GIS coupled Water Budget for Spatial and Temporal Analysis of Water Resources: Horowhenua, New Zealand

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Abstract

A spatially and temporally enhanced water budget was developed for and applied for Horowhenua, New Zealand for 2007, 2008 and 2009. High resolution daily precipitation surfaces were generated in ArcGIS along with daily actual evapotranspiration derived from the FAO Penman-Monteith equation. Gauged stream flows were used to derive surface outflows from a regression model. Groundwater level data were used to derive potentiometric surfaces, storage change from water table fluctuations (WTF) and hydraulic gradient which in turn were used to calculate groundwater outflow via Darcy's Law. Annual groundwater storage change varied significantly with mean estimated values of -53.8, -23.9 and 42.5 Million cubic meters (Mcm) estimated for 2007, 2008 and 2009 respectively. Monthly storage had a higher variability, with values being greater in magnitude than net annual change. Total volume and rainfall pattern were identified as key explanations for the storage change behaviour. A low rainfall year inclines towards a negative storage change and a high rainfall year towards a positive storage change. However, a high rainfall year may have a negative recharge if storms occur whereby rainfall intensity is increased resulting in larger surface outflow as a percentage of rainfall. Resolution of GIS surfaces is very important for evapotranspiration which is affected by landuse, thus retention of spatial integrity with an appropriate resolution is important. Data availability was a major limitation to the potential of the GIS-coupled water budget technique, specifically a nonexistence of stream gauging stations in locations near the coast. Likewise, an absence of daily information for groundwater levels and water consumption data impeded the temporal resolution that could be achieved. This research has displayed the potential of a water budget coupled with high resolution GIS data to provide valuable information for water resources.

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Chapter 1 Introduction

1.1 Introduction

Groundwater is the water resource that exists underneath the land surface (Krešić, 2009) accounting for a mere 1.81 % of the total global water resources. However, groundwater makes up 98.4 % of the world's unfrozen fresh water (Table 1)(Mather, 1984) making it of particular interest for life on this planet. Humans require a lot of water, not only for direct consumption and food production but also for industry, manufacturing and recreational purposes (Healy, Winter, LaBaugh, & Franke, 2007). The necessity of fresh water for life, together with the large reserves, broad geographical distribution and general good quality of the water puts a huge impetus on the study and management of groundwater as one of our most important resources (Krešić, 2009; McCarron & Zarour, 2005).

Table 1 Water in the hydrosphere from Mather (1984) and Montgomery (2006)

Reservoir	Percentage of Total Water	Percentage of Fresh Water	Percentage of Unfrozen Fresh Water
Oceans	97.54	-	-
Ice	1.81	73.9	-
Ground Water	0.63	25.7	**98.4**
Lakes and Streams			
Salt	0.007	-	-
Fresh	0.009	0.36	1.4
Atmosphere	0.001	0.04	0.2

The study of groundwater is broad and encompasses many different avenues of research. One of the most important aspects of groundwater studies is the understanding of groundwater recharge and replenishment. Groundwater recharge is essentially the inflow or replenishment to the groundwater reservoir working off the basic water budget principle of '*Flow In – Flow Out = Change In Storage*'. Understanding the relationship between groundwater recharge, groundwater discharge and the change in storage within the water budget is fundamental for effective sustainable management and quantification of groundwater resources and hence is often used as measure of sustainable yield of a groundwater system (Healy, et al., 2007; Zarour, 2008).

Groundwater cannot not be viewed in isolation, for increased understanding the groundwater hydrology is viewed within wider hydrological cycle. The groundwater behaviour is dependent on many variables drawing from an interdisciplinary approach for evaluation. Climate, geology, soils, landuse and vegetation, drainage patterns and anthropogenic abstraction are all key variables that are needed for groundwater behaviour understanding (Healy, et al., 2007). A variety of literature has suggested the importance of distributed approaches in improving spatial estimates incorporating spatially varied components already mentioned and the increase in high resolution spatial data has made this desirable (Batelaan & De Smedt, 2007; Freeze, 1969; Jackson, 2002)

The Horowhenua Region is subject increasing focus due to the propagation of hydrological issues surrounding management of water resources and the health of Horowhenua Lake which is replenished from both groundwater and surface water. Horowhenua also displays interesting surface drainage patterns, geological conditions and significant relationships between surface water, groundwater and its replenishment. In order to accurately develop and assess management of the water resources and understanding of the catchment hydrology is needed. These facets along with the easily manageable catchment size of Horowhenua provide an ideal study location to initiate research from which future research can build on and be developed.

1.2 Aim and objectives

This research aims to improve understanding of water resources in Horowhenua, New Zealand, by carrying out a spatially and temporally distributed GIS coupled water budget. This will be accomplished with the following objectives:

- Identify the key components affecting recharge and discharge in Horowhenua
- Implement a high resolution spatially sensitive GIS technique to calculate and create daily evapotranspiration layers for 2007, 2008 and 2009
- Create high resolution spatially sensitive GIS layers for daily rainfall for 2007, 2008 and 2009
- Create monthly potentiometric surfaces from well data and use it for calculation of hydraulic gradient and subsequent calculation of groundwater outflow for 2007, 2008 and 2009
- Estimate surface water outflow from extrapolation of gauged stream flow data for 2007, 2008 and 2009
- Calculate monthly and annual water budget components to establish storage change behaviour for 2007, 2008 and 2009
- Analyse and discuss relationships between calculated water budget components and groundwater level behaviour and implications produced for management
- Critically appraise the technique utilized along with limitations and aspects that can improve the technique for future use

1.3 Literature review

Groundwater recharge is an important part of the hydrological cycle but it remains as one of the more poorly constrained and difficult components of the cycle to identify (Lee, Chen, & Lee, 2006; Lerner, Issar, & Simmers, 1990). Recharge rates have been the subject of study for hydrologists for many years for the purpose of deducing long-term yields of groundwater systems e.g. Theis (1937, 1940). Various methods have been developed to quantify recharge which produce estimates over various temporal and spatial scales incorporating a range of complexity and expenses (Healy & Cook, 2002).

Groundwater analysis techniques include many different methods ranging from direct to numerical. Lysimeters are used as a direct point measurement of recharge involving an in-situ soil and vegetation containment which collects percolated water beneath to measure recharge and deduce actual evapotranspiration. Frequently they are used as accuracy test for less direct modes e.g. Liu & Luo (2010), Vaughan, Trout & Ayars (2007), and Xu & Chen (2005). The devices are generally representative but isolated from surrounding environment and generally need to be large in order to reduce edge effects causing them to be expensive in construction (Lerner, et al., 1990). Assortments of Darcian approaches rely on Darcy's law in governing the movement of water through the unsaturated zone. It is very widely applicable and easy to apply when hydraulic conductivity and hydraulic gradient information is available (Scanlon, Healy, & Cook, 2002). The individual technique is used in numerous studies and forms the basis for numerical ground flow models dependent largely on the accuracy of the hydraulic conductivity information which can give rise to large uncertainties due to the large order of magnitude ranges that exist for its values. Zero flux plane methods, also stemmed from Darcian equations, are first described by Richards, Gardner & Ogata (1956) as a soil-moisture budget approach equating recharge to changes in soil-water storage below the zero flux plane working best when large fluctuations exist (Scanlon, et al., 2002).

Tracer techniques such as isotopic (Stewart, Mehlhorn, & Elliott, 2007; Taylor, Brown, Cunliffe, & Davidson, 1992; Taylor, et al., 1989), thermal (Anderson, 2005; Hatch, Fisher, Revenaugh, Constantz, & Ruehl, 2006) and historic and environmental involving tracking movement of the tracer through the groundwater system. Tracers can be very expensive and can provide difficultly in actual quantification (Scanlon, et al., 2002). Water table fluctuation (WTF) provides another physical technique based around the rise and fall of the water table. The technique encounters problems when other variables are responsible for water table fluctuations, particularly that of water extraction e.g. Healy & Cook, (2002).

Water Budget techniques are based around calculation of inputs and outputs to deduce change in storage. The technique is widely used by government bodies for safe groundwater withdrawal levels, although possibly applied and interpreted inappropriately (Bredehoeft, 2002; Zhou, 2009) . Major components usually consist of precipitation, evapotranspiration, surface water outflow and inflow, groundwater inflow and outflow. This method is highly flexible as its not dependent on mechanism understanding and can be applied at a range of spatial and time scales, however can encounter large uncertainties especially in arid environments (Lerner, et al., 1990; Scanlon, et al., 2002). Numerical models have increased the potential for water budget application by increasing spatial and temporal accuracy e.g Batelaan & De Smedt (2007) and with ever increasing computing capability potential is likely to increase further.

A brief overview of groundwater recharge estimation techniques was given above as the topic is very large and beyond the scope of this paper to discuss every technique in detail. Detailed discussions and literature for many of the techniques can be found in de Vries & Simmers (2002), Lerner et al. (1990), Scanlon et al. (2002), and Simmers (1988). The water budget forms the central methodology to this papers research and forms the basis for the following literature discussion.

Estimation techniques based on water balance formed the early conventional method for estimating recharge. The technique was based on the work of Penman (Penman, 1948, 1950a, 1950b, 1951) and Grindley (Grindley, 1967, 1969, 1970) researching water balance to determine evaporation and soil moisture deficits; recharge estimates were just a by-product (Rushton & Ward, 1979). Rushton and Ward (1979) state that the Penman-type method tends to underestimate the recharge due to the Penman equation assuming unlimited water supply for evaporation. However, this is not the case as actual evaporation frequently falls below the potential evaporation. Lysimeter investigation over 3 years found 175 % more recharge than the conventional method in Bunter Sandstone, Nottinghamshire (Kitching & Bridge, 1974; Kitching, Shearer, & Shedlock, 1977). Rushton and Ward (1979) state that similar findings were found using tracer studies (Downing, Smith, & Warren, 1978; Smith, Wearn, Richards, & Rowe, 1970), stream analysis (Headworth, 1970) and groundwater hydrographs (Headworth, 1972; K. Ward, 1976).

Howard and Lloyd (1979) indicate that the water balance equation was frequently applied uncritically with little regard for significant errors that can arise due to sensitivity of equation parameters. Often this is due to inaccurate or unrepresentative input data being used and a lack of appreciation for the influence of time increments on the estimation. The study showed strongly that daily recharge calculations should be adopted as opposed to ten day or monthly calculations to avoid large errors arising. Furthermore, Howard and Lloyd (1979) note that topography is seen to strongly influence precipitation and evapotranspiration and hence a need for a series of recharge nodes to establish the areal distribution. Rushton (1988) states the need to consider location and flow mechanisms critically; success of one model at one location doesn't guarantee the successful implementation of the same model at another location. If not considered properly, models may ignore the difference between potential recharge from the soil zone and the actual recharge which enters the aquifer (Rushton, 1988).

Increasing realization for the need of reliable accurate recharge estimation caused groundwater recharge studies to experience a relative explosion in the mid-1980s demonstrated by numerous publications that emerged from international conferences (de Vries & Simmers, 2002). Several articles arose comparing multiple recharge techniques. Johansson (1988) demonstrated the need for comparative due to uncertainty in estimates experienced between six methods of recharge estimation in sandy till in southwest Sweden. The water budget methods (one allowing recharge only when no soil moisture deficit occurred and the other allowing a fraction of the precipitation to provide recharge while deficit occurred) showed similar annual values to the one dimensional soil water flow model (SOIL), although some individual years deviated by 25 – 30 % primarily due to underestimation during summer. Allowing recharge to occur with soil moisture deficit proved to more accurately reproduce the dynamics expressed from fluctuations in groundwater level. Uma (1988) showed that as a percentage of total rainfall, water balance methods were consistently higher (13.15 % to 43.19 %) compared to baseflow regression (10.86 % to 28.59 %) and a groundwater stage technique (20.67 % to 37.23 %) but overall showed good correlation, during a study of numerous watersheds in Nigeria. Sinha (1988) compared empirical methods, hydrological budgeting and groundwater fluctuations, and stated the need for quantification of the water resources of different administrative units in a realistic basis in order to have proper management of those resources. Groundwater fluctuation (WTF) method was recommended as the most suitable except for where monitoring was insufficient, in which case hydrological budgeting should be used.

The water budget approach has a great advantage of being flexible, applicable to the full spectrum of areal and temporal scales, and makes use of readily available meteorological data that is rapid to apply over study areas (Lerner, et al., 1990; Scanlon, et al., 2002). The accuracy of the water budget approach depends on the accuracy of the other components in the water balance equation as shown by Senarath (1988), where accurate precipitation and evapotranspiration made realistic estimation possible. The accuracy becomes more of a significant issue when the recharge rate is small relative to the other variables (Scanlon, et al., 2002). On that basis the water balance technique is best suited for temperate or humid areas where there are few periods of less than potential evaporation, as opposed to semi-arid environments where precipitation and evaporation are close to equal and large errors arise (Allison, 1988). Groundwater recharge studies have predominantly focused on semiarid and arid locations whereas groundwater recharge studies for humid areas have generally been incorporated into overall water balance studies and not explicitly studied well (de Vries & Simmers, 2002).

The need to understand the spatial and temporal variation of groundwater recharge has been put forward by some authors, e.g (Batelaan & De Smedt, 2007; Jackson, 2002; Sophocleous, 1992). The non-linearity nature of groundwater recharge and high spatial and temporal variability needs an established practical method to regionalize recharge estimates; requiring continuous monitoring and appropriate representative input data (Sophocleous, 1992). However, most direct measurements of the hydrological variables are point measurements and are not adequately integrated over space and time (Sophocleous, 1992). Geographic Information Systems (GIS) and remote sensing have been identified as a solution to offer improvements to spatial and temporal estimates (Jackson, 2002; Sophocleous, 1992; Tilahun & Merkel, 2009). This argument for application of GIS is enhanced due to increases in technology leading to significant increases in high resolution spatial data for surface characteristics such as land cover, vegetation and soil characteristics, and providing a means to directly account for spatial variability of variables that affect recharge (Batelaan & De Smedt, 2007). Sophocleous (1992) used a limited number of site specific but year-round measurements to regionalize and analyse recharge in central Kanasas over spatial and temporal time scales. Study revealed recharge events typically last for 5 – 7 days with 2 – 4 being precipitation days. The overlaid GIS layers showed that recharge zonation agreed well with site recharge estimates.

Batelaan & De Smedt (2007) use a spatially distributed water balance model to simulate long-term average recharge depending on landcover, soil texture, topography and hydrometeorology parameters through a GIS-based iterative process. Results showed that recharge varied significantly spatially in a complex pattern depending to a large extent to soil texture and land cover, with negative recharge occurring where there is a shallow water table. Tilahun & Merkel (2009) also used a GIS-based distributed water model in Ethiopia which was critical in determining that recharge was less than had been presumed. The rainfall volume was apportioned as: 75% evapotranspiration, 20% runoff, and 5% recharge.

A variety of studies have been conducted in the Horowhenua area, either singularly or within the context of wider regional studies, although until recently most of the focus has been on the water quality as opposed to quantity. The geology of the region has been relatively well studied with early literature by the likes of (Adkin, 1910) accounting for the varied physical features and geological formations of the area around the Ohau River. Wider studies of the regional groundwater system have had a misappropriation of mathematically modelled layers as genuine hydrostratigraphic units whereby vertical subdivision of the groundwater resource into a 5-aquifers system was conceptualized and adopted (Zarour, 2008). The 5-aquifer model however misconceives the regional geology and lacks understanding of channelled deposition and vaguely utilizes principles of sequence stratigraphy (Zarour, 2008). The regional groundwater system is thus conceptualized as a single, heterogeneous, anisotropic, hydraulically interconnected groundwater system that occurs in hydraulic continuity with surface water regimes (Zarour, 2008). Furthermore, Zarour (2008) states that the regional groundwater system is limited to the uppermost veneer of the sedimentary sequence.

Begg et al. (2005) conducted a geological synopsis of the hydrogeology of the entire Manawatu-Horowhenua region. They stated the quaternary units (Q1 to Q4) as being the most significant for groundwater resources; and the importance of tectonic activity in effecting the distribution of aquifers and sediment transport. Begg et al. (2005) also mapped groundwater quality domains indicating the potential for direct recharge and interaction with surface water features indicated by valley fill and unconfined zones (Figure 1)

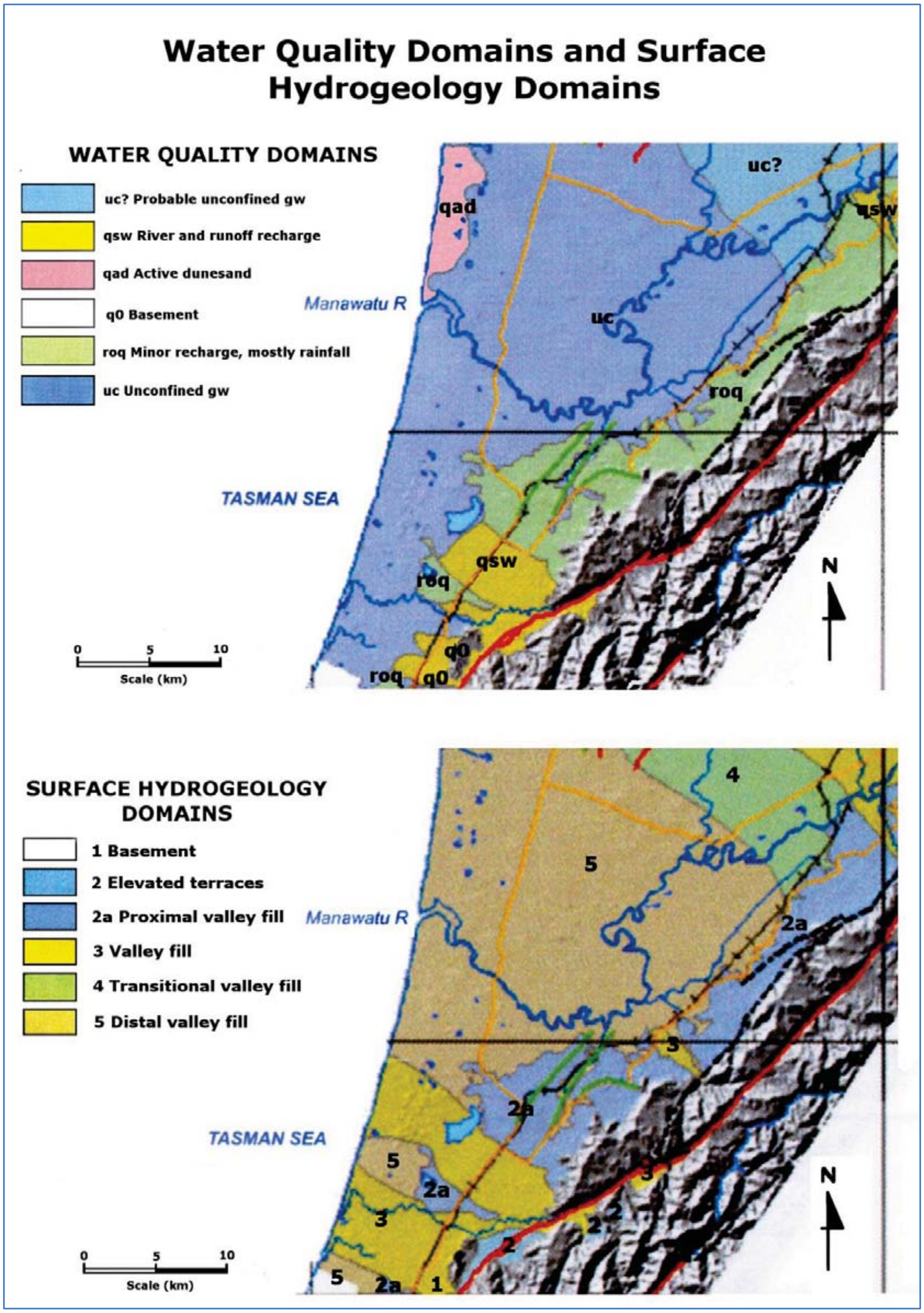

Figure 1 Top: water quality domain of Manawatu showing Horowhenua has basement rock in the Tararua Range; predominately river and runoff recharge (qsw) in the plains and flanked by minor recharge (roq); and unconfined groundwater between Lake Horowhenua and the coast. Bottom: surface hydrogeology domains of the Manawatu showing Horowhenua is predominately valley fill (3), proximal valley fill (2a) and distal valley fill (5). Adapted from (Begg, Palmer, & Gyopari, 2005).

Work by Bekesi (1996) into nitrate concentrations of the groundwater in the region displayed the model (from earlier work of (Bekesi, 1989)) of the Horowhenua whereby the assumption of displacement of the greywacke along the north-east orientated Levin fault (Figure 2). The fault was thought to result in upward movement of the groundwater due to the low hydraulic conductivity of the basement greywacke and upward rising deep groundwater (Bekesi, 2001). This fault however is not documented in the qmap series (Begg & Johnston, 2000). A development of a regional rainfall recharge model was displayed by (Bekesi & McConchie, 1999) by calculating mean annual recharge at each rainfall station and using the Monte Carlo technique to randomize soil moisture parameters and subsequently produce a recharge surface (Figure 3). There was good agreement with the groundwater levels obtained as well as the spatial recharge pattern was that is expected due to physical processes and influences of the Tararua Range (Bekesi & McConchie, 1999).

Recently, a steady state water budget approach was used in Horowhenua by White, et al. (2010) to understand the average hydrological conditions and the surface water and

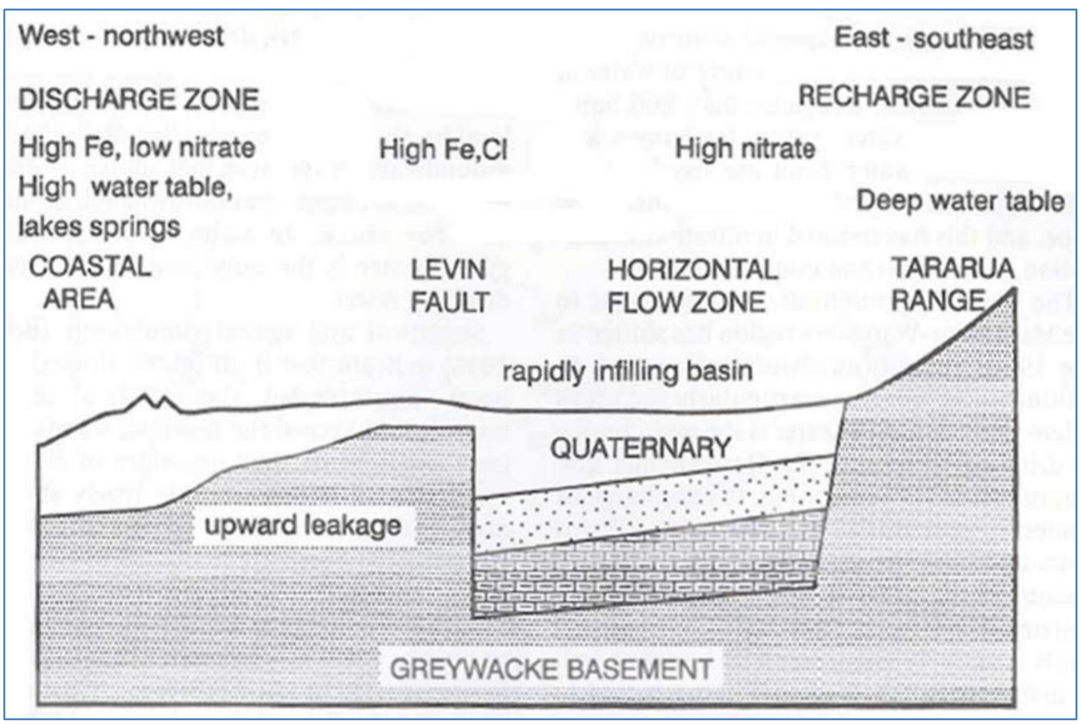

Figure 2 Conceptual hydrogeological cross section for Horowhenua showing proposed Levin Fault as a controlling structure to groundwater movement. From (Bekesi, 1996)

groundwater interactions occurring in the plains of the Horowhenua area. The study used multiple data sources for different variables, resulting in 7 different calculations showing high uncertainty and irregularity of data values that have been estimated for some of the constraining variables. The water budget was calculated for sub-catchments within the Horowhenua area with their preferred values presented in Table 2. It was elucidated from this study that rainfall is the largest recharge component to the groundwater system, of which most discharges to the sea. Significant groundwater recharge occurs from the rivers (Figure 4) which probably re-enter the surface water near the coast. The Lake Horowhenua groundwater catchment outflows almost entirely to Lake Horowhenua and is essential in the Lake water budget.

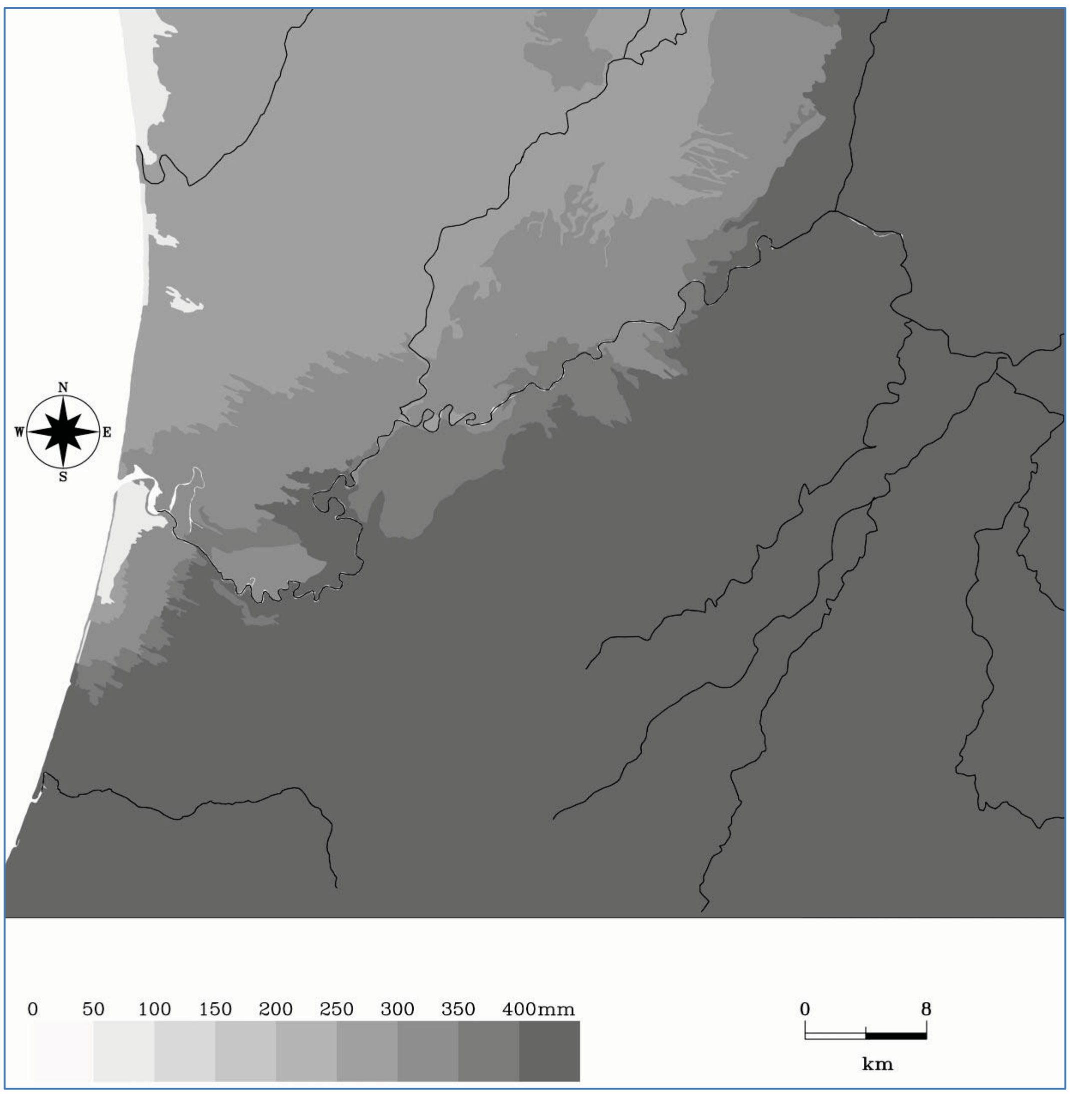

Figure 3 Recharge map from (Bekesi & McConchie, 1999) showing highest recharge occurring over the greywacke ranges

Table 2 Preferred values for Horowhenua annual water budget from White et al. (2010)

Groundwater Catchment	Components of the Water Budget (Mcm)*						
	P_{Spasmo}	$Q_{in\ Horizons/Niwa}$	E_{Spasmo}	$Q_{out\ Spasmo}$	G_{out}	U_{GW}	U_{surf}
Waitarere North	16.2	0	13.4	0	2.8	0	0
Poroutawhao	24	0	19.2	0.4	3.9	0.5	0
Koputaroa West	32.9	0	20.6	1.4	9.7	1.2	0
Koputaroa East	49.0	7.2	28.4	9.3	17.4	1.1	0
Waitarere South	24.9	20.6	19.7	20.8	4.5	0.5	0
Lake Horowhenua	63.8	0	41.1	20.6	0	2.1	0
Waiwiri	24.2	0	17.8	0.6	5.5	0.3	0
Ohau	61.0	229.4	36.9	227.5	23.3	2.7	4.7
Waikawa	41.7	43.8	26.2	45.6	12.2	1.5	0.1
Total	**337.7**	**301.1**	**223.3**	**326.2**	**79.3**	**9.9**	**4.8**

*P_{spasmo} = Precipitation, $Q_{in\ Horizons/Niwa}$ = Surface water flow, E_{Spasmo} = Evapotranspiration, $Q_{out\ Spasmo}$ Surface water outflow, G_{out} = groundwater outflow, U_{GW} Groundwater use, U_{surf} = Surface water use. Water balance equation: $P_{spasmo} + Q_{in\ Horizons/Niwa} – E_{Spasmo} – Q_{out\ Spasmo} – G_{out} – U_{GW} – U_{surf}$

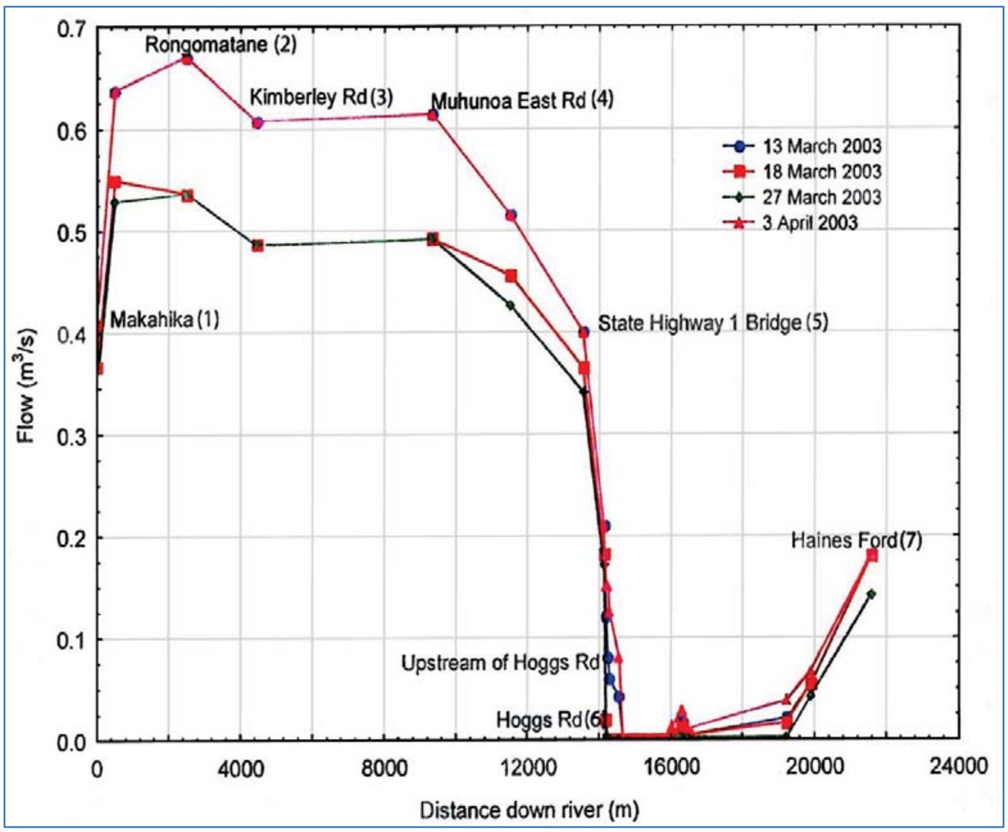

Figure 4 Low-flow gauging's of Ohau River showing river losses and gains from White et al. (2010)

1.4 Study Area

The study location is located in the south west of the North Island of New Zealand (Figure 5) in a region known as Horowhenua. The major township in the area is Levin with a population of ≈20 000 and covering an 8 km^2 area; total population of Horowhenua is ≈30,000. The site falls within the Manawatu-Wanganui region which is divided into seven Groundwater Management Zones (GMZ) by Horizons Regional Council. The boundary of the study area is therefore based on the management boundary zone known as Horowhenua Groundwater Management Zone, (HGMZ) (Zarour, 2008). The area totals 392.57 km^2 and has been divided into a series of sub-catchments by White et al. (2010) and for purposes of comparison these will be used in the results and discussion (Figure 6). Three hill catchments have been allocated (Figure 6) based on River Environment Classification (REC) data produced by (Snelder, Briggs, & Weatherhead, 2004).

The climate of Horowhenua is temperate and tends to be quite windy due to the exposure to weather systems from the Tasman Sea although there are few climatic extremes. The prevailing airflows are North-westerly. Summers have the most settled weather and are warm, typically ranging from 19°C to 24°C seldom exceeding 30°C (Mackintosh, 2001). Conversely, winters tend to have the most unsettled weather, with maximum air temperatures ranging from 10°C to 14°C. Sunshine hours average around 2000 hours annually (Mackintosh, 2001). Average annual rainfall for Levin is 1095 mm (Manawatu-Wanganui Regional Council, 1998).

Table 3 Calculated area for designated sub-catchments in Horowhenua

Catchment	Area (km^2)
Waitarere North	19.50
Waitarere South	28.52
Lake Horowhenua	61.54
Waiwiri	25.88
Ohau	54.18
Waikawa	38.48
Plains Total	**228.09**
Hill 1	1.21
Hill 2	125.33
Hill 3	39.44
Hillslope Total	**165.99**
Total*	**392.57**

***NB: Total area may not match a summation of the sub-catchments due to a slight variation in boundaries between GNS catchments and the boundaries derived in this study.**

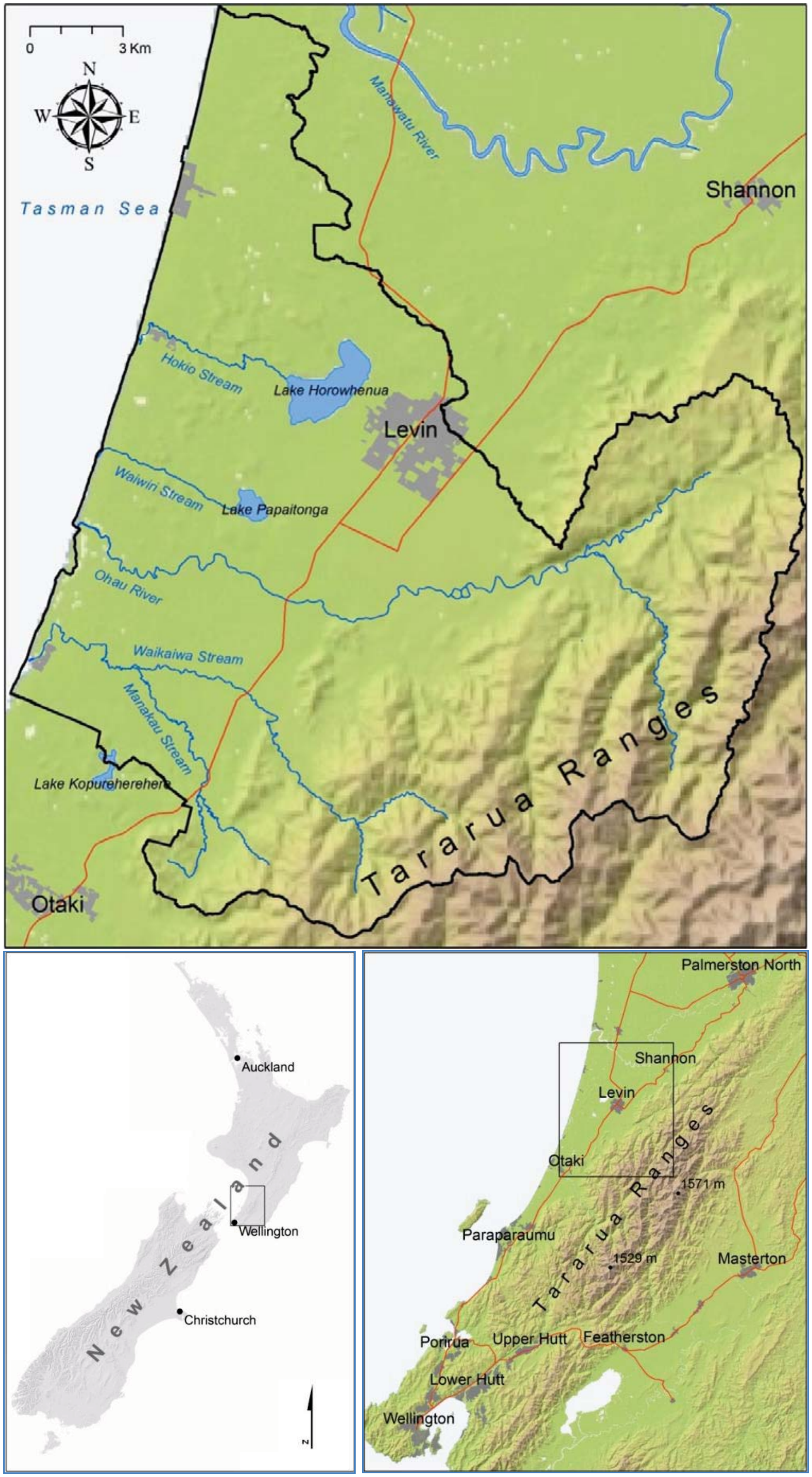

Figure 5 Bottom left: map of New Zealand showing study region. Bottom right: Horowhenua study site shown within region. Top: Horowhenua study area showing key cultural features

The lowlands of Horowhenua are dominated by fixed and mobile sand dunes with the formation of dune ridges aligned northwest/southeast (Hesp & Shepherd, 1978). Longshore drift brings the sediment from the north and wind action subsequently blows it inshore to form the largest dune field in New Zealand. Drainage of the Tararua Range has resulted in coalescing fans of alluvial debris forming inland from the dunes with a distinct margin separating sand country and alluvial plains (Begg, et al., 2005). Peat swamps occur quite extensively where dunes have impeded drainage particularly around Lake Horowhenua. The Tararua Range forms a northeast/southwest aligned axial range which sets the back drop to Horowhenua as the dominate feature to the east (Figure 5). The highest elevation in the Tararua Range is 1504 m asl however within the study area the maximum elevation is in the order of ≈1000 m asl but most of the catchment doesn't exceed 500 m asl (Figure 7).

