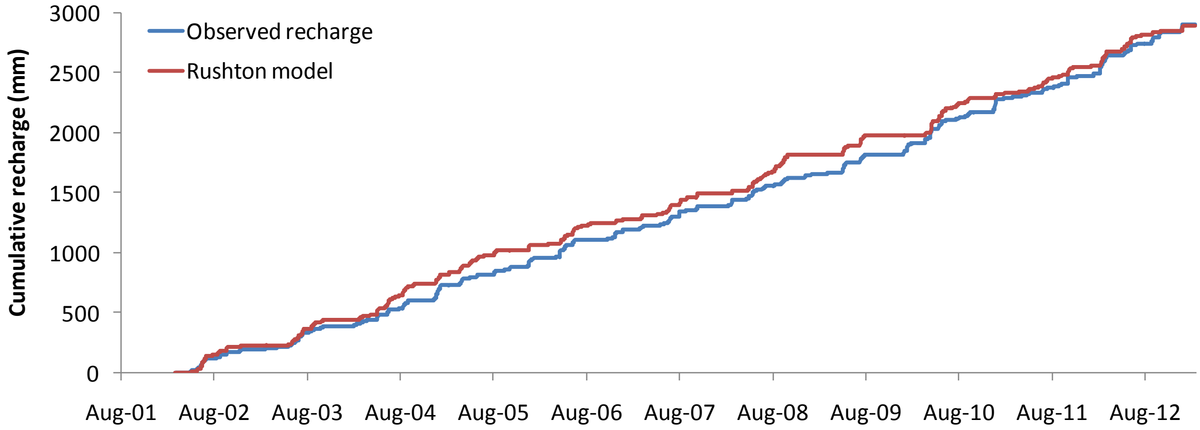

Figure 16: Comparison of observed (blue) and modelled (red) recharge, Five Rivers aquifer

The results of the annual recharge estimation from the WTF method for each of the three sites are listed in Table 6. The model seems to perform best for the Central Plains aquifer, with only 35 mm/year of difference in average, compared to 40 mm for the Castlerock aquifer and 88 mm for the Five Rivers aquifer. The discrepancy between predicted and observed recharge is a result of the model performance, although it is expected that the uncertainty linked to the WTF method and the associated assumptions are a significant source of error.

The Central Plains aquifer is the site which seems the most suited to the WTF method. It is isolated from any river recharge and the hydrograph exhibits sharp peaks followed by logarithmic recessions after each recharge event. Pumping is also thought to be minimal in the vicinity of this site.

The Castlerock aquifer is not quite as suitable, as the water table does not seem to respond in a sharp fashion to the new recharge events, due to a difference in geology and depth to the water table.

Finally, the Five Rivers site was selected for this study because isotopic values suggested that the river influence was limited at this site. However, this aquifer is known to have major interactions with river stage in places. It is likely that the major discrepancy between modelled and observed recharge can be attributed to the poor performance of the WTF method in this aquifer rather than a poor model performance.

Table 6: Comparison of observed and modelled aquifer recharge (mm)

Year	Castlerock		Central Plains		Five Rivers	
	Observed	Modelled	Observed	Modelled	Observed	Modelled
2003	-	-	-	-	300	219
2004	-	-	-	-	377	331
2005	270	322	-	-	376	289
2006	347	259	242	287	306	216
2007	391	223	272	277	343	218
2008	366	358	276	258	380	326
2009	224	172	203	163	192	153
2010	456	393	301	226	461	314
2011	294	316	270	176	402	259
2012	359	340	262	193	374	308
Average	**338**	**298**	**261**	**226**	**351**	**263**
Difference	**-40**		**-35**		**-88**	

7.3 The vertical travel time model

A total of 34 sites were age dated throughout Southland, using the tritium method. However, only 19 sites were assigned a mean residence time (MRT) with acceptable level of confidence. The remaining sites require the collection of additional tritium samples in order to provide more reliable age estimates.

For these 19 sites, the groundwater MRT derived from tritium samples ranged from 1 to 44 years, with most samples between 1 and 10 years. The "modelled age" ranged from 3 to 10.5 years, with most results between 3 and 5 years.

