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A tool to estimate ground water recharge from hillslopes to shallow foothill aquifers

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A tool to estimate ground water recharge from hillslopes to shallow foothill aquifers

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Summary

Our aim in this study was to develop a simple and reliable tool to estimate potential hillslope recharge to shallow unconfined valley aquifers common throughout New Zealand. In particular, the study has provided estimates of hillslope recharge as inputs into the three-dimensional FEFLOW model developed to inform the recommending of water allocation limits in the upper Motueka and Motupiko valleys (Gusyev et al. 2012; Fenemor & Thomas 2013).

A model to estimate transient phreatic levels in a shallow hillslope recharging a down-gradient valley aquifer was developed using empirical relationships between rainfall, drainage and phreatic levels. The model was then calibrated using measured rainfall and transient phreatic levels at the Korere hillslope in the upper Motueka valley. Validity of the model was successfully tested by comparing the predicted and measured phreatic levels at 4 hillslope sites in the catchment up to 39 km away from the hillslope where the model was initially calibrated. Recharge to the valley aquifer across the foothill – ground water interface was calculated by assuming the flow is directly proportional to the hydraulic gradient of the transient phreatic level at the interface.

The model was better at predicting peak recharge rates than recharge rates during flow recessions, because the phreatic recessions were not steep enough at some sites. More work is needed on this aspect of the model if it is to be used elsewhere.

1 Introduction

A research project was launched by the Integrated Catchment Management (ICM) of Landcare Research with the collaboration of the Tasman District Council (TDC) and the Institute of Geological and Nuclear Science Ltd (GNS) to understand the relationships and interaction between shallow foothill water tables and valley ground water aquifers in the upper Motueka catchment.

The aim was to understand and model the relative contribution of hillslope recharge to total recharge of alluvial valley aquifers common throughout New Zealand. In particular, the study provided estimates of hillslope recharge as inputs into the three-dimensional FEFLOW model (Gusyev et al. 2012) developed to inform the setting of water allocation limits in the upper Motueka and Motupiko valleys (Fenemor 2007; Fenemor & Thomas 2013) and understanding the effects of groundwater interaction as a major driver of stream behaviour (Davie et al. 2008).

The Motueka Catchment is located in the north-west of the South Island, New Zealand, at the western margin of the Moutere Depression and drains an area of 2180 km^2 (Figs 1 & 2). The catchment is dominated by mountains and hill country, showing that about 67% of the catchment has slopes greater than 15^O.

The study area corresponds to the upper part of the Motueka Catchment (Fig. 2), which is 887 km^2 wide. The Upper Motueka Catchment is composed of three main river valleys: Motupiko River (344 km^2), Tadmor River (124 km^2), and Motueka River (419 km^2). The main stem of the Upper Motueka Catchment flows north to the sea for about 110 km. The river is joined from the west by a series of generally much larger tributaries, which drain both hill terrain on Moutere gravel (Motupiko, Tadmor) and mountainous terrain underlain by a complex assemblage of sedimentary and igneous rocks. The main features of the river system in this area include: (1) steep, narrow headwater channels; and (2) broad floodplain and terrace systems within hilly Moutere gravel terrain, which flow below the upper Motueka Gorge to the Wangapeka confluence (Fig. 2).

Mean annual rainfall for the catchment is estimated at 1600 mm. However, there is a strong spatial pattern of rainfall variation, primarily related to topography. There is a gradient both north and southwards away from the Tapawera gauge. At the farthest point of the model grid the annual precipitation is approximately 14% higher than at Tapawera. Annual open pan evaporation at Motueka is 1106 mm and is strongly seasonal, with mean monthly values ranging form 27 mm in July to 179 mm in January. While annual evaporation is less than annual rainfall, soil moisture deficits are common in summer when evaporation exceeds rainfall and irrigation is required on many crops in the catchment.