The most significant drainage features of Horowhenua are the Ohau River, and the Waikawa Stream to the south which confluences with the Manakau Stream as it drains to the coast (Figure 5 and Figure 8). The watercourses are relatively steep in gradient due to the narrow distance between the Tararua Range to the east and the Tasman Sea, in the order of ≈10 km. Tributaries feeding the Ohau River and Waikawa Stream are very numerous in the Tararua Range and provide significant inflow to the catchment (Figure 8). The Hokio and Waiwiri Streams serve as drainage to the Tasman Sea from lakes, Horowhenua and Papaitonga respectively. Lake Horowhenua has an area of 2.9 km^2, maximum depth is less than 2 m and is fed by groundwater and surface runoff (mostly in the form of drains) from the surrounding rural and urban area (Manawatu-Wanganui Regional Council, 1998). The number of lakes is high in the coastal sand zone and with the addition of swamps and a high water table exhibit a discharging zone (Bekesi, 2001). The area is widely believed to have significant surface-ground water interactions exhibited by the series of coastal lakes which are maintained by the discharging groundwater (Bekesi, 2001) and also by a drop in stream flow when water courses are flowing over the gravel plains (White, et al., 2010).

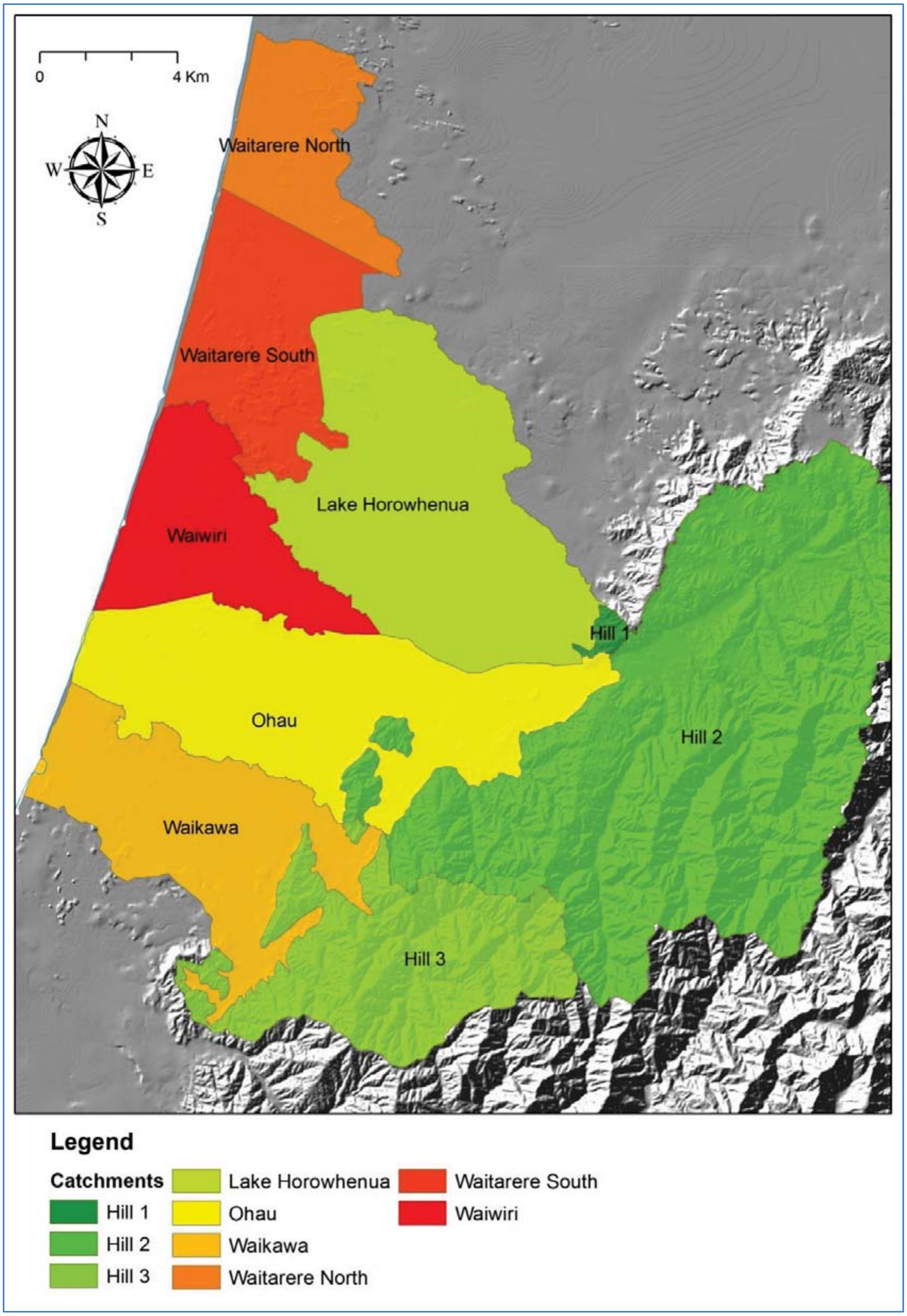

Figure 6 Horowhenua sub-catchments adapted from White et al. (2010) with the three additional 'Hill' catchments

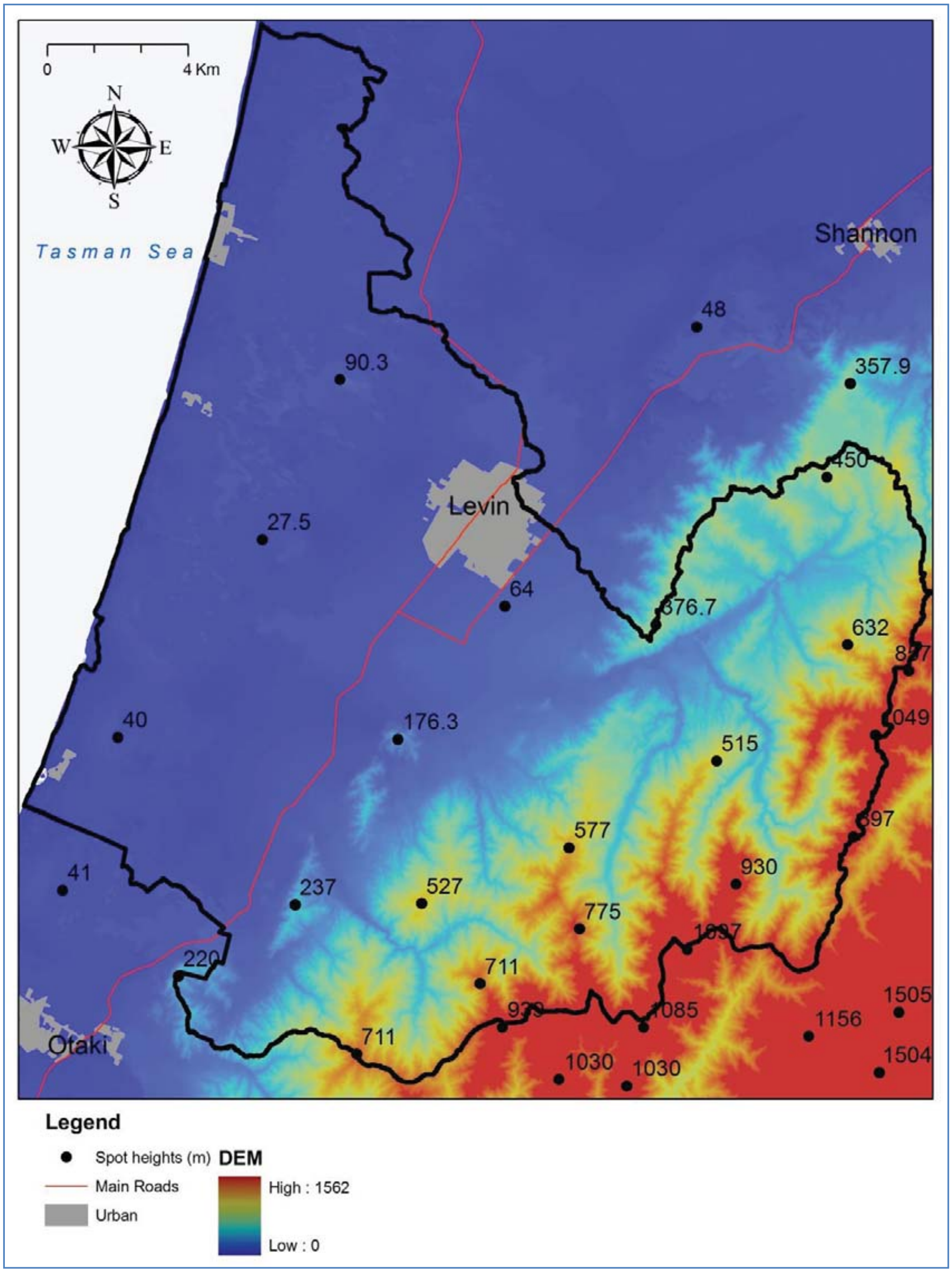

Figure 7 Digital elevation model and spot height values of Horowhenua (Data from Land Information New Zealand)

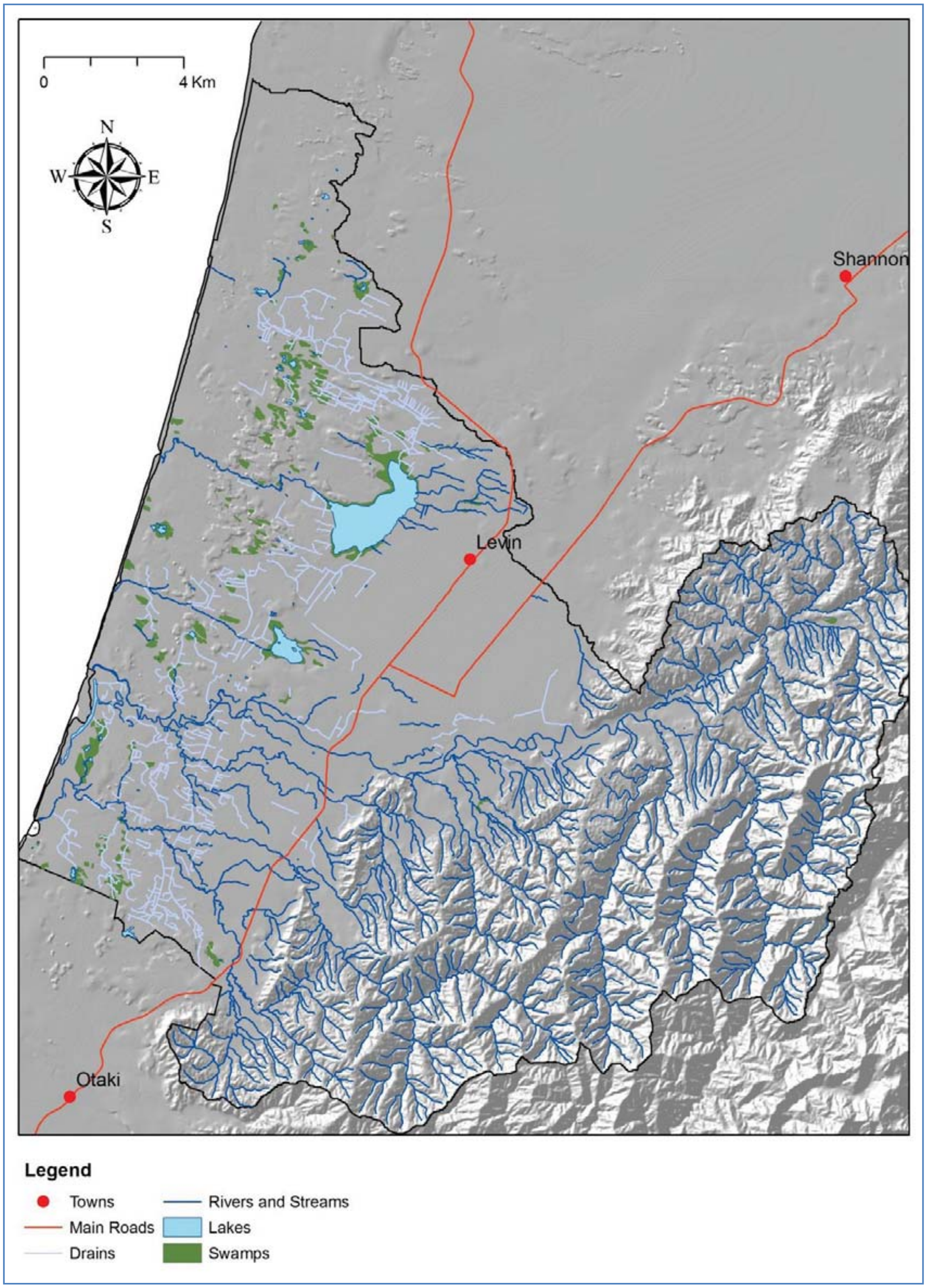

Figure 8 Hydrology features of Horowhenua showing extensive hydrological network in the upper catchments. Significant swamp type areas are shown in the coastal area west of the lakes as well as extensive drainage systems that have been constructed. An absence of surface water channels occurs in the immediate area to the east of the lakes. (Data from Land Information New Zealand)

Horowhenua has 4 main geological areas; greywacke-argillite hill and steep slopes; late Quaternary gravel plains, early Quaternary beach deposits and Holocene coastal sand. The Tararua Range is part of the Torlesse Supergroup dated to the Triassic-Jurassic period. The Rock type is predominately quartzofeldspathic greywacke with alternating greywacke-argillite sequences and poorly bedded greywacke with minor coloured argillite, conglomerate, basalt, chert and limestone. Groundwater storage capacity is limited; some joints and non-mineralized shear and fault plane may provide the rocks with some storage potential (Begg, et al., 2005). Eroded rock from the Tararua Range provides the source for younger gravels in the floodplains.

Extensive outcrops of Pleistocene aggradational gravel exist in the Horowhenua between the foothills of the Tararua Range to the line of coastal dune lakes which make up the alluvial plains (Zarour, 2008). The Last Glacial gravel deposits from Porewa (OIS 4) Rata (OIS 3) and Ohakea (OIS 2) are the result of aggradational deposits that have filled incised channels into the Tokomaru Marine Surface; the upper surface of the Otaki Formation (OIS 5) which consists of marine gravel and sand, commonly underlying loess and fan deposits (Figure 9 & Figure 10). The Q2al and Q3al gravels are poorly to moderately sorted with minor sand and silt underlying aggradational terraces (Begg & Johnston, 2000; Hughes, 2005; Zarour, 2008). Remnants of older more weathered gravels (Q6al and Q8al) are also found on the western flanks of the Tararua Range that correlate to Marton, Burnand and Aldworth surfaces (Figure 9). A NNE- to NE trending dextral fault known as the Northern Ohariu runs just inland and parallel to the western flank of the northern portion of the Tararua Range between Otaki and Palmerston (Palmer & Van Dissen, 2002).

Holocene gravels (Q1al) flank the current channels as well-sorted flood plain gravels. The Horowhenua lowlands east of the gravels are characterised by Holocene dune sands; inactive Aeolian dunes (Q1ds) and active dunes (Q1dm) (Figure 9). Areas surrounding lakes Horowhenua and Papaitonga are surrounded by over-bank sands, silts and clays (Q1as) due to swamp deposits composed of poorly consolidated silt, mud, peat and sand.

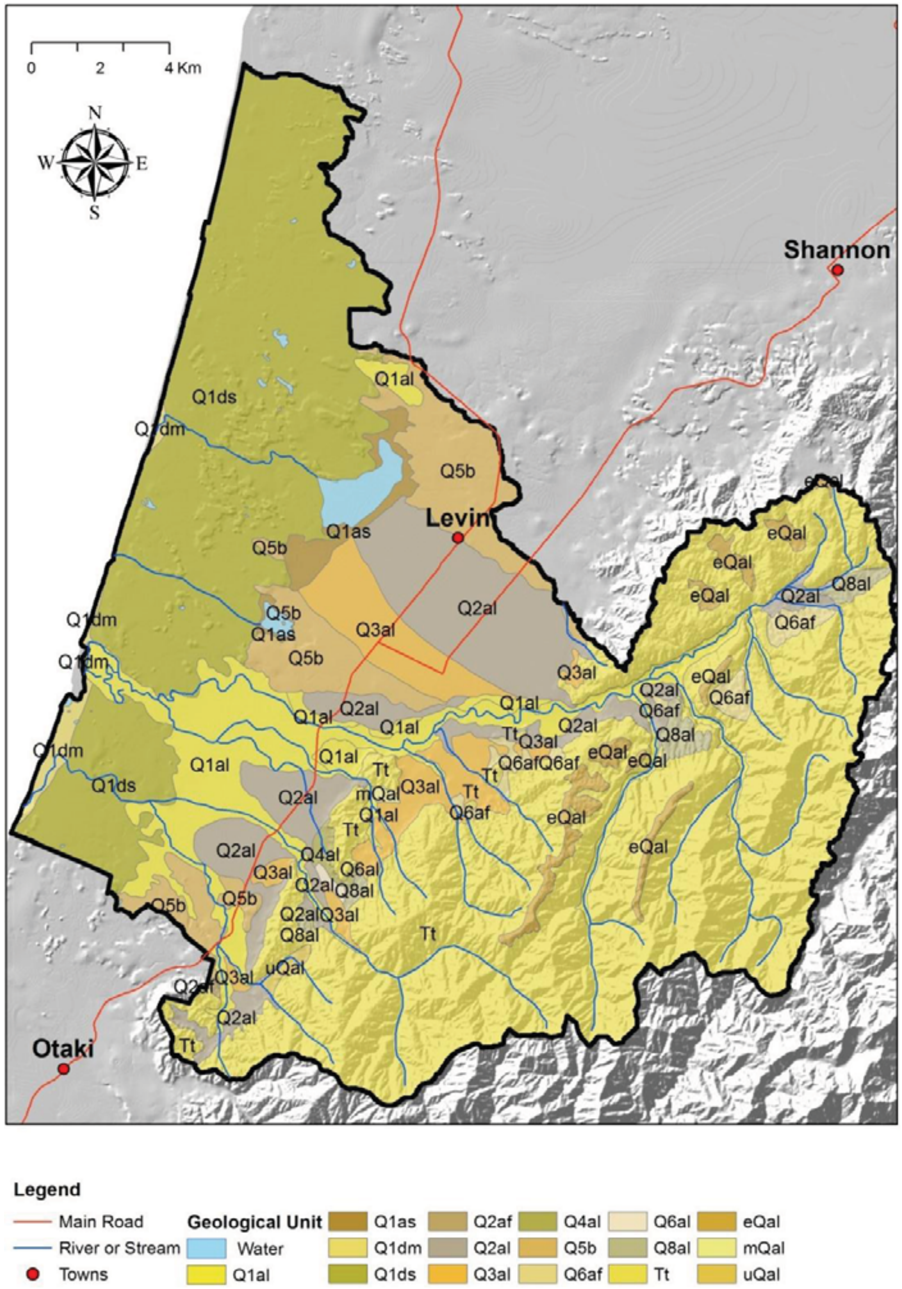

Figure 9 Geological units in the Horowhenua, adapted from (Begg & Johnston, 2000); **Q1as =** Holocene swamp deposits consisting of poorly consolidated silt, mud, peat and sand; **Q1dm =** Holocene active dunes; **Q1ds =** Holocene aeolian dunes; **Q2af & Q6af** = Quaternary poorly sorted steep fan deposits; **Q2al =** Quaternary poorly to moderately sorted gravel with minor sand or silt underlying aggradational terraces**; Q3al, Q4al, Q6al & Q8al =** Quaternary weathered, poorly to moderately sorted gravel underlying loess covered, commonly eroded aggradational surfaces; **Q5b =** Quaternary beach deposits consisting of marine gravel with sand commonly underlying loess and fan deposits; **Tt =** Rakaia terrane grey sandstone/mudstone sequences and poorly bedded sandstone; **eQal =** Quaternary undifferentiated weathered, poorly sorted loess-covered fan gravel, alluvial gravel and lacustrine silt deposits including Te Muna formation; **mQal =** Quaternary undifferentiated weathered, poorly sorted loess-covered fan gravel deposits including Ahiaruhe formation; **uQal =** Quaternary undifferentiated weathered, poorly sorted loess-covered alluvial gravel deposits

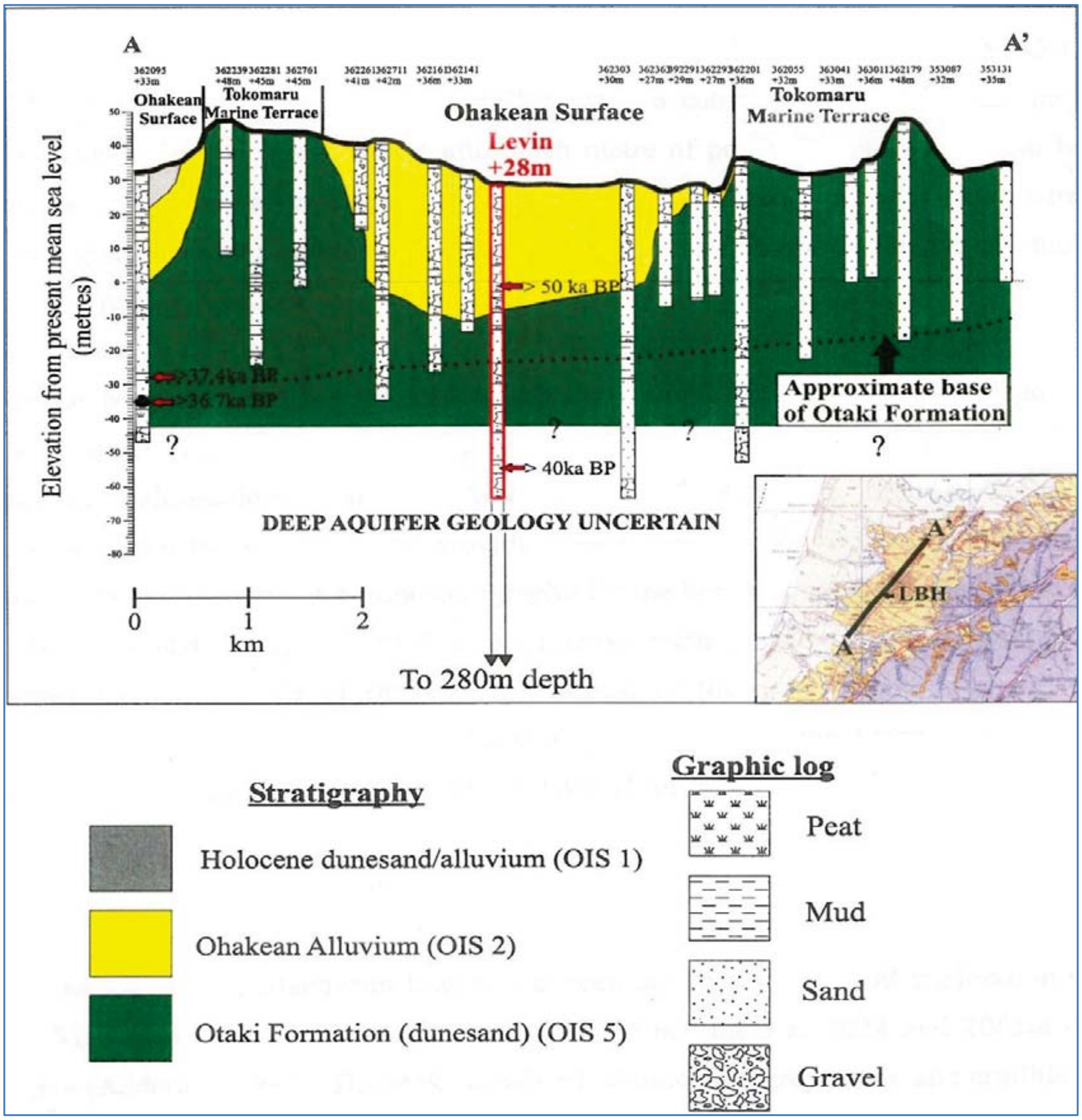

Figure 10 Cross section running northeast-southeast along the Horowhenua plain between Ohau and Heatherlea. Also shows the surface outcrops of the Last Interglacial marine strata (green), Last Glacial Maximum strata (yellow) and Holocene deposits (grey). From (Hughes, 2005)

Chapter 2 Methods

2.1 Introduction

The components of this research and water budget calculations encompass a wide range of data sources covering multiple subject disciplines. This results in a wide range of approaches that lead to complexity and estimates which are prone to large errors. Lerner et al. (1990) explains the main sources for error come from; 1). Incorrect conceptual model, an incorrect understanding of the recharge process is the most common error and also the most serious due to the understanding forming the foundation for methods used. The range of aspects to which wrong conceptual errors can arise are site specific and thus critical understanding and evaluation was needed key simplifications. 2). Neglecting spatial and temporal variability arises because recharge processes are generally non-linear in both temporal and spatial dimensions, thus variability needs to be accounted for as best possible, i.e. greater resolution. 3). Measurement error is a product of the equipment which is used to take the initial measurement and can often be estimated mathematically. 4). Calculation error arises from being careless during calculations. The study attempts to limit errors as much as possible with assumptions made done so with a clear conceptual model and understanding in mind. This section is broken into the methods for each key component of the water budget equation.

2.2 Precipitation

Rainfall data was collected from weather stations belonging to National Institute of Water and Atmospheric Research (NIWA) and Horizons Regional Council (Table 4). Distribution was spread much further than the Horowhenua area in order to include as much data for interpolation as possible and to show the regional gradient, particularly the east west pattern which is affected by the Tararua Range (Figure 11). Missing rainfall data was simply left blank. Missing data may have some implications on the recorded data immediately following the missing values which in some instances appear to be relatively large and possibly is an accumulated value of rainfall over the days that had missing values. Natural neighbour was used for interpolation which provides a simple method, guaranteeing values are within the sample range. It doesn't infer trends in the data, nor does it show features such ridges or valleys which is ideal for phenomena such as climate. The interpolation creates a rainfall map on a daily scale for 2007, 2008 and 2009.

Table 4 Weather station names and observing authority for collected rainfall data

Station Name	Observing Authority
Te Horo, Longcroft	NIWA 3308
Te Horo, Jonelle	NIWA 7387
Manakau	NIWA 3302
Levin AWS	NIWA 3275
Muhunoa East, Waima	NIWA 3282
Moutoa	NIWA 3269
Bainesse	NIWA 3253
Opiki	NIWA 3255
Putara	NIWA 2395
Paraparaumu	NIWA 12442
Waitatapia	NIWA 3207
Reikorangi	NIWA 3327
Waikanae Waterworks	NIWA 3307
Upper Mangahao, No.1 Dam	Horizons
Manawatu, Moutoa	Horizons
Mangaone, Milson Line	Horizons
Kahuterawa, Scotts Road	Horizons
Forest Rd Drain, Drop Structure	Horizons

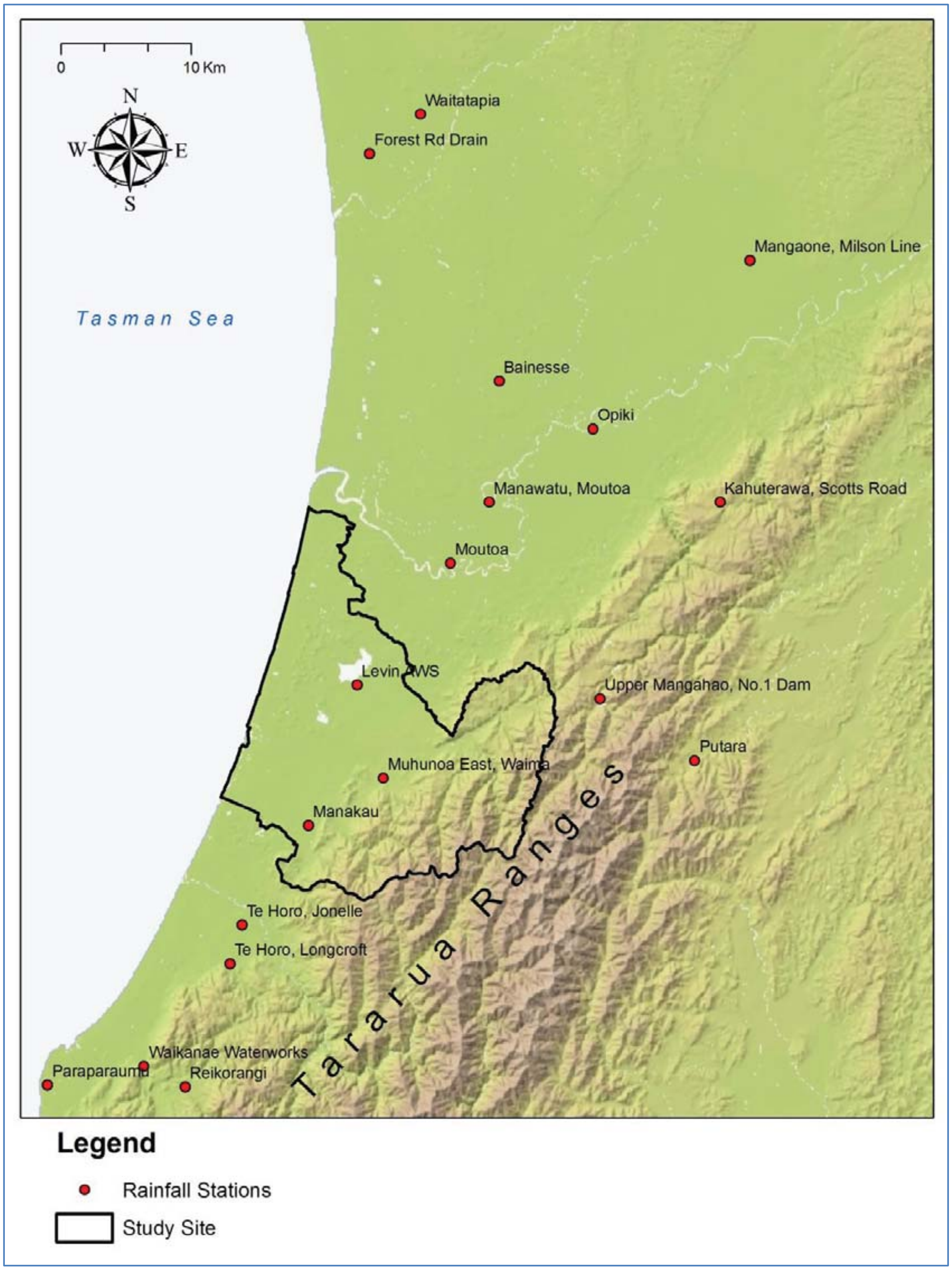

Figure 11 Locations of weather stations (Table 4) for collected rainfall data

2.3 Evapotranspiration

2.3.1 Introduction

Actual evapotranspiration under non-standard conditions($ET_{c\ adj}$) was estimated using the Food and Agricultural Organization (FAO) method of the Penman-Monteith equation (Allen, et al., 1998). The method (Equation 1) involves calculation of a reference evapotranspiration which is multiplied by crop transpiration, water stress and evaporation coefficients. Most of the following equations relating to evapotranspiration are adapted from Allen et al. (1998). The following sections are based around these four components of Equation 1.

Equation 1

$$ET_{c\,adj} = (K_s K_{cb} + K_e)ET_0$$

Where;

$ET_{c\ adj}$ = the actual evapotranspiration (mm day^{-1})

K_s = water stress coefficient

K_{cb} = basal crop coefficient for transpiration

K_e = soil evaporation

ET_0 = reference evapotranspiration (mm day^{-1})

$ET_{c\ adj}$ will be referred to simply as ET_c from hereon in.

2.3.2 Reference evapotranspiration (ET_0)

The reference evapotranspiration (ET_0) is based on an unambiguous definition of a reference surface "A hypothetical reference crop with an assumed crop height of 0.12 m, a fixed surface resistance of 70 s m^{-1} and an albedo of 0.23." which resembles an extensive surface of green well-watered grass of uniform height, actively growing and completely shading the ground (Allen, et al., 1998) p. 23). The ET_0 was calculated from meteorological data consisting of solar radiation, air temperature, air humidity and wind speed. Data was collected from NIWA’s CliFlo; the national climate database. Not

all stations have the relevant data recorded, in which case the nearest station which had the data was used (
Table 5 and Figure 12). Atmospheric pressure, relative humidity and wind speed were extracted at hourly timescales and then averaged to give daily averages. Radiation and minimum/maximum air temperatures were extracted at a daily timescale from the CliFlo database. Missing data was filled using an average between the two data values either side of the day or days that had missing data once the data was converted to a daily timescale.

The suitability for using east coast weather stations was considered carefully bearing in mind a strong east – west variability in climate. The main meteorological variable which was obtained from east coast weather stations was atmospheric pressure. Atmospheric pressure is a synoptic scale variable. This variable is relatively stable across regions and actually has little effect on the reference evapotranspiration values. Elevation of Tararua Range is factored in using a digital elevation model as the 'z' value in Equation 6. The argument for using the remaining spatially spread weather stations is simply that they were the closest weather stations with the required data and their perimeter extents were needed to encompass the study catchment. The natural neighbour interpolation method however limits the impact of these distant variables in applying weights based on 'area-stealing' effect which closest data points have on the interpolated values (Sibson, 1981). For this reason the Martinborough weather station, which occurs on the eastern side of Tararua range (Figure 11), is used for the other meteorological components. The components of the evapotranspiration equation also exhibit varying levels of influence on the evapotranspiration value. For instance, the solar radiation component has the largest influence (relative to temperature and humidity) on the reference evapotranspiration value. It also is the least variable component over a wider regional scale, thus geographically dispersed weather stations can still be deemed adequate for inclusion. In comparison, wind speed shows the greatest local variation where restricting data retrieval to west coast weather stations was most suitable. However, even wind speed, unless extreme, does not cause large variations in evapotranspiration values.

Table 5 Weather stations collected evapotranspiration calculation data, not all weather stations were used for all data

Station Name	Radiation, air temperature, relative humidity, wind speed	Atmospheric pressure
Levin Aws	Yes	Yes
Palmerston North Aws	Yes	Yes
Wanganui,Spriggens Park Ews	Yes	Yes
Paraparaumu Aero Aws	Yes	Yes
Wellington, Kelburn Aws	Yes	Yes
Martinborough Ews	Yes	No
Dannevirke Ews	No	No
Ohakea Aero	No	Yes
Castlepoint Aws	No	Yes

Equation 2

$$ET_0 = \frac{0.408\Delta\,(R_n - G) + \gamma\frac{900}{T+273}U_2(e_s - e_a)}{\Delta + \gamma(1 + 0.34U_2)}$$

Where;

ET_0 = reference evapotranspiration (mm day^{-1})

R_n = net radiation at crop surface (MJ m^{-2} day^{-1})

G = soil heat flux density (MJ m^{-2} day^{-1})

T = mean daily air temperature at 2 m height (°C)

U_2 = wind speed at 2 m height (m s^{-1})

e_s = saturation vapour pressure (kPa)

e_a = actual vapour pressure (kPa)

$e_s - e_a$ = saturation vapour pressure deficit (kPa)

Δ = slope vapour pressure curve (kPa $°C^{-1}$)

γ = psychometric constant (kPa $°C^{-1}$)

Radiation and wind speed was extracted direct from CliFlo and missing data dealt with according to the method prescribed above. The Soil heat flux value is small compared to radiation especially when the ground is covered by vegetation and time steps are 24hrs or longer. For Daily time periods FAO technique suggest that soil heat flux can be ignored and thus,

Equation 3

$$G_{day} \approx 0$$

The Psychometric constant is calculated using the following the equation

Equation 4

$$\gamma = \frac{C_p P}{\epsilon \lambda}$$

Where;

γ = psychometric constant (kPa °C^{-1})

P = atmospheric pressure (kPa)

λ = latent heat of vaporization =2.45 (MJ kg^{-1})

C_p = specific heat at constant pressure = 1.013 10^{-3} (MJ kg^{-1} °C^{-1})

ε = ratio molecular weight of water vapour/dry air = 0.622

Equation 4 simplifies to;

Equation 5

$$\gamma = 0.665 \times 10^{-3} P$$

Where;

γ = psychometric constant (kPa °C^{-1})

P = atmospheric pressure (kPa)

Atmospheric pressure (P) was calculated from Equation 6. In Equation 6 the value 101.3 is given as the standard atmospheric pressure and is scaled according to elevation. In this study 101.3 was substituted for the actual atmospheric pressure value which was reduced to atmospheric pressure at sea level for each weather station. The highest elevation for the weather stations was 207 m for the pressure data whereas elevation in Horowhenua reaches ≈ 1000 m on the flanks of the Tararua Range. Thus the elevation was factored in spatially via a digital terrain model layer in ArcGIS where elevation was substituted for the *'z'* value Equation 6. The resultant psychometric value was a map layer which gave a more accurate account for the pressure distribution over the area.

Equation 6

$$P = 101.3\left(\frac{293 - 0.0065z}{293}\right)^{5.26}$$

Where;

P = atmospheric pressure (kPa)

z = elevation above sea level (m)

Maximum and minimum air temperatures to calculate $e^0(T)$ according to Equation 7 which was used to calculate mean saturation vapour pressure (e_s) (Equation 8) and mean vapour pressure (e_a) (Equation 9).

Equation 7

$$e^0(T) = 0.6108exp\left[\frac{17.27T}{T + 237.3}\right]$$

Where;

e°(T) = saturation vapour pressure at the air temperature T (kPa),

T = air temperature (°C),

exp[..] = 2.7183 (base of natural logarithm) raised to the power [..]

Equation 8

$$e_s = \frac{e^0(T_{max}) + e^o(T_{\min})}{2}$$

Where;

e_s = mean saturation vapour pressure

$e°(T_{max})$ = saturation vapour pressure at maximum air temperature from Equation 7

$e°(T_{min})$ = saturation vapour pressure at minimum air temperature from Equation 7

Equation 9

$$e_a = \frac{e^0(T_{min})\frac{RH_{max}}{100} + e^0(T_{max})\frac{RH_{min}}{100}}{2}$$

Where;

e_a = actual vapour pressure (kPa)

$e°(T_{min})$= saturation vapour pressure at daily minimum temperature (kPa)

$e°(T_{max})$= saturation vapour pressure at daily maximum temperature (kPa)

RH_{max} = maximum relative humidity (%)

RH_{min} = minimum relative humidity (%)

Slope of saturation vapour pressure curve (Δ) was calculated using the mean daily temperatures derived from the daily minimum and maximum temperatures.