As a general observation, the modelled travel times appear to be more consistent throughout the region than the measured age. This might be due to the complexity of processes and flow path contributing to the MRT of a groundwater sample, which our model is not replicating. Even for wells screened shallow in the upper portion of the aquifer, the saturated mixing time is the major component of the travel time, contributing 80% of the total travel time on average.

Of the 19 sites used for validation of the model, 8 sites had modelled ages within 1.5 years of the tritium age, 4 sites had modelled ages within 3 years of the tritium age and 2 sites within 4.5 years (Table 7). Considering the regional scale of the modelling exercise and the margin of error in tritium age interpretation, these results are considered acceptable (Figure 17).

Table 7: Comparison of modelled travel time and mean residence time determined from Tritium concentration (years)

Site name	Mean Residence Time	Modelled vadose zone travel time	Modelled saturated mixing time	Total modelled travel time
D45/0006	3.5	0.4	3.4	3.8
D45/0164	6	2.2	2.9	5.1
E44/0007	1	0.6	3.3	3.9
E44/0008	8	1.3	2.5	3.8
E44/0377	3	1.4	2.7	4.1
E45/0010	3.5	0.4	3.3	3.7
E45/0055	2	0.2	3	3.2
E46/0093	5	0.2	2.8	3
E46/0104	12	0.1	3	3.1
F44/0139	14	7	3.5	10.5
F46/0194	4	2.2	3.1	5.3
F46/0195	85	2.9	3.1	6
F46/0693	44	0.3	4	4.3
F47/0252	1.5	0	3	3
F47/0258	2	0.1	2.9	3

One of the major sources of discrepancy between tritium age and modelled age is thought to be depth of the well within the aquifer. Several investigations have shown that the mixing of groundwater from different flow paths increases with depth, and that less mixing occurs in the very upper part of the aquifer. Consequently, the MRT determined from tritium concentration for wells screened deep within the aquifer is the result of the mixing of waters of different ages, and this level of complexity is not incorporated in the saturated mixing model. Additionally, if the well has been pumped for production, or simply during the sampling, pumping-induced mixing can occur and draw older water to the well.