Groundwater in the Upper Motueka Catchment is abstracted from shallow, unconfined alluvial aquifers that occur in the Quaternary river terrace formations and modern river deposits. These are (from oldest to youngest) the Moutere Gravel, Manuka, Tophouse, Speargrass, and modern river gravel formations. The Quaternary Gravels are underlain by the Moutere Gravel Formation throughout the whole study area (Stewart et al. 2005).

The Moutere Gravel Formation consists of rounded greywacke clasts up to 0.6 m diameter (most less than 0.2 m diameter) in a yellowish-brown, silty, clay matrix. The formation contains minor clasts of very weathered ultramafics in the Motueka River upstream of the

Motupiko River confluence. Moutere gravel is widespread throughout the upper Motueka River catchment and forms the hill country between the valleys of the Motueka, Motupiko, and Tadmor Rivers (Stewart et al, 2005).

The Speargrass Formation is widespread in upper reaches of the valleys but absent in the lower reaches. An aggradation surface occurs on the Speargrass Formation terrace that is approximately one to two metres higher than the degradation surface. Groundwater is abstracted from the formation within the study area. The average saturated thickness of the Speargrass Formation is estimated to be between 5 and 8.5 m.

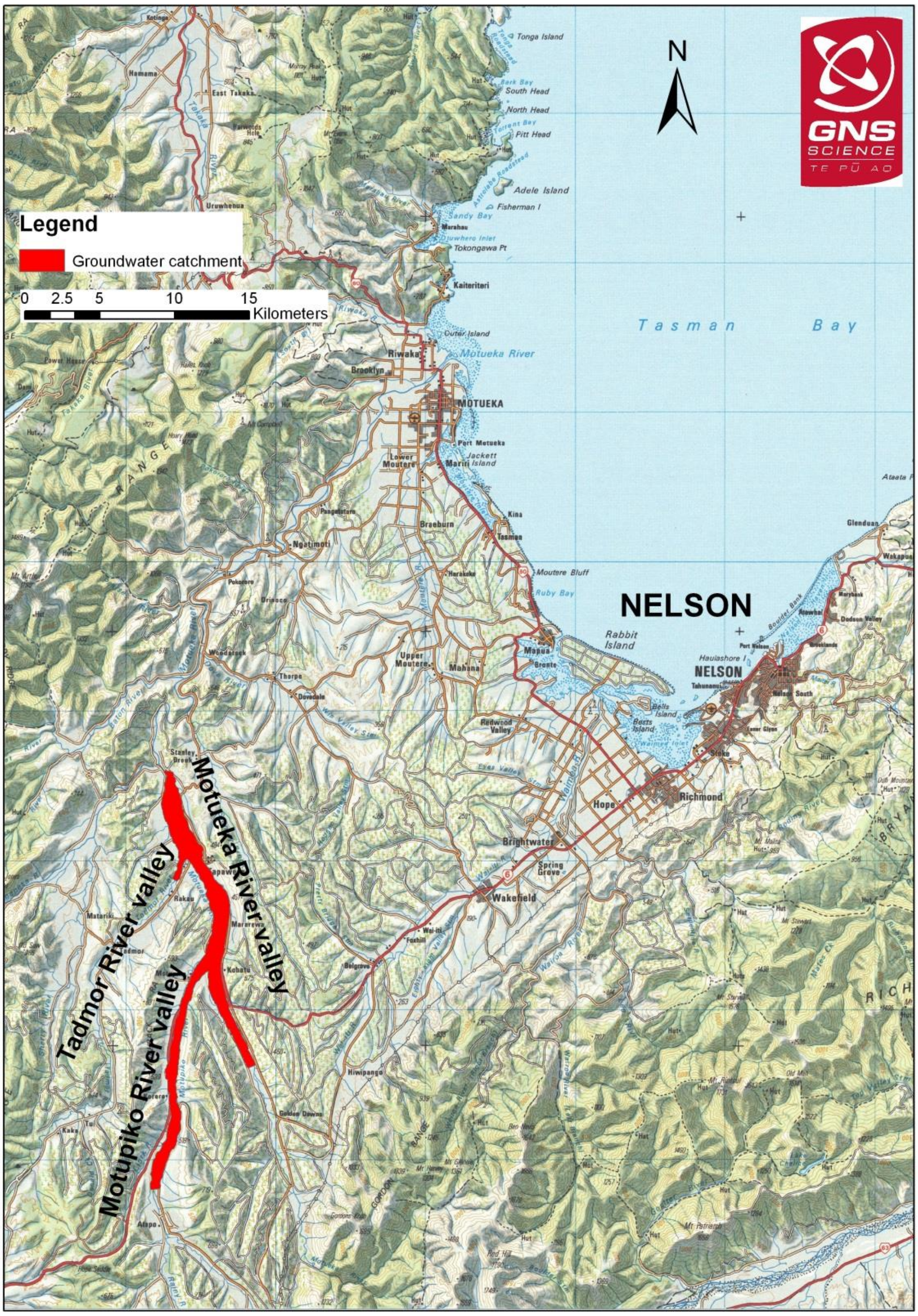

Figure 1 Map of Upper Motueka Catchment.