Equation 10

$$\Delta = \frac{4098\left[0.6108\exp\left(\frac{17.27T}{T+237.3}\right)\right]}{(T+237.3)^2}$$

Where;

Δ = slope of saturation vapour pressure curve at air temperature T (kPa $°C^{-1}$)

T = air temperature (°C)

exp[..] = 2.7183 (base of natural logarithm) raised to the power [..]

The ET_0 equation was simplified to Equation 11 with each variable representing a map layer of interpolated values. This was done due to the psychometric constant already being in map form. All of the variables except γ were interpolated using a 'natural neighbour' interpolation as described for the precipitation method, and produced a series of thematic map layers that were combined according to Equation 11 using map algebra in ArcGIS.

Equation 11

$$ET_0 = \frac{\alpha + \gamma\beta}{\Delta + \gamma\delta}$$

Where;

α = $0.408\Delta(R_n - G)$

β = $(900/(T+273)) * U_2 * (e_s - e_a)$

γ = Psychometric constant

δ = $(1+0.34U_2)$

Δ = slope vapour pressure curve

2.3.3 Crop Coefficient (K_{cb})

The crop coefficient (K_{cb}) (Equation 12) is the ratio of ET_c/ET_0 and typical values for K_{cb} were obtained from Allen et al. (1998) using the mid vegetation growth stages (Table 6). The mid growth stage K_{cb} value was used due to the noticeable dominance of perennial vegetation such as grass pasture in the study area. Additionally, using just the K_{cb} value simplifies the calculation that would otherwise involve considerable extra time, information and calculations pertaining to individual growing schedules and crop guides. The K_{cb} value is adjusted for environments which deviate from a sub humid climate where RH_{min} differs from 45 % and wind speed differs from 2 m s^{-1} each day. Wind speed and relative humidity data was taken from NIWA Cliflo climate database.

Mean plant height values are taken from Allen et al. (1998) where possible or by approximation based on the dominate vegetation and plant growth heights (Table 6).

The approximations made for the land cover in the form of crop coefficients is a potential source for significant error. The crop coefficients from Allen et al. (1998) are for agriculturally managed crops, subject to human influence in growth patterns and pristine conditions. The appropriateness of assigning these crop values to naturally grown vegetation is not the most desirable technique. This is mainly an issue with the indigenous bush and forest areas which represent a more complicated structure than that of a plantation forest from which their crop coefficient values originate. Additionally, coefficients are based on a list of a select few crops, which is by no means exhaustive. Many vegetation types not currently provided with an indexed coefficient value and hence values are approximated for indigenous vegetation. There is provision by Allen et al. (1998) for pristine vegetation environments to be adjusted for non-pristine but this was unable to be investigated due to project constraints. Overall, however the crop coefficients themselves are similar in value for most vegetation types in Horowhenua (within 0.1).

Equation 12

$$K_{cb} = K_{cb\,(Tab)} + [0.04(U_2 - 2) - 0.004(RH_{min} - 45)]\left(\frac{h}{3}\right)^{0.3}$$

Where;

$K_{cb\,(Tab)}$ = value for K_{cb} mid vegetation growth stage taken from Allen et al. (1998) (Table 6)

U_2 = mean value for daily wind speed at 2 m height (m s^{-1})

RH_{min} = mean value for daily minimum relative humidity (%)

h = mean plant height (m) (Table 6).

Table 6 Land Cover Database (LCDB) 2 names showing assigned $K_{cb\ (tab)}$, mean plant height and root depth values

Land Cover Database (LCDB) 2 Name	Simplified vegetation type	$K_{cb\ (tab)}$	Mean plant Height (m)	Assigned root depth(m)
Afforestation (imaged, post LCDB 1)	Forest	0.95	2	1
Afforestation (not imaged)	Forest	0.95	2	1
Broadleaved Indigenous Hardwoods	Scrub	0.9	10	1
Built-up Area	Urban	NA	0	NA
Coastal Sand and Gravel	Exposed	NA	0	NA
Deciduous Hardwoods	Forest	0.95	20	1
Estuarine Open Water	Water	NA	0	NA
Fernland	Scrub	0.95	3	1
Flaxland	Sedge/marsh	1.05	1.2	0.5
Forest Harvested	Forest	0.95	2	1
Gorse and Broom	Scrub	0.95	2	1
Herbaceous Freshwater Vegetation	Sedge/marsh	1.15	1.5	0.5
High Producing Exotic Grassland	Grassland	0.9	0.225	0.5
Indigenous Forest	Forest	0.95	30	1
Lake and Pond	Water	NA	0	NA
Landslide	Exposed	NA	0	NA
Low Producing Grassland	Grassland	0.9	0.225	0.5
Major Shelterbelts	Forest	0.95	20	1
Manuka and or Kanuka	Scrub	0.95	6	1
Mixed Exotic Shrubland	Scrub	0.95	5	1
Orchard and Other Perennial Crops	Horticulture	0.83	3.5	1
Other Exotic Forest	Forest	0.95	30	1
Pine Forest - Closed Canopy	Forest	0.95	50	1
Pine Forest - Open Canopy	Forest	0.95	20	1
River	Water	NA	0	NA
River and Lakeshore Gravel and Rock	Exposed	NA	0	NA
Short-rotation Cropland	Horticultural	0.95	0.4	0.5
Sub Alpine Shrubland	Scrub	0.95	4	0.5
Surface Mine	Exposed	NA	0	NA
Transport Infrastructure	Urban	NA	0	NA
Urban Parkland/ Open Space	Turf	0.9	0.1	0.5

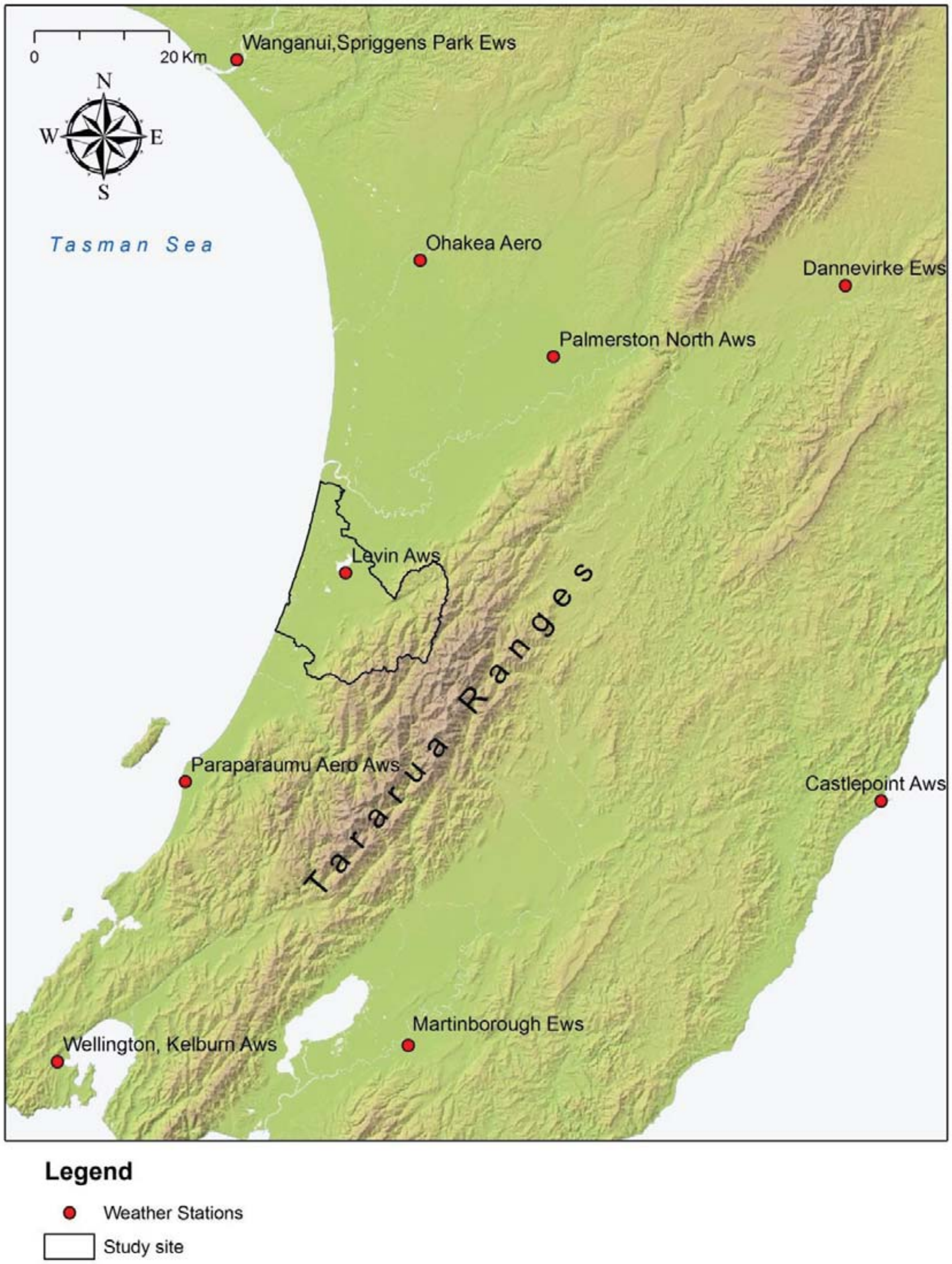

Figure 12 Location of weather stations used for collect radiation, air temperature, wind speed, relative and humidity air pressure evapotranspiration calculations

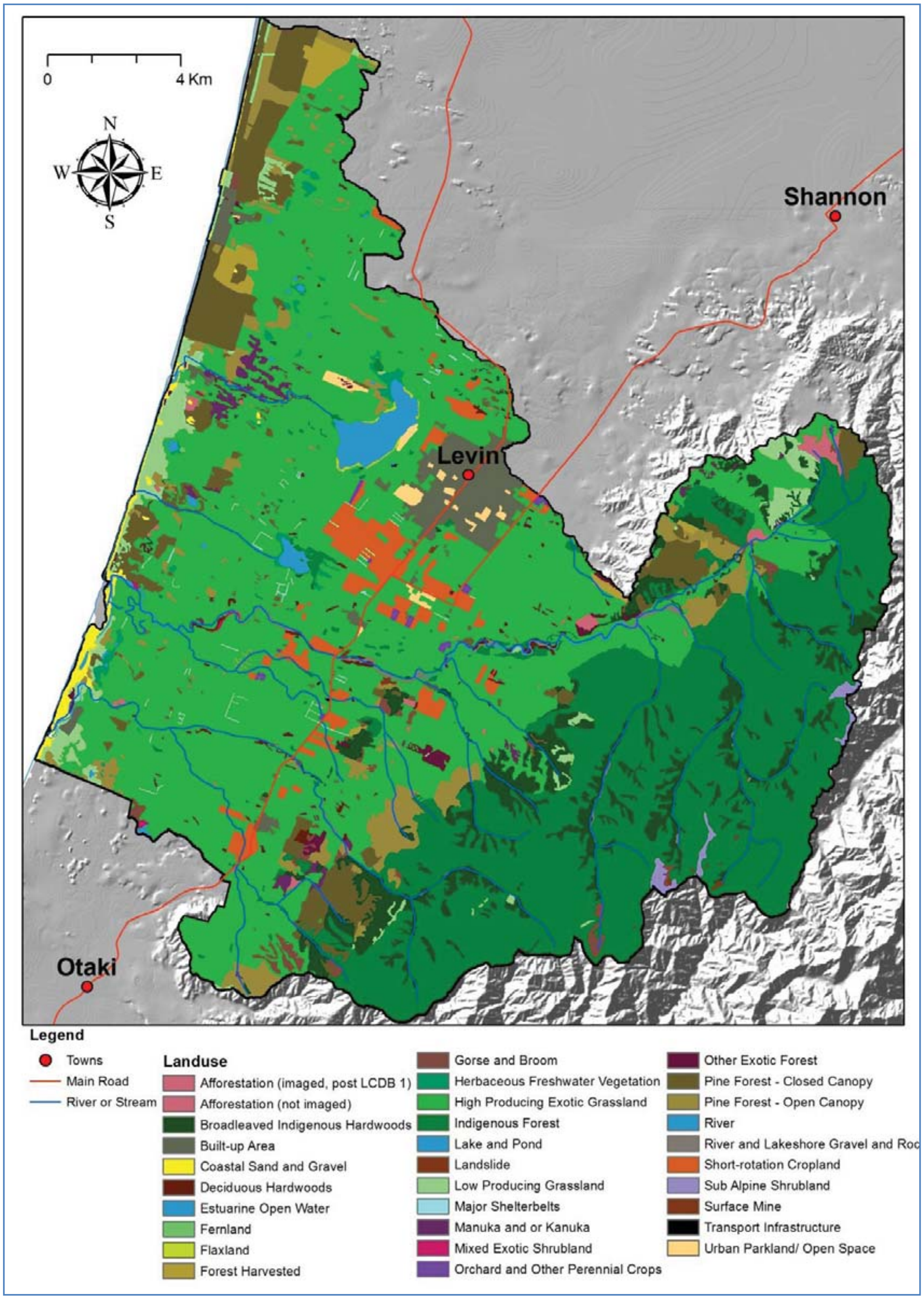

Figure 13 Landuse map for Horowhenua (50 m resolution) derived from New Zealand Land Cover Database 2 (LCDB2) administered by Ministry for the Environment (MfE) and described by Thompson (2004), and Thompson, Grüner & Gapare (2003).

2.3.4 Water Stress Coefficient K_s

Water Stress is given by a water stress coefficient K_s, and is multiplied by the crop coefficient to account for less than full potential evapotranspiration (Equation 13). The water stress is dependent on soil water availability which refers to the ability for a soil to retain water for use by vegetation. The water stress point is dependent on the inherent soil characteristics and time from when the last wetting event occurred (Figure 14).

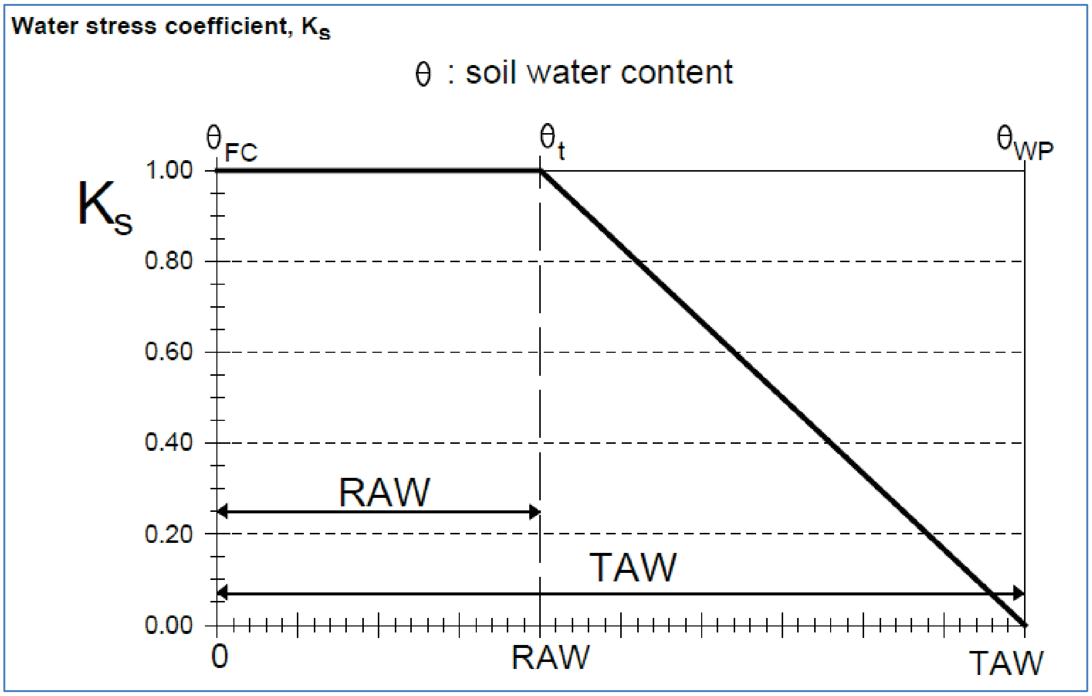

Figure 14 Change in soil water content over time after Allen *et al*. (1998). When soil is at field capacity (Θ_{FC}) the water content is not limiting evapotranspiration and thus K_s is equal to 1. Providing no wetting event has occurred, the water content of the soil will decrease until it hits a threshold (Θ_t) or stress point at which the rate of transpiration begins to reduce as water content approaches wilting point (Θ_{wp}).

Equation 13

$$K_s = \frac{TAW - D_r}{TAW - RAW} \; when \; D_r > RAW$$

$$K_s = 1 \; when \; D_r < RAW$$

Where;

K_s = dimensionless transpiration reduction factor dependent on available soil water (0 – 1)

D_r = root zone depletion (mm)

TAW = total available soil water in the root zone (mm)

RAW = the readily available soil water in the root zone (mm)

Soil moisture (RAW and TAW)

Soil moisture content, TAW and RAW (Equation 14, Equation 15), was derived from previous field measurements using fabric-related analysis by Palmer & Wilde (unpublished) for soils in Horowhenua. Water content for field capacity, stress point and wilting point was taken at tensions of 10 kPa, 100 kPa and 1500 kPa respectively. The soil moisture values were matched with the soil classification data from the Land Resource Information (LRI) GIS dataset provided by Landcare Research (Figure 15 and Table 7). However, the soil samples are not a direct match with the LRI classification data so the values used for RAW and TAW calculation were assigned based on soil characteristics and local expert knowledge (Palmer, Personal Comm.).

Table 7 The pairing of the soil sample from Palmer & Wilde (unpublished) with the 'gensoi' classification assigned in Land Resource Inventory (LRI) classification

Soil sample	LRI Soil Classification
Manawatu Silt and Fine Sandy Loam	1
Rangitikei Fine Sandy Loam	1c
Kairanga Silt Loam	2,2b
Paruhau Silt Loam	12, 35b
Motuiti Sand 45% Puke Puke Peaty Loam (45%) Omanuka Peat (10%)	23, 23b, brock
Koputaroa fine sandy loam	24
Ashhurst Stony Silt Loam/Takapau	75, 76b,77, 122,123,124, 46H, Msoil
Levin Silt Loam	76, 76a, 78b
Kopua Silt Loam	78
Makerua Peat	107
Omanuka Peat	107j, 108b
Dealt with separately	Lake
	Town

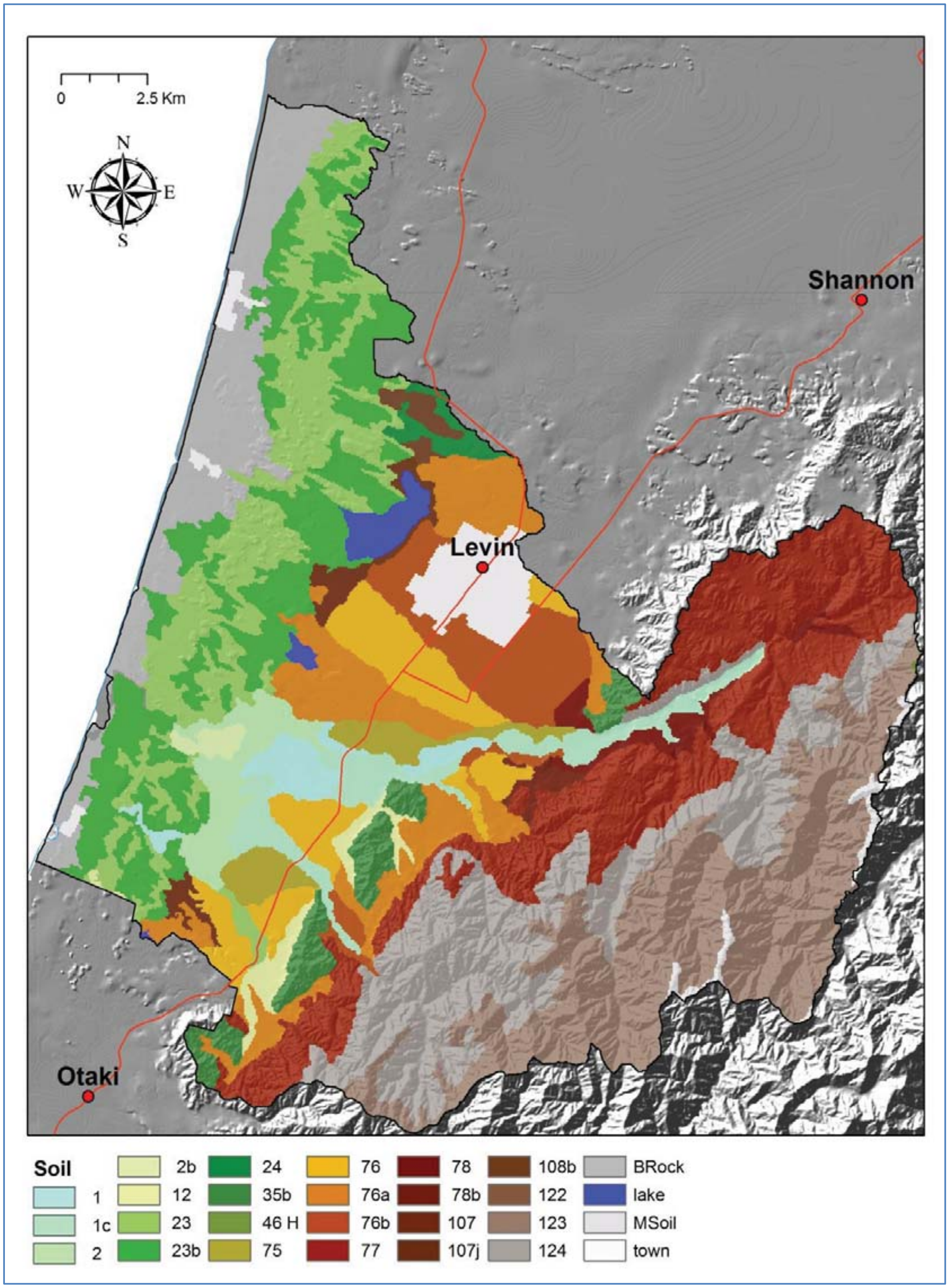

Figure 15 Soil map of Horowhenua derived from Land Resource Inventory (LRI) data. Soil types are clarified in Table 7

Soil values in the coastal area are complicated by the non-uniformity of the soil characteristics expressed by the presence of associations (Cowie, Fitzgerald, & Owers, 1967). In this instance the values used were a combination of Motuiti Sand (45 %), Puke Puke Peaty Loam (45 %) and Omanuka Peat (10 %) (Palmer, Personal Comm.). LRI data prescribed as 'town' was reclassified to the surrounding soil type which depended on the town. Lake data was assigned a value of zero. Two horizon depths were used to express shallow and deep rooting crops; 0.5 m was chosen as a depth for the shallow rooting crops, particularly to account for the grasslands; and, 1.0 m was chosen as the rooting depth for deeper rooting crops such as forest (Table 6). The final TAW and RAW values are displayed in Table 8 and Figure 16.

Equation 14

$$TAW = (\theta_{FC} - \theta_{WP})Z_r$$

Where;

TAW = the total available soil water in the root zone (mm)

θ_{FC} = the water content at field capacity (% v/v)

θ_{WP} = the water content at wilting point (% v/v)

Z_r = the rooting depth (mm)

Equation 15

$$RAW = (\theta_{FC} - \theta_{SP})Z_r$$

Where;

RAW = the readily available soil water in the root zone (mm)

θ_{FC} = the water content at field capacity (% v/v)

θ_{SP} = the water content at stress point (% v/v)

Z_r = the rooting depth (mm)

Table 8 Calculated Readily Available Water (RAW) and Total Available Water (TAW) and the prescribed Land Resource Inventory (LRI) soil classes for a depth of 0.5 m and 1 m.

LRI SOIL Classes	Soil Moisture (mm)			
	RAW		TAW	
	0.5 m	1 m	0.5 m	1 m
1	40	75	95	157
1c	42	90	107	217
2,2b	34	63	136	265
12, 35b	26	44	73	130
23, 23b, brock	38	77	81	161
24	34	54	62	115
75, 76b,77, 122,123,124, 46H, Msoil	23	45	91	182
76, 76a, 78b	43	68	103	193
78	31	62	85	169
107	115	216	272	558
107j, 108b	54	120	161	389
Lake	**0**	**0**	**0**	**0**
Town	**0**	**0**	**0**	**0**

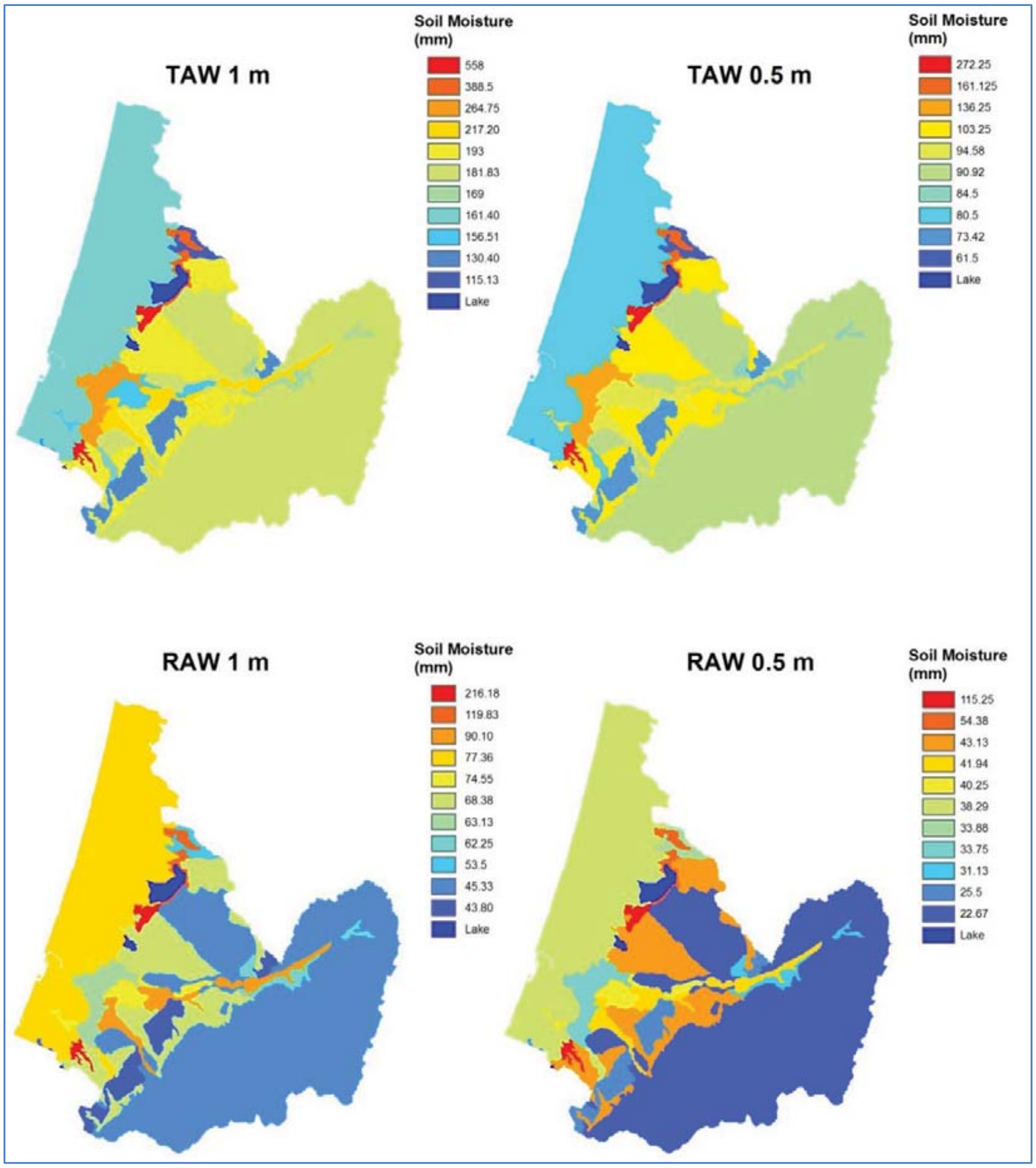

Figure 16 Spatial pattern of Readily Available Water (RAW) and Total Available Water (TAW) soil moisture content for 0.5 m and 1 m root depth in Horowhenua

Root zone depletion (D_r)

The depletion factor is required to calculate the water stress coefficient and requires the calculation of a soil moisture balance using Equation 16. The calculation was run at a daily time scale.

Equation 16

$$D_{r,i} = D_{r,i-1} - (P - RO)_i - I_i - CR_i + ET_{c,i} + DP_i$$

Where;

$D_{r,I}$ = root zone depletion at the end of day i (mm)

$D_{r,i-1}$ = water content in the root zone at the end of the previous day, i-1 (mm)

P_i = precipitation on day i (mm) (section 2.2)

RO_i = runoff from the soil surface on day i (mm)

I_i = net irrigation depth on day i that infiltrates the soil (mm)

CR_i = capillary rise from the groundwater table on day i (mm)

$ET_{c,i}$ = crop evapotranspiration on day i (mm)

DP_i = water loss out of the root zone by deep percolation on day i (mm)

Capillary rise (CR_i) is generally considered to be zero when water table is more than 1 m below the surface, and will be ignored for simplification purposes (Allen, et al., 1998). Irrigation values are also ignored and precipitation is calculated from the section 2.2. ET_c is determined from Equation 1 using K_s value corresponding to the depletion value at the start of the day. Initial depletion was estimated based on deficit data from the surrounding weather stations at spot sites for the month prior to January 2007, taken from the Clifo dataset from NIWA. Several of the sites show values ≈ 10 mm on the 31st December 2006 (Figure 17).

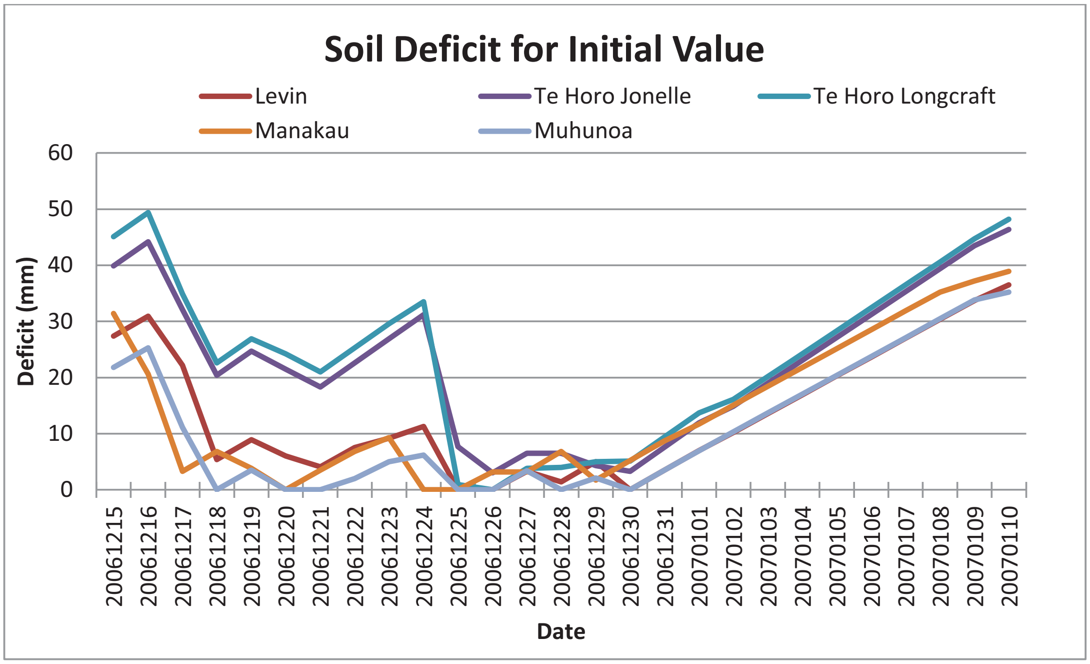

Figure 17 Soil deficit values from 5 weather stations (Figure 11 & Figure 12) during the time period of 15th December 2006 to 10th January 2007. The soil deficit approaches zero a few days prior to start of 2007 and reaches approximately 10 mm on the 31st December 2006.

Deep percolation occurs when field capacity is exceeded and depletion (D_r) is equal to zero. When that condition is reached, the following equation can be used to calculate deep percolation, at all other times deep percolation is equal to zero.

Equation 17

$$DP_i = (P_i - RO_i) + I_i - ET_{c,i} - D_{r,i-1} \qquad \geq 0$$

Where;

DP_i = water loss out of the root zone by deep percolation on day i (mm)

P_i = precipitation on day i (mm)

RO_i = runoff from the soil surface on day i (mm)

I_i = net irrigation depth on day i that infiltrates the soil (mm)

$ET_{c,i}$ = crop evapotranspiration on day i (mm)

$D_{r,i-1}$ = water content in the root zone at the end of the previous day, i-1 (mm)

Runoff

Runoff (RO) was calculated using the Soil Conservation Service (SCS) method (e.g. Somashekar, Ravikumar, Sowmya, Dar, & Ravikumar, 2011) utilizing runoff Curve Numbers (CN) to determine runoff that factors in soil characteristics and land cover (Equation 18 and Equation 19). A runoff curve number is an empirical hydrological parameter used for predicting approximate runoff that considers landuse and soil conditions. Soil type is categorized (Table 9) into four hydrological soil groups (Loucks, van Beek, Stedinger, Dijkman, & Villars, 2005; Zhan & Huang, 2004). The land cover was assigned appropriate land use categories from Zhan & Huang (2004) which were then intersected with the hydrological soil groups to achieve unique landcover–soil group dependent curve numbers (Table 10). A high curve number means high runoff whereas as a low curve number means low runoff.

Equation 18

$$RO = \frac{(rainfall - 0.2S)^2}{rainfall + 0.8S}$$

Where;

RO = Runoff (mm)

S = potential maximum soil retention

Equation 19

$$S = \left(\frac{25400}{CN}\right) - 254$$

Where;

S = potential maximum soil retention

CN = Curve Number

Table 9 Definition for the Soil Conservation Method (SCS) hydrological soil groups from Loucks et al. (2005) and the assigned soils to the hydrological soil group.

Soil Type	Definition	Soil Group assigned
A **(Low runoff potential)**	High infiltration rates even when thoroughly wetted. Chiefly deep, well toe excessively drained sands or gravels. High rate of water transmission.	
B	Moderate infiltration rates when thoroughly wetted. Chiefly moderately deep to deep, moderately-well to well drained soils with moderately-fine to moderately-coarse textures. Moderate rate of water transmission.	1, 1c, 24, 46 H, 75, 76b, 77, 122, 123, 124, Msoil
C	Slow infiltration rates when thoroughly wetted. Chiefly solids with layer that impedes downward movement of water, or soils with a moderately-fine to fine texture. Slow rate of water transmission.	2, 2b, 12, 23, 23b, 35b, 76, 76a, 78, 78b, Brock
D **(high runoff potential)**	Very slow infiltration and transmission rates when thoroughly wetted. Chiefly soils that are; clay soils with higher welling potential, have a permanent high water table, have clay pans or clay layer near the surface, or over nearly impervious material	107, 107j, 108b, lake, town

Table 10 Curve Numbers (CN) assigned to landcover and intersected hydrological soil groups

Landcover database name	Hydrological Soil Group	Curve number (CN)
Afforestation	B	56
	C	70
Broadleaved Indigenous Hardwoods	B	66
	C	77
	D	85
Built-up Area	B	72
	C	81
	D	86
Coastal Sand and Gravel	C	50
	D	50
Deciduous Hardwoods	B	60
	C	73
	D	79
Estuarine Open Water	C	0
Fernland	C	70
Flaxland	B	56
	C	70
	D	77
Forest Harvested	B	56
	C	70
	D	77
Gorse and Broom	B	56
	C	70
	D	77
Herbaceous Freshwater Vegetation, Rivers and lakes	B	0
	C	0
	D	0
High/Low producing Exotic Grassland	B	61
	C	74
	D	80
Indigenous Forest	B	66
	C	77
	D	85
Landslide	B	56
Major Shelterbelts	B	66
	C	77
	D	85
Manuka and or Kanuka	B	56
	C	70
	D	77
Mixed Exotic Shrubland	C	70
	D	77
Orchard and Other Perennial Crops	B	66
	C	77
Other Exotic Forest	B	63
	C	75
	D	82
Pine Forest – closed/open Canopy	B	66
	C	77
	D	85
River and Lakeshore Gravel and Rock	B	86
	C	91
Short-rotation Cropland	B	78
	C	85
	D	89
Sub Alpine Shrubland	B	69
Surface Mine	B	86
	C	91
Transport Infrastructure	B	89
	C	92
Urban Parkland/ Open Space	B	61
	C	74
	D	80

2.3.5 Evaporation Coefficient (K_e)

Evaporation occurs at the maximum rate when the soil is wet and is calculated as a separate component to transpiration. The equation is based on a minimum of two expressions so as to never exceed the K_{cmax} .

Equation 20

$$K_e = K_r(K_{c\,max} - K_{cb}) \ \le f_{ew}K_{c\,max}$$

Where;

K_e = soil evaporation coefficient

K_{cb} = basal crop coefficient

$K_{c\,max}$ = maximum value of K_c following rain

K_r = dimensionless evaporation reduction coefficient dependent on the cumulative depth of water depleted from the topsoil

f_{ew} = fraction of the soil that is both exposed and wetted

The basal K_{cb} value is taken from the K_{cb} value calculated from Equation 12. The maximum K_c value ($K_{c\,max}$) is calculated from the equation below using the same relative humidity and windspeed data that is used in Equation 12. K_e values for water covered areas was assigned a value of 1.10 following values from (Allen, et al., 1998).

Equation 21

$$K_{c\,max} = max\left\{\left\{1.2 + [0.04(u_2 - 2) - 0.004(RH_{min} - 45)]\left(\frac{h}{3}\right)^{0.3}\right\}, \{K_{cb} + 0.05\}\right\}$$

Where;

h = mean maximum plant height during the period of calculation

K_{cb} = basal crop coefficient

RH_{min} = the mean value for daily minimum relative humidity during the (%)

The exposed soil fraction (f_{ew}) values were considered to be 0.1 for all landuse types except the horticultural areas which were considered to be 0.2, and lakes where a value of 1.0 was given. These values were assigned somewhat arbitrarily with minimal effect following information from Allen et al. (1998). The soil reduction coefficient (K_r) (Equation 22) factors in reduction of available water over time between wetting events (Figure 18).