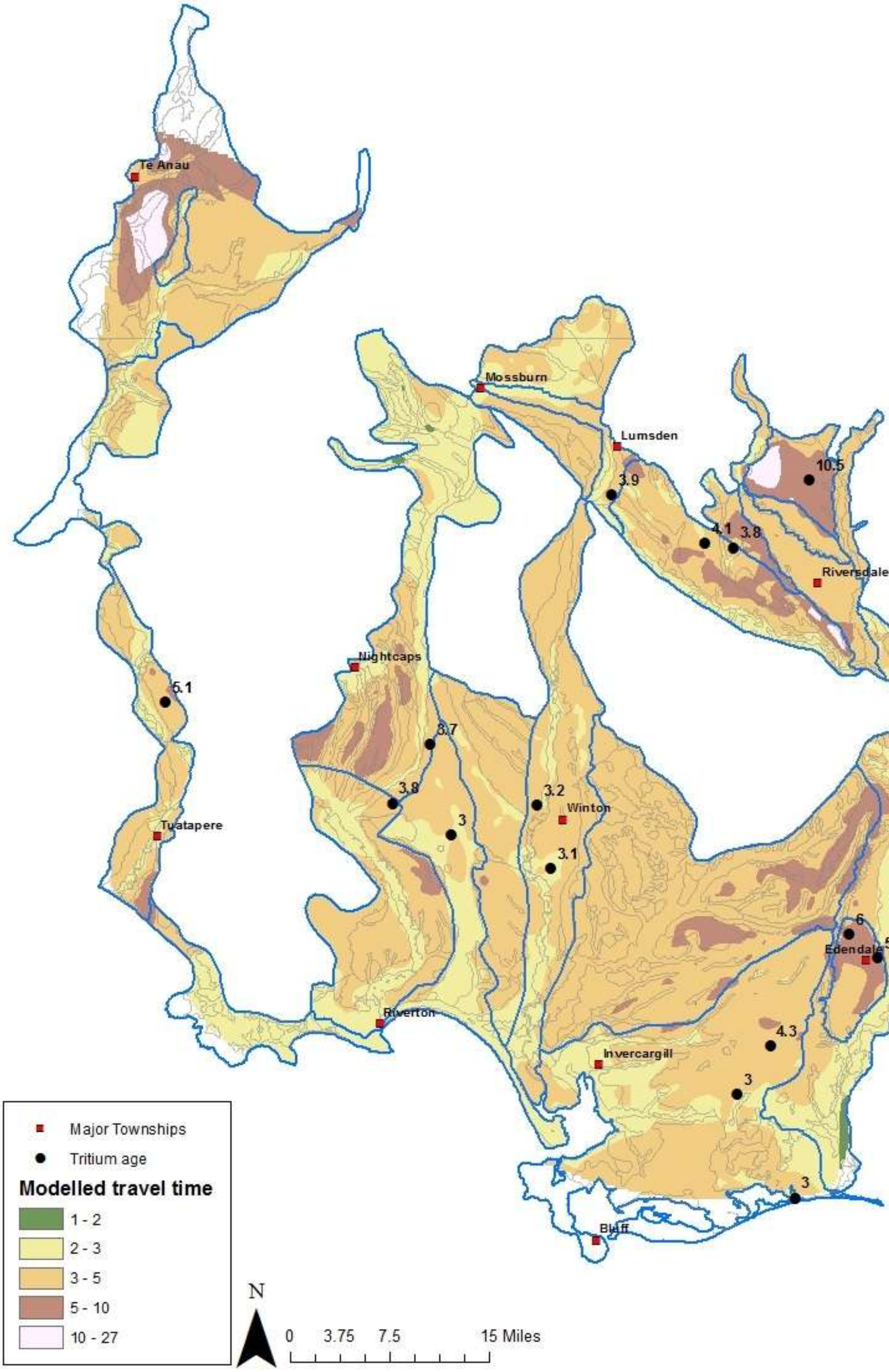

Figure 17: Comparison of the modelled travel time for the region, with tritium derived Mean Residence Time (years)

A study completed by Environment Southland has been investigating the changes in groundwater type and age with depth, using a nested piezometer. The piezometers has screens at 3 m, 6 m, 9 m, 12 m and 15 m (just above bottom of the aquifer) in a Quaternary gravel aquifer of Central Southland. Tritium age dating for each of these depths indicated that groundwater age changes abruptly between 12 m and 15 m. The MRT is estimated around one year at each screen depth, except for at 15 m where the MRT is around 60 years. These results suggest that old water is found at the base of the aquifer, while the rest of the aquifer contains mainly younger water and most likely well mixed. This old water is thought to flow along nearly horizontal flow paths on the base of the aquifer.

A similar process is likely to produce some of the high discrepancies observed between modelled and tritium groundwater age, provided that our model is limited to vertical mixing. For instance, F46/0693 is 12.5 m deep with a MRT of 44 years, while the neighbouring well F47/0252 is only 7 m deep and has a MRT of 1.5 years. The model gives a rather acceptable age for the shallower well (F47/0252: 3 years), while it fails at predicting the age for the deeper one (F46/0693: 4.3 years). It is likely that the deeper sample is drawing old water (85 years old) from a deeper part of the aquifer, perhaps under the effect of pumping.

Another source of discrepancy between modelled travel time and tritium age is the influence of streams and rivers on aquifer recharge. The model predicts vertical travel time and mixing time of vertically infiltrating water, but if river recharge has occurred, the tritium age reflects the mixing of water from land surface recharge with stream water. As stated previously, the major assumption made for the validation of the model against tritium measurement is that the tritium age is equivalent to the vertical travel time. We consider to be working under this assumption as long as vertical infiltration is the major source of aquifer recharge, which is not the case when river water recharges the aquifer. Additionally, the path followed by river water can be highly complex (e.g. Precipitation → recharge to aquifer 1 → baseflow to stream → recharge to aquifer 2 etc…) hence the variability in the age of river water.