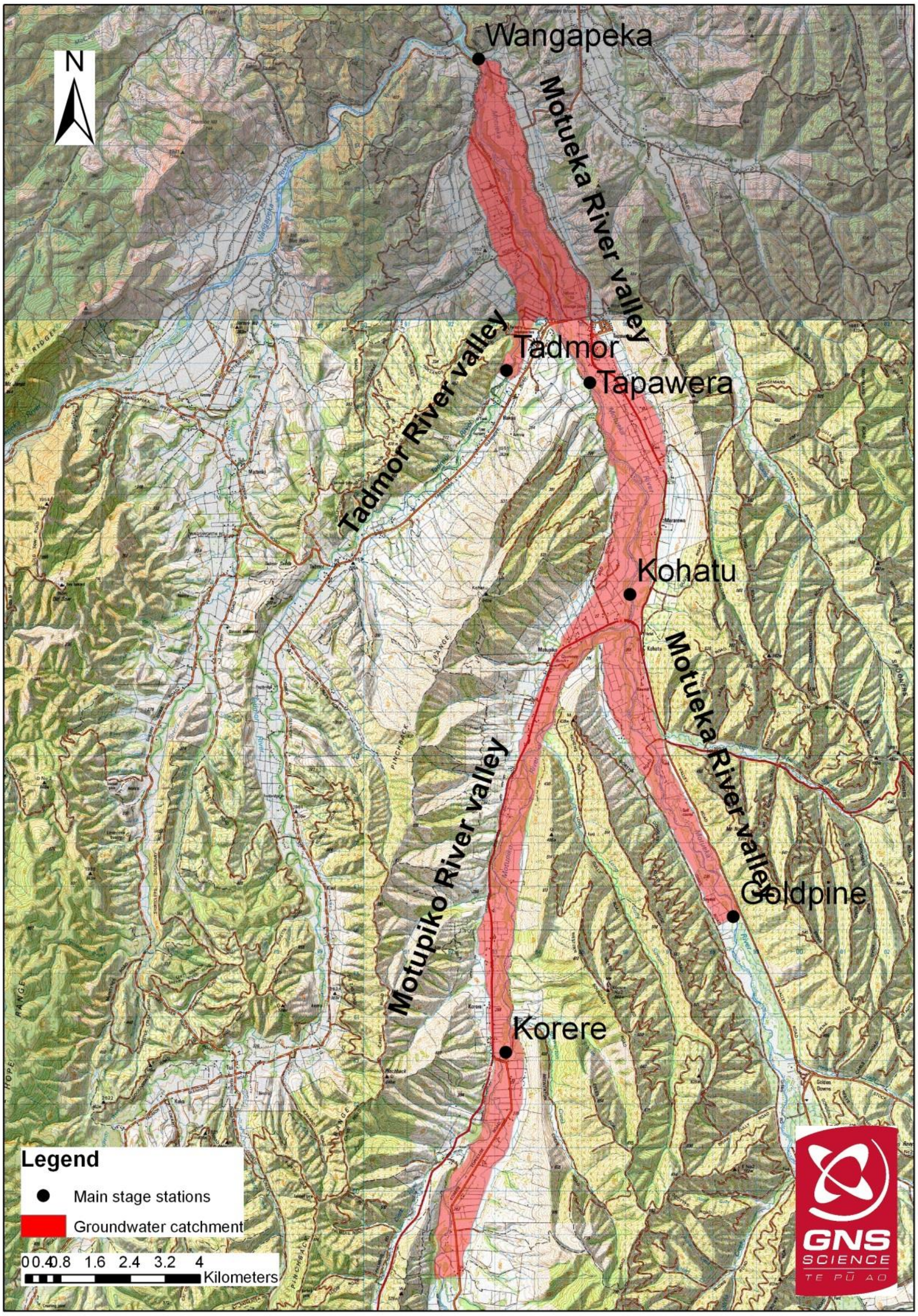

Figure 2 Upper Motueka Catchment with the river monitoring stations and main irrigated areas.

The development of the hillslope recharge model to estimate potential recharge to the shallow ground water aquifer and calibration of the model was completed in 2010 (Hong et al. 2010). The objective of this study is to validate this transient hillslope recharge model using field measured data at three foothill sites.