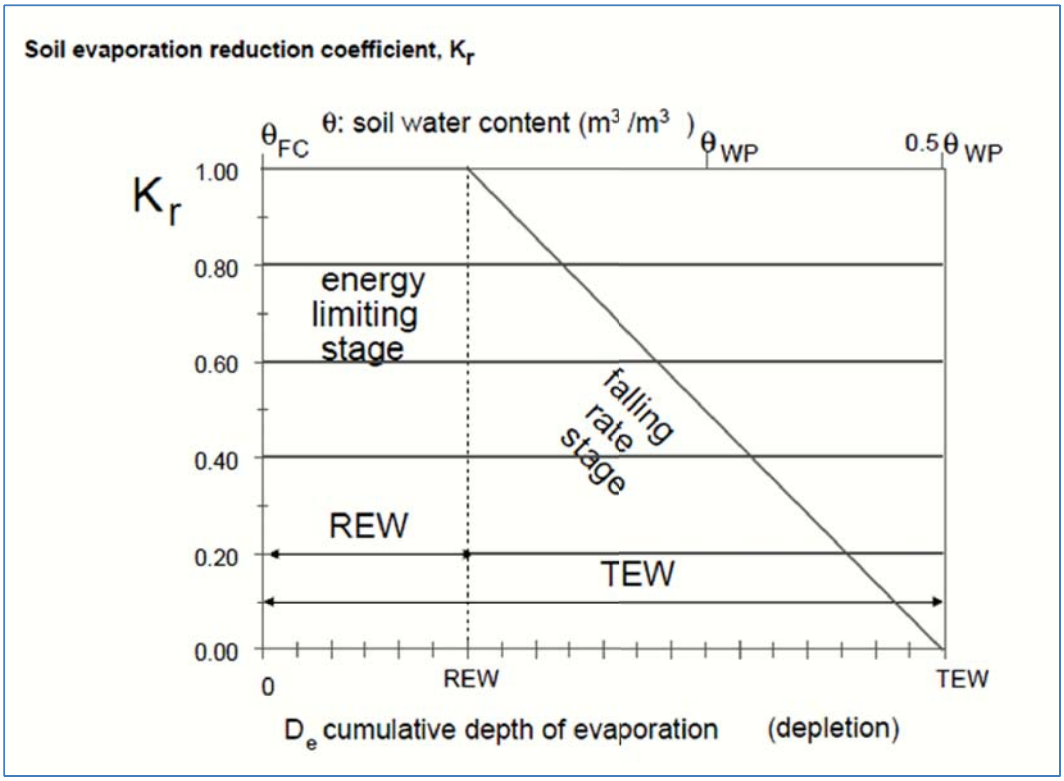

Figure 18 Change in soil water content over time after Allen et al. (1998). When soil is at field capacity (Θ_{FC}) the water content is not limiting evapotranspiration and thus K_r is equal to 1. Providing no wetting event has occurred, the water content of the soil will decrease and between REW and TEW evaporable water will be reduced.

Equation 22

$$K_r = \frac{TEW - D_{e,i-1}}{TEW - REW} \qquad D_{e,i-1} > REW$$

Where;

K_r = dimensionless evaporation reduction coefficient dependent on the soil water depletion from the topsoil layer (K_r = 1 when $D_{e,i-1} \leq$ REW)

$D_{e,i-1}$ = cumulative depth of evaporation from the soil surface layer at the end of day i-1 (mm)

TEW = maximum cumulative depth of evaporation from the soil surface layer when K_r = 0 (mm)

REW = cumulative depth of evaporation (mm) at the end of stage 1

The $D_{e,\,i\text{-}1}$ was calculated using a soil moisture budget shown in Equation 23. The Total Evaporable Water (TEW) and Readily Evaporable Water (REW) was calculated from the same information as RAW and TAW using a shallower depth of 150 mm using Equation 24 from Allen et al. (1998).

Equation 23

$$D_{e,i} = D_{e,i-1} \; - \; (P_i - RO_i) + \frac{E_i}{f_{ew}} + T_{ew,i} + DP_{e,i}$$

Where;

$D_{e,i-1}$ = cumulative depth of evaporation following complete wetting from the exposed and wetted fraction of the topsoil at the end of day i-1 (mm)

$D_{e,i}$ = cumulative depth of evaporation following complete wetting at the end of day i (mm)

P_i = precipitation on day i (mm)

RO_i = precipitation runoff from the soil surface on day i (mm)

E_i = evaporation on day i (i.e., $E_i = K_e \, ET_o$) (mm)

$T_{ew,i}$ = depth of transpiration from the exposed and wetted fraction of the soil surface layer on day i (mm)

$DP_{e,\,i}$ = deep percolation loss from the topsoil layer on day i if soil water content exceeds field capacity (m)

f_{ew} = exposed and wetted soil fraction (0.01 – 1)

Equation 24

$$TEW = (\theta_{FC} - 0.5\theta_{WP})Z_e$$

Where;

TEW = maximum depth of water that can be evaporated from the soil when the topsoil has been initially completely wetted (mm)

θ_{FC} = soil water content at field capacity (m^3 m^{-3})

θ_{WP} = soil water content at wilting point (m^3 m^{-3})

Z_e = depth of the surface soil layer that is subject to drying by way of evaporation (0.10 - 0.15 m)

Equation 25

$$REW = (\theta_{FC} - \theta_{WP})Z_e$$

Where;

REW = cumulative depth of evaporation (depletion) at which point evaporation is limited by water content (mm)

θ_{FC} = soil water content at field capacity (m^3 m^{-3})

θ_{WP} = soil water content at wilting point (m^3 m^{-3})

Z_e = depth of the surface soil layer that is subject to drying by way of evaporation (0.10 - 0.15 m)

2.4 Urban precipitation and evapotranspiration

One important approximation which applies to precipitation and evapotranspiration is an approximation of precipitation and evapotranspiration values for urban areas. Urban areas exert influences on the water pathway that occur as well as influencing the land area available for evapotranspiration. Rainfall that falls in urban areas either falls on open ground or falls on impermeable surfaces such as roofs or pavement. If the rainfall lands on an impermeable surface the water flows into waste water outflows, and out to sea, which is regarded as a loss to the system. Evapotranspiration occurs on vegetated areas such as lawns, parks and gardens and is assumed to be negligible on impermeable surfaces. An arbitrary coefficient of 0.5 has been chosen as a fraction of permeable land areas. This approximation is considered as a rough value that needs refinement which is outside the restraints on this research. The application of this variable is predominantly for spatial interpretation and probably not necessary for quantitative calculations. However, the effects of this variable on total quantitative values are limited due to total urban area being several orders of magnitude lower than total non-urban area in Horowhenua. This is applied to the K_c value for evapotranspiration and also applied to the precipitation during calculation.

2.5 Groundwater Level and Outflow

Darcy's law is a fundamental equation for groundwater flow through porous mediums and is used in this study to estimate the groundwater outflow to the sea through the coastal boundary.

Equation 26

$$G_{out} = KIA$$

Where;

G_{out} = volumetric groundwater flow ($m^3\ s^{-1}$)

K = hydraulic conductivity ($m\ s^{-1}$)

I = hydraulic gradient (unitless)

A = Cross sectional area (m^2)

The value for hydraulic conductivity (K) is obtained through standard values for fine sands (Domenico & Schwartz, 1990) at a value of 2×10^{-4} $m\ s^{-1}$ according to White et al. (2010). Cross sectional area was established by measuring the coastline for each coastal sub-catchment and multiplied by a depth for aquifer thickness of 40m; an upper maximum as mentioned by White et al. (2010). The hydraulic gradient (I) is estimated from a potentiometric map that is created from well data obtained from Horizons Regional Council. The well data was for 22 well sites (3 from outside the study site to the north) which had almost complete monthly records for the three year study period (Figure 20). Water levels for lakes; Horowhenua, Papaitonga and Kopureherehere were also used in the potentiometric surface construction though they remain static. Lake Horowhenua was assigned average water levels based on data from Williams (2002). The average lake level values are taken from September 1980 to May 1981. The values are also attributed to Lake Papaitonga, assuming that they have similar behaviour and levels. Lake Kopureherehere was assigned a value of 18 m. The potentiometric map was interpolated using natural neighbour interpolation and with the coastline acting as a zero elevation barrier. From the potentiometric map, hydraulic gradients were calculated over a 1 km buffer of the coastline and then zonally averaged for their respective groundwater catchment.

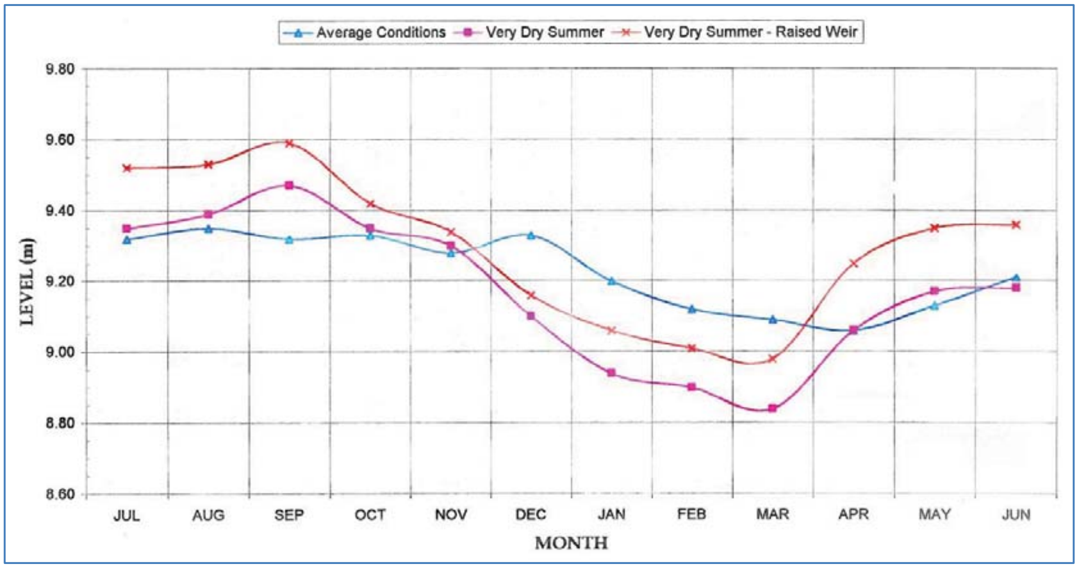

Figure 19 Lake Horowhenua monthly lake levels for September 1980 to May 1981 from (Williams, 2002)

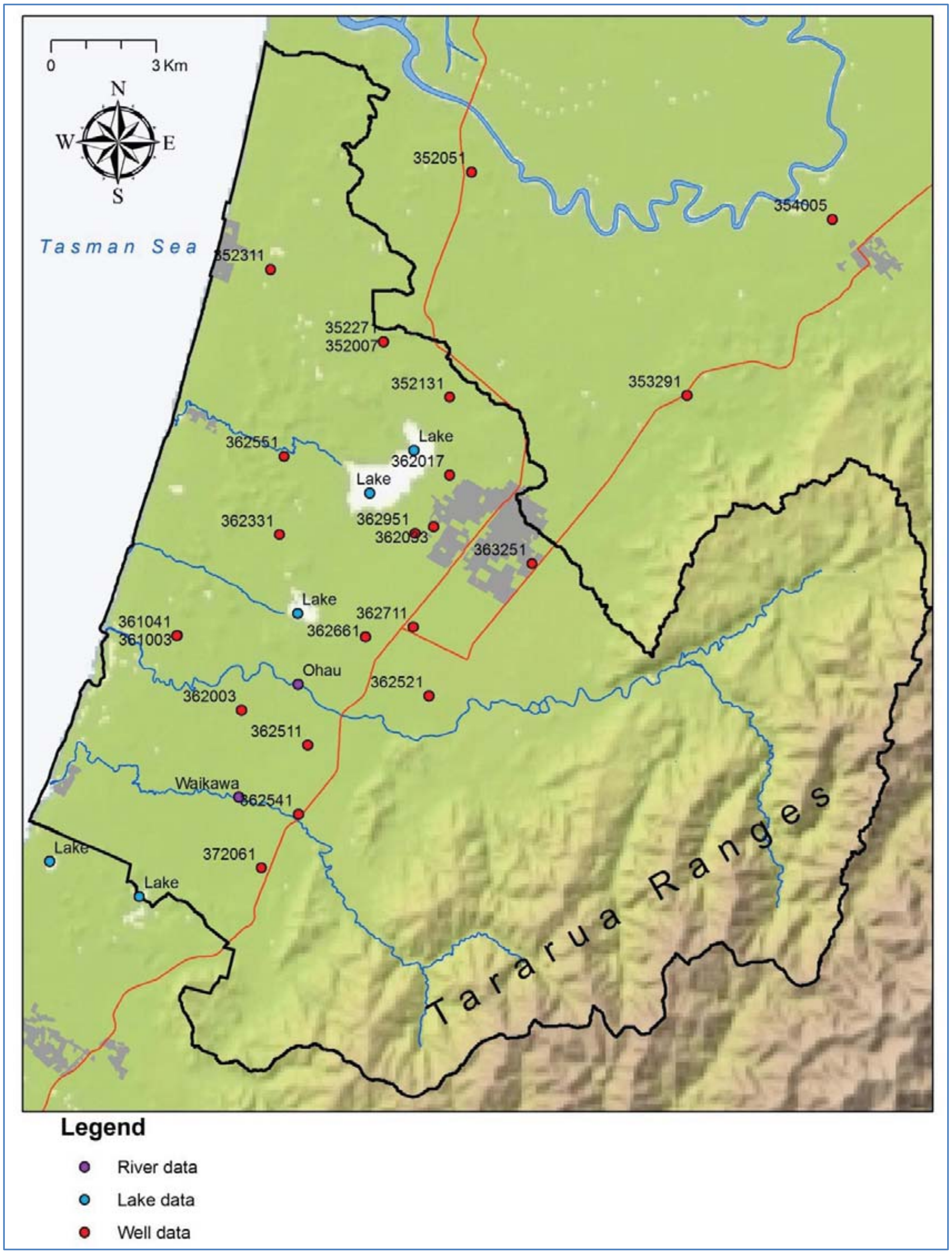

Figure 20 Groundwater well data locations in Horowhenua obtained from Horizons

2.6 Consumptive Use

The values for consumptive use of groundwater (including abstraction/withdrawals) have been observed from White et al. (2010). The water consumption volumes are not actually used in this study's water budget calculation due to the assumed return flow that would occur. Nevertheless it is useful in knowing the consumption of water to understand the demand on the water resources, particularly when groundwater levels are low. Water consumption is not metered in Horowhenua making accurate estimates difficult. Nevertheless White et al. (2010) have assigned a ± 10 % error these figures.

Table 11 Groundwater consumption in Horowhenua obtained from White et al. (2010)

Catchment	Consumption (Mcm)
Waitarere North	0.0
Waitarere South	0.5
Lake Horowhenua	2.1
Waiwiri	0.3
Ohau	7.4
Waikawa	1.6
Plains Total	**11.9**

2.7 Surface Water outflow

Surface water outflow is difficult to calculate due to the presumed major interactions between groundwater and surface water in the Horowhenua and lack of flow monitoring in the coastward channels of Horowhenua Rivers. This component is approached by firstly; analysing direct stream flow at the three gauging stations upstream; and secondly, estimating surface water outflow to the coast using an extrapolation technique.

The measured channel flow of the three main channels is obtained from three gauging stations at the base of the hillslopes. Ohau flow data was missing for 2007 and 2008, as was the December data for the Waikawa in 2009. The Ohau flow data was plotted against the Waikawa and Manakau data to deduce the best fit relationship. A best fit trend line was plotted and associated equation was used to extrapolate and estimate the missing values in the Ohau River.

The method used to estimate total surface outflow is from Woods, Hendrikx, Henderson & Tait (2006). The method uses a simple equation to calculate mean flow of un-gauged streams and rivers (Equation 27).

Equation 27

$$Q^{*} = \frac{P}{Err + 1} - AE$$

Where;

Q^{*} = corrected runoff

P = Precipitation over catchment

AE = Evapotranspiration

Err = error value

The error value (Equation 28) is calculated using gauging stations in the Ohau, Manakau and Waikawa streams (Figure 21), as well as mean precipitation and mean evapotranspiration for the upstream catchment from the gauging station to calculate bias.

Equation 28

$$Err = \frac{\left(P - (Qm + E)\right)}{(Qm + E)}$$

Where;

Err = error value

P = precipitation over catchment

E = evapotranspiration

Q_m = measured catchment runoff

Application of the equation is for the entire catchment of the Ohau River and the entire catchment of the Waikawa/Manakau catchment. The Waikawa and Manakau streams converge downstream from the gauging station therefore the error values are combined based on the proportion of land area for each individual catchment; Waikawa, 68 %; Manakau, 32 %. The calculation takes into account the two largest catchments which account for ≈75 % of the rainfall in Horowhenua. This is a simple equation and is cautiously used in this application whereby it is extrapolated over a greater area to which the measured catchment runoff may apply, particularly if groundwater/surface water interactions occur.

The Hokio stream outflow from the Lake Horowhenua is estimated at an average of 1.12 $m^3 s^{-1}$ (White, et al., 2010). This figure is attached to the total surface outflow. Although the location is just downstream of Lake Horowhenua, the area east of the lake is a discharging zone and flow is assumed to be stable over a short distance to the coast. Similarly, mean flow from Lake Papaitonga via Waiwiri stream is estimated at 0.018 $m^3 s^{-1}$ and is added onto the surface outflow calculation (White, et al., 2010).

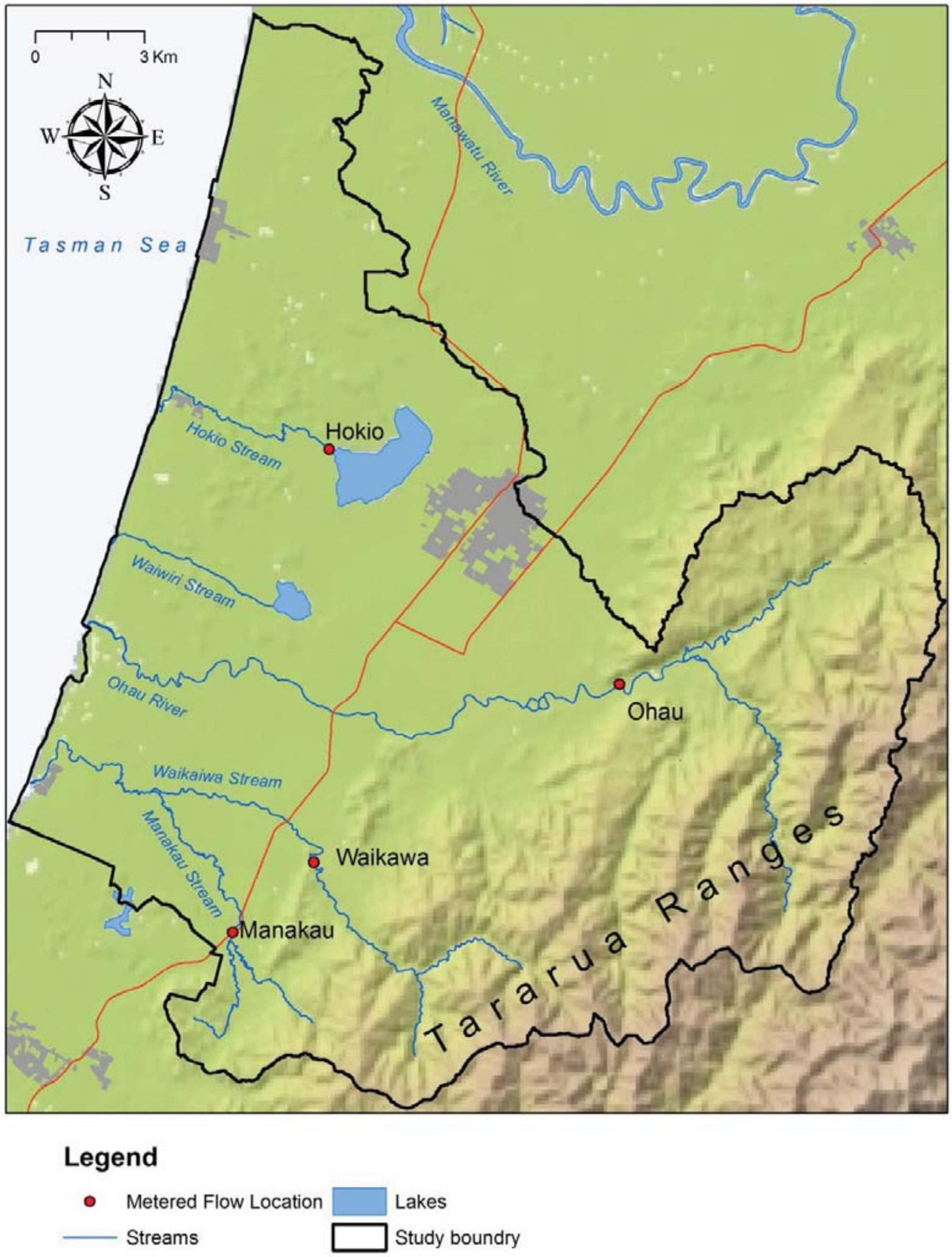

Figure 21 Horowhenua study location showing the flow meter locations on the Manakau, Waikawa and Ohau River

2.8 Water Budget and Storage Change

Calculation of the water budget uses the values obtained from the method mentioned above; precipitation (P), evapotranspiration (ET_c), groundwater outflow (G_{out}) and surface water outflow (S_{out}), which are used to calculate change in storage (ΔS) according to Equation 29.

Equation 29

$$P - ET_c - G_{out} - S_{out} = \Delta S$$

Where;

P = precipitation

ET_c = actual evapotranspiration

G_{out} = groundwater outflow

ΔS = change in storage

S_{out} = surface water outflow

In addition, change in storage (ΔS) is estimated using the difference in monthly potentiometric surfaces, similar to the water table fluctuation method (WTF). Values for average changes in groundwater levels for the catchment were estimated using monthly potentiometric surfaces which subtracted from one averaged to get the mean change in groundwater level. This was only able to be done for the area covered by the potentiometric surface, which does not include the entire lowland plains and results in an extrapolation of mean values to the unmeasured area of the catchment. The averaging of the differences was done individually for each catchment (Figure 6) from the plains and subsequently multiplied by the respective total catchment areas to get the volumetric change in storage. Calculating via each individual catchment helps to give the most appropriate mean values to the unmeasured areas. Annual lake levels have a net effect of no change as lake level data was not present. Assigning the storage values uniquely to each month also presents issues due to not having the measurement taken on the same day each month and also varying slightly for each well station. For purposes of simplicity the storage change between, for example, January 07 and February 07 is assigned to the month of January as it is assumed the influential climate factors of January are exhibited in the following month.

The estimation of change in storage is additionally used reversely in Equation 30 to calculate the value of surface outflow to get an idea of accuracy of the independent surface water outflow.

Equation 30

$$P - ET_c - G_{out} - \Delta S \; = \; S_{out}$$

Where;

P	= precipitation
ET_c	= actual evapotranspiration
G_{out}	= groundwater outflow
ΔS	= change in storage
S_{out}	= surface water outflow

It is difficult to use this method to estimate the monthly water outflow due to changes in storage between groundwater measurements not fitting neatly into a calendar month and thus the budget doesn't calculate properly for each individual month.

Chapter 3 Results

3.1 Introduction

This section displays the results according to each of the key water budget components. Results are mostly summarised into monthly and annual time intervals and also according to sub-catchments.

3.2 Precipitation

Total rainfall was calculated to be 447.19 Mcm, 642.26 Mcm and 621.52 Mcm for the years of 2007, 2008 and 2009 respectively. Average rainfall for 2007, 2008, and 2009 was 1232.28, 1739.57, and 1690.62 mm respectively; 2007 was evidently a drier year, while both 2008 and 2009 were wetter. A clear West-East gradient exists with higher rainfall occurring in the east as the elevation rises due to the Tararua Range. This is also shown in Figure 22 where the mean rainfall for each catchment is shown and averages increase in the more eastern catchments. A breakdown of rainfall into each assigned catchment (Table 12) shows the greater rainfall component that comes from the hill slope catchments despite having a lower area in comparison to the plains.

Rainfall appears to be quite sporadic from month to month and year to year (Figure 23). The 2007 year has overall lower monthly rainfall, particularly February and December appear to be considerably low. June, July and September also appear to be low in each year (Figure 25). The 2008 year clearly has some very wet months, notably July and August, but also, June and October are wetter than in other years showing a very wet winter period. It also has a very wet summer period shown in Figure 26 by the significant red distribution into the catchment. The 2009 year has very wet summer period shown within January and February. January has the wettest month for the year as well as one of the wettest months of the 2007 – 2009 periods. February of 2009 also has the wettest February out of the 2007 – 2009 periods (Figure 27). Rainfall increases substantially in the later months of the year of 2009, causing the total rainfall for the year to rise to above average.

Table 12 Calculated rainfall volumes for each sub-catchment and for years 2007, 2008, 2009

Catchment	Area (km^2)	Rainfall Mcm		
		2007	2008	2009
Waitarere North	19.50	15.74	22.09	20.97
Waitarere South	28.52	23.34	33.18	31.29
Lake Horowhenua	61.54	60.63	86.07	80.39
Waiwiri	25.88	21.45	31.12	29.20
Ohau	54.18	53.23	79.19	74.75
Waikawa	38.48	34.80	52.89	50.44
Plains Total	**228.09**	**209.19**	**304.54**	**287.04**
Hill 1	1.21	1.78	2.55	2.46
Hill 2	125.33	220.26	300.05	299.35
Hill 3	39.44	52.53	75.76	74.84
Hillslopes Total	**165.99**	**274.57**	**378.36**	**376.65**
Total	**392.57**	483.75	682.90	663.69

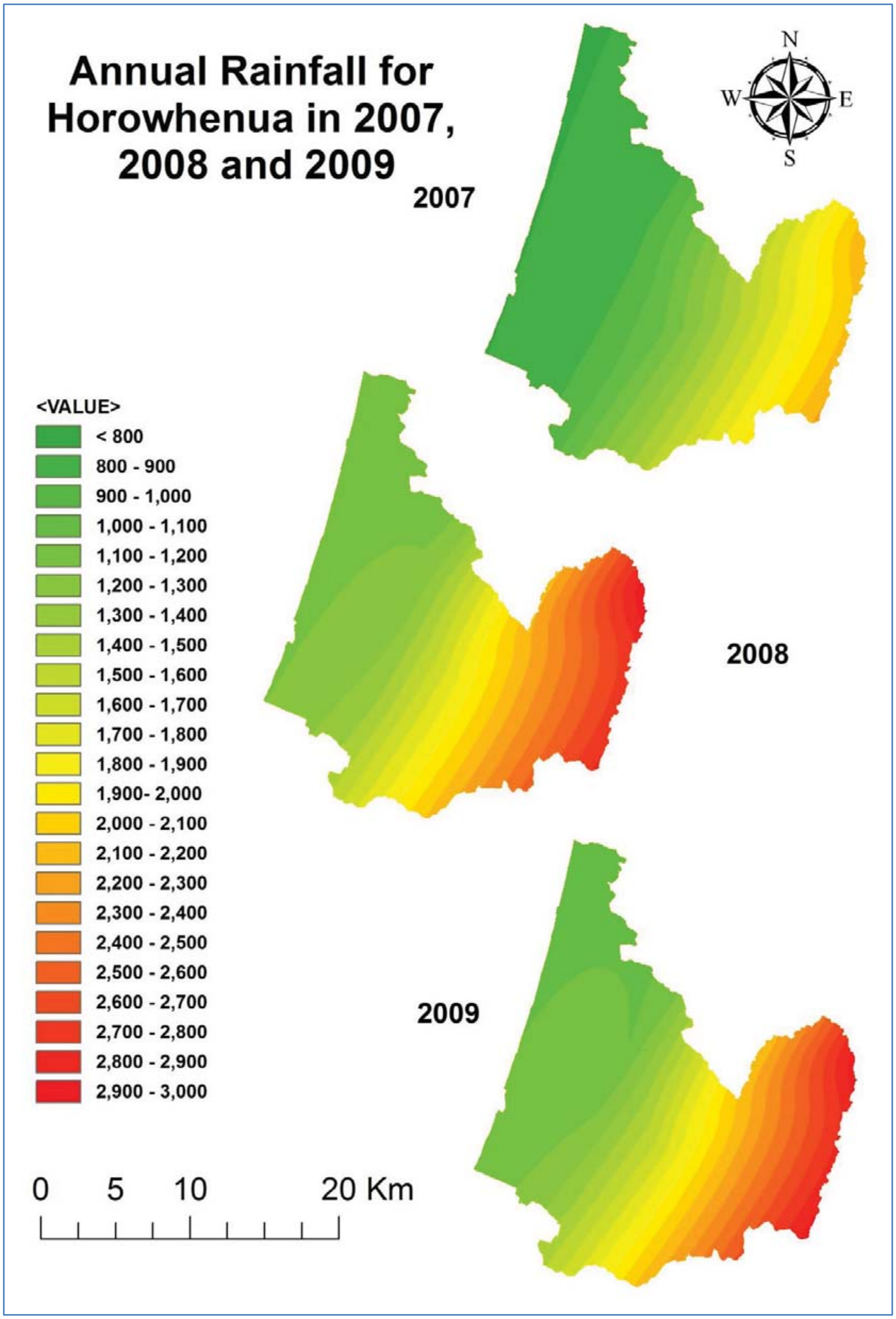

Figure 22 Annual rainfall for 2007, 2008 and 2009

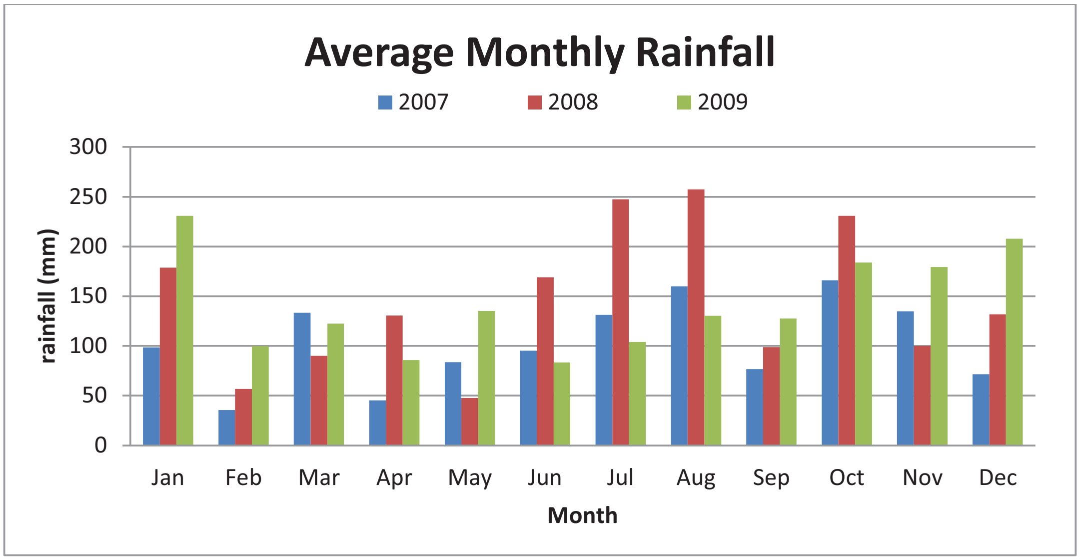

Figure 23 Graph of average monthly rainfall for 2007, 2008 and 2009

Table 13 Average monthly and annual rainfall for 2007, 2008 and 2009

Year	Monthly Rainfall (mm)												
	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec	Total
2007	98.5	35.7	133.2	45.3	83.8	95.3	131.3	160.0	76.7	166.0	134.9	71.6	1232.3
2008	178.8	56.7	90.1	130.6	47.7	169.1	247.5	257.5	98.9	230.9	99.9	131.9	1739.6
2009	230.9	99.9	122.3	85.9	135.2	83.3	103.9	130.3	127.7	183.9	179.4	207.9	1690.6

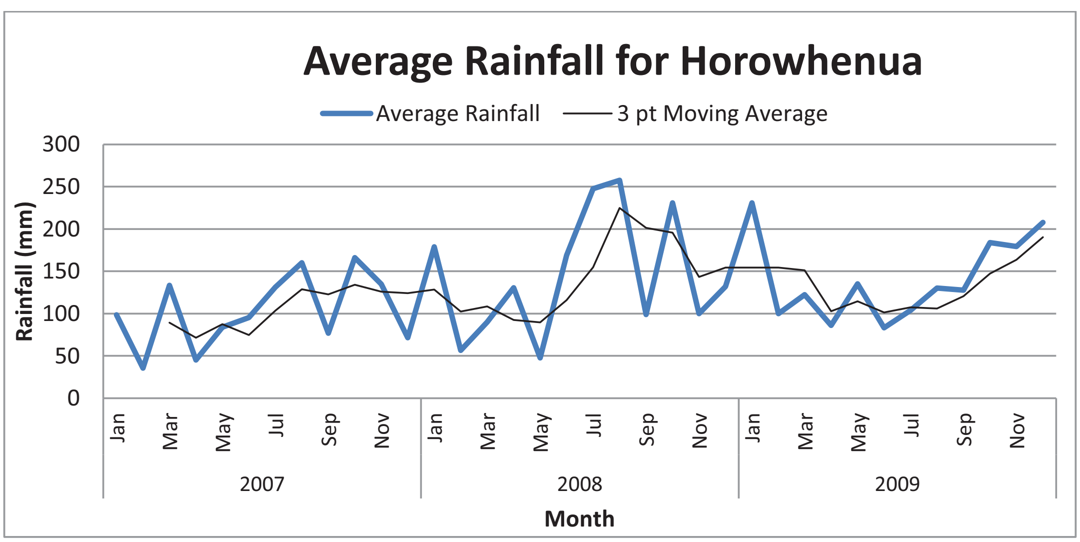

Figure 24 Average monthly rainfall for Horowhenua showing smoothing trendline for 2007, 2008 and 2009