8. Conclusion

The objective of this report has been to answer two questions:

1. If land use practices are changed, how long will it take before there is a change in shallow groundwater quality?
2. Have groundwater nitrate concentrations reached an equilibrium, or can we expect a "time bomb" sometime in the future

The results of our modelling show that the majority of Southland's shallow groundwater quality would show some response to a change in farming practices within five years. This indicates that Southland's recent dairy farming boom will have already impacted shallow groundwater quality for most of the region. Conversely, if nitrate leaching losses are reduced, we can expect to see an improvement in shallow groundwater quality beneath properties within five years for most of the region.

Longer time lags tend to be associated with lower permeability sediments beneath the mid-Pleistocene (Q6-Q10) outwash gravels. These older terraces tend to be elevated higher above the main rivers. The combination of low permeability and a deeper water table contribute to time lags that are longer than the younger sediments. Time lags of five to ten years can be expected in parts of the Edendale, Knapdale, Longridge, east Makarewa, northwest Te Anau, south Upper Aparima, east Waimea Plain and Wendonside groundwater zones. Longer times lags can be expected in localised areas of the Knapdale, Longridge, Te Anau, and Wendonside zones.

We have attempted to verify our transit time model by comparing the results with tritium mean residence times. Unfortunately the tritium data that is available is from production bores screened at depths that are too deep to compare with our modelled results. The initial model verification for shallow bores does show encouraging results. However, we recommend that our time lag predictions be considered as preliminary estimates until the results have been verified with better data. We have good confidence in the relative values of the time lag estimates, which enable responsive areas to be managed differently to areas of longer travel times. We have less confidence in the absolute values of the travel times. This is reflected in the standard errors associated with the calculations, which are up to 50% in some areas. To improve our estimates, we intend to compare our modelled predictions with environmental tracer data from the top of the water table. This will help to constrain the absolute and relative time lag predictions made by the analytical models.

From a perceptive of vertical percolation of water, the question of a nitrate "time bomb" has to be answered in the negative. For most of Southland, shallow groundwater can be expected to have reached equilibrium nitrate concentrations for past land use. However, from the perspective of a cumulative environmental effect on a water supply bore, the question is yet to be answered. The reason for this is that the horizontal and cumulative effect of land drainage within groundwater needs to be addressed on a 2-D basis. Nitrate concentrations will tend to accrue along a flow path, as recharge from the land surface is mixed with flow from up-gradient parts of the aquifer. As a result, we can expect a continued increase in nitrate concentrations in deeper groundwater.

A similar logic applies for the cumulative effect of leaching on a surface water body. The nitrate discharged to a stream or wetland from groundwater is the result of many different flow paths, and the nitrate concentration will tend to accrue along these flow paths. It is hoped that the results of this study will provide the baseline input data to a regional scale assessment that considers horizontal flow towards surface water bodies.

9. References

Allen, R., Pereira, L.S., Raes, D., Smith, M., 1998. Crop evapotranspiration: guidelines for computing crop water requirements, *Irrigation and Drainage Paper 56*. FAO, Rome, 300p.

Apello, C.A.J. & Postma, D. 2005. *Geochemistry, groundwater and pollution* (2nd ed.). Balkema, Rotterdam, 649 p.

Brooks, R.H., & Corey, A.T. 1964. Hydraulic properties of porous media. *Hydrology Paper 3*. Colorado State University, Fort Collins.

Campbell, G.S., 1974. A simple method for determining unsaturated conductivity from moisture retention data. *Soil Science 117 (2)*, 311-314.

Cannavo, P., Richaume, A., & Lafolie, F., 2004. Fate of nitrogen and carbon in the vadose zone: in situ and laboratory measurements of seasonal variations in aerobic respiratory and denitrifying activities. *Soil Biology & Biochemistry (36)*, 463–478.

Charbeneau, R.J. & Daniel, D.D., 1992. Contaminated transport in unsaturated flow. In: Maidment, D.R. (ed.), 1992. *Handbook of Hydrology*. McGraw-Hill, New York.

Close, M., 2010. Critical review of contaminant transport time through the vadose zone. Technical report prepared for Environment Canterbury, Environment Canterbury Report no. R10/113.