Eleven hillslope sites (A–K shown in Figs 3 & 4) were identified as potential areas of hillslope recharge to the shallow unconfined aquifer of the upper Motueka valley and tributaries. Selected hillslopes have identical features of planar gentle slopes with streams running along the foothills.

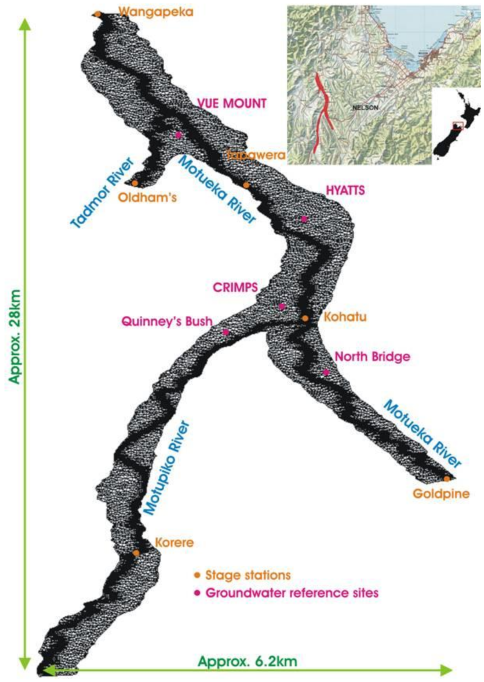

Figure 3 Two-dimensional view of the Upper Motueka groundwater model domain (Figure courtesy Timothy Hong, GNS).