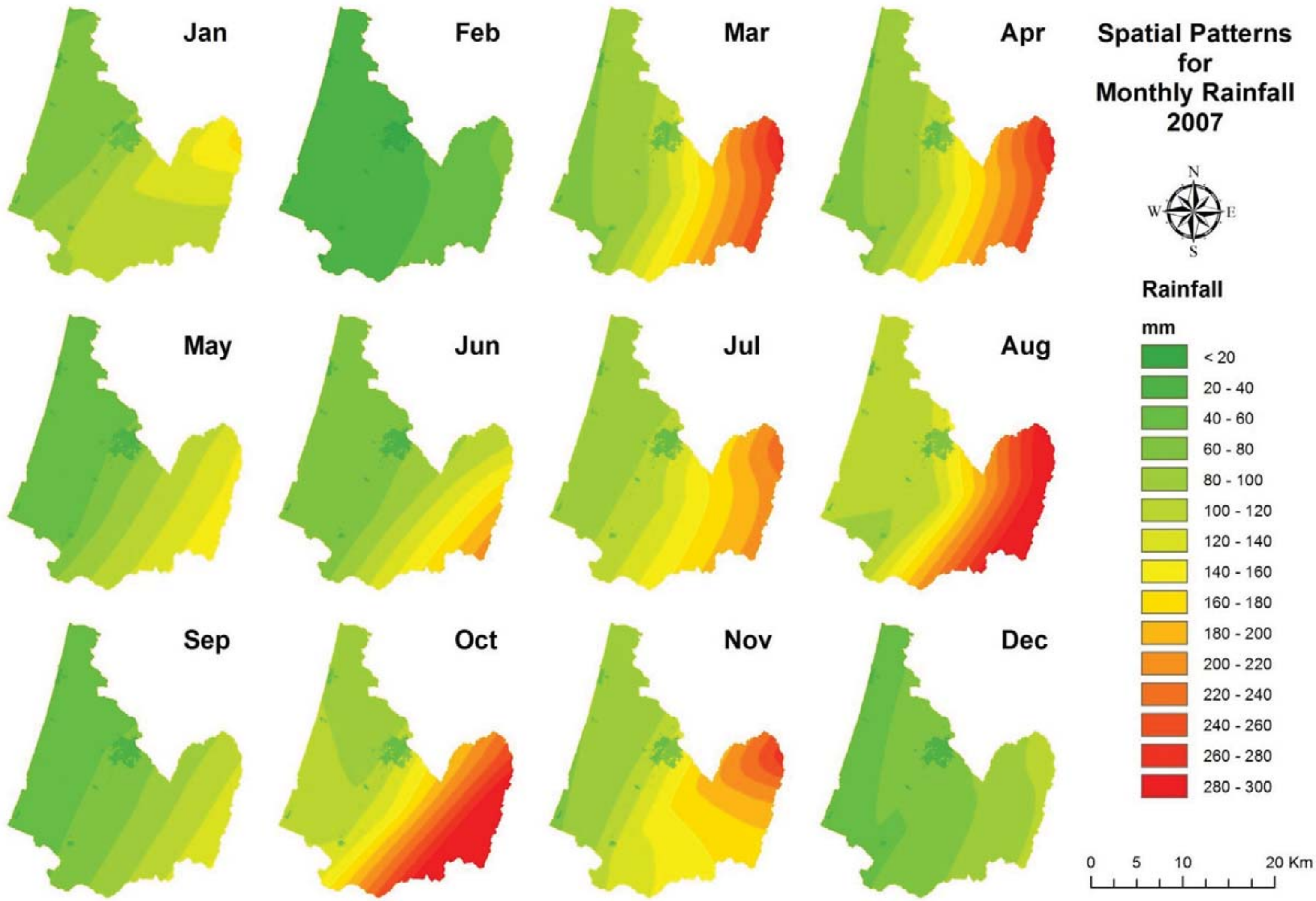

Figure 25 Spatial patterns for monthly rainfall in 2007

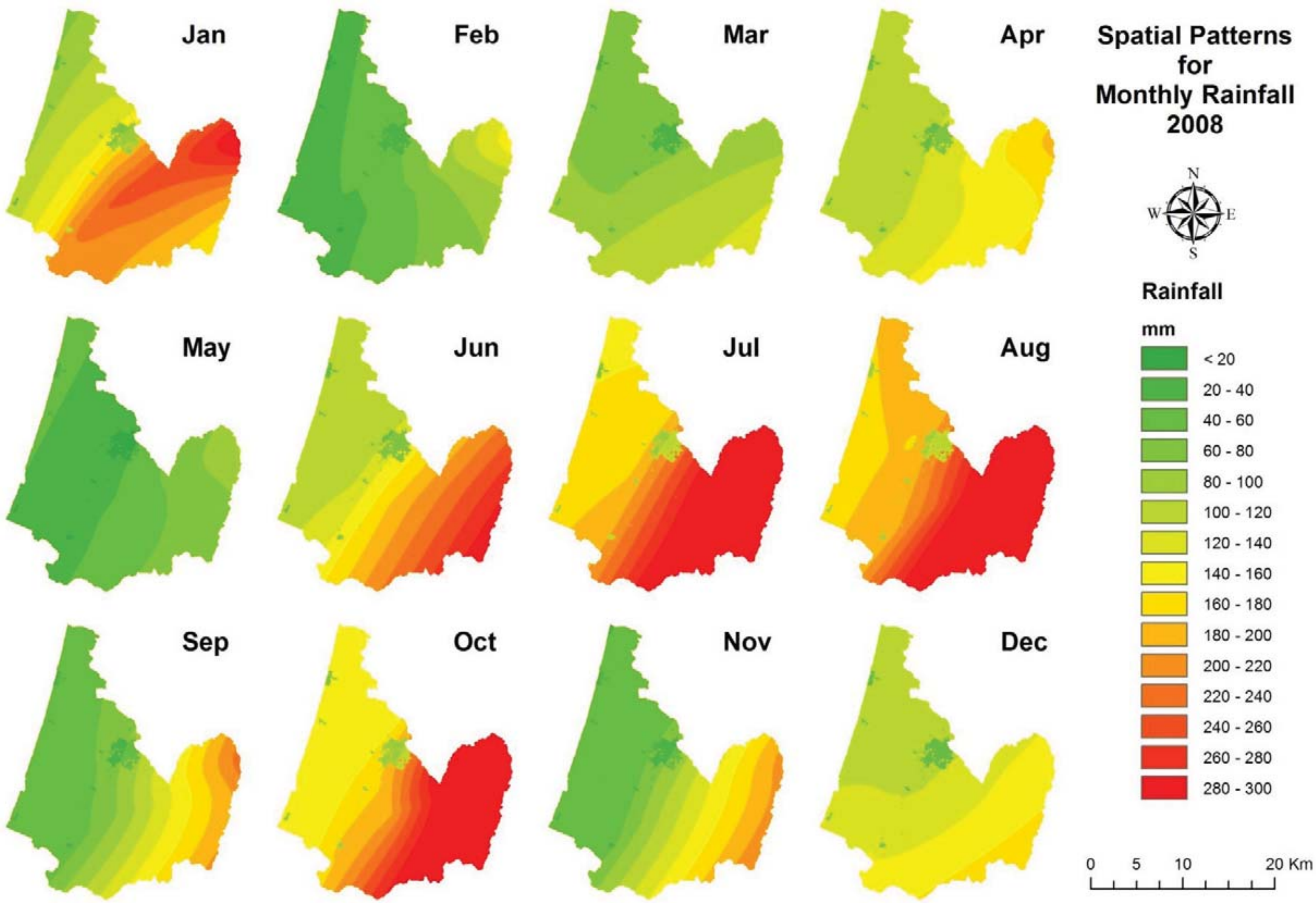

Figure 26 Spatial patterns for monthly rainfall in 2008

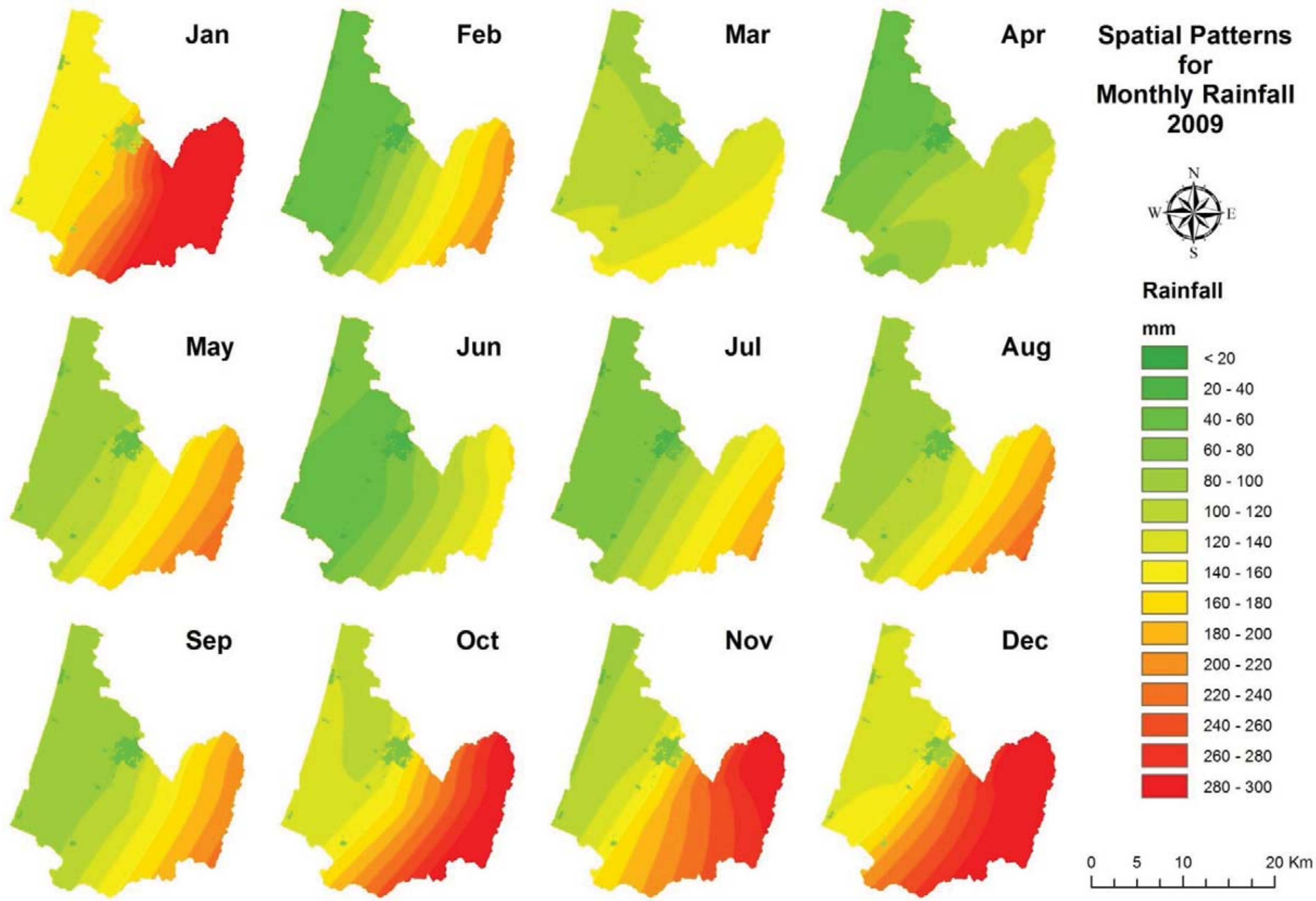

Figure 27 Spatial patterns for monthly rainfall in 2009

3.3 Evapotranspiration

Evapotranspiration rates for the area show a similar pattern for each year with the lowest evaporation occurring in central Horowhenua in an ellipsoid type pattern which rises eastwards slightly although is not attributed to elevation effects. ET_c is approximately 10 Mcm higher than ET_o in 2007, most of the difference arising from the hillslope catchments (Table 14). In 2008 the difference is approximately 20 Mcm but this time the difference arises from the plains. The 2009 values are actually very similar, and only differ by ≈ 4 Mcm. Month to month there doesn't seem to be any explicit pattern of whether the ET_c is higher than the ET_o or not except that the latter half of each year typically tends to result in an increase in value when changing to ET_c values. The opposite pattern is possibly shown in the first half of 2007 and 2008. The pattern itself is markedly affected by the landuse with clear visualization of the landuse pattern (Figure 31).

Monthly analysis of the evapotranspiration shows considerable seasonal trends for each year (Figure 32). The highest evapotranspiration occurs in the summer months of December, January and February (Figure 33, Figure 34, and Figure 35), with the lowest values occurring in May, June, July and August. The 2007 year has the highest evapotranspiration overall (Table 14) with the highest monthly values in 7 out of 12 months. The 2008 year has the lowest evapotranspiration and 2009 was between 2007 and 2008. There is a loose relationship with rainfall, where the highest rainfall year is the lowest evapotranspiration year and vice versa. The variability of the evapotranspiration between years and even within years is a lot less pronounced than the rainfall variability.

Spatially there are clear patterns exerted on the evapotranspiration pattern from the calculation components. Clearly, the landuse pattern (Figure 13) can be seen exerted in the evapotranspiration images (Figure 31, Figure 33, Figure 34 and Figure 35). Similarly, an east-west trend is seen for evapotranspiration although it's not clear what main variable causes this pattern. The effect of different soils is also visible in the images, although it is harder to see amongst the landuse patterns. The soil effects are seen clearest in the drier months such as February (Figure 33). The soil patterns can be seen quite clearly with 76b soils, which Levin sits on, and the 76/76a soils to the south where an increase in evapotranspiration is seen. The other strong differences are between the 75 soils at the base of the foot hills and the 78 soils in the upper reaches of the Ohau River.

Table 14 Comparison of ETo and ETc values for sub-catchments in 2007, 2008 and 2009

	Evaporation (Mcm)					
Catchment	2007		2008		2009	
	ETo	ETc	ETo	ETc	ETo	ETc
Waitarere North	11.93	12.37	12.44	11.42	11.57	11.71
Waitarere South	17.04	17.35	17.60	15.97	16.69	16.65
Lake Horowhenua	35.36	34.51	36.35	32.07	34.49	33.01
Waiwiri	15.40	15.87	15.74	14.50	15.14	15.28
Ohau	32.42	33.32	33.23	30.69	31.77	32.10
Waikawa	23.55	23.80	23.97	21.70	23.14	23.07
Plains Total	**135.69**	**137.22**	**139.32**	**126.35**	**132.80**	**131.82**
Hill 1	0.77	0.79	0.81	0.75	0.73	0.73
Hill 2	83.95	89.73	88.29	87.48	78.91	82.17
Hill 3	24.87	26.55	25.91	25.21	24.45	25.49
Hillslopes Total	**109.60**	**117.07**	**115.00**	**113.44**	**104.09**	**108.39**
Total	**245.29**	**254.29**	**254.32**	**234.747**	**236.90**	**240.21**

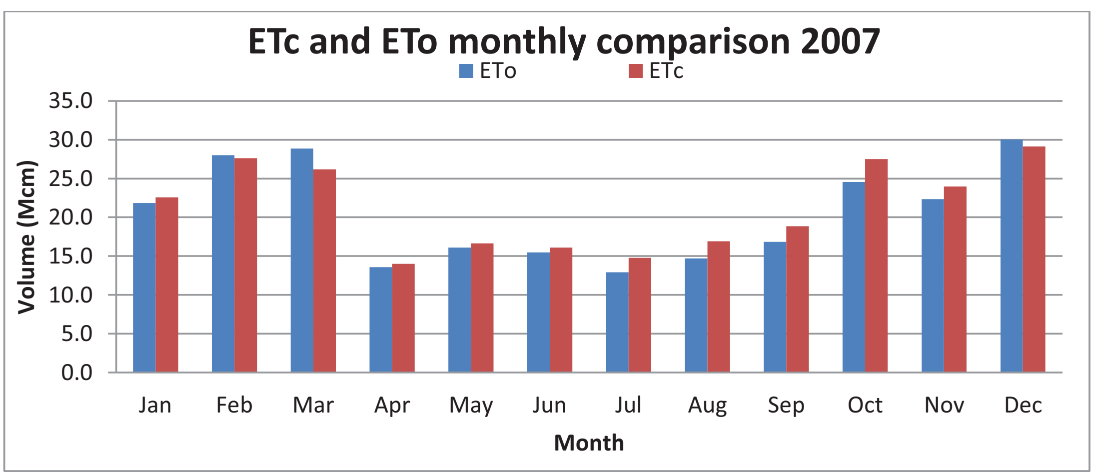

Figure 28 Monthly ETc and ETo comparison in 2007

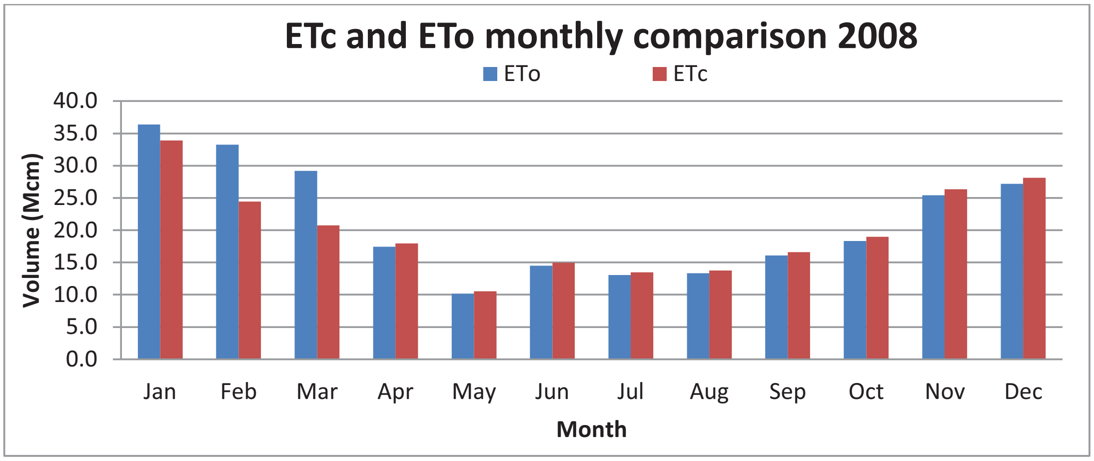

Figure 29 Monthly ETc and ETo comparison in 2008

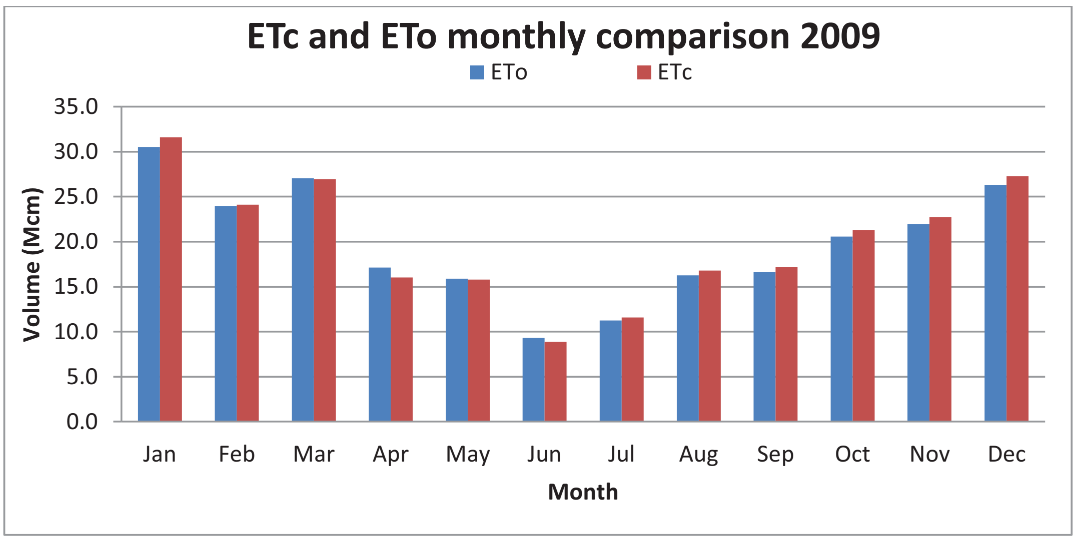

Figure 30 Monthly ETc and ETo comparison in 2009

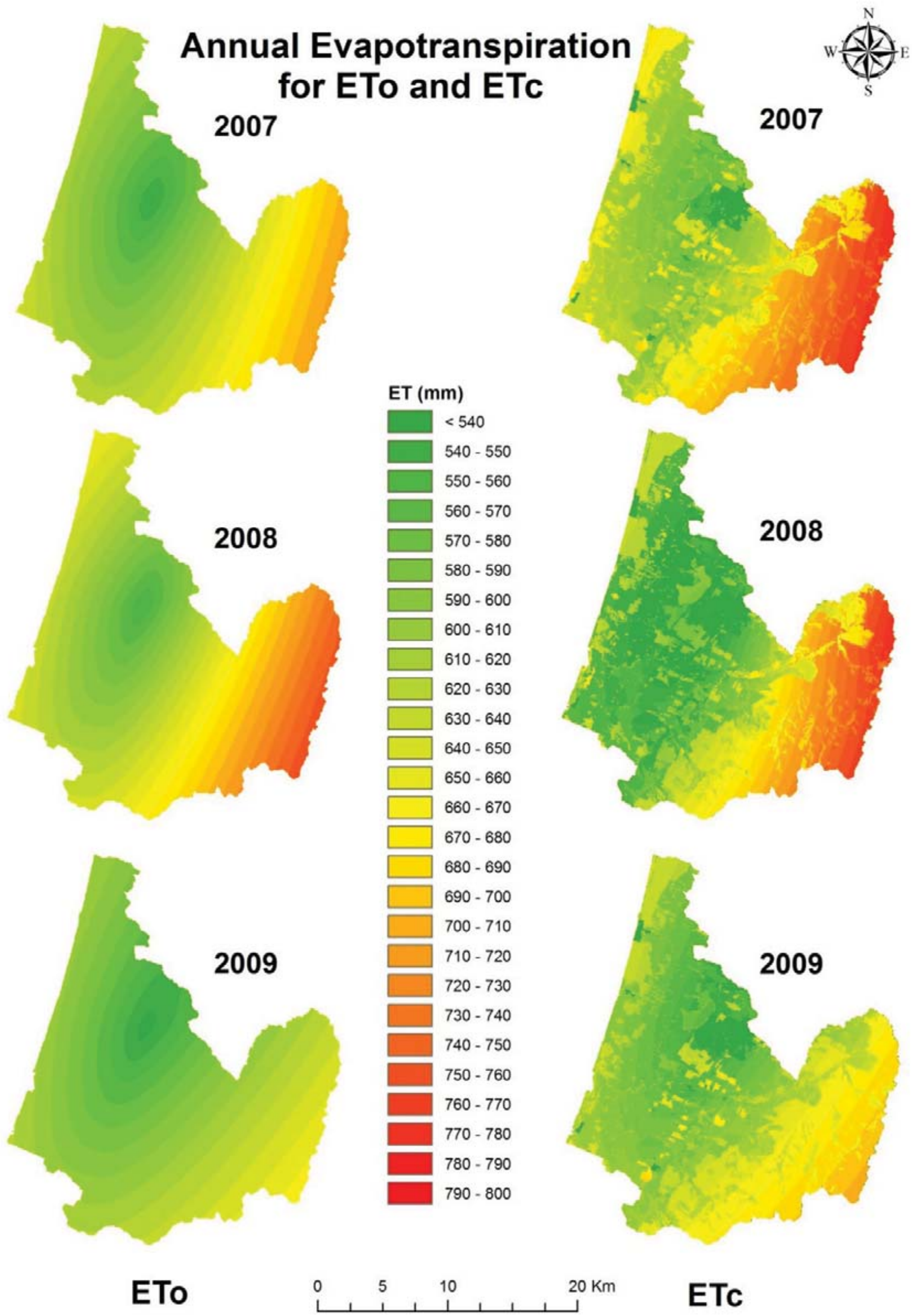

Figure 31 Annual evapotranspiration for ETo and ETc in 2007, 2008 and 2009

Table 15 Average monthly evapotranspiration values for 2007, 2008 and 2009

Year	Monthly Evapotranspiration (mm)											
	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec
2007	57.5	70.3	66.7	35.7	42.4	41.0	37.6	43.1	48.1	70.1	61.0	74.2
2008	86.4	62.3	52.8	45.7	26.8	38.1	34.4	35.0	42.3	48.4	67.1	71.6
2009	80.5	61.4	68.6	40.8	40.2	22.6	29.5	42.8	43.7	54.3	58.0	69.5

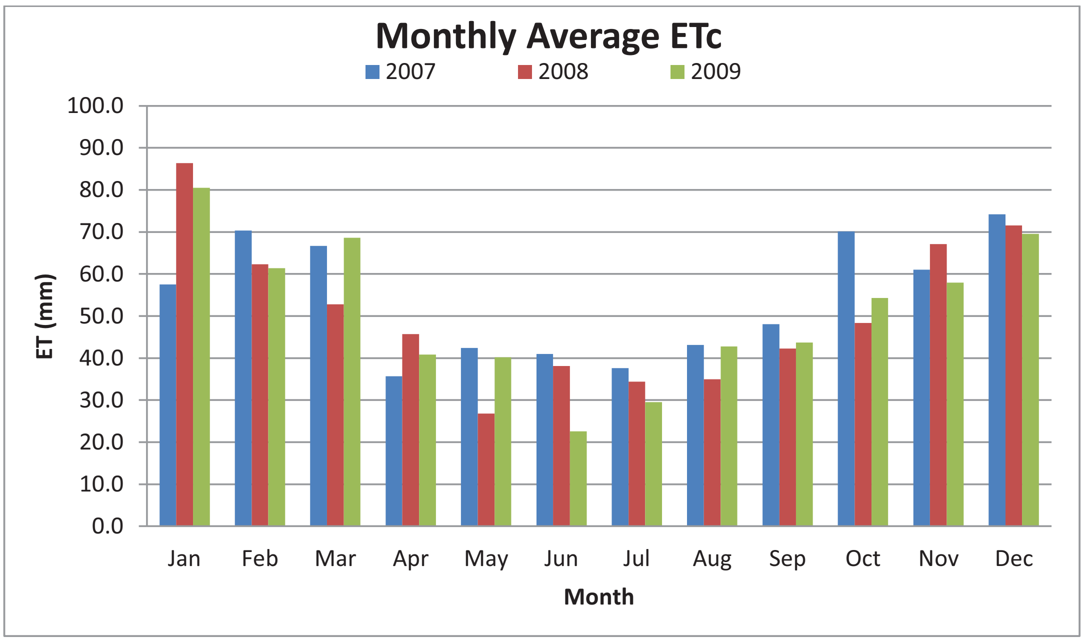

Figure 32 Average monthly evapotranspiration values for 2007, 2008 and 2009

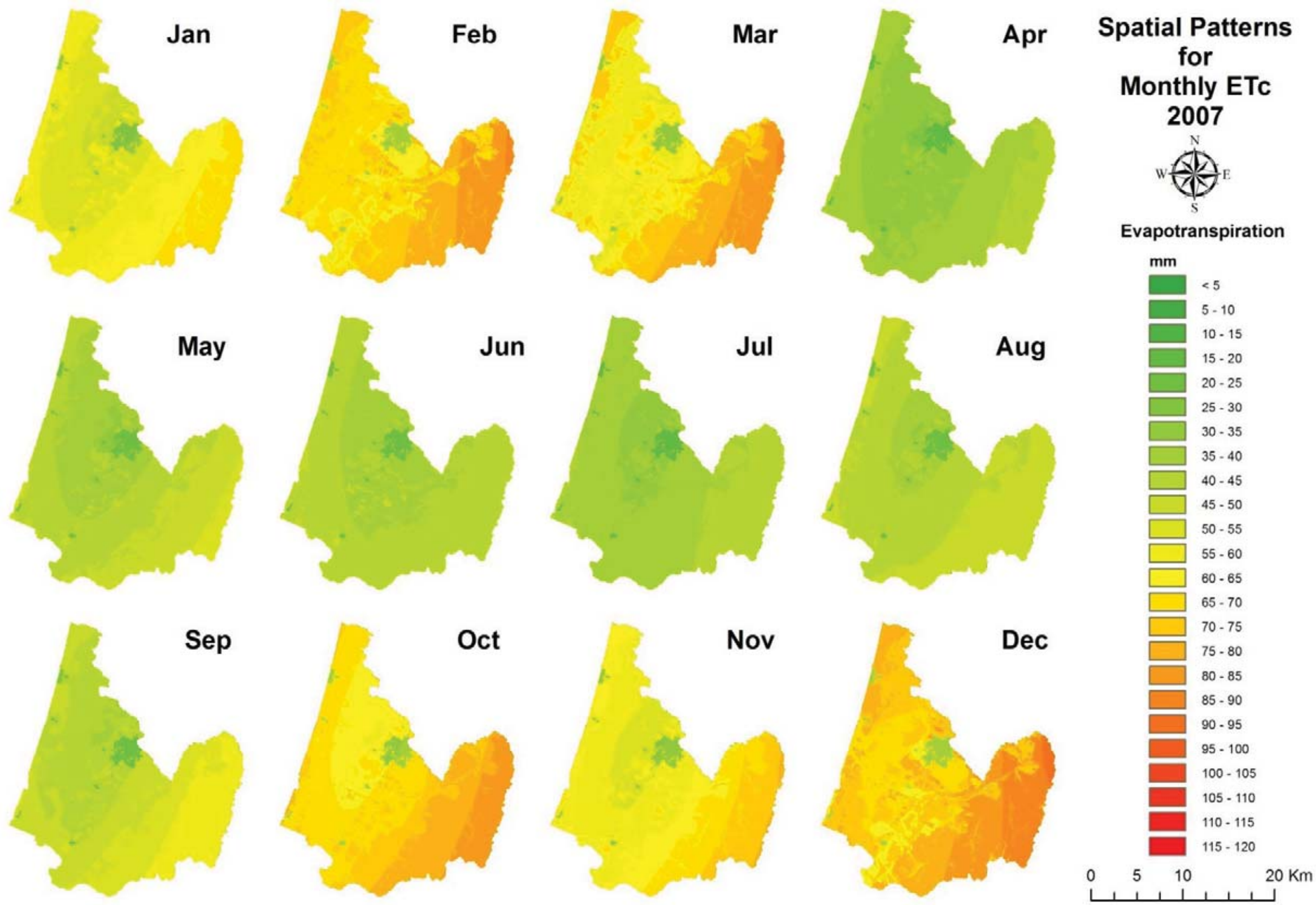

Figure 33 Spatial pattern of monthly evapotranspiration for 2007

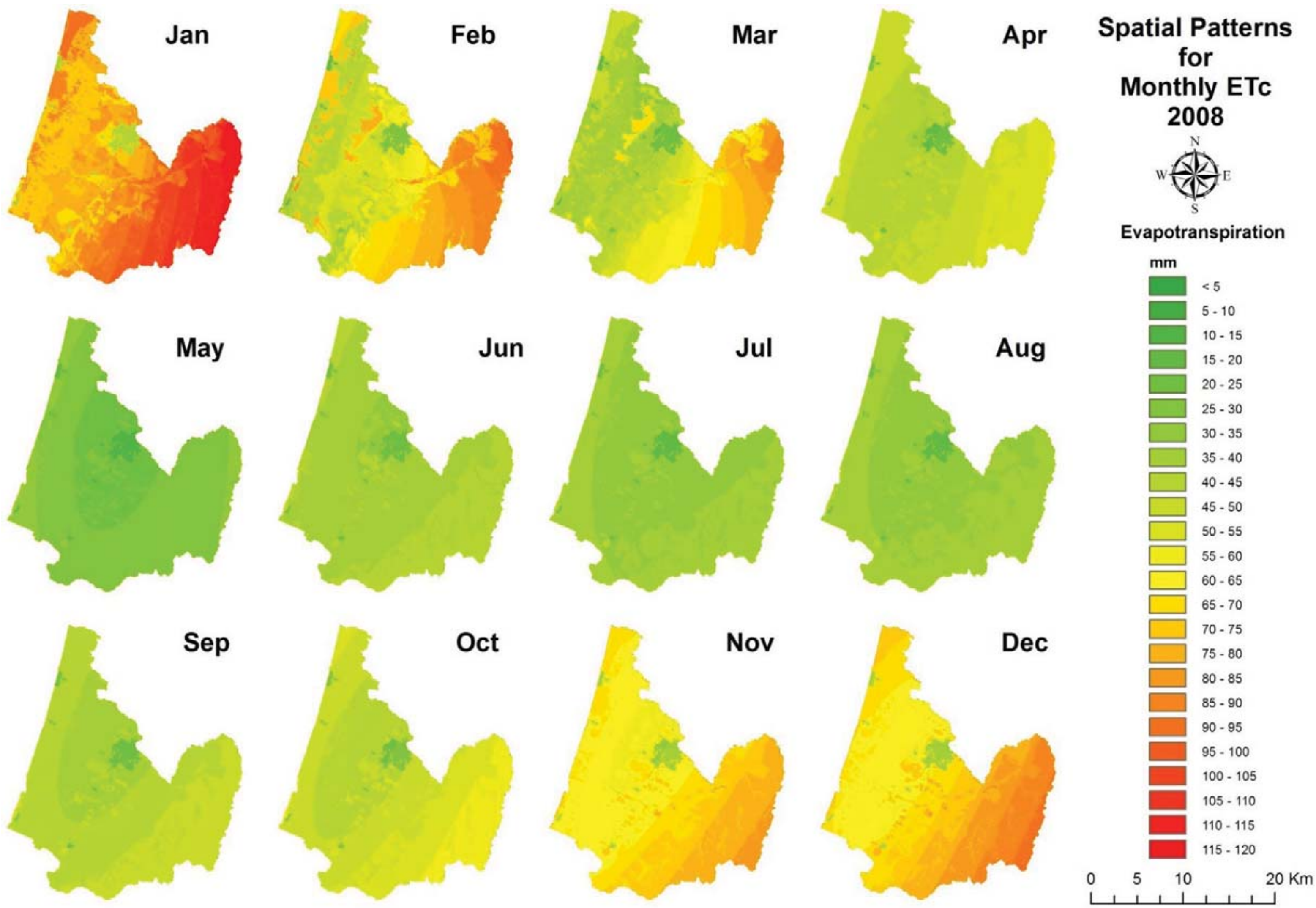

Figure 34 Spatial pattern of monthly evapotranspiration for 2008

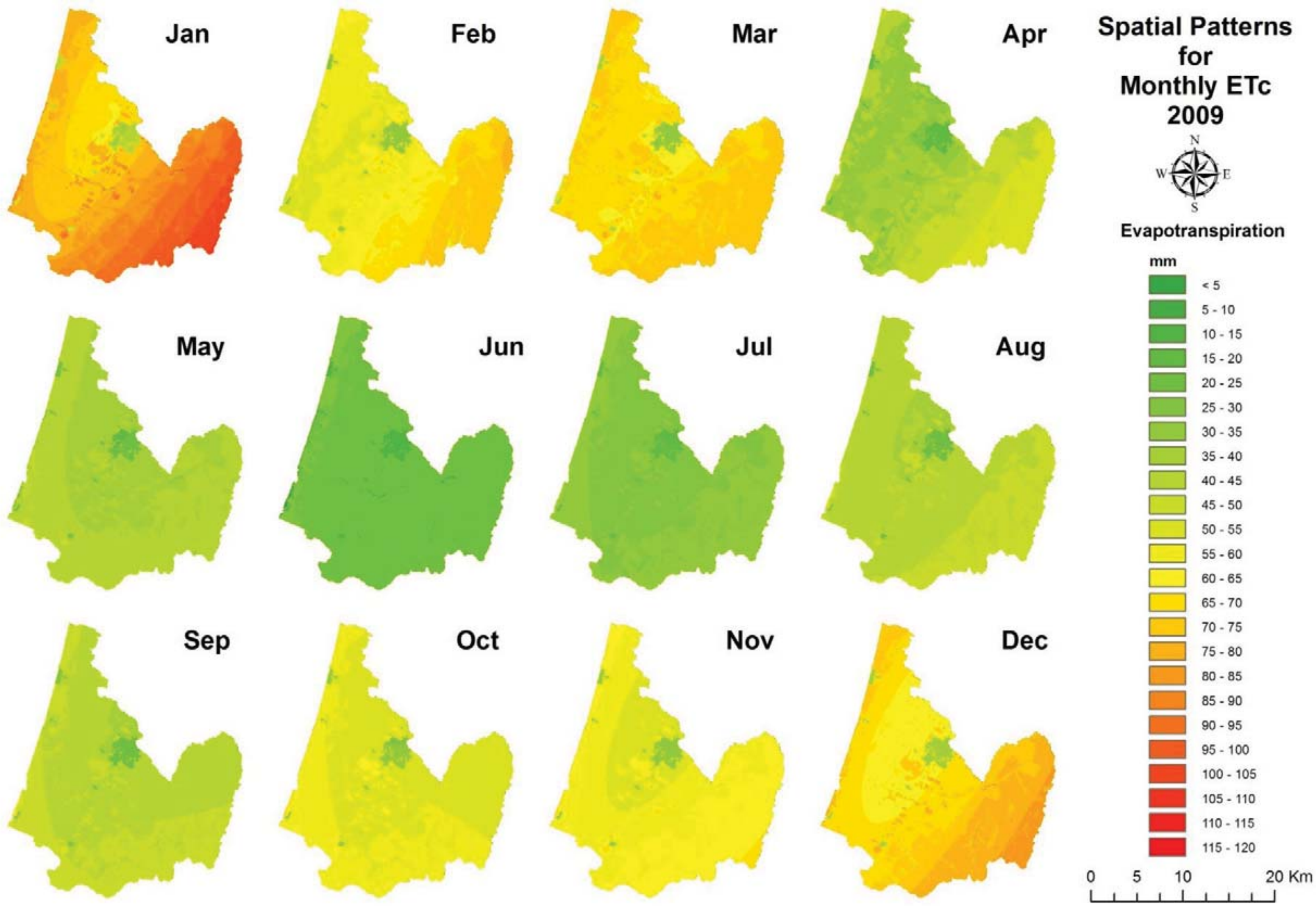

Figure 35 Spatial pattern of monthly evapotranspiration for 2009

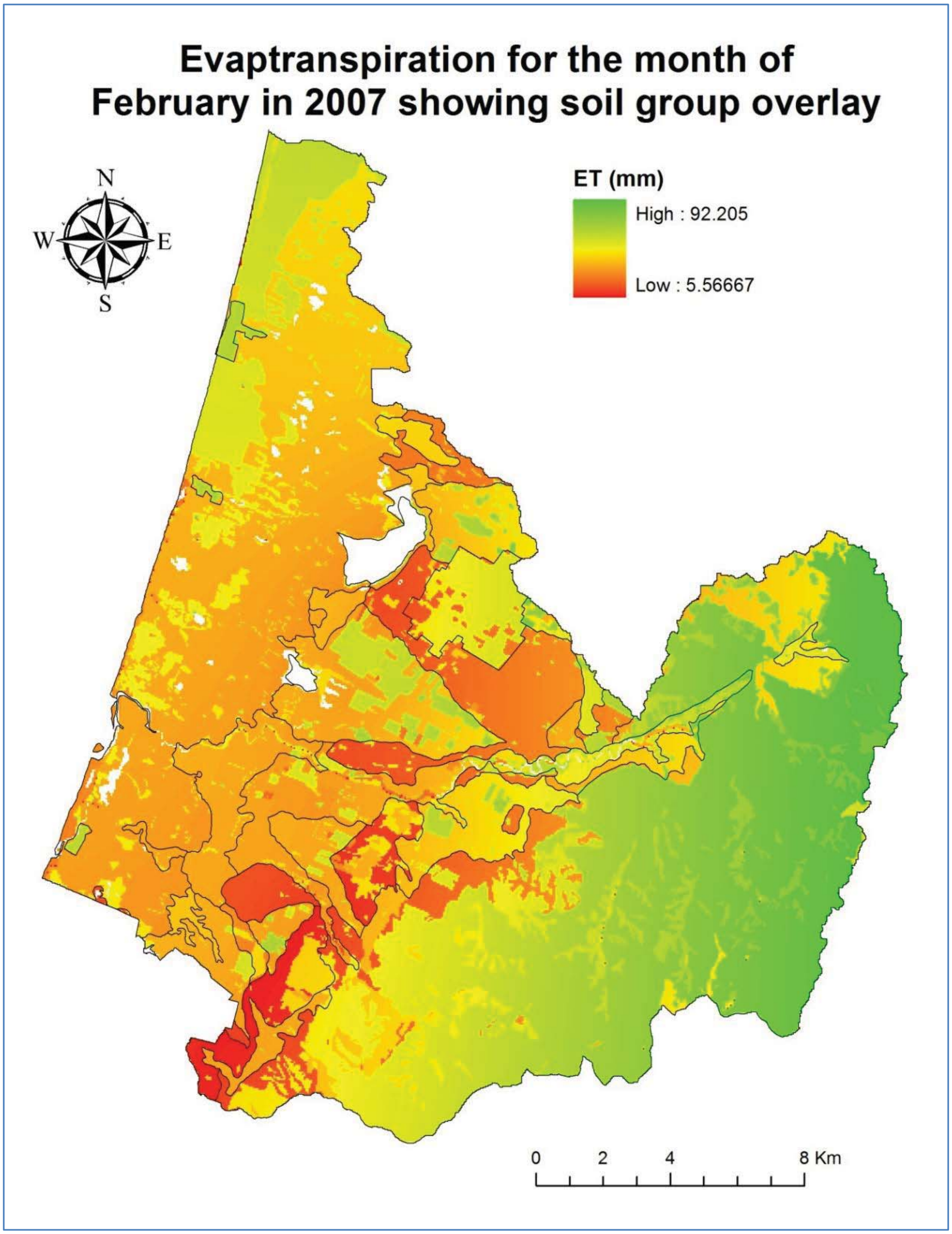

Figure 36 Evapotranspiration for February 2007 with the soil group overlay

3.4 Excess rainfall

Excess rainfall is referred to simply as rainfall – evapotranspiration and gives an indication of the water available for other components and thus the maximum potential outflow of a catchment if it had a steady state balance (White, et al., 2010). Figure 37 shows excess water is most apparent in the higher rainfall periods. However if the high rainfall months occur in the summer season then the effect of evapotranspiration loss has a greater impact and effectively subdues the excess rainfall value for that month.

The excess rainfall comes mostly from the hill slopes (Figure 38, Figure 39 and Figure 40), which is where the highest rainfall occurs. However the 'Hill 2' catchment for 2007 also has the largest deficit in rainfall in February. The Ohau and Lake Horowhenua catchments also experience overall higher excess rainfall volumes. February for all catchments typically had the lowest excess, sometimes going negative, except in 2009 where the Hill 2 and Hill 3 catchments did not have the lowest excess rainfall (Figure 40).