Dann R., Close M., Flintoft M., Hector R., Barlow H., Thomas S., Francis G., 2009. Characterization and estimation of hydraulic properties in an alluvial gravel vadose zone. *Vadose Zone Journal 8*: 651–663.

De Silva, C.S. & Rushton, K.R., 2007. Groundwater recharge estimation using improved soil moisture balance methodology for a tropical climate with distinct dry seasons. *Hydrological Sciences Journal 52:5*, 1051-1067.

Domenico, P.A., & Schwartz, F.W., 1998. Physical and chemical hydrogeology (2nd ed.). Wiley, New York, 506p.

Fetter, C.W., 2001. *Applied Hydrogeology* (4th ed.). Prentice Hall, New Jersey.

Giltrap, D.J., 2002. Pedotransfer functions for moisture porosity and bulk density. Unpublished Landcare research report.

McWorter, D.B., & Sunada, D.K., 1977. Groundwater hydrology and hydraulics. Water Resources Publications, Ft. Collins, 290 p.

Meals, D.W., Dressing, S.A., & Davenport, T.E., 2010. Time lag in water quality response to best management practices: A review. *Journal of Environmental Quality 39*: 1, 85–96.

Mualem, Y., (1976). A new model for predicting the hydraulic conductivity of unsaturated porous media. *Water Resources Research 12*: 513-522.

Peck, R.B., Hanson, W.E., Thornburn, T.H., 1974. Foundation Engineering. Wiley, New York, 514p.

Posavec, K., Bačani, A., Nakić, Z, 2006. A Visual Basic spreadsheet macro for recession curve analysis. *Groundwater 44(5). DO: 10.1111/j.1745-6584.2006.00226.x*

Rawls, W.J., Ahuja, L.R., Brakensiek, D.L., Shirmohammadi, A., 1992. Infiltration and soil water movement. In: Maidment, D.R. (ed.), 1992. *Handbook of Hydrology*. McGraw-Hill, New York.

Rawls, W.J. & Brakensiek, D.L., 1985. Prediction of soil water properties for hydraulic modelling. *Watershed Management in the Eighties*, ASCE, p293-299.

Rossi, C. & Nimmo, J.R., 1994. Modeling of soil water retention from saturation to oven dryness. *Water Resources Research 30*: 701-708.

Rushton, K.R., Eilers, V.H.M., Carter, R.C., 2006. Improved soil moisture balance methodology for recharge estimation. *Journal of Hydrology 318*, 379-399.

Scotter, D., Heng, L., 2003. Estimating reference crop evapotranspiration in New Zealand. *Journal of Hydrology (NZ) 42*, 1-10.

Sousa, M.R., Jones, J.P., Frind, E.O., Rudolph, D.L., 2013. A simple method to assess unsaturated zone time lag in the travel time from ground surface to reactor. *Journal of contaminant hydrology 144*, 138-151.

Turnbull, I.M., 2000. Geology of the Wakatipu area. Institute of Geological & Nuclear Sciences 1:250,000 geological map 18. Lower Hutt, Institute of Geological & Nuclear Sciences. 1 sheet and 72 p. Lower Hutt, New Zealand. Institute of Geological & Nuclear Sciences Limited.

Turnbull, I.M. & Allibone, A.H. 2003. Geology of the Murihiku area. Institute of Geological & Nuclear Sciences 1:250,000 geological map 20. 1 sheet and 74 p. Lower Hutt, New Zealand. Institute of Geological & Nuclear Sciences Limited.

Turnbull, I.M., Allibone, A.H., Jongens, R., 2010. Geology of the Fiordland area. Institute of Geological & Nuclear Sciences 1:250,000 geological map 17. 1 sheet + 97p. Lower Hutt, New Zealand. GNS Science.

Van Genuchten, M., 1980. A closed-form equation for predicting the hydraulic conductivity of unsaturated soils. *Soil Science Society Journal 44(5)*: 892-898.

Wilson, S.R.; and Lu, X. 2011: Rainfall recharge assessment for Otago groundwater basins. Otago Regional Council technical report.