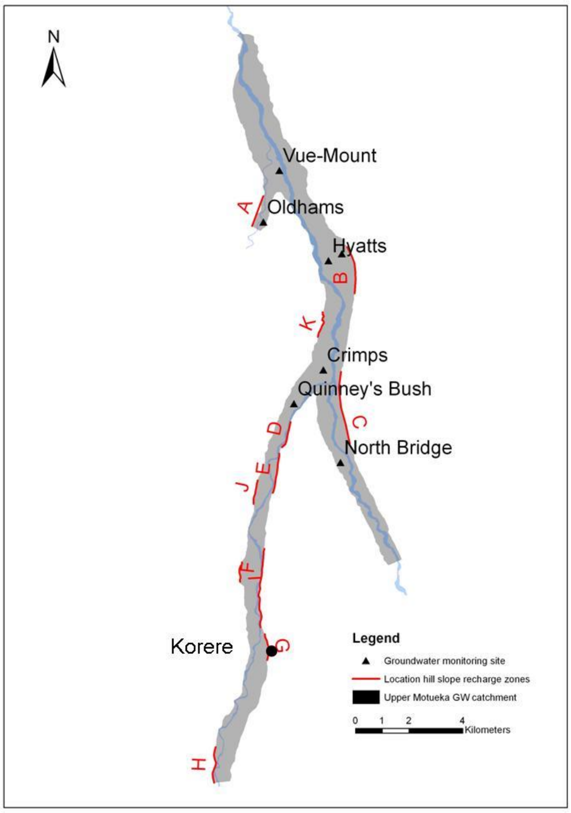

Figure 4 Eleven hillslopes were identified as potential recharge zones. Hillslopes B, C, E, I and J were installed with water level recorders to record phreatic levels along the slope (Figure courtesy Timothy Hong, GNS).

Selected hillslopes, which faced west and east, drain to the Motueka and Motupiko rivers respectively. Streams, which are usually dry or have very low flow in dry period, come to life during rainfall events. Overland flow is also common during high intensity rainfall events. In order to model the hillslope recharge to the shallow aquifer, it is essential to understand the hillslope hydrology process.

2 Hillslope hydrology (Previous work)

Hillslopes have commonly been regarded as sources of predominantly surface runoff rather than sources of aquafer recharge. Understanding the mechanism of hillslope recharge should help groundwater managers to better quantify and manage recharge from hillslopes.

Maimai catchment in the Tawhai State Forest, near Reefton, North Westland, South Island, has been one of the major research areas for some well-known and well-resourced research teams around the world. Pioneering research work to understand processes of rainfall, stream flow, and hillslope hydrology has been carried out in this catchment since New Zealand Forest Research Institute laid the foundation for multi-catchment research in 1974.

One of the early research studies in the Maimai catchment (Mosley 1979) suggested rain water rapidly moves vertically in the down slope direction in a form of saturated wedge altered by the shape of the bedrock. Water was found to be moving considerable distances through the soil on the rising limb of the hydrograph. Mosley's observations suggested that seepage and macro-pore flow were the predominant mechanisms of storm water flow generation. The preferential flow path effect on the process of hillslope hydrology has been another topic of debate for many years. Mosley (1982) reported that the existence of preferential flow paths along live and dead roots and cracks were the main transport mechanisms for the generation of runoff. Profile wetness increased downslope, and verticality and thickness of the saturated zone decreased in an upslope direction. This is a typical sign of the building up of a phreatic surface closer to the stream edge.

Pearce's (1986) studies in the same catchment suggested that old water is mainly responsible for hydrograph generation. Electrical conductivity and chloride experiments suggested there was only a small contribution from rain to storm flow. Sklash et al. (1986) suggested that saturated wedges on the lower slopes were formed by fast infiltrating rain water that forms a phreatic surface. They also discounted the early pipe and macro-flow observed by Mosley. However, macro-pore flow has been measured by other researchers, for example, McDonnell (1990a), and Woods and Rowe (1996).

McDonnell (1990b) observed that development of matric potential in the slope varies with rainfall intensity, magnitude, and antecedent conditions. Through isotope analysis, it was also found that old water could dominate subsurface flow up to 85% by volume (McDonnell et al. 1991).

Comprehensive research work by Woods and Rowe (1996), using a subsurface collection system at Maimai catchment, showed that subsurface flow from hill slopes is notoriously variable in magnitude and timing. Woods and Rowe developed a topographic index to describe the temporal and spatial variation of subsurface flow on a trenched face of a hill slope. Woods et al. (1997) modified the TOPMODEL index of Bevan and Kirkby (1979) to accommodate varying time source in their Maimai experiment area. They developed their model by assuming rainfall, soil depth, and saturated hydraulic conductivity are uniform in time and space. The model assumes: 1) Soil lies above an impermeable surface; 2) Saturated hydraulic conductivity is constant for the entire soil depth. The modified model suggested a simple nonlinear function of catchment saturated zone thickness to estimate the pattern of recharge, and also provided an index to estimate saturated zone thickness with many assumptions.