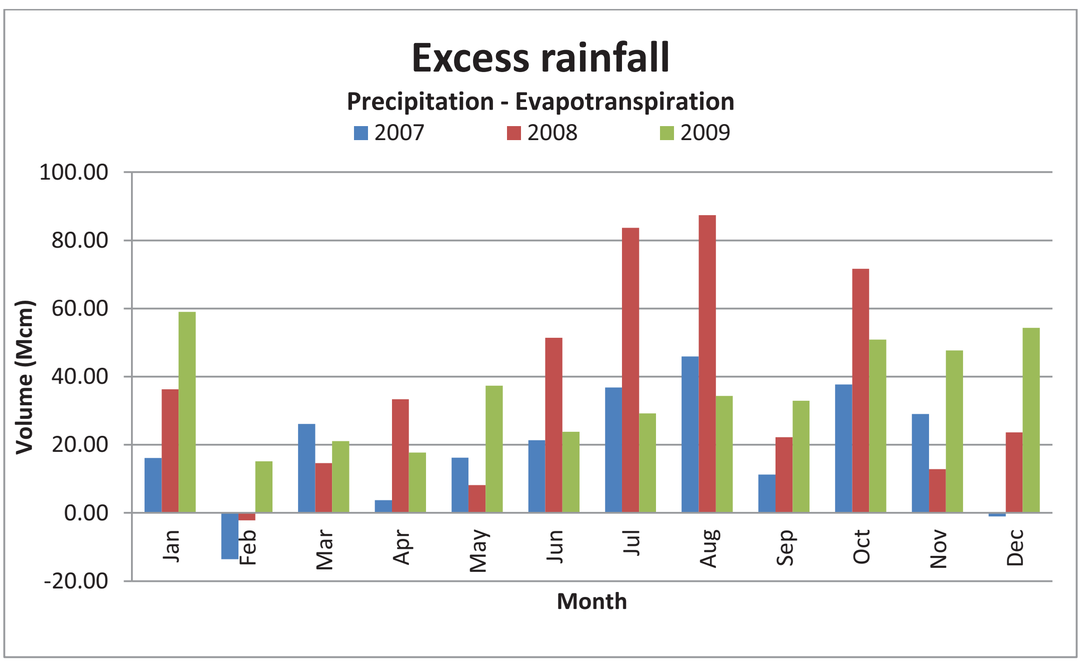

Figure 37 Monthly excess rainfall for 2007, 2008 and 2009

Table 16 Excess rainfall shown for sub-catchments in 2007, 2008 and 2009

Catchment	Excess Rainfall Volumes		
	2007	2008	2009
Waitarere North	3.37	10.67	9.26
Waitarere South	5.99	17.20	14.64
Lake Horowhenua	26.12	54.00	47.38
Waiwiri	5.58	16.62	13.92
Ohau	19.91	48.50	42.65
Waikawa	11.00	31.19	27.37
Plain Total	**71.96**	**178.19**	**155.22**
Hill1	1.00	1.80	1.73
Hill2	130.53	212.57	217.18
Hill3	25.97	50.55	49.35
Hillslopes Total	**157.50**	**264.92**	**268.26**
Total	**229.46**	**448.15**	**423.48**

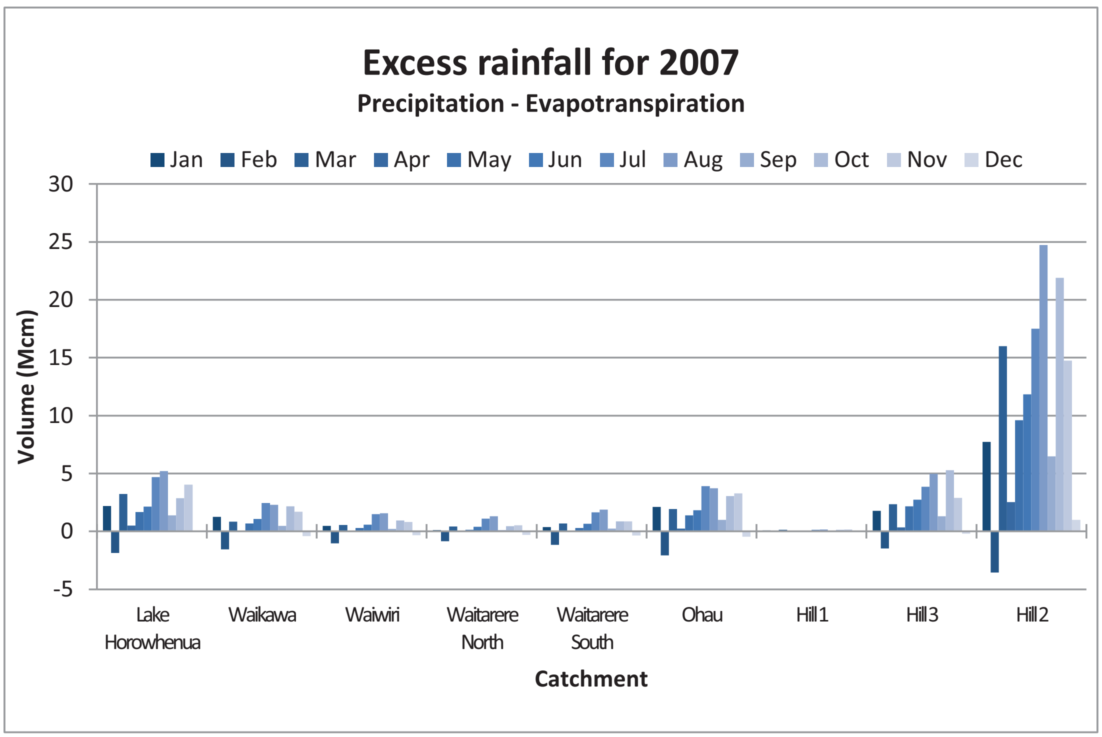

Figure 38 Excess rainfall for each sub-catchment and month in 2007

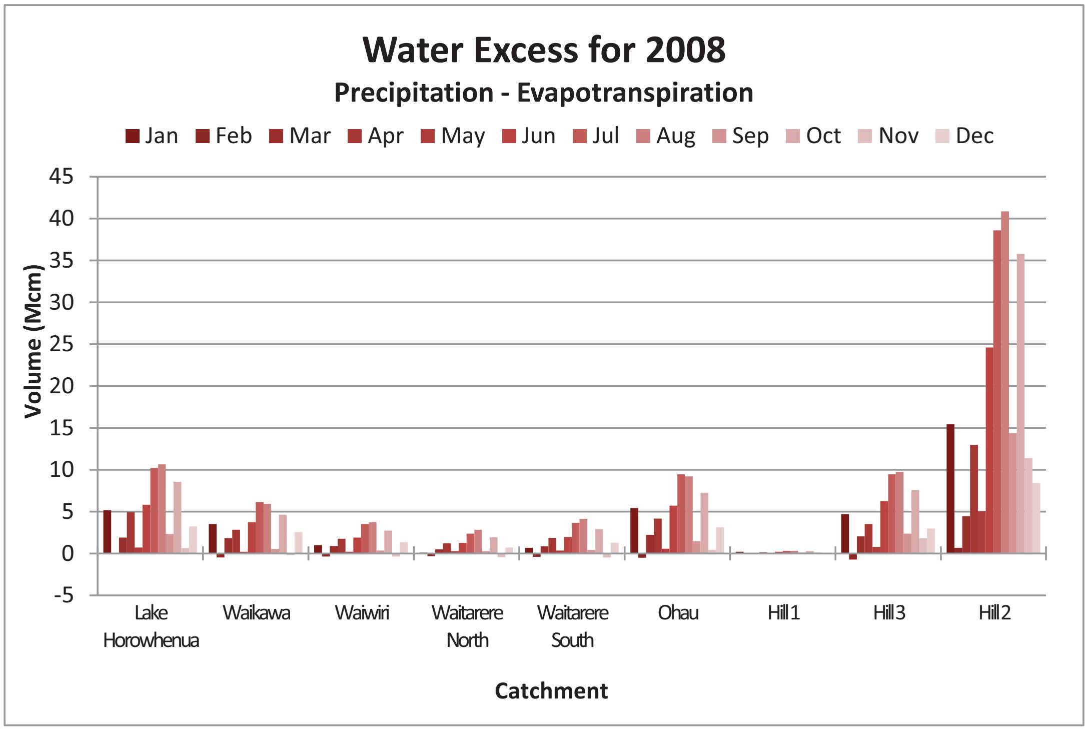

Figure 39 Excess rainfall for each sub-catchment and month in 2008

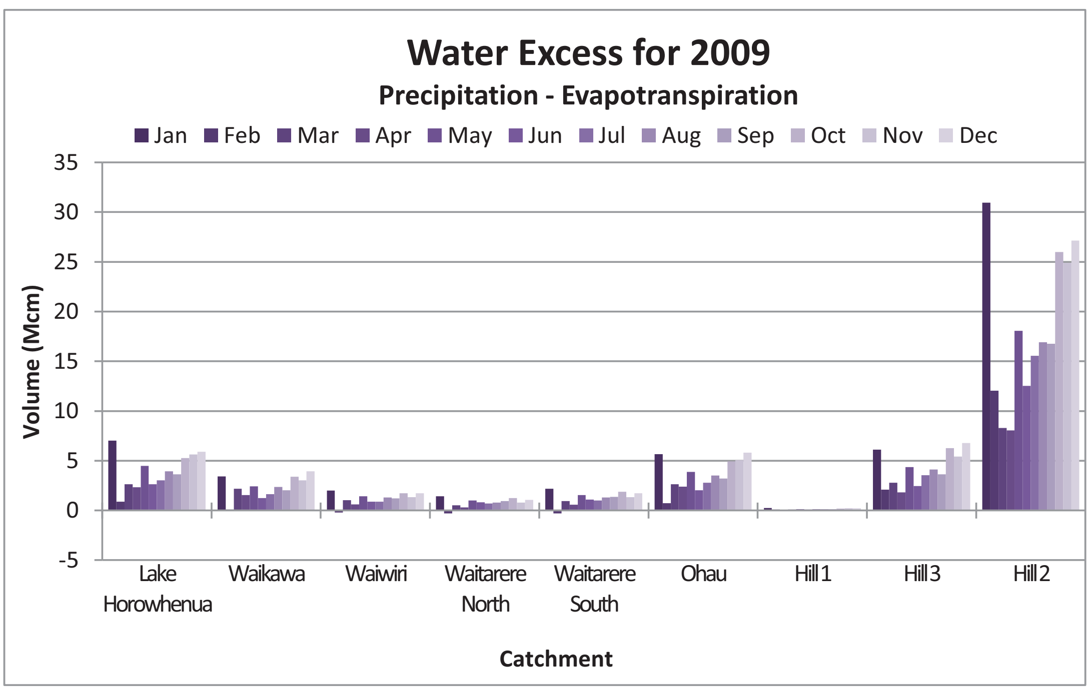

Figure 40 Excess rainfall for each sub-catchment and month in 2009

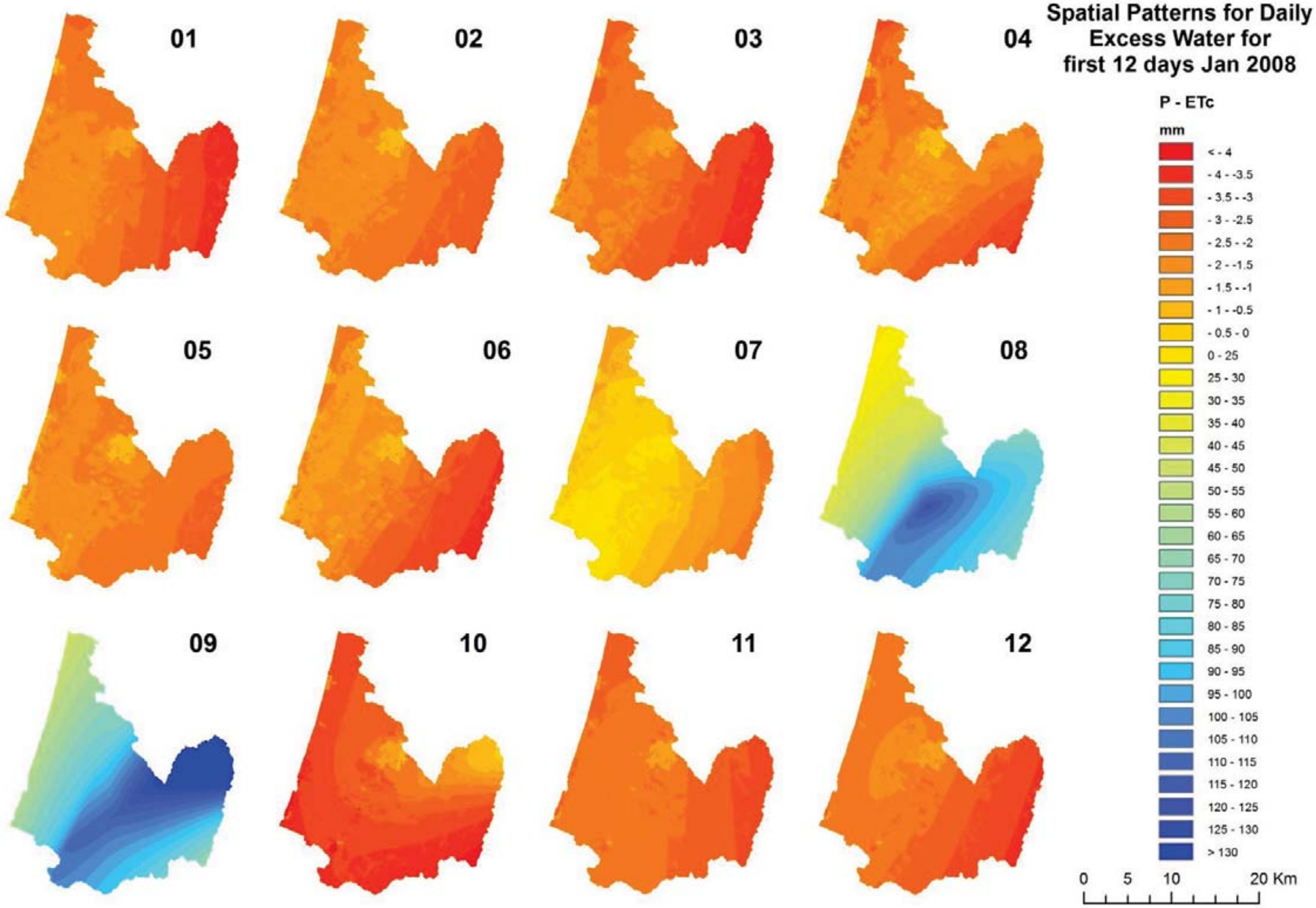

Figure 41 Daily excess rainfall over the first 12 days of January 2008

3.5 Groundwater levels and groundwater outflow

3.5.1 Introduction

This section is separated into three segments; potentiometric surface and resultant hydraulic gradient; the groundwater outflow calculation; and analysis of the water table fluctuations and calculation of change in storage using water table fluctuation method.

3.5.2 Hydraulic gradient and potentiometric surface

The three year averaged potentiometric surface (Figure 42) shows a gradient from east to west. The elevation of the groundwater is 44.72 m at the most eastern point and steadily drops to zero at the coastal boundary. The gradient appears to be steeper in the southern part of the catchment, shown by the narrower contour intervals, and gentler in the north. The existence of the lakes appears to have a significant effect on the pattern of the potentiometric surface, which brings a lower groundwater elevation farther to the east. The groundwater catchment overlay shows the areas of the plains that are not included on the potentiometric map.

The calculated hydraulic gradients from monthly potentiometric surfaces show similar trends to the fluctuations which occur in groundwater well data with a seasonal influence in each year; a steeper gradient in the winter period and shallower gradient in the summer (Figure 43). The variation of the gradients is generally quite low with only a few sporadic fluctuations. The Waikawa catchment has the highest gradient followed by Waitarere South, Waiwiri, Ohau and Waitarere North (Table 17). The variation of hydraulic gradient along the coast is quite significant when comparing the Waikawa catchment in the south, to the Waitarere North catchment to the north.

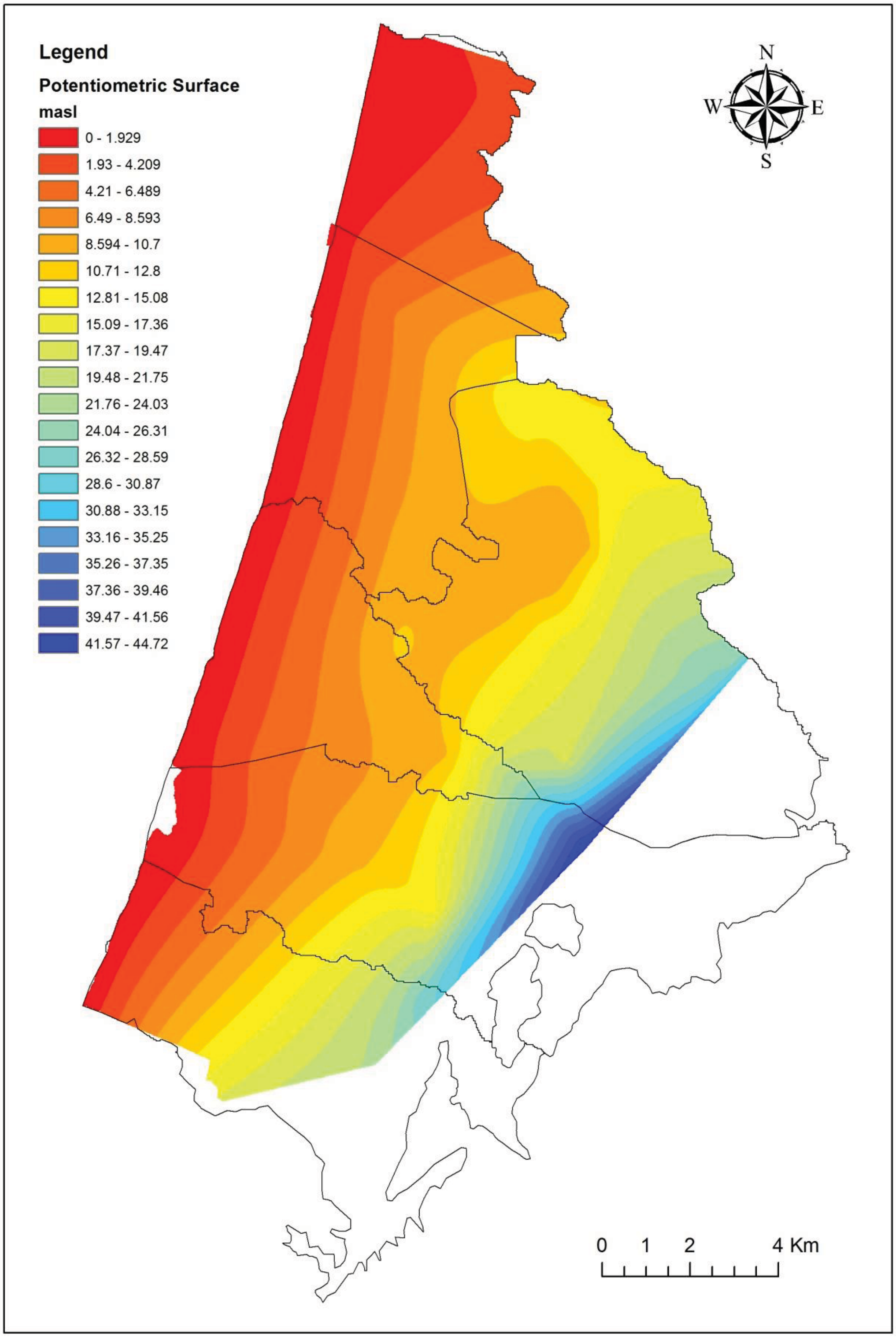

Figure 42 Averaged three year potentiometric surface for Horowhenua with overlaid sub-catchments.

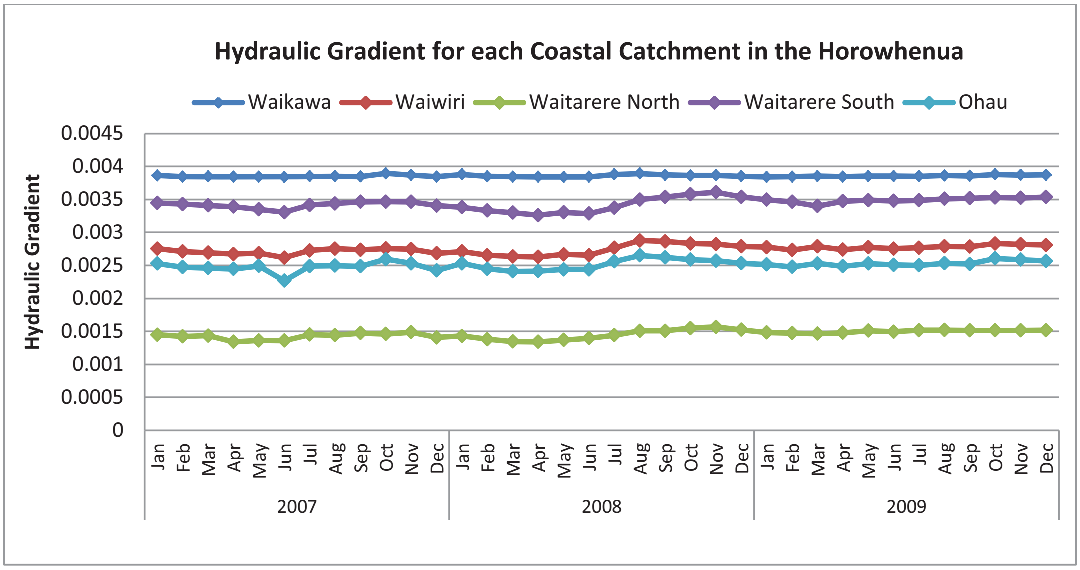

Figure 43 Hydraulic gradient for each of the coastal catchments at a monthly scale (Table shown in Appendix 1)

Table 17 Hydraulic gradient for each coastal catchment (from appendix 1)

Year	Average Hydraulic Gradient			Coast length (m)
	2007	2008	2009	
Waitarere North	0.0014	0.0014	0.0015	4,632.24
Waitarere South	0.0034	0.0034	0.0035	6,625.61
Waiwiri	0.0027	0.0027	0.0028	6,302.12
Ohau	0.0025	0.0025	0.0025	2,166.10
Waikawa	0.0039	0.0039	0.0039	3,573.75

3.5.3 Groundwater outflow

Total annual outflow volumes are highest for 2009 with 16.9 Mcm followed by 2008 and 2007 with 16.7 Mcm and 16.5 Mcm respectively (Table 18). The outflows for each month show a seasonal trend, with higher loss in general over the winter months from July to November (Figure 44). The year 2007 has an unusually low outflow in the month of June but a relatively high January outflow. The 2008 year has a significant increase around June; the previous months had the lowest outflow out of the three years tested, and then increases to have the highest outflow for the remainder of the year. The 2009 year shows a steadily increasing outflow through the year.

Table 18 Average daily outflow per month to coast for 2007, 2008 and 2009

Year	Daily Outflow to coast per month (Mcm)												
	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec	Total
2007	0.046	0.045	0.045	0.045	0.045	0.044	0.045	0.046	0.046	0.046	0.046	0.045	**16.5**
2008	0.045	0.044	0.044	0.044	0.044	0.044	0.046	0.047	0.047	0.047	0.047	0.047	**16.7**
2009	0.046	0.046	0.046	0.046	0.046	0.046	0.046	0.046	0.046	0.047	0.047	0.047	**16.9**

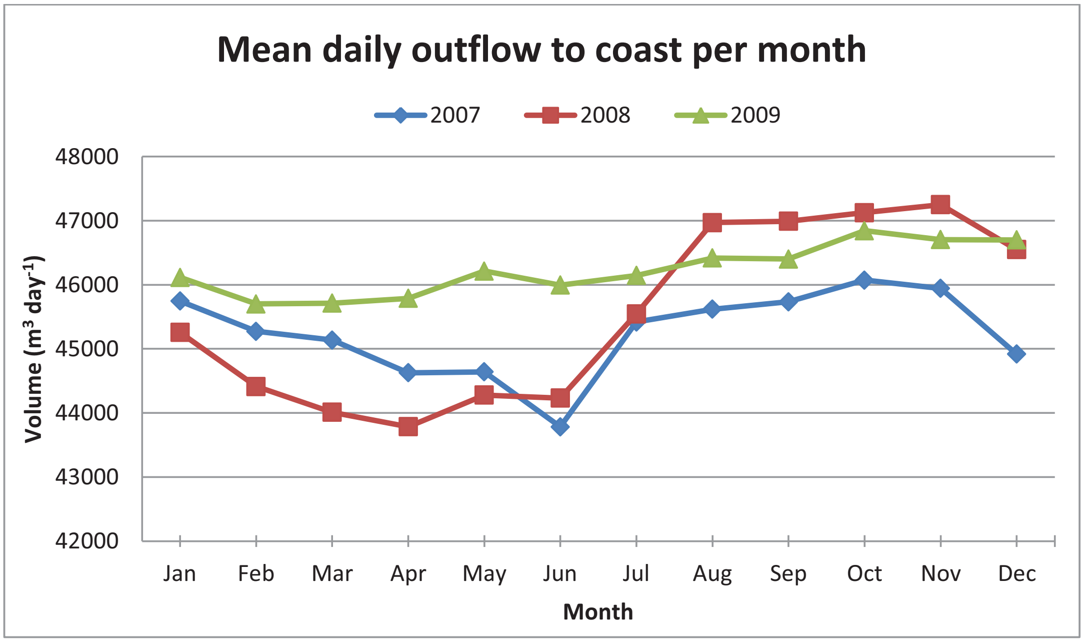

Figure 44 Average daily outflow to the coast for each month for 2007, 2008 and 2009

3.5.4 Groundwater level analysis and water table fluctuation (WTF)

Groundwater table

The average groundwater level for the Horowhenua plains increases from 2007 to 2009 (Figure 45). The winter months of each year are noticeably higher than the summer months. Each year shows a different degree of variability; 2008 shows the greatest variability, with the lowest summer levels followed by the highest winter levels. The lowest winter water table levels are shown in 2007 and the higher summer water levels are displayed in 2009 (Figure 45). Groundwater fluctuation (Figure 45) shows a similarity to the rainfall patterns that are observed (Figure 24). This is emphasised by the 3 point moving average trendline showing smoothed data where the summer and winter trends are exhibited in each. Specifically, the highest water table in winter and lowest water table in summer of 2008 correlating well with the dry summer and wet winter observed from rainfall analysis.

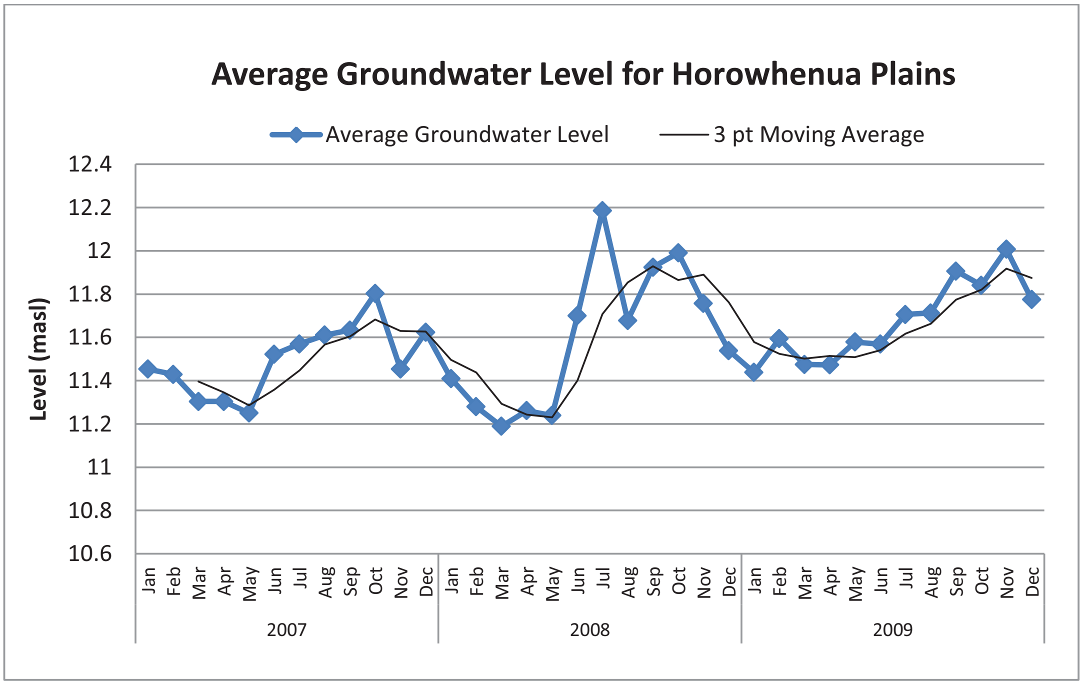

Figure 45 Averaged groundwater level for Horowhenua with 3 point moving mean

The groundwater levels show monthly fluctuations and the magnitude of variation differs from well to well (Figure 46). The shallower wells, located in the coastal zone with groundwater levels of ≈ 5 masl and ≈ 15 masl show little monthly fluctuation in level and are shown to have a standard deviation of ≈ 0.2 m (Figure 47). Wells with groundwater levels between ≈ 15 masl and ≈ 30 masl tend to be east of the lakes and show a greater degree of variation which occurs monthly with standard deviations around ≈ 1 m (Figure 47) and individual fluctuations being as much as 5 m (Figure 46).

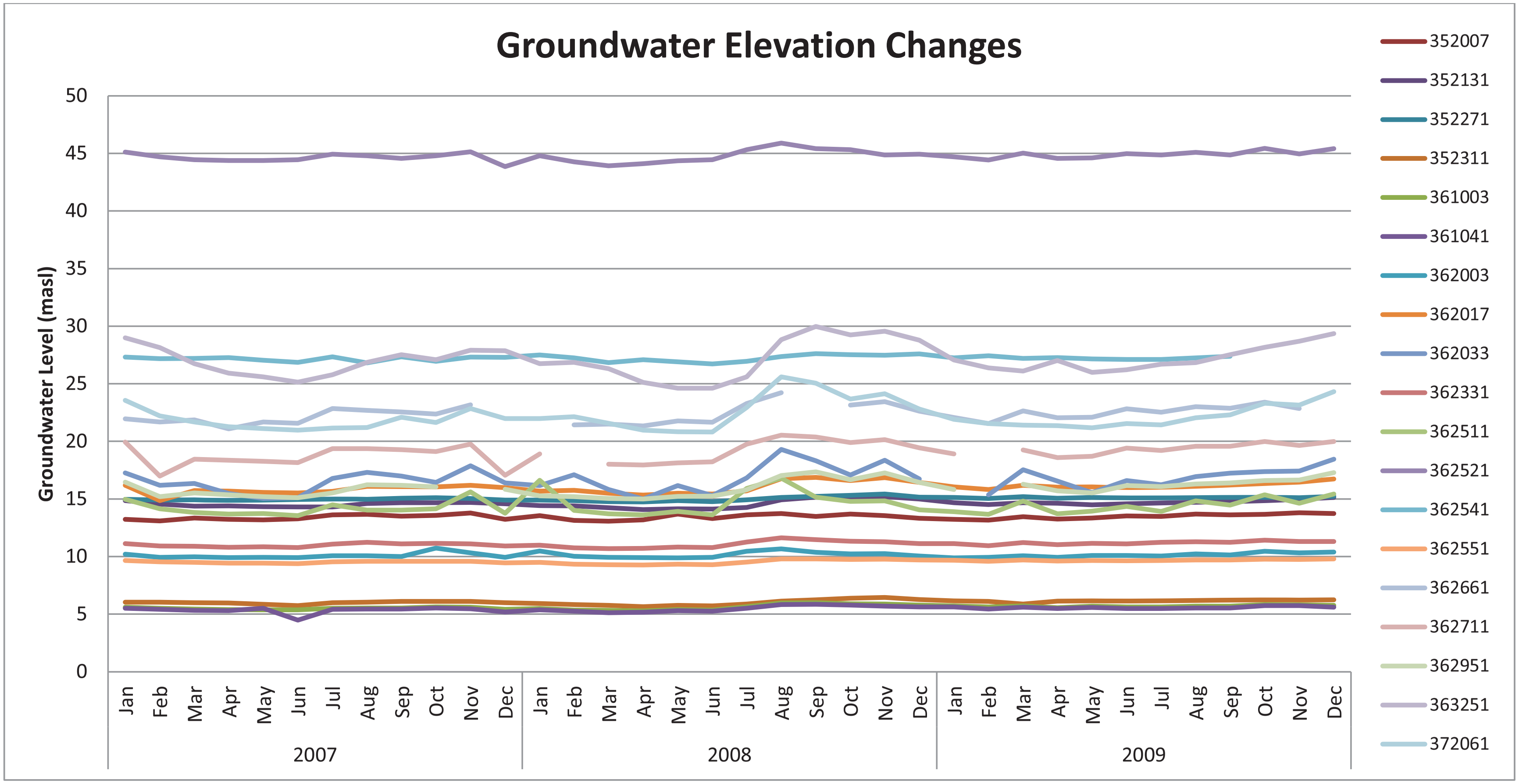

Figure 46 Monthly groundwater level changes for the monitored wells in Horowhenua (Figure 20) from 2007 to 2009

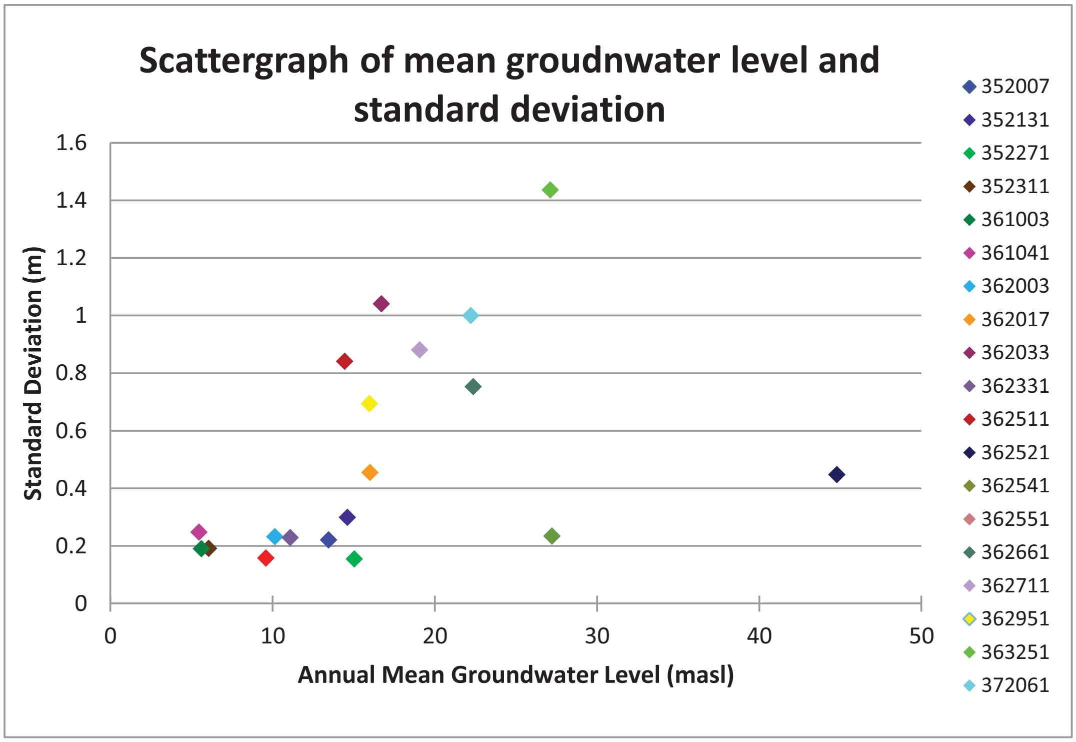

Figure 47 Scatter graph between mean and standard deviation of the groundwater level (2007 – 2009) for the monitored wells in Horowhenua (Figure 20)

Water table fluctuation

The water table fluctuation (WTF) method of differenced monthly potentiometric surfaces give a spatially sensitive indication of the change occurring in groundwater volume. High variability is shown between each month (Figure 48) which can be related to the averaged groundwater level changes (Figure 45). Significant increase in groundwater volume is seen in June and July of 2008 (+105.1 and +110.9 Mcm respectively) followed by a significant loss of groundwater volume (-115.7); this pattern is also seen in Figure 45. Summer months show a generally negative change except for February 2009 which displays a positive change coherent with a rainfall event. Positive change in groundwater volume is mostly observed in winter months (Figure 48), with the notable outlier being August 2008 with -115.7 Mcm change (Table 19). The extent of the monthly fluctuations is much more subdued in 2007 and 2008 compared to 2009. Monthly variations are significantly larger (positive or negative) than net annual change for example between January and November 2007 groundwater volumes were double the decrease experienced for the net change (Table 19). Total change in storage volumes for 2007, 2008 and 2009 are estimated at -35.3 Mcm, -19.3 Mcm and 53.8 Mcm respectively (Table 19).

Table 19 Monthly changes in groundwater volume for 2007, 2008 and 2009

Year	Monthly changes in groundwater volume (Mcm)												
	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec	Total
2007	-74.1	-5.7	-28.5	0.0	-12.2	61.8	11.0	9.4	5.3	38.5	-79.5	38.7	**-35.3**
2008	-48.8	-29.7	-20.7	16.5	-5.2	105.1	110.9	-115.7	56.1	15.2	-53.4	-49.6	**-19.3**
2009	-23.1	35.5	-27.1	-0.5	24.2	-2.3	31.3	1.5	44.2	-14.9	37.9	-52.9	**53.8**

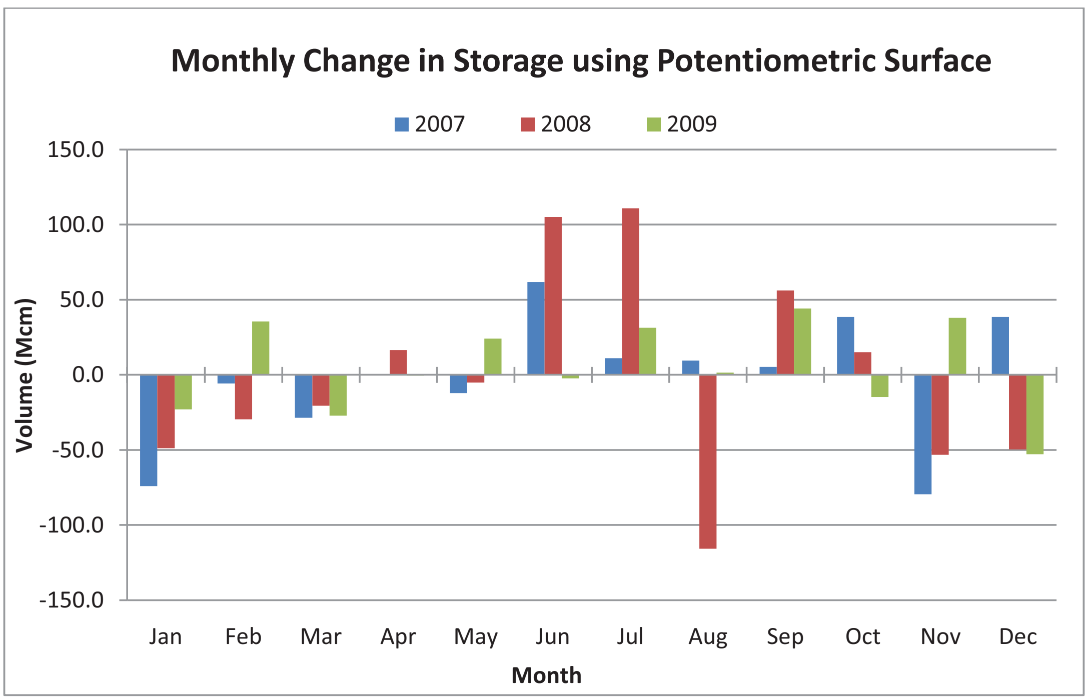

Figure 48 Monthly volumetric changes in storage for Horowhenua (Mcm)

Table 20 Annual groundwater storage change for each plain sub-catchment using water table fluctuation method

Sub Catchment	Change in Storage Mcm		
	2007	2008	2009
Lake Horowhenua	-35.44	8.29	31.70
Waikawa	-8.95	-4.57	16.20
Waiwiri	-0.89	1.19	-4.37
Waitarere North	-0.41	1.43	0.33
Waitarere South	-2.28	3.96	0.69
Ohau	12.69	-29.62	9.30

3.6 Surface water

3.6.1 Introduction

Results for surface water outflow give indications of surface outflow from the hill slopes into the plains and show two estimated coastal surface outflows reported in three sections; measured channelled flow including extrapolated Ohau flow; surface outflow to the coast using stream flow extrapolation; and stream outflow to the coast using the water table fluctuation change in storage value inputted in the water budget equation.