Zemansky, G, & White, P.A. 2012. Physical, chemical and isotopic methods for characterising shallow groundwater rainfall recharge. *New Zealand Hydrological Society Conference, Nelson 2012*.

Appendix 1: Summary of Rushton soil moisture balance model

The land-surface recharge model selected for this report is a simple spread-sheet model, based on the soil moisture balance model described by Rushton et al. (2006). This model has been verified with lysimeter data from Canterbury (REF). The Rushton model was found to predict lysimeter recharge accurately, as long as the soil hydraulic properties are known.

A key assumption made in the recharge calculations is that water is able to freely drain from the soil profile into underlying permeable geology. In many areas, the estimated rainfall recharge can only be considered to be potential recharge. For groundwater recharge to occur, there needs to be a suitable receiving medium underlying the soil profile.

For example, in the Manuherikia and Ida Valleys, much of the valley floor consists of Tertiary mudstones. These mudstones tend to form a barrier to drainage, so that rainfall moving through the soil profile is impeded. In these areas, rainfall infiltration either moves laterally towards the nearest surface water body, or surface ponding occurs once the soil moisture reaches field capacity. Likewise, we assumed that recharge would not be intercepted by drainage before reaching the water table.

Spread-sheet calculations for soil moisture balance follow the algorithms provided in Rushton et al. (2006). The Rushton model consists of a two-stage process - calculation of near-surface storage and calculation of the moisture balance in the subsurface soil profile. The near-surface soil storage reservoir provides moisture to the soil profile once all the near-surface outputs have been accounted for. If there is no moisture deficit in the soil profile, recharge to groundwater occurs.

The Rushton model has been adapted for this report to incorporate run-off, which was calculated using the US Department of Agriculture, Soil Conservation Service (SCS) run-off curve number model. The SCS run-off model is described in Rawls et al. (1992).

There are three steps to describe the soil moisture balance:

1. ***calculation of infiltration to the soil zone (In), and near-surface soil storage for the end of the current day (SOILSTOR)*** - note that infiltration (In), as specified by the Rushton algorithms, is not just infiltration (rainfall-run-off), it also includes SOILSTOR from the previous day.

2. ***estimation of actual evapotranspiration (AET)*** - PET is derived by the Penman-Monteith equation (Allen et al., 1998). A crop co-efficient is not applied here since the crop is assumed to be pasture, which is the reference crop for the Penman-Monteith equation. Most pastures in New Zealand behave like the reference crop for most of the year (Scotter and Heng, 2003).

3. ***calculation of soil moisture deficit and groundwater recharge*** - recharge occurs only when the soil moisture deficit is negative, i.e. when there is surplus water in the soil moisture reservoir. Note that Rushton model is usually started in winter when the initial soil moisture deficit can be safely assumed to be nil.

The three steps outlined above partition near-surface soil storage between near-surface soil storage for the following day, AET, and the soil moisture deficit/reservoir respectively.

In addition to rainfall and PET data, the soil moisture balance model requires input values for four different soil input parameters to calculate the daily soil moisture deficit. These parameters are described below:

- ***Curve Number*** - a curve number is estimated for each soil, which is then used to calculate maximum soil retention of run-off. The lower the curve number, the greater the soil retention threshold, which results in reduced run-off. Pasture in good condition on free-draining soil has a low curve number (39). Pasture in poor condition on a poorly drained soil has a high curve number (89). Additional values are given in Table 5.5.1 of Rawls et al. (1992).

- ***Profile Available Water*** - PAW or TAW is calculated from field capacity, wilting point and rooting depth data. Typical values for field capacity and wilting point are given in Table 19 of Allen et al. (1998), and many values for New Zealand soils can be found in the literature (e.g. McLaren and Cameron, 1996). It is more difficult to determine appropriate values for rooting depth. Values quoted in the literature are usually for uninhibited root penetration. Some knowledge of the soil profile is required to estimate rooting depth, because root penetration at a particular site may be limited by the presence of a resistive layer such as a loess or clay pan. If rooting depth is not known, it may be estimated from the profile thickness for thicker soil units. However, because rooting depth is also a function of water capacity and aeration properties of the soil, some caution is needed in using profile thickness as a proxy.