Bedrock topography could play a major role in hill slope hydrology. Discussions were initiated between McDonnell , Woods, and Rowe, after Woods and Rowe's subsurface flow experiments led them to conclude that bedrock topography is more suitable than surface topography to estimate the drainage area. Contradicting early findings, Graham et al (2010) argued that major control on subsurface flow generation is not the measurable standard parameters ,such as soil depth, permeability, texture or surface topography, but the parameters that are difficult to measure, such as micro-scale bedrock topography and bedrock permeability, a theory supported by recent work by Hale and McDonnell (2016). However, numerous tracer experiments in the Maimai catchment could not find a correlation between soil depth, water table response, and topography, suggesting other factors control the processes of hillslope hydrology (McDonnell et al. 1998). Saturation depths observed on densely instrumented hillslope in the Panola mountain research watershed in Georgia showed that temporal and spatial variability of saturation depths was highly variable due to variation in drainable pores (Weiler & McDonnell 2004).

Mathematical models to analyse hillslope hydrological processes described by Kirkby (1978) and Anderson and Brooks (1996) were based on complex numerical integration of the 3-D Richard's equation generated for the subsurface. These nonlinear diffusion equations are very difficult to parameterise in order to reflect the complex physical geometry of the bedrock surface and heterogeneity of subsoil, and surface topography. The Hillslope Storage Boussinesq (HSB) equation has also been used to analyse hill slope storage and runoff, which also demands precise knowledge of the spatial variation of physical soil and surface parameters that control the hillslope hydrology.

Extensive laboratory and field experiments led researchers to believe use of Richard equation with piecewise continuous water retention and hydraulic functions could account for the increase in hydraulic conductivity near saturation caused by macropores (Schaap & van Genuchten 2005; Borgesen et al. 2006).

Well-planned and well-resourced research programmes equipped with modern technology will continue to search for new hillslope hydrological activities to improve understanding hillslope hydrology processes.

The following key factors from all research to date, may or may not influence the process of hillslope hydrology and lead to the recharge of ground water and runoff:

1. Macro- and micro-scale surface topography
2. Macro- and micro-scale bedrock topography
3. Spatial variation of soil layer depth and soil heterogeneity
4. Spatial variation of soil hydraulic parameters (hydraulic conductivity)
5. Map of bedrock fractures and soil-bedrock interface leakage
6. Bedrock permeability
7. Preferential flows as a result of dead/live roots and macro-pores
8. Changes in the volume of soil properties (shrinkage cracks)

3 Hillslope recharge model development

Resources required for collecting sufficient spatial data to build a model capable of simulating hillslope hydrology addressing all the key factors listed above are beyond most research organisations. Simple models are in any case needed for resource management purposes.

Due to the complexity, no simple tools have been developed to estimate rainfall runoff and hillslope recharge using rainfall and a handful of easily measurable parameters. Rather than spending resources to "understand" hillslope hydrology, perhaps it is time to put what is already learnt into practise to develop simple but reliable tools to simulate hillslope recharge and runoff. Our aim in this study is to develop a simple and reliable tool to estimate potential hillslope recharge to a shallow unconfined aquifer.

3.1 Theory

Observations of a saturated wedge over bedrock created by infiltration could be related to the development of the phreatic surface along the hillslope (Mosley 1979; Sklash et al. 1986). A hydraulic gradient created by the transient phreatic level is responsible for subsurface water moving across an interface perpendicular to the slope.

In this report we assume the hydraulic gradient created by the transient phreatic levels along the foothill-aquifer interface with alluvial gravel is responsible for hillslope recharge to the shallow aquifer. Therefore the key to developing the recharge model is an ability to estimate transient phreatic levels at the foothill-aquifer interface. Our major concern is not how the rain water is transported to the foothill interface or the process of development of the transient phreatic surface at the foothill; rather, we are interested in the behaviour of climatic-induced transient phreatic levels at the foothill interface. As long as we can predict the transient phreatic level at the foothill interface with alluvial gravel, we do not need to know the key parameters listed above.