3.6.2 Surface channel flow

Ohau flow extrapolation

A scatter graph of the Ohau river flow with the flows from the Manakau and Waikawa rivers revealed the Waikawa catchment to have a higher correlation with an R^2 value of 0.84 (Figure 50) compared with 0.70 (Figure 49) from Manakau flow. Thus, the Waikawa was used for extrapolating the Ohau river values using the trendline equation (Figure 50).

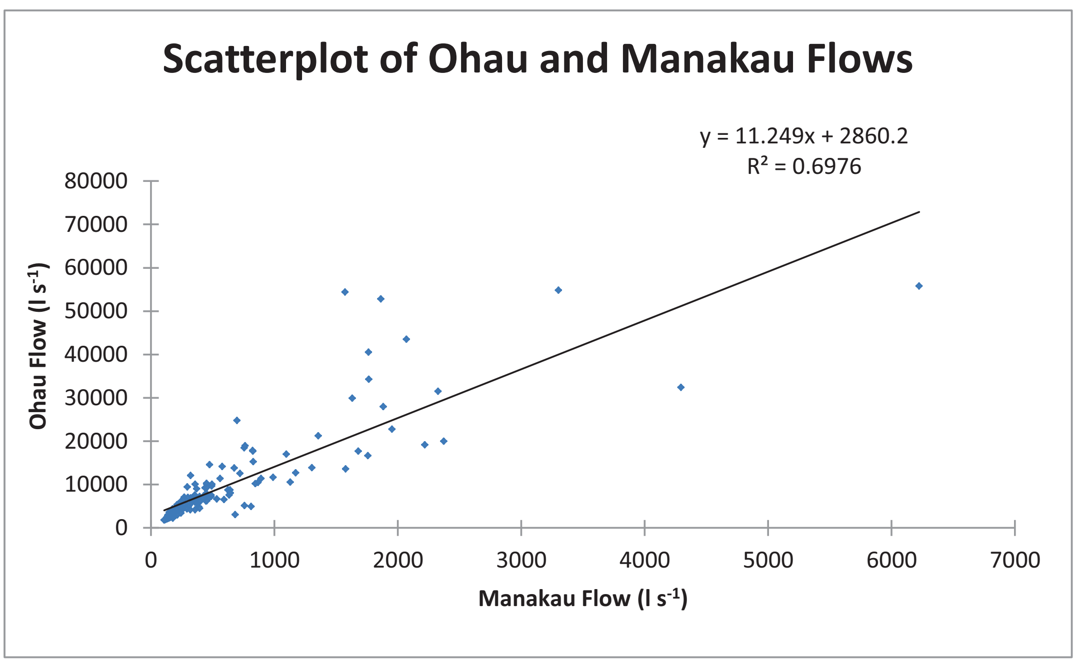

Figure 49 Scatterplot correlation of Ohau River and Manakau Stream flows showing line of best fit, R^2 value and equation

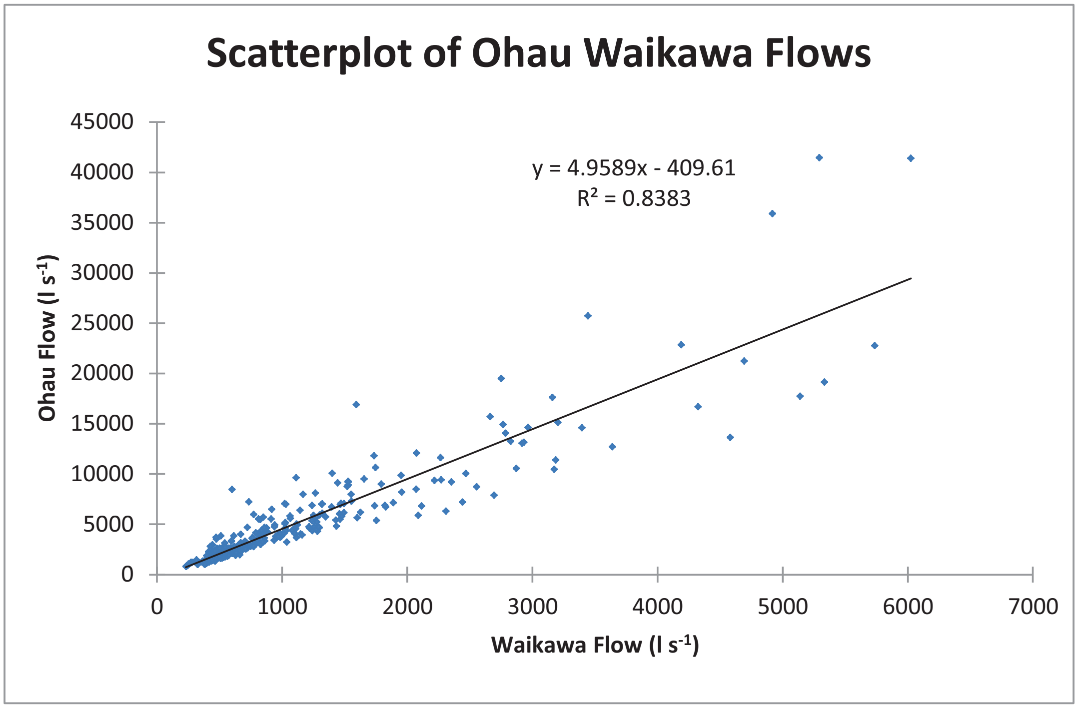

Figure 50 Scatterplot correlation of Ohau River and Waikawa stream flows showing line of best fit, R^2 value and equation.

Channel flow at gauging stations

The Ohau River has a much greater flow than the Waikawa and Manakau streams (Figure 51). Flow was greater in the second half for the year over the winter months in 2007, 2008 and 2009 (Figure 52). 2007 experienced overall lower flows throughout the year relative to 2008 and 2009, but also more evenly distributed flow throughout the year. January 2008, uncharacteristically, had a very high flow from all streams (Figure 51) but was relatively dry until winter where flow increased substantially (Figure 53). 2009 had significant increase in flow in autumn and unusually high flows in early summer (Figure 52 and Figure 53). Total stream flow at the gauging stations was calculated as 189.6 Mcm in 2007, 296.5 Mcm for 2008 and 249.6 Mcm in 2009 (Table 21). The Hokio stream outflow from Lake Horowhenua is estimated at 35.3 Mcm per year based on an average flow of 1.12 $m^3 s^{-1}$, giving adjusted total gauged flows of 224.93 Mcm, 331.83 Mcm, and 284.93 Mcm for 2007, 2008 and 2009 respectively.

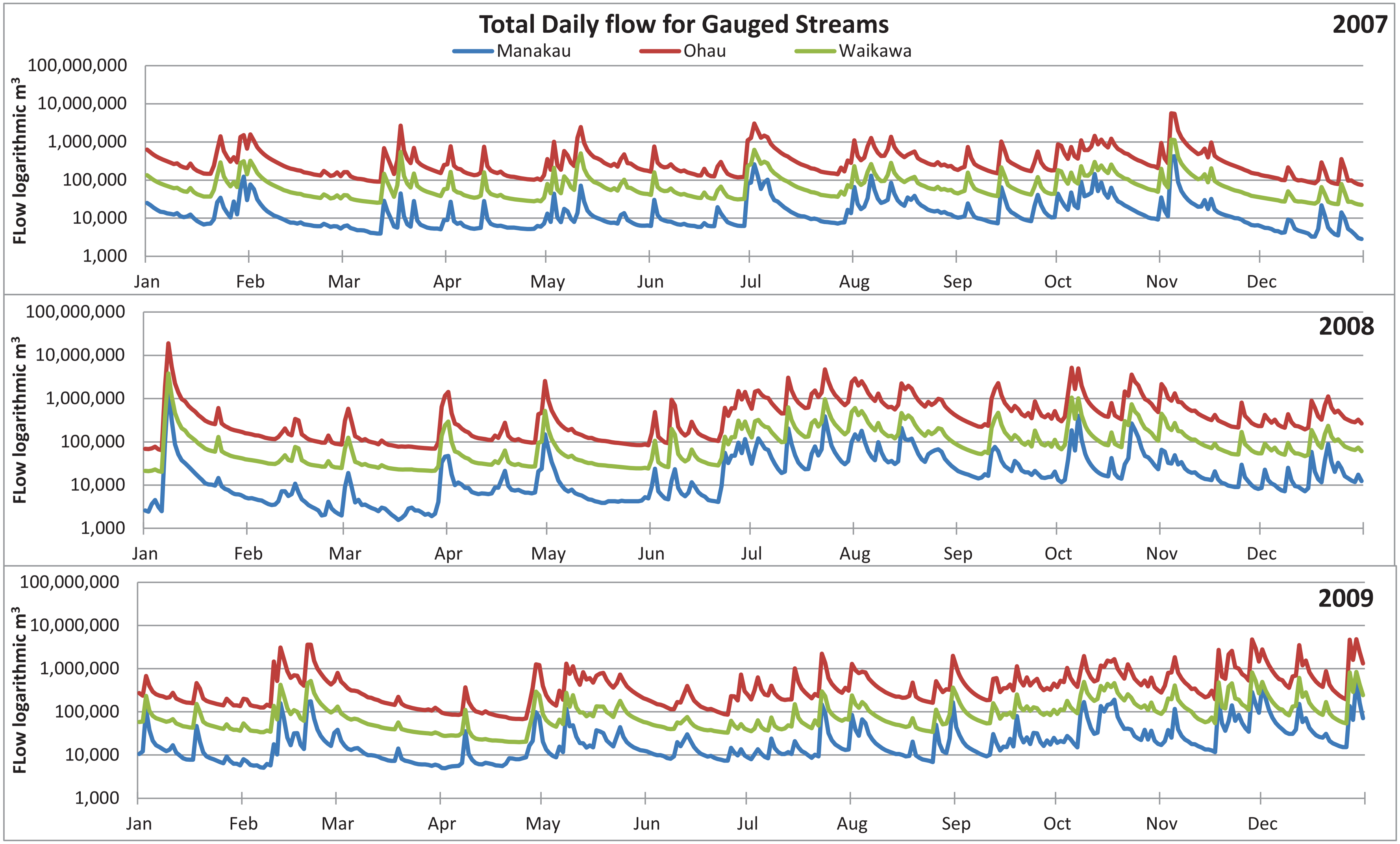

Figure 51 Logarithmic stream flows at gauging stations of Ohau River, Waikawa Stream and Manakau stream comparing 2007 (top), 2008 (middle) and 2009 (bottom)

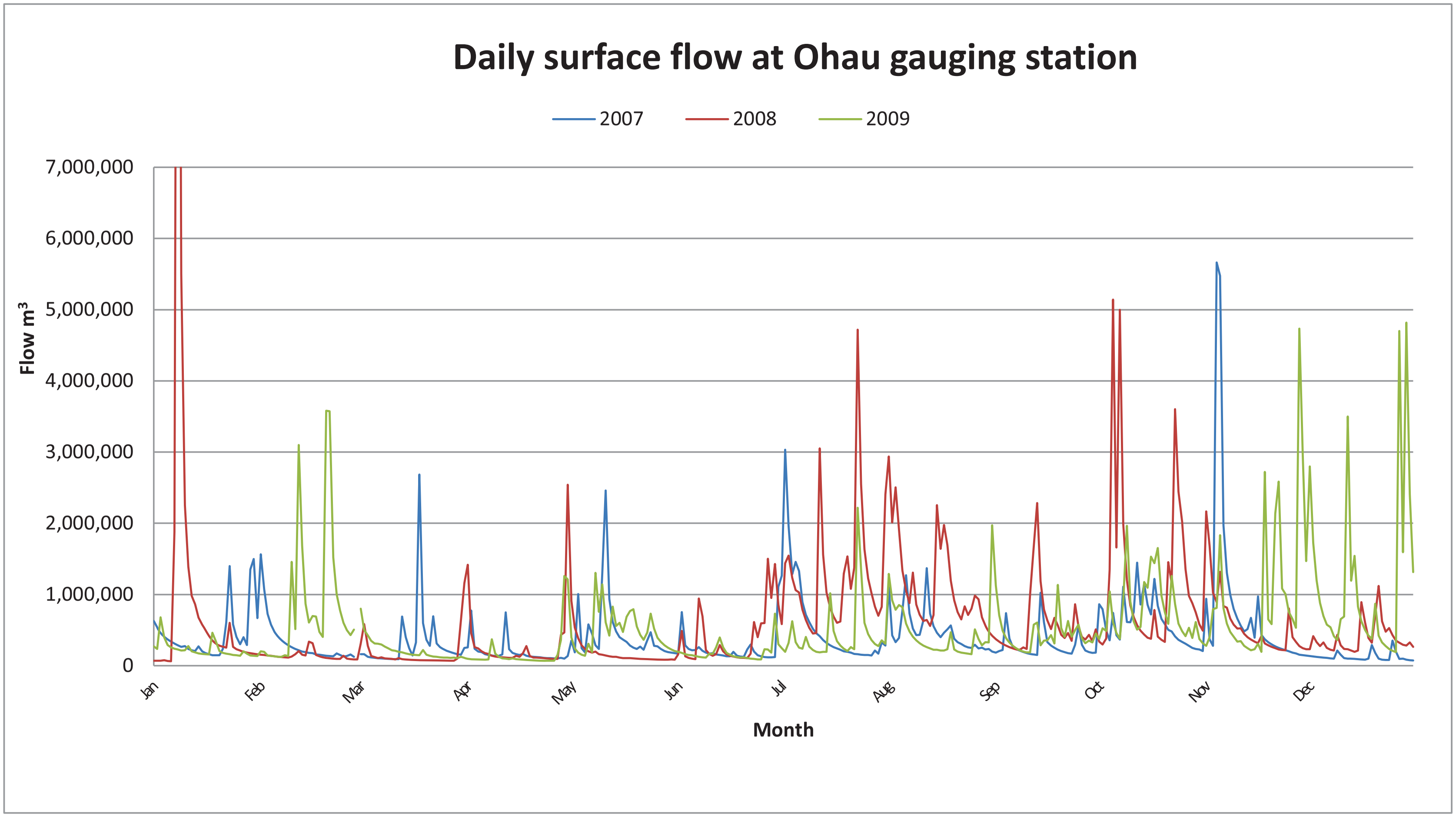

Figure 52 Daily flow for Ohau River comparing 2007, 2008 and 2009

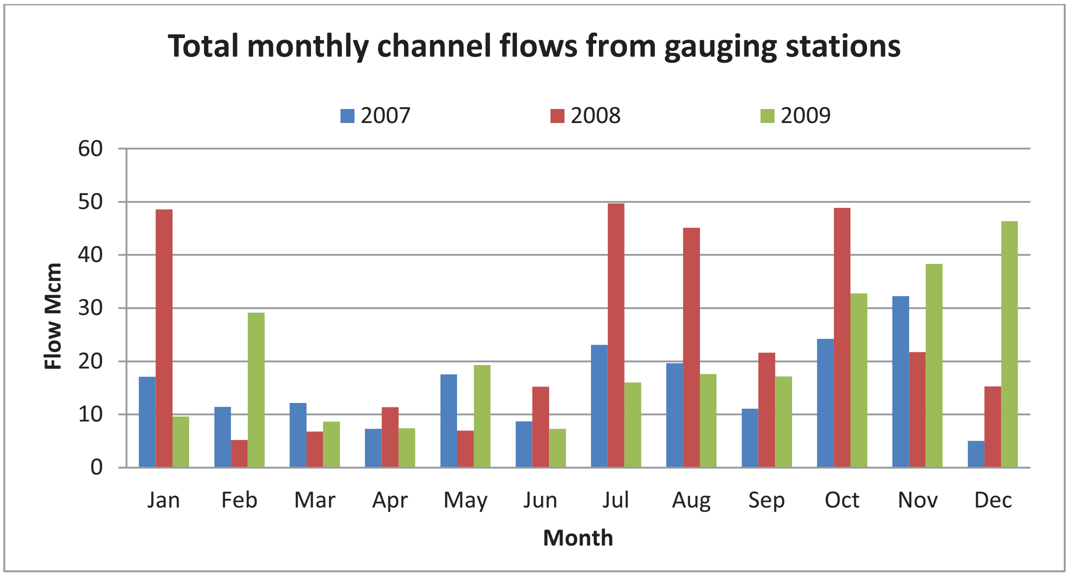

Figure 53 Total monthly channel flows from all gauging stations for 2007, 2008 and 2009

Table 21 Monthly and annual surface channel flow for Manakau Stream, Waikawa Stream and Ohau River

Catchment	Volume (Mcm)												
	2007												
	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec	Total
Man	0.6	0.4	0.3	0.2	0.4	0.3	1.1	1.0	0.5	1.1	1.4	0.2	**7.5**
Ohau	13.5	9.0	9.7	5.7	14.1	6.8	18.1	15.3	8.7	19.1	25.5	3.9	**149.2**
Wai	3.0	2.0	2.2	1.4	3.1	1.6	3.9	3.3	2.0	4.1	5.4	1.0	**32.8**
Tot	**17.1**	**11.4**	**12.2**	**7.3**	**17.5**	**8.7**	**23.1**	**19.6**	**11.1**	**24.2**	**32.2**	**5.0**	**189.6**
	2008												
Man	2.6	0.1	0.2	0.5	0.3	0.7	2.2	2.2	0.7	2.0	0.6	0.6	**12.8**
Ohau	38.0	4.0	5.3	8.9	5.4	11.9	39.3	35.5	17.2	38.8	17.4	12.0	**233.7**
Wai	7.9	1.0	1.3	2.0	1.3	2.6	8.2	7.4	3.7	8.1	3.7	2.7	**49.9**
Tot	**48.6**	**5.2**	**6.8**	**11.4**	**7.0**	**15.2**	**49.7**	**45.1**	**21.6**	**48.9**	**21.7**	**15.3**	**296.5**
	2009												
Man	0.5	1.0	0.3	0.4	0.8	0.4	0.6	0.8	0.7	1.8	1.8	2.6	**11.6**
Ohau	7.1	24.3	6.8	5.6	15.3	5.5	13.2	14.1	13.3	24.4	30.4	36.9	**196.9**
Wai	2.0	3.8	1.6	1.4	3.2	1.4	2.2	2.7	3.2	6.6	6.1	6.9	**41.1**
Total	**9.6**	**29.2**	**8.7**	**7.4**	**19.3**	**7.3**	**16.0**	**17.6**	**17.2**	**32.8**	**38.3**	**46.4**	**249.6**

3.6.3 Surface Outflow from Stream Extrapolation

The method used for the estimation of runoff shows a negative error bias between observed (gauged) and estimated values for most of the catchment calculations except 2008 and 2009 Manakau catchments (Table 22). Applying the estimated bias error (Equation 27) to the whole catchment results in a coastal outflow of 250.0 Mcm, 419.7 Mcm, and 340.2 Mcm for 2007, 2008 and 2009 respectively (Table 23).

Table 22 Components of error calculation for each gauged upstream catchment where P = precipitation; E = Evapotranspiration; Qm = measured catchment runoff; Err = error term for each catchment

Year	Catchment from gauging	Values for error calculation			
		P(Mcm)	E(Mcm)	Qm(Mcm)	Err
2007	Ohau	191.81	75.66	149.25	-0.147
	Manakau	16.12	9.44	7.47	-0.046
	Waikawa	43.29	21.18	32.84	-0.199
2008	Ohau	258.72	74.23	233.65	-0.160
	Manakau	24.59	8.78	12.82	0.138
	Waikawa	61.80	20.19	49.90	-0.118
2009	Ohau	259.12	68.88	196.65	-0.024
	Manakau	23.89	9.20	11.58	0.150
	Waikawa	61.20	20.28	41.01	-0.001

Table 23 Components of outflow calculation for total catchments using error values derived from Table 22 where P = precipitation; E = evapotranspiration, and Q^* = corrected runoff

Year	Catchment	Values for catchment flow calculation				
		P (Mcm)	E (Mcm)	Err	Q*(Mcm)	Q* total (Mcm)
2007	Ohau	273.5	123.1	-0.147	197.6	250.0
	Waikawa/Manakau	87.3	50.4	-0.150	52.4	
2008	Ohau	379.2	118.2	-0.160	333.1	419.7
	Waikawa/Manakau	128.7	46.9	-0.036	86.6	
2009	Ohau	374.1	114.3	-0.024	269.1	340.2
	Waikawa/Manakau	125.3	48.6	0.047	71.1	

3.6.4 Surface outflow from Change in Storage

The surface water outflow estimated from the water budget equation and water table fluctuation (WTF) method give values of 248.2 Mcm, 445.8 Mcm and 352.8 Mcm for 2007, 2008 and 2009 respectively. This represents considerable variability in surface water outflow from year to year; with 2008 almost double the outflow of 2007. The values are not dissimilar from estimated stream flow values calculated above. The 2007 WTF values give a very similar outflow component by 1.77 Mcm, whereas the 2008 values give a higher outflow component by 26.05 Mcm and 2009 varying by 12.56 Mcm (Table 23 & Table 24).

Table 24 (Method 1) Water budget components using Water Table Fluctuation (WTF) method where; P = precipitation; E = evapotranspiration; ΔS = change in storage; G_{out} = groundwater outflow; C = groundwater consumption; S_{out} = surface water outflow (runoff)

Year	Components of Water budget (Mcm)				
	P	E	ΔS	G_{out}	S_{out}
2007	483.8	254.3	-35.3	16.5	248.2
2008	682.9	239.8	-19.3	16.7	445.8
2009	663.7	240.2	53.8	16.9	352.8

3.1 Water Budget calculation

The water budget calculation results show significant differences in change in storage values between years as well as between the two calculation methods. Both methods show a negative recharge occurring in 2007. Water table fluctuation calculation (method 1) gives ΔS values of -35.3 Mcm for 2007, -19.3 Mcm for 2008 and 53.8 Mcm for 2009 (Table 24). Method 2, using stream flow extrapolation plus Hokio stream outflow, gives values of -72.4 Mcm, - 28.6 Mcm, and 31.1 Mcm for years 2007, 2008 and 2009 (Table 25). They all have the same order with 2007 being the largest deficit, and 2009 having positive ΔS. Method 2 shows overall a greater surface outflow which result in the ΔS decreasing (Table 26). A more preferred value for this study possibly lies somewhere in between these values. This particularly assumes that one method overestimates and the other underestimates change in storage values. Therefore a mean of method 1 and method 2 gives values of -53.8 Mcm, -23.9 Mcm, and 42.5 Mcm for 2007, 2008 and 2009 respectively (Table 26). Monthly change in storage values using method 2 (Table 26) show overall lower net monthly volumes (Table 27 and **Error! Reference source not found.**) compared to monthly volumes encountered from the water table fluctuation volumes (
Table 19). Negative volumes are shown for all 2007 and all months in 2008 except April, July and August. Positive values are shown for all 2009 months except for February.

Table 25 (Method 2) Components of water budget calculation using Woods et al. (2006) stream flow calculation plus Hokio and Waiwiri stream outflow

Year	Components (Mcm)				
	P	E	ΔS	G_{out}	S_{out}
2007	483.8	254.3	-72.35	16.51	285.83
2008	682.9	239.8	-28.55	16.67	455.63
2009	663.7	240.2	31.13	16.87	376.13

P = precipitation; E = evapotranspiration; ΔS = change in storage; G_{out} = groundwater outflow; C = groundwater consumption; S_{out} = surface water outflow (runoff)

Table 26 Water budget component showing the two different values for streamflow and change in storage as well as the averaged preferred values

	Methods (Mcm)					
Year	Water Table Fluctuation Method (1)		Woods *et al.* (2006) + Hokio and Waiwiri outflow Method (2)		Preferred value mean	
	S_{out}	ΔS	S_{out}	ΔS	S_{out}	ΔS
2007	248.2	-35.3	285.8	-72.35	267.0	-53.8
2008	445.8	-19.3	455.6	-28.55	450.7	-23.9
2009	352.8	53.8	376.1	31.13	364.5	42.5

S_{out} = surface outflow; and ΔS = change in storage

Table 27 Monthly change in storage (ΔS) volumes based on surface outflow using Woods et al. (2006) plus Hokio and Waiwiri stream outflow

	Volume (Mcm)												
ΔS	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec	Total
2007	-5.6	-6.9	-7.2	-3.8	-5.9	-5.6	-3.8	-5.8	-5.1	-11.0	-6.0	-6.2	**-72.4**
2008	-5.2	-5.2	-2.1	0.8	-2.7	-1.9	1.3	2.2	-4.7	-1.2	-8.5	-2.0	**-28.6**
2009	5.9	-3.0	2.0	0.8	3.8	2.0	1.5	2.9	2.9	4.1	3.4	4.1	**31.1**

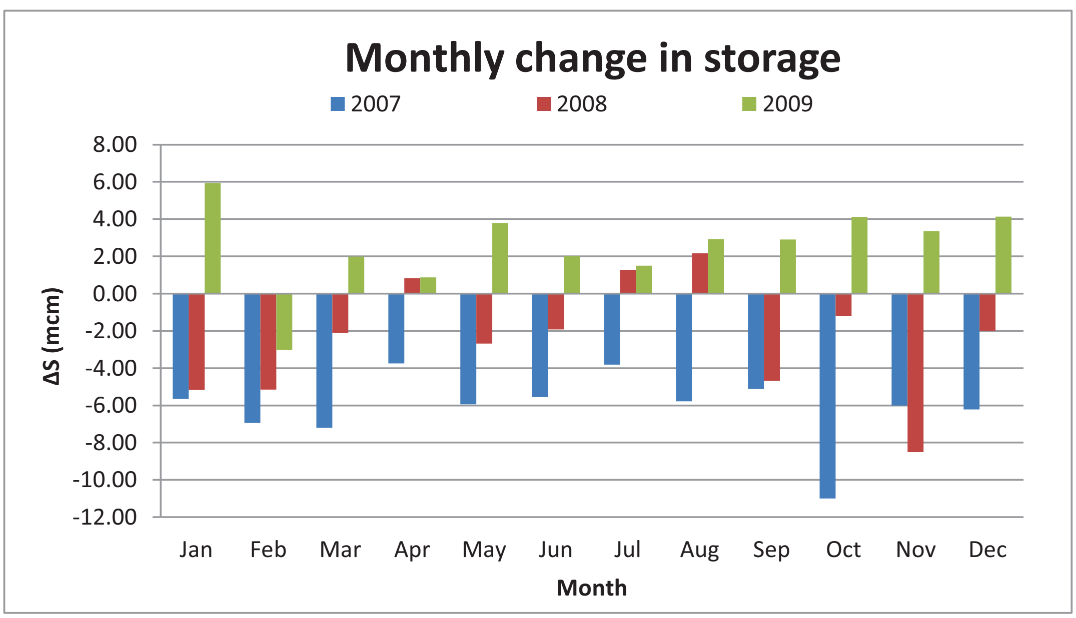

Figure 54 Monthly change in storage (ΔS) volume based on surface outflow from woods et al. (2006) plus Hokio and Waiwiri stream outflow (method 2)

Chapter 4 Validation

4.1 Introduction

The calculations, methods and model implemented in this research are susceptible to potentially large errors which are inherently difficult to quantify. Sources of error in this type of study typically lie in uncertainties in the complexity of the model-data space such as observation errors and uncertainties in land cover, soil classifications and precipitation maps (Batelaan & De Smedt, 2007). Within the constraints of the conceptual model and its resolution, the main sources for error lie in calculation simplifications and measurement errors.

4.2 Rainfall

Error in rainfall value is associated with measurement of data and from the interpolation algorithms. Measurement errors exist from missing values which impair the accuracy to measure the temporal variability. However the missing values are very few in number and only affect a couple of weather stations which is a small error considering the thousands of individual data values. The natural neighbour interpolation lessens the error due to the missing value being interpolated just the same as any other unmeasured point in the interpolated surface and on a number of occasions missing data coincided with a dry period and the value was zero anyway. However, following missing values, a few days had an uncharacteristically high rainfall value indicating a possible summation of the previous day's rainfall. This causes a concern for daily rainfall values if the said day was looked at explicitly but doesn't present a significant issue for monthly rainfall values which are collated from daily data. Greater errors are suspected to occur in the mountainous areas at higher elevations where observations can be sparse, include missing values and encounter greater extremes.

Tait & Sturman (2008) exhibit average rainfall maps over a period from 1971 to 2000. Rainfall values calculated in this study are in good accordance with their values. The pattern of rain isohyets was very similar. Rainfall in 2007 showed comparatively lower rainfall than 2008 and 2009 and was below the average values expressed by Tait & Sturman (2008) (Figure 22). The 2008 year which displayed a high amount of rainfall was above the average values shown by Tait & Sturman (2008) as too was 2009 which, although not as wet as 2008, still had above average annual rainfall. Rainfall from White et al. (2010) for total annual values also show in contrast the variability of actual individual years compared with long term averages. The 2007 year experienced 22.61 Mcm less rainfall than the average value from White et al. (2010). In the same way, 2008 and 2009 experienced higher rainfall values by 72.74 Mcm and 55.25 Mcm respectively (Table 28) which demonstrate the considerable differences that may occur from year to year.

Table 28 Comparison of sub-catchment annual rainfall with annual mean values from White et al. (2010) for Horowhenua plains.

	Rainfall Mcm			
Sub-catchment	This Study			White et al. (2010)
	2007	2008	2009	Annual Mean
Waitarere North	15.74	22.09	20.97	16.2
Waitarere South	23.34	33.18	31.29	24.9
Lake Horowhenua	60.63	86.07	80.39	63.8
Waiwiri	21.45	31.12	29.20	24.2
Ohau	53.23	79.19	74.75	61.0
Waikawa	34.80	52.89	50.44	41.7
Plains total	**209.19**	**304.54**	**287.04**	**231.8**

4.3 Evapotranspiration

As with rainfall, the main error arising for evapotranspiration is calculation and measurement error. The reference evapotranspiration is calculated from basic meteorological components. The temporal scale for the variables was of high resolution, being extracted at hourly values for atmospheric pressure, relative humidity and wind speed and extracted daily for radiation and maximum/minimum air temperatures which serves to minimize temporal scale error.

Evapotranspiration values calculated in this study are overall lower than values calculated by White et al. (2010). White et al. (2010) display two values (derived from NIWA and E_{SPASMO}) for evapotranspiration which differ by ≈ 11 % of each other, giving an uncertainty value of 5.5 %. Using this error tolerance the E_{SPASMO} value is within the error tolerance for four catchments in 2007 but not for 2008 and 2009 values (Table 29). The larger variations occur in Lake Horowhenua, Waiwiri and Ohau sub-catchments. Annual variation is expected to deviate from a long term averaged value from White et al. (2010). Considering the acceptable agreement and the rationale of the method, the evapotranspiration values are reasonable.

Table 29 Comparison of annual average evapotranspiration values from White et al. (2010) with annual evapotranspiration values for 2007, 2008 and 2009. Where, E_{NIWA} and E_{SPASMO} are two evapotranspiration values used by White et al. (2010) and ET_c is the actual evapotranspiration from this study

Groundwater sub-catchment	Evapotranspiration (Mcm) White et al. (2010)		ET_{SPASMO} - ET_C as a percentage of E_{SPASMO}		
	E_{NIWA}	E_{SPASMO}	2007	2008	2009
Waitarere North	14.4	13.4	7.7	14.8	12.6
Waitarere South	21.1	19.7	11.9	18.9	15.5
Lake Horowhenua	46.1	41.1	16.0	22.0	19.7
Waiwiri	19.4	17.8	10.8	18.5	14.2
Ohau	40.7	36.9	9.7	16.8	13.0
Waikawa	29.3	26.2	9.2	17.2	11.9

4.4 Groundwater outflow, potentiometric surface and hydraulic gradient

The potentiometric surface used in this study shows good agreement with the values and pattern presented by White et al. (2010) (Figure 42 Averaged three year potentiometric surface for Horowhenua with overlaid sub-catchments.Figure 42). The prominent patterns that correspond between the potentiometric surfaces are the pattern around Lake Horowhenua as well as the groundwater high point at the northern end of the Lake Horowhenua catchment. The area west of the lakes however appears to be rather simplified compared with the surface shown by White et al. (2010). The comparison of catchment hydraulic gradients with those of White et al. (2010) do vary quite considerably in some catchments, particularly for Waitarere South, Ohau and Waikawa where higher hydraulic conductivities were calculated in this study (Table 30). The difference is perhaps due to the use of additional data by White et al. (2010) in potentiometric surface construction, such as surface water measurements, other groundwater measurements beside the monitoring wells, and piezometers in the Waikawa catchment. The hydraulic gradient values appear conceptually logical considering topography and the distance between hillslopes and the coast, shorter distances between the coast and the foothills (as seen in the southern catchments) are expected to have a steeper gradient.

Table 30 Comparison of average annual sub-catchment hydraulic gradient calculated in this study for 2007, 2008 and 2009 with the long term average from White et al. (2010)

Year	Average annual sub-catchment hydraulic gradient			
	2007	2008	2009	White et al. (2010)
Waitarere North	0.0014	0.0014	0.0015	0.002
Waitarere South	0.0034	0.0034	0.0035	0.002
Waiwiri	0.0027	0.0027	0.0028	0.003
Ohau	0.0025	0.0025	0.0025	0.001
Waikawa	0.0039	0.0039	0.0039	0.001

The groundwater outflow values calculated are subject to high uncertainty given by the range of possible hydraulic conductivities that could be assigned to the heterogeneous sand substrate of the coast and to a lesser extent the cross sectional area of the aquifer. The uncertainty is not expected to extend by orders of magnitude and makes up a small component of the water budget calculation even with generous cross sectional area and hydraulic conductivity for sand. Annual groundwater outflows calculated by White et al. (2010) are slightly lower, calculated as 12.8 Mcm using a 40 m cross section. However, this figure is scaled up significantly to equate water budget to a steady state condition ranging in value from 41.7 Mcm to 70 Mcm for groundwater outflow via the coast.

4.5 Surface water outflow

The values for the gauged stream catchments show similar values to the excess water associated with the hillslope catchments that are upstream of the gauging stations (Figure 21). With the excess water representing 83.1 %, 89.3 %, and 107.0 % for 2007, 2008 and 2009 respectively. However for 2007 and 2008 it goes against expectation to have the stream flow out of the sub-catchment to be higher than the excess water value. This would point towards either an overestimation of evapotranspiration, underestimation of rainfall values or overestimation of stream flow. Overestimation of evapotranspiration is not likely since the comparison of evapotranspiration values with those of White et al. (2010) showed a probable underestimation of evapotranspiration values. Underestimation of the rainfall values could have occurred given the greater potential for error in areas of higher elevation, due to absence of weather stations and possible extreme values not been captured. Additionally, stream flow calculations for the Ohau catchment, which were estimated for 2007 and 2008, could be overestimated, particularly for larger flow volumes. Despite this discrepancy the values for stream flow appear to be in good agreement with those observed by White et al. (2010), who give multiple values for inflow into the plain sub-catchments and encounter a similar issue of a lower excess rainfall than the excess rainfall allows (Table 31). The long term averaged excess rainfall value from White et al. (2010) shows good agreement with the annual excess rainfall values for 2008 and 2009, but not 2007 which was the drier year (Table 12**).**

Table 31 Comparison of annual excess rainfall values with gauged stream flows for upstream catchments from the stream gauging locations in 2007, 2008 and 2009; and with the Ohau, Waikawa and totalled sub-catchment values in White et al. (2010) .

Variable	Annual values					
	White et al. (2010)			This study		
	Ohau	Waikawa	Total	2007	2008	2009
Surface outflow	229.4	43.8	273.2	189.6	296.5	249.6
Excess rainfall	192.5	41.9	234.4	156.5	263.1	266.5

Difference	-36.9	-1.9	-38.8	-33.1	-33.4	16.9
% Difference	-19.2	-4.5	-16.6	-21.2	-12.7	6.3

4.6 Change in storage

The two methods used for monthly change in storage show a contrast of values. The water budget calculated volumes using method 2 are low, whereas the values from the WTF (method 1) method are much larger. The differences are likely due to the types of measurements that each of them portray and to a lesser extent, the timing of the measurements and the associated errors from the values. In the second method values are based on the water budget for the month which would be the net change for that month, however, this would have smoothing effect on any extremes that take place throughout the month. The WTF method from differenced potentiometric surfaces is based on monthly point measurements of groundwater elevation. This can potentially capture quite varied data, depending on the timing of the two dates compared to occurrence of water table fluctuations. The dates of the groundwater level data do not align directly with the start and end of each calendar month, so the time period may straddle two months; hence the assigned monthly label is not strictly appropriate when used in comparison with the other variables. The large values do however show the extent of the larger rainfall events, notably the 2008 June – August rainfall maximums (Figure 23) which are reflected in the storage values quite clearly with ≈ 100 Mcm increases followed by a ≈100 Mcm decrease (Figure 48).

Chapter 5 Discussion

5.1 Introduction

This research has undertaken a water budget calculation for the Horowhenua region, analysing the main components over three years. This provides both quantitative and qualitative outcomes that give greater understanding of what is occurring with the geohydrology of Horowhenua. The aspects of the discussion apply to several areas of groundwater research and for purposes of clarity have been divided into two sections. The first section focuses on the patterns, relationships, trends and generalizations from the water budget calculation that contribute to geo-hydrological understanding of groundwater recharge, discharge and change in storage. The second section pertains to an evaluation and implications arising from the spatial and temporal GIS technique used and considerations needed for improvement.