- ***Profile Readily Available Water*** - PRAW or RAW is related to PAW by a depletion factor, p. The depletion factor is the average fraction of RAW that can be depleted from the root zone before moisture stress (reduction in ET). For NZ conditions, p should be around 0.4 to 0.6, typically 0.5 for pasture. (See Table 22 of Allen et al. (1998) for more values.)

- ***Fracstor*** - this is the near-surface soil retention. Values are estimated for Otago soils based on values obtained elsewhere. The contribution of Fracstor to the soil moisture balance is small, so errors resulting from estimations are considered to be negligible. Typical values are 0 for a coarse sandy soil, 0.4 for a sandy loam and 0.75 for a clay loam (Rushton, 2006, p. 388). Appropriate values can be estimated from field observations (de Silva and Rushton, 2007). If the soil dries quickly, then Fracstor will be less than 0.3. If the soil surface remains wet after heavy rainfall, so that it is not possible to work the soil for several days, then Fracstor is likely to be in the range 0.6–0.8.

Appendix 2: Regional depth to groundwater

Brydon Hughes, February 2013

Objective

- To produce a spatial coverage representing a best estimate of the average depth to groundwater across the Southland region.
- To produce a spatial coverage quantifying the typical magnitude of seasonal groundwater level variation across the Southland region.

Compilation of input data

- Groundwater level data recorded in the Wells database was exported to excel and a pivot table created to calculate average groundwater levels at all sites with >5 groundwater level measurements. Where possible this data was utilised to guide inclusion of other static water level data and the generation of interpolated groundwater levels. A similar procedure was used to utilise data for telemetered groundwater levels monitoring sites stored on the Hilltop archive (utilising the PSum function in Hydro). These data comprised the initial data added to the input data shapefile.

- Results from piezometric surveys undertaken in various parts of the region were plotted in ArcGIS and manually added to the input data shapefile. Each record was compared to surrounding measurements and baseline water level data (taking into account topographical variations identified from a combination of the NZMS260 coverages and the DEM) to exclude any data points influenced due to semi-confined/confined conditions or localised effects (e.g. pumping or unexplained variations including possible transcription errors).

- Static water levels recorded in the Wells database were exported, plotted in ArcGIS and manually added to the input data shapefile. Comparison was made with baseline monitoring, piezometric survey and nearby Wells static levels to try and exclude spurious data points which were not consistent with surrounding levels and/or topographic variations. Where available bore logs were utilised to try and identify if individual bores were screened at a depth likely to encounter semi-confined or confined aquifers and this data excluded (particularly in areas such as the Oreti Basin, Wendonside, Eastern Southland, the Lower Aparima (Gunnies Bush) and the Upper Aparima catchment where the deeper aquifers are known to exist). In many areas, particularly west of the Oreti River, the Wells static water level data (with relatively sparse baseline sites) is the only data available to quantify groundwater depth. It is noted that in some areas (especially the Te Anau Basin) it is very difficult to identify bores exhibiting some degree of confinement.

- Dummy points were added along the major surface water features. Dummy values of either 0.5 m or 1 m were entered for rivers/streams where there is a known hydraulic connection. Dummy points of 2 m were entered along those streams where direct hydraulic connection was uncertain but groundwater depths clearly decrease due to topographical variations. Surface water points were generally spaced relatively closely to ensure continuity in the final contour output (total of around 3,500 surface water points added).

- In order to replicate the influence of topography on groundwater depths, infill areas with no other data and ensure the final output reflected the actual geometry of the water table a large number of interpolated data points (~3,000) were added to the input data file. The location of these data points is largely arbitrary but topically reflects topography (i.e. terrace margins, valley margins and areas with little/no data). Interpolated groundwater depth were estimated from existing static water level data or estimated from topographical variations. For example, interpolated groundwater depths were assumed to be relatively deep under elongate terrace features e.g. moraine terraces in Te Anau, ridges at the eastern end of the Waimea Plains, where the topography rises rapidly from surrounding lower lying terraces. Interpolated points were added in an iterative manner in areas with limited data to produce a final coverage which replicated the estimated geometry of the water table (based on available static water levels, topographic variations and similar areas elsewhere).