Flow net analysis to estimate water flow across the interface under steady state

Two-dimensional steady-state flow under a hydraulic gradient, described by the Laplace equation, is most conveniently solved by graphical construction of a flow net. Because data were limited, flow net analysis described in groundwater hydrology (U.S. Army Corps of Engineers 1999) was applied to estimate the hill slope water flow.

A flow net is a network of stream lines and equipotential lines. In an isotropic porous medium, stream lines are perpendicular to the equipotential lines (Fig. 5).

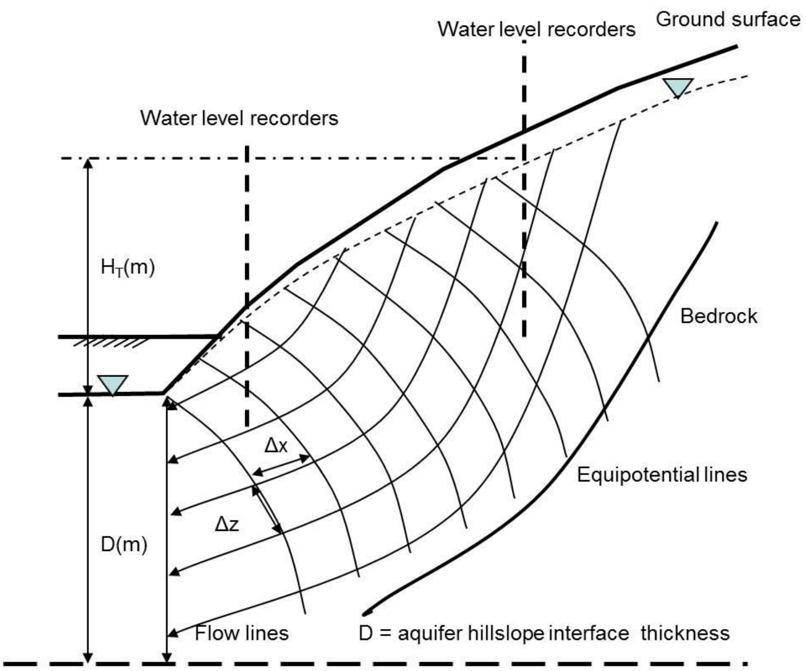

Figure 5 Approximation of the water flow at the interface of the foothill and shallow aquifer boundary by a two-dimensional flow network.

3.2 Assumptions

1. Flow is two dimensional
2. Steady state flow conditions are assumed
3. Porous medium is homogenous and isotropic
4. Flow nets are drawn as approximate squares
5. Bottom boundary is taken as impermeable (bedrock).

Number of equipotential drops = N_d

Number of flow tubes = N_f

Flow through a unit thickness of one square element in m^3/day is given by Eq-1

$$q = k\frac{\Delta h}{\Delta x}\Delta z \qquad [1]$$

where $\Delta h = H_t/N_d$ and H_t is the total head drop (m), and k is the hydraulic conductivity (m/day), Δz is the depth of an element, and Δx is the potential drop across an element.

Since the flow elements were constructed to have approximate square shape, $\Delta z \sim \Delta x$

$$q = k\frac{H_t}{N_d} \qquad [2]$$

the total flow (m^3/boundary length/day) $Q = q \times N_f = K H_t N_f/N_d$

$$Q = qN_f = \frac{kH_tN_f}{N_d} \qquad [3]$$

N_f and N_d were selected to form approximate square shapes of flow elements. According to Eq-3, hill slope recharge is governed by the hydraulic head gradient, and the uniform saturated hydraulic conductivity of the porous media. Parameters N_f and N_d are selected to form approximate square elements in the flow net.