5.2 Geo-hydrological understanding

The water budget equation shows annually that there is a high degree of variability in the recharge that occurs. In this study an estimated change in storage of -53.8 Mcm, -23.9 Mcm, and 42.5 Mcm for 2007, 2008 and 2009 is observed. This shows that individual years do not experience a net storage change of zero, nor are uniformly positive or negative in net storage change. These are considerable volume changes given for the size of the Horowhenua water catchment making up - 12.0 %, - 3.7 % and + 6.8 % of the total rainfall for 2007, 2008 and 2009 respectively. At the start of 2009 the storage was 77.7 Mcm less than at the start of 2007 (Table 26) and could have considerable implications for management of the resources which are based around mean groundwater reservoir capacity and traditionally averaged recharge values. A steady state system is often observed for annual studies where change in storage is assumed to be zero. This may be valid for long term analysis, management strategies and getting a crude understanding of key components, but the appropriateness for in-depth studies on more immediate time scales is questionable. The annual variability in groundwater storage fluctuation needs to be acknowledged as being significant in value and consequential to any management planning around groundwater resource allocation. The long term stability of the water resource or long term trends cannot be evaluated from just three years of data especially with the disconnection of values between the sequential years.

Variability is shown further at a monthly time scale of storage change (Figure 48) where monthly values fluctuate considerably, which is not shown in annual values. The annual net change in groundwater storage does not indicate this variability because both positive and negative change can occur between data sampling and effectively cancel each other out. The magnitude of the change in storage in some months is very much greater in volume than the net storage change occurring annually and can be of the opposite value (Table 19). Again this is of considerable importance due to the nature in which management of groundwater resources is often implemented. If negative change in groundwater volumes can occur in one month that is greater in volume than the volume of net groundwater volume change in an annual year or over an averaged long term change, then available water for the specified month could fall short of the required needs and speculated availability of the groundwater for that period. Therefore the importance of the time period between data observations is imperative in properly evaluating the full spread of volume fluctuations and effects of individual events on groundwater reservoir behaviour. In this study the groundwater level data were provided at monthly scale which put a limitation on evaluating the relationship between the recharge events and the response time of the groundwater reservoir to rainfall events.

This variability of the storage values can be predominantly attributed to rainfall which is the only water input into the Horowhenua catchment due to accepted no-flow boundaries. High variability in rainfall (Figure 23), which is echoed in the groundwater level fluctuations (Figure 45), represents a very important component in the nature of the recharge source. The importance of the rainfall quantity is seen visibly with the total annual rainfall for 2007, which only has 69.2 % and 72.0 % of the total rainfall in 2008 and 2009 respectively (Table 12). Subsequently 2007 has the largest negative change in annual storage of -53.8 Mcm (Table 26). This is 225.1 % of 2008 change in storage and -79.0 % of 2009 change in storage (Table 26). It is unfortunate that daily groundwater data was not available for the three year period to accurately calculate rainfall-groundwater response times. The relationship cannot be appraised to accuracies above a month and most likely only exhibits responses to 'rainfall clusters' as described by Wu et al. (1996). It is likely however, that due to the shallow-intermediate groundwater table, response times are quick, and shown in the groundwater system after a short duration; this appears to be substantiated in this research. According to Wu et al. (1996) however, the 'critical interval' of time to exist between adjacent rainfall events to generate distinct peaks approaches a monthly scale when groundwater depth is greater than 6 m. It is likely that any individual rainfall events are missed entirely and even extended rainfall durations may be missed, or partially missed given Sophocleous's (1992) 5 – 7 day recharge events. Thus, if enough time with appropriate weather conditions passes before the next measurement date then the event may not be observed at all.

The overall quantity of the rainfall entering the catchment is not the only key variable to consider for change in storage values. This point is clearly illustrated in the differences between the annual rainfall quantities of 2008 and 2009. The total rainfall for 2008 and 2009 are significantly higher than 2007, with 2008 being more or less the same (2008 is 2.9 % higher) (Table 12). This would presumably result in similar change in storage values for 2008 and 2009 and furthermore the above average annual rainfall would suggest a positive change in storage to be expected. However, a positive change in storage of 42.5 Mcm occurs for 2009 yet a negative change in storage of -23.9 Mcm occurs for 2008 (Table 26) even though total 2008 and 2009 rainfall volumes are similar. An explanation for this discordance suggests that the characteristics of the rainfall must have an influence on the percentage that actually equates to recharge. Insight to this occurrence is found by looking at the stream outflow from the three gauging stations (Figure 51 and Figure 52). It is quite apparent large peak flows, over and arbitrary flood level of 3 Mcm, occurred more frequently during 2008 and were of greater magnitude. The peak flows indicate the occurrence of flood events pertaining to storms. This is verified by the very notable flood event which occurred on the 8th of

January 2008 (Figure 52), which resulted in the partial wash-out of the Kirkaldie Bridge crossing the Ohau River (Figure 55). Storms characteristically have high rainfall intensity and accordingly a high volume of rainfall enters the catchment. However the consequence of high rainfall intensity is a high rate of runoff as infiltration capacity is either exceeded and/or saturation point is reached. In addition, Spring soils tend to be more saturated and rainfall falling around this time is more likely to generate higher rates of runoff. Subsequently, higher rates of surface flow and discharge to the coast occur, resulting in a lower overall percentage reaching the groundwater reservoir. Essentially the rainfall input bypasses the catchment and heads straight to the coast where it discharges into the ocean.

The significance of a lower percentage of rainfall entering the catchment from heavy rainfall events could be amplified in the context of global climate trends. Recent work has shown a global increase in climatic extremes (Alexander, et al., 2006), possible change in precipitation patterns from widening of tropical belts (Seidel, Fu, Randel, & Reichler, 2008) and an increase in probability of intense rainfall events for many extra-tropical areas (Groisman, et al., 2005). If storms become more frequent then the percentage of rainfall that runs off and bypasses the catchment is likely to increase, putting pressure on water resources.

Figure 55 Image of the wash-out of Kirkaldie Bridge from the 8th January 2008 storm (Kete Horowhenua, 2008)

In addition to rainfall temporal patterns, it is important to consider the spatial occurrence of the rainfall since highest intensity occurs in the upland areas (Figure 22). Around ≈ 56 % of the rainfall volume that enters the catchment does so in the upper hill country, with the area of the hill country accounting for only c.45 % (Table 12). Regardless of the discrepancy between the flow value and excess rainfall value, the relatively close values show that the majority of the rainfall does in fact flow out of the hillslope catchments and into the plains via river channels. This places importance on the upstream rainfall and hydrology patterns associated with hillslopes. The quantity of the rainfall decreases towards the coastline (Figure 22) and thus the coastal zone east of the coastal lakes is comparatively less important both as an area of groundwater recharge and as a reservoir for storing groundwater. The health of the groundwater reservoir and maintaining replenishment is dependent more so on the upper catchment areas of the plains and hillslopes. This emphasizes the importance of hill slope management and landuse in influencing the nature of the hydrological health of the area. The fraction of rainfall that bypasses the catchment in storm events discussed above is greatly affected by landuse. The effects of landuse and vegetation change are well documented for their effects on increasing or decreasing runoff and modifying catchment hydrographs in flood events e.g. Knighton (1998) and Ward & Robinson (2000), whereby runoff coefficients for forest are lower than grassland.

Lake Horowhenua is solely fed by the Lake Horowhenua and 'Hill 1' sub-catchments dominated by aggradational gravels and an absence of sub-catchment feeding from the Tararua Range. Absence of surface drainage features indicate excess rainfall infiltrates down to the water table where it flows through the porous Q2al gravels and discharges into Lake Horowhenua, rather than forming channelled surface runoff. The excess rainfall amounts to 27.1, 55.8 and 49.1 Mcm of excess rainfall for 2007, 2008 and 2009 respectively. The removal of surface water outflow via Hokio stream, estimated at 35.3 Mcm per annum, from excess rainfall, results in storage change values of -8.2, 20.5 and 13.8 Mcm for 2007, 2008 and 2009 respectively. Water table fluctuation values show different values but similar patterns of -35.4, 8.3 and 31.7 Mcm for 2007, 2008 and 2009 respectively (Table 20). Either way, the results show the impacts of variability in recharge quantity conditioning the change in storage values for the catchment, and by extension the lake behaviour. The average outflow of Hokio stream which discharges Lake Horowhenua is estimated at 35.3 Mcm per annum. Based on the estimated Hokio outflow, a minimum of 35.3 Mcm must be entering the lake on average each year. A negative storage change for the entire catchment in 2007 would suggest the required inflow is put in jeopardy, although presumably lower outflows would be experienced and in turn lower inflows would be required to maintain water levels. This is conceptually sensible when considering average lake depth is ≈ 2 m and an inflow/outflow deficit of ≈ 2.9 Mcm would result in a 1 m decrease in lake level based on average outflow. This means lake level is very sensitive to groundwater levels and has a low tolerance for fluctuations. Given the low tolerance together with a weir, which maintains a steady lake level, and the large deficit from the water table fluctuation, it is probable the replenishment of Lake Horowhenua is maintained at the expense of the groundwater storage. Thus the ability of the storage reservoir to buffer dry periods becomes very important for maintaining a healthy lake system.

Estimated groundwater consumption makes up a very small fraction of the total volume of water coming into the catchment, ranging from c.1.7 to c.2.5 % of total rainfall. Although this is a small fraction in terms of the total water volumes in the Horowhenua catchment, it could have significant effects in drier conditions when groundwater levels are at their lowest. Change in storage varies considerably annually, ranging from large negative fluctuations to equally positive changes, thus in single year; the water consumption could be the difference between a positive or negative net change. The effects of water consumption however could be somewhat negated by lowering the hydraulic gradient thereby reducing groundwater outflow following Darcy's equation. The effects would also be negated with water returned via return flow into the system. If the change in groundwater outflow from virgin state conditions is equal to water consumption values then the groundwater reservoir can remain in good health (Bredehoeft, 2002).

5.3 Critical appraisal of approach

The water budget coupled with high resolution data provides a powerful tool for groundwater resource studies. However, it is reliant on crucial aspects which need to be considered for improvement of future implementations. The quality of the data is an essential consideration for the development of this technique which required a large amount of data processing especially for the evapotranspiration component. Attention to evapotranspiration is very important, forming the largest water loss for most catchments and is a key parameter for most physical processes in the soil-crop-climate domain (Eitzingera et al. 2002). However, accurate calculation of evapotranspiration is not simple due to high temporal and spatial variations. The pattern observed for actual evapotranspiration (Figure 31) is shown to be mainly inherited from landuse (Figure 13), soils (Figure 15) and climate shown by the reference evapotranspiration (Figure 31). The resolution required is that which adequately retains the detail of the spatial features. Climate trends do not necessarily impose the need for high resolution due to the smoothing effects of interpolation, which itself is subject to greater error than resolution can correct for. Required resolution is also dependent on the range of values encountered. Thus, steep land tends to require higher resolution due to the high gradient encountered for most many climatic variables used in evapotranspiration calculations.

Landuse and vegetation have greatest spatial change with landuse varying from paddock to paddock very abruptly, a good example are orchards which are surrounded by pasture. The resolution consequently needs to be at a scale which accounts for these abrupt changes in significant landuse areas and preserves the macro pattern of landuse. This study used a resolution of 50 m, which from visual examination preserves the landuse pattern. This resolution is in theory able to encapsulate a 0.25 ha rectangular block, which seems to adequately account for most agricultural subdivisions, such as cropping land, pasture land and orchards, even some major shelter belts are able to be shown. External limitations such as computer processing capabilities may however form the restricting variable for assigning resolution in which case the catchment size will also contribute to the resolution that can be used.

The extent to which increased resolution modifies actual evapotranspiration values will vary depending on the inherent landuse characteristics of the study area which means that a 'one size fits all' approach is not appropriate. Assumptions that are made in one catchment cannot necessarily be made in other catchments. If the catchment is dominated by one or two types of landuse then increased resolution will not necessarily increase evapotranspiration accuracy significantly. This is possibly the case in Horowhenua which is dominated by forested areas on the hillslopes and grassland in the plains. The nature of the dominant species will also determine the extent to which landuse adjusts the evapotranspiration due to variation of the crop coefficient and uniformity of the landuse throughout the seasons. The dominant landuse experienced in Horowhenua varies crop coefficient values by 5 % and tends to be relatively constant in growth phase throughout the year, ignoring events such as re-sowing grass. If the landuse was dominated by orchards or any landuse which deviates significantly from the grass reference evapotranspiration then the effects of crop coefficients will be more important as will the accuracy of the landuse data. Landuses which vary significantly from season to season also need to be carefully considered, as the evapotranspiration coefficients can vary noticeably for different growth phases e.g. brassicas and other small vegetable crops. This has implications for temporal trends, particularly at monthly or shorter analysis timeframes. Thus, an understanding of the catchment landuse characteristics is necessary in order to know what assumptions are valid and how some factors are increased or lessened in importance.

The soil patterns also have an exerted effect on evapotranspiration patterns and tend to become significant in the drier months over summer limiting evapotranspiration (Figure 33, Figure 34 and Figure 35). The limitation arises from the soil moisture deficit values and the stress that this puts on vegetation when the deficit is large and limits the water that plant roots can draw up. This is also why the grassed areas in the plains are more affected than the forested areas, due to the difference in rooting depths. The relationship between the evapotranspiration pattern and soil patterns is exemplified in February 2007 (Figure 36) where soils that have lower readily available soil moisture also have lower evapotranspiration values.

Data availability is a critical factor which has limited the full effectiveness of this research; a limitation that was alluded to be Senarath (1988). This has occurred in the form of direct measurement data which are absent altogether and also for data which do not have the desired temporal scales. Surface outflow is the most significant component in this research that requires improvement to available data. Spatial distribution of surface flow is not well constrained due to an absence of gauging stations in strategic positions. A catchment scaled water budget calculation would ideally have surface flow outputs gauged from the exit points of the catchment i.e. within or near the coastal boundary zone and for each channel. Although values estimated within this study are both internally consistent and align with other research there is considerable room for improving accuracy by deploying a comprehensive network of gauging stations. Thus, gauging stations for Ohau River, Waikawa Stream and to a lesser extent the Hokio and Waiwiri Streams would have been very useful in constraining the surface outflow volumes and by extension the change in storage values. This would also provide daily values for surface outflow. Further gauging stations would be useful upstream and downstream of the gravel plains to provide data to further explore the interactions between surface water and groundwater along the river course. Alternatively, an improved and more comprehensive surface runoff model would have been useful in giving greater confidence to the surface outflow values. This would also have the advantage of giving a spatial representation of the surface runoff and could be verified with gauged stream flows. In addition, groundwater outflow to the coast needs to be explored further to identify the actual aquifer cross sectional area that is discharging to the ocean as well as improved monitoring of water consumption volumes and at a higher temporal resolution.

In comparison with other groundwater analysis techniques, the water budget has been used as a somewhat coarse method of water resource investigations due to its simplicity. However, coupled with high resolution temporal and spatial component data it has the ability to be much more powerful and capable of exploring many groundwater phenomena. This is important given that many groundwater techniques such as lysimeters and isotopic tracers are expensive (Lerner, et al., 1990) whereas this technique remains inexpensive to implement pending data collection systems are already in place. When using daily information it has the ability to be executed iteratively as Batelaan & De Smedt (2007) showed and also scaled up to any time period that is desired.

Chapter 6 Conclusion

This research implemented a GIS based high resolution spatial and temporal water budget in Horowhenua, New Zealand. Rainfall volume varied annually with 2008 and 2009 experiencing considerably more rainfall than 2007 as well as high variability experienced seasonally. Seasonal trends of evapotranspiration were relatively stable and total evapotranspiration volumes were relatively consistent annually. High or low annual rainfall volumes didn't define positive or negative storage change. Spatial distribution and intensity have an important impact on whether positive or negative storage change will occur. This was explained by the presence of storms where high rainfall intensity resulted in substantial surface runoff and stream outflow allowing water to effectively bypass the catchment, producing larger rainfall volumes, but less rainfall that is available for recharge.

Annual groundwater storage change was not steady state in an independent year, but varied considerably between each of the studied years as well as experiencing significant monthly storage change which sometimes was of greater magnitude than the net annual change. This is important for groundwater management where groundwater storage deficits in a dry month may exceed annually derived values. Lake levels for Lake Horowhenua have a greater susceptibility to groundwater storage deficits in drier phases due to being fed only by the 'Lake Horowhenua' catchment which has a lack of upstream catchment from Tararua Range. The capacity of the reservoir to buffer drier periods may be compromised by lowering of the water table.

Although water budget components are agreeable, uncertainty remains over some components due to absence of appropriate data which impeded the extent to which daily analysis could be achieved. This pertains to surface water outflow where gauging stations were not in appropriate positions to measure outflow of the Horowhenua catchment. Increased gauging stations in coastal locations would provide much improved accuracy in constraining this significant variable and improving the reliability of the findings. Alternatively, a suitably accurate surface runoff model would also provide increased accuracy to surface outflow values with the benefit of spatial representation. Increased monitoring of groundwater consumption and daily well data information would provide additional increases to validity of the results and help to draw better-quality relationships between rainfall and groundwater storage responses.

This implementation of this technique has shown significant promise in producing beneficial research and, pending additional data, has potential to become a very robust technique for budgeting catchment water resources. The technique also has the prospect of being implemented in real-time, becoming a management tool that can calculate water budgets daily and very simply once the initial GIS model is created. The capabilities of the water budget approach are expanded substantially with the aid of high resolution spatial and temporal GIS data, providing an inexpensive, widely applicable technique that is capable of providing valuable research to groundwater studies.

Reference List

Adkin, G. L. (1910). The post-Tertiary geological history of the Ohau River and the adjacent coastal plain, Horowhenua Country, North Island. *Transactions of the New Zealand Institute, XI11*, 496-520.

Alexander, L. V., Zhang, X., Peterson, T. C., Caesar, J., Gleason, B., Klein Tank, A. M. G., et al. (2006). Global observed changes in daily climate extremes of temperature and precipitation. *J. Geophys. Res., 111*(D5), D05109. doi: 10.1029/2005jd006290

Allen, R. G., Pereira, L. S., Raes, D., & Smith, M. (1998). Crop evapotranspiration: guidelines for computing crop water requirements, Irrigation and Drainage Paper 56. Rome, Italy: FAO.

Allison, G. B. (1988). A review of some of the physical, chemical and isoptic techniques available for estimating groundwater recharge. In I. Simmers (Ed.), *Estimation of natural groundwater recharge* (Vol. 222, pp. 49 - 72). Dordrecht, Holland: D. Reidel Publishing Company.

Anderson, M. P. (2005). Heat as a Ground Water Tracer. *Ground Water, 43*(6), 951-968. doi: 10.1111/j.1745-6584.2005.00052.x

Batelaan, O., & De Smedt, F. (2007). GIS-based recharge estimation by coupling surface-subsurface water balances. [doi: DOI: 10.1016/j.jhydrol.2007.02.001]. *Journal of Hydrology, 337*(3-4), 337-355.

Begg, J. G., & Johnston, M. R. (Cartographer). (2000). Geology of the Wellington area.

Begg, J. G., Palmer, A., & Gyopari, M. (2005). Geological synopsis of the Manawatu-Horowhenua area for a review of the region's hydrogeology. New Zealand: Geological & Nuclear Sciences.

Bekesi, G. (1989). *Geophysical studies over the North Horowhenua Sedimentary Basin.* Unpublished MSc Thesis, Victoria University, Wellington.

Bekesi, G. (1996). Nitrate-Nitrogen in Horowhenua Groundwater: Manawatu-Wanganui Regional Council Report No 96/EXT/243, 27.

Bekesi, G. (2001). Manawatu-Wanganui. In M. R. Rosen & P. A. White (Eds.), *Groundwaters of New Zealand* (pp. 387 - 396). Wellington, New Zealand: New Zealand Hydrological Society Inc. & Claxton Press.

Bekesi, G., & McConchie, J. (1999). Groundwater recharge modelling using the Monte Carlo technique, Manawatu region, New Zealand. [doi: DOI: 10.1016/S0022-1694(99)00128-6]. *Journal of Hydrology, 224*(3-4), 137-148.

Bredehoeft, J. D. (2002). The water budget myth revisited: why hydrogeologists model. (Issue Paper). *Ground Water, 40*(4), 340(346).

Cowie, J. D., Fitzgerald, P., & Owers, W. (1967). Soils of the Manawatu-Rangitikei *Soil Bureau Bulletin, 29*.

de Vries, J., & Simmers, I. (2002). Groundwater recharge: an overview of processes and challenges. [10.1007/s10040-001-0171-7]. *Hydrogeology Journal, 10*(1), 5-17.

Domenico, P. A., & Schwartz, F. W. (1990). *Physical and chemical hydrogeology*. New York: John Wiley and Sons Inc.

Downing, R. A., Smith, D. B., & Warren, S. C. (1978). Seasonal Variations of Tritium and Other Constituents in Groundwater in the Chalk Near Brighton, England. *Journal of the Institution of Water Engineers and Scientists, 32*(2), 123 - 136.

Eitzingera, J., Marinkovica, D., & Höschb, J. (2002). *Sensitivity of different evapotranspiration calculation methods in different crop-weather models.* Paper presented at the Proceedings of the International Environmental Modelling and Software Society (iEMSs) Meeting—Integrated Assessment and Decision Support, June 24–27, 2002, Lugano, Switzerland.

Freeze, R. A. (1969). The Mechanism of Natural Ground-Water Recharge and Discharge 1. One-dimensional, Vertical, Unsteady, Unsaturated Flow above a Recharging or Discharging Ground-Water Flow System. *Water Resour. Res., 5*(1), 153-171. doi: 10.1029/WR005i001p00153

Grindley, J. (1967). The estimation of soil moisture deficits. *The Meteorological magazine, 96*(1137), 97 - 108.

Grindley, J. (1969). The calculation of actual evaporation and soil moisture deficit over specified catchment areas *Hydrological memorandum - no. 38*. Bracknell, Berkshire: Meteorological Office, Hydrological Services.

Grindley, J. (1970). Estimation and mapping of evaporation *Symposium on World Water Balance* (Vol. 1: Pub. No. 92 IASH-UNESCO, pp. 200 - 213): International Association of Hydrological Sciences, pub. no. 92.

Groisman, P. Y., Knight, R. W., Easterling, D. R., Karl, T. R., Hegerl, G. C., & Razuvaev, V. N. (2005). Trends in Intense Precipitation in the Climate Record. *Journal of Climate, 18*(9), 1326-1350. doi: doi:10.1175/JCLI3339.1

Hatch, C. E., Fisher, A. T., Revenaugh, J. S., Constantz, J., & Ruehl, C. (2006). Quantifying surface water-groundwater interactions using time series analysis of streambed thermal records: Method development. *Water Resour. Res., 42*(10), W10410. doi: 10.1029/2005wr004787

Headworth, H. G. (1970). The selection of root constants for the calculation of actual evaporation and infiltration for Chalk catchments. *Journal of the Institution of Water Engineers, 24*, 431 - 446.

Headworth, H. G. (1972). The analysis of natural groundwater level fluctuations in the Chalk of Hampshire. *Journal of the Institution of Water Engineers, 26*, 107 - 124.

Healy, R. W., & Cook, P. (2002). Using groundwater levels to estimate recharge. [10.1007/s10040-001-0178-0]. *Hydrogeology Journal, 10*(1), 91-109.

Healy, R. W., Winter, T. C., LaBaugh, J. W., & Franke, O. L. (2007). Water budgets: Foundations for effective water-resources and environmental management: U.S. Geological Survey Circular 1308, 90 p.

Hesp, P. A., & Shepherd, M. J. (1978). Some aspects of the late Quaternary geomorphology of the lower Manawatu Valley, New Zealand. *New Zealand Journal of Geology and Geophysics, 21*, 403 - 412.

Howard, K. W. F., & Lloyd, J. W. (1979). The sensitivity of parameters in the Penman evaporation equations and direct recharge balance. [doi: DOI: 10.1016/0022-1694(79)90069-6]. *Journal of Hydrology, 41*(3-4), 329-344.

Hughes, G. (2005). Paleoenvironmental analysis and chronology of strata encountered within Levin groundwater investigation bore, main road south, Levin (S. o. E. Sciences, Trans.). Wellington: Victoria University of Wellington.

Jackson, T. (2002). Remote sensing of soil moisture: implications for groundwater recharge. [10.1007/s10040-001-0168-2]. *Hydrogeology Journal, 10*(1), 40-51.

Johansson, P.-O. (1988). Methods for estimation of natural groundwater recharge directly from precipitation - comparative studies in sandy till. In I. Simmers (Ed.), *Estimation of natural groundwater recharge* (pp. 239 - 270). Dordrecht, Holland: D. Reidel Publishing Company.

Kitching, R., & Bridge, L. R. (1974). Lysimeter installations in sandstone at Styrrup, Nottinghamshire. [doi: DOI: 10.1016/0022-1694(74)90004-3]. *Journal of Hydrology, 23*(3-4), 219-232.

Kitching, R., Shearer, T. R., & Shedlock, S. L. (1977). Recharge to Bunter Sandstone determined from lysimeters. [doi: DOI: 10.1016/0022-1694(77)90036-1]. *Journal of Hydrology, 33*(3-4), 217-232.

Knighton, D. (1998). *Fluvial forms and processes*. London: Arnold.

Krešić, N. (2009). *Groundwater Resources: Sustainability, Management, and Restoration.* United States of America: McGraw-Hill.

Lee, C.-H., Chen, W.-P., & Lee, R.-H. (2006). Estimation of groundwater recharge using water balance coupled with base-flow-record estimation and stable-base-flow analysis. [10.1007/s00254-006-0305-2]. *Environmental Geology, 51*(1), 73-82.

Lerner, D. N., Issar, A. S., & Simmers, I. (1990). Groundwater Recharge a guide to understanding and estimating natural recharge. *International Contributions to Hydrogeology* (Vol. 8). Hanover: Verlag Heinz Heise.

Liu, Y., & Luo, Y. (2010). A consolidated evaluation of the FAO-56 dual crop coefficient approach using the lysimeter data in the North China Plain. *Agricultural Water Management, 97*(1), 31-40. doi: DOI: 10.1016/j.agwat.2009.07.003

Loucks, D., van Beek, E., Stedinger, J. R., Dijkman, J. P. M., & Villars, M. T. (2005). *Water Resources Systems Planning and Management: An Introduction to Methods, Models and Applications*. Paris: United Nations Education, Scientific and Cultural Organization.

Mackintosh, L. (2001). The Climate of New Zealand: SW North Island Retrieved 6th January, 2011, from http://www.niwa.co.nz/education-and-training/schools/resources/climate/overview/map_sw_north

Manawatu-Wanganui Regional Council. (1998). Lake Horowhenua and Hokio Stream Catchment Management Strategy. Palmerston North: Manawatu-Wanganui Regional Council. Report No 98/EXT/315.

Mather, J. R. (1984). *Water Resources: Distribution, use, and management*. New York: John Wiley & Sons, Inc.

McCarron, C., & Zarour, H. (2005). State of the environment of the Manawatu-Wanganui Region-2005. Technical report five-groundwater. . Palmerston North, New Zealand: Horizons Regional Council.

Montgomery, C. W. (2006). *Environmental Geology* (7th ed.). New York: McGraw-Hill.

Palmer, A., & Van Dissen, R. J. (2002). Northern Ohariu Fault : earthquake hazard assessment of a newly discovered active strike-slip fault in Horowhenua. Wellington: Earthquake Commission Reseach Foundation.

Penman, H. L. (1948). Natural Evaporation from Open Water, Bare Soil and Grass. *Proceedings of the Royal Society of London. Series A, Mathematical and Physical Sciences, 193*(1032), 120-145.

Penman, H. L. (1950a). The dependence of transpiriation on weather and soil conditions. *European Journal of Soil Science, 1*(1), 74-89.

Penman, H. L. (1950b). The water balance of the Stour catchment area. *Journal of the Institution of Water Engineers, 4*, 457 – 469.

Penman, H. L. (1951). The water balance of catchment areas *General Assembly in Brussels in 1951: Reports and minutes of meetings of the Committee on Water Surface* (Vol. 3: Pub. No. IAHS, pp. 434 - 442). Leuven, Belgium: International Association of Hydrological Sciences (IAHS).

Richards, L. A., Gardner, W. R., & Ogata, G. (1956). Physical Processes Determining Water Loss from Soil1. *Soil Sci. Soc. Am. J., 20*(3), 310-314. doi: 10.2136/sssaj1956.03615995002000030004x

Rushton, K. R. (1988). Numerical and conceptual models for recharge estimation in arid and semi-arid zones. In I. Simmers (Ed.), *Estimation of natural groundwater recharge* (pp. 222 - 238). Dordrecht, Holland: D. Reidel Publishing Company.

Rushton, K. R., & Ward, C. (1979). The estimation of groundwater recharge. [doi: DOI: 10.1016/0022-1694(79)90070-2]. *Journal of Hydrology, 41*(3-4), 345-361.

Scanlon, B., Healy, R. W., & Cook, P. (2002). Choosing appropriate techniques for quantifying groundwater recharge. [10.1007/s10040-001-0176-2]. *Hydrogeology Journal, 10*(1), 18-39.

Seidel, D. J., Fu, Q., Randel, W. J., & Reichler, T. J. (2008). Widening of the tropical belt in a changing climate. [10.1038/ngeo.2007.38]. *Nature Geosci, 1*(1), 21-24.

Senarath, D. C. H. (1988). Estimation of recharge of sand aquifer of the island of Mannar, Sri Lanka. In I. Simmers (Ed.), *Estimation of natural groundwater recharge* (pp. 423 - 434). Dordrecht, Holland: D. Reidel Publishing Company.

Sibson, R. (1981). A Brief Description of Natural Neighbor Interpolation. Chapter 2 In *Interpolating multivariate data* (pp. 21 - 36). New York: John Wiley & Sons.

Simmers, I. (Ed.). (1988). *Estimation of natural groundwater recharge*. Dordrecht, Holland: D. Reidel Pubishing Company.

Sinha, B. P. C. (1988). Natural ground water recharge estimation methodologies in India. In I. Simmers (Ed.), *Estimation of natural groundwater recharge* (pp. 301 - 311). Dordrecht, Holland: D. Reidel Publishing Company.

Smith, D. B., Wearn, P. L., Richards, H. J., & Rowe, P. C. (1970). Water movement in the unsaturated zone of high and low permeability strata by measuring natural tritium. In International Atomic Energy Agency (IAEA) (Ed.), *Symposium on the Use of Isotopes in Hydrology* (pp. 73 - 87). Vienna.

Snelder, T., Briggs, B., & Weatherhead, M. (2004). New Zealand River Environment Classification User Guide. Wellington: Ministry for the Environment.

Somashekar, R., Ravikumar, P., Sowmya, S., Dar, M., & Ravikumar, A. (2011). Runoff Estimation and Morphometric Analysis for Hesaraghatta Watershed Using IRS–1D LISS III FCC Satellite Data. *Journal of the Indian Society of Remote Sensing, 39*(1), 95-106. doi: 10.1007/s12524-011-0074-6

Sophocleous, M. (1992). Groundwater recharge estimation and regionalization: the Great Bend Prairie of central Kansas and its recharge statistics. [doi: DOI: 10.1016/0022-1694(92)90051-V]. *Journal of Hydrology, 137*(1-4), 113-140.

Stewart, M. K., Mehlhorn, J., & Elliott, S. (2007). Hydrometric and natural tracer (oxygen-18, silica, tritium and sulphur hexafluoride) evidence for a dominant groundwater contribution to Pukemanga Stream, New Zealand. *Hydrological Processes, 21*(24), 3340-3356. doi: 10.1002/hyp.6557

Tait, A., & Sturman, J. (2008). Annual rainfall maps for the Horizons-Manawatu Region: National Institute of Water and Atmospheric Reseach Limited. NIWA Client Report WLG2008/69.

Taylor, C. B., Brown, L. J., Cunliffe, J. J., & Davidson, P. W. (1992). Environmental tritium and 18O applied in a hydrological study of the Wairau Plain and its contributing mountain catchments, Marlborough, New Zealand. *Journal of Hydrology, 138*(1-2), 269-319. doi: Doi: 10.1016/0022-1694(92)90168-u

Taylor, C. B., Wilson, D. D., Brown, L. J., Stewart, M. K., Burden, R. J., & Brailsford, G. W. (1989). Sources and flow of north Canterbury plains groundwater, New Zealand. *Journal of Hydrology, 106*(3-4), 311-340. doi: Doi: 10.1016/0022-1694(89)90078-4

Theis, C. V. (1937). Amount of ground-water recharge in the southern High Plains. *Transactions - American Geophysical Union, 18*, 564 – 568.

Theis, C. V. (1940). The source of water derived from wells: essential factors controlling the response of an aquifer to development. *Civil engineering, 10*, 277 - 280.

Thompson, S. (2004). New Zealand Land Cover Database 2: User Guide (pp. 24). Wellington: Ministry for the Environment

Thompson, S., Grüner, I., & Gapare, N. (2003). New Zealand Land Cover Database Version 2. Illustrated guide to target classes (pp. 127). Wellington Ministry for the Environment.

Tilahun, K., & Merkel, B. (2009). Estimation of groundwater recharge using a GIS-based distributed water balance model in Dire Dawa, Ethiopia. [10.1007/s10040-009-0455-x]. *Hydrogeology Journal, 17*(6), 1443-1457.

Uma, K. O. (1988). Groundwater recharge from three cheap and independent methods in the small watersheds of the rain forest belt of Nigeria. In I. Simmers (Ed.), *Estimation of* *natural groundwater recharge* (pp. 435 - 447). Dordrecht, Holland: D. Reidel Publishing Company

Vaughan, P. J., Trout, T. J., & Ayars, J. E. (2007). A processing method for weighing lysimeter data and comparison to micrometeorological ETo predictions. *Agricultural Water Management, 88*(1-3), 141-146. doi: DOI: 10.1016/j.agwat.2006.10.008

Ward, K. (1976). *The calculation of recharge from meteorological data, with particular reference to the South Humberside Chalk aquifer.* M. Sc. Thesis, University of Birmingham, Birmingham.

Ward, R. C., & Robinson, M. (2000). *Principles of hydrology* (4th ed.). London, UK: McGraw-Hill.

White, P. A., Raiber, M., Zarour, H., Meilhac, C., & Green, S. (2010). Horowhenua water resources: water budget and groundwater-surface water interaction *GNS Science Conustancy Report 2010/22* (pp. 120p).

Williams, G. J. (2002). Hokio Weir Investigation: Investigation of the effect of the weir on the water quality and cultural & amenity values of Lake Horowhenua. Palmerston North, New Zealand: Horizons Manawatu 2002/EXT/544.

Woods, R., Hendrikx, J., Henderson, R., & Tait, A. (2006). Estimating mean flow of New Zealand rivers. *Journal of Hydrology (NZ), 45*(2), 95 - 110.

Wu, J., Zhang, R., & Yang, J. (1996). Analysis of rainfall-recharge relationships. *Journal of Hydrology, 177*(1-2), 143-160. doi: Doi: 10.1016/0022-1694(95)02935-4

Xu, C. Y., & Chen, D. (2005). Comparison of seven models for estimation of evapotranspiration and groundwater recharge using lysimeter measurement data in Germany. *Hydrological Processes, 19*(18), 3717-3734. doi: 10.1002/hyp.5853

Zarour, H. (2008). Groundwater Resources in the Manawatu-Wanganui Region: Technical Report to Support Policy Development. Palmerston North, New Zealand: Horizons Regional Council.

Zhan, X., & Huang, M.-L. (2004). ArcCN-Runoff: an ArcGIS tool for generating curve number and runoff maps. *Environmental Modelling & Software, 19*(10), 875-879. doi: DOI: 10.1016/j.envsoft.2004.03.001

Zhou, Y. (2009). A critical review of groundwater budget myth, safe yield and sustainability. *Journal of Hydrology, 370*(1-4), 207-213. doi: DOI: 10.1016/j.jhydrol.2009.03.009

Appendix 1: Hydraulic gradient

Hydraulic Gradient													
Catchment	2007												
	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec	Ave
Waitarere North	0.0015	0.0014	0.0014	0.0013	0.0014	0.0014	0.0015	0.0014	0.0015	0.0015	0.0015	0.0014	**0.0014**
Waitarere South	0.0035	0.0034	0.0034	0.0034	0.0034	0.0033	0.0034	0.0035	0.0035	0.0035	0.0035	0.0034	**0.0034**
Waiwiri	0.0028	0.0027	0.0027	0.0027	0.0027	0.0026	0.0027	0.0028	0.0028	0.0028	0.0028	0.0027	**0.0027**
Ohau	0.0027	0.0027	0.0026	0.0026	0.0027	0.0024	0.0027	0.0027	0.0027	0.0028	0.0027	0.0026	**0.0026**
Waikawa	0.0047	0.0047	0.0047	0.0047	0.0047	0.0047	0.0047	0.0047	0.0047	0.0047	0.0047	0.0047	**0.0047**
	2008												
	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec	Ave
Waitarere North	0.0014	0.0014	0.0014	0.0013	0.0014	0.0014	0.0014	0.0015	0.0015	0.0016	0.0016	0.0015	**0.0015**
Waitarere South	0.0034	0.0033	0.0033	0.0033	0.0033	0.0033	0.0034	0.0035	0.0036	0.0036	0.0036	0.0036	**0.0034**
Waiwiri	0.0027	0.0027	0.0027	0.0026	0.0027	0.0027	0.0028	0.0029	0.0029	0.0029	0.0028	0.0028	**0.0028**
Ohau	0.0027	0.0026	0.0026	0.0026	0.0026	0.0026	0.0027	0.0028	0.0028	0.0028	0.0027	0.0027	**0.0027**
Waikawa	0.0047	0.0047	0.0047	0.0047	0.0047	0.0047	0.0047	0.0047	0.0047	0.0047	0.0047	0.0047	**0.0047**
	2009												
	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec	Ave
Waitarere North	0.0015	0.0015	0.0015	0.0015	0.0015	0.0015	0.0015	0.0015	0.0015	0.0015	0.0015	0.0015	**0.0015**
Waitarere South	0.0035	0.0035	0.0034	0.0035	0.0035	0.0035	0.0035	0.0035	0.0035	0.0035	0.0035	0.0036	**0.0035**
Waiwiri	0.0028	0.0027	0.0028	0.0028	0.0028	0.0028	0.0028	0.0028	0.0028	0.0029	0.0028	0.0028	**0.0028**
Ohau	0.0027	0.0027	0.0027	0.0027	0.0027	0.0027	0.0027	0.0027	0.0027	0.0028	0.0028	0.0027	**0.0027**
Waikawa	0.0047	0.0047	0.0047	0.0047	0.0047	0.0047	0.0047	0.0047	0.0047	0.0047	0.0047	0.0047	**0.0047**