Derivation of depth to groundwater coverage

- The depth to groundwater coverage was derived from the input data shapefile using ArcGIS spatial analyst. The depth to water coverage was interpolated using the inverse distance weighted (IDW) method. Default settings were used except the number of points utilised in the interpolation was reduced to 4, which seemed to produce a relatively even coverage particularly along rivers and streams.

- A mask was applied to the interpolation to include only those areas with data included in the input data shapefile (essentially the extent of groundwater zones defined in the RWP modified to remove the area between Ohai and Eastern Bush where these is significant topographical and geological complexity and no static water level data). A separate mask was used to model the Upper Mataura groundwater zone.

Seasonal groundwater level variation

- Seasonal groundwater level variations were derived from evaluation of temporal groundwater level data from baseline monitoring sites typically having more than three years record. Various statistics were utilised (i.e. 10 percentile, 90 percentile, standard deviation) to quantify the range of natural seasonal variation. However, due to inter-seasonal variation such rigid procedures were potentially influenced by the period and length of record (i.e. if the record included particularly wet years (e.g. 2004/05) or dry periods (2003). Ultimately, an estimate of 'normal' seasonal variation was derived from visual inspection of each individual hydrograph taking into account the nature of the seasonal response (e.g. riparian vs terrace-type variation), and the magnitude of variations in specific seasons. In a majority of cases, most bores showed a relatively consistent magnitude of seasonal variation except in exceptionally wet/dry years.

- Seasonal variations assigned to individual monitoring sites were plotted in ArcGIS. It was originally intended to either contour the data or assign various levels of 'typical' seasonal fluctuation to sub-areas of individual groundwater zones. However, the distribution of data was insufficient to reliably undertake this process.

- As a general rule of thumb, the data show seasonal variations are typically:
  - >1 metre in Riparian aquifers (i.e. those exhibiting a stream-like hydrograph) and areas along the margins of hydraulically connected rivers and streams in Lowland aquifers;
  - between 1 to 2 metres in Lowland aquifers (away from stream margins) and along downgradient margins of terrace aquifers;
  - up to 3 metres in central and upper areas of Terrace aquifers; and
  - treater than 4 metres only in the limestone aquifer at Isla Bank (likely to reflect the low secondary porosity in this fractured/jointed system). This observation may also apply to other limestone systems in the Lower Aparima and Forest Hill areas.

Comments on preliminary coverages

- As it currently stands, the regional depth to groundwater coverage represents an approximation of the likely 'average' depth to groundwater across the region. As previously noted, this probably doesn't vary too much seasonally across a majority of the region where typical seasonal variations are less than 2 metres.

- In some areas e.g. the Oreti Basin, western end of the Waimea Plain, Riversdale, Knapdale, Edendale and areas of the Central Plains, depth to groundwater is relatively well constrained by the availability of data and relatively flat topography. However in other areas e.g. along most basin margins, the eastern end of the Waimea Plain, Hedgehope/Titipua/Croydon, the Lower Waiau catchment and Te Anau Basin there is greater uncertainty due to a lack of static water level data and topographical complexity (in many cases bores are typically drilled in lower-lying areas so depth to groundwater is often poorly constrained on higher ridges).

- It is often difficult to reliably determine the depth to the shallowest water table in areas of geological complexity. For example, in the Te Anau Basin spatially discrete perched aquifers are likely to occur under some of the larger terraces (as evidenced by springs/seeps on terrace risers) but the distribution of these features is unknown. A majority of bores tend to be drilled to intercept the deeper 'regional' piezometric surface. Similarly, in the area of the Kepler Swamp, the shallow water table in the swamp may be perched over a deeper regional piezometric surface. In this area, depth to groundwater has been assumed to reflect the perched water table.

- The current project is set up so it can be modified as additional data becomes available. This process is relatively simple involving adding new data points to the input data shapefile (and modifying any existing interpolated levels in the vicinity) then interpolating a revised depth to groundwater coverage using Spatial Analyst as described above.