Hillslope recharge to the unconfined aquifer takes place through an interface with a water table depth above the bedrock, which is typically c. 3–4 m deep for most hill slopes in the area (Timothy Hong, GNS, pers. comm.). Recharge from the hillslope remains active as long as hydraulic gradient is present at the foothill interface. Daily water flow across the hillslope interface can be estimated (Eq-3) if transient phreatic level records are available.

Recharge driven by the transient phreatic level gradient includes both surplus recharge and the base flow component. Rainfall is a short time event but the piezometric water level is transient in nature, which is similar to a well hydrograph. The transient nature of the phreatic surface is controlled by the 7 key parameters listed before. The challenge we face now is to find a way to model hillslope recharge by considering the key findings of previous research. We take a different, but simple, approach to create a hillslope recharge tool (model) without discarding any of the key findings of previous research work of hillslope hydrology. Our idea is to create a deterministic type model where all the key findings (parameters) found to be involved in the process of transporting rain water towards the foothill interface are encapsulated in a “Black Box” (Fig. 6).

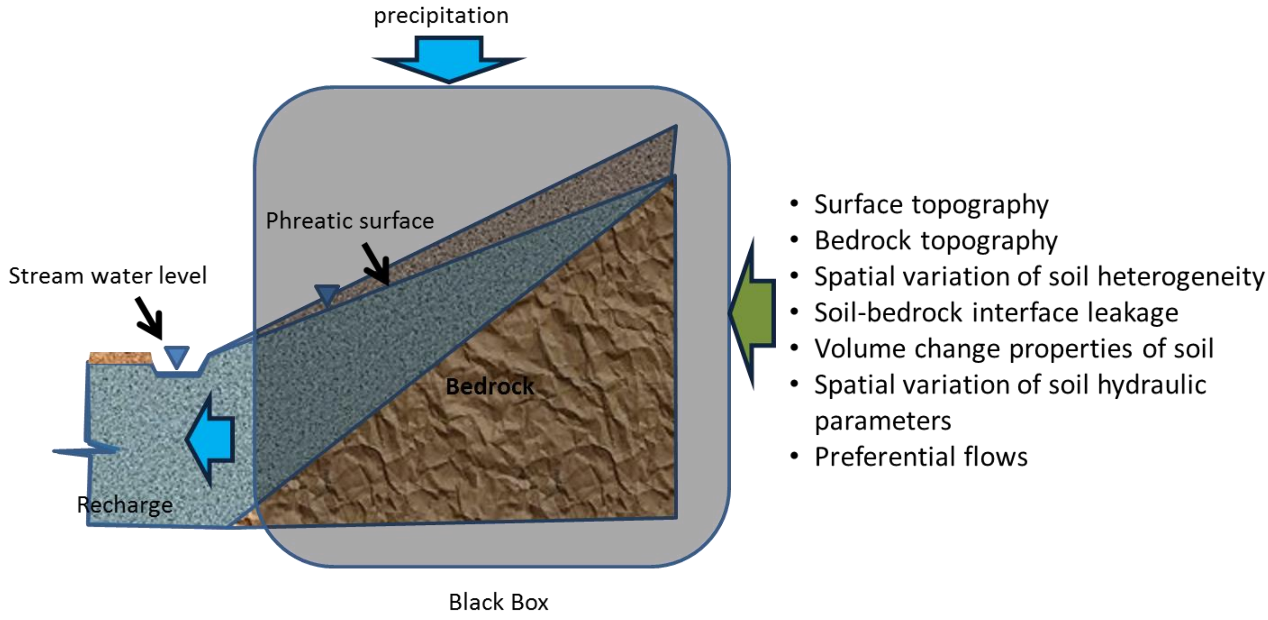

Figure 6 Key parameters which are responsible for complex hydrological process to transport rain to foothill are encapsulated in the black box.

We do not attempt to create a universal tool to estimate hillslope recharge for every hillslope. Our black box deterministic model approach will be most applicable to the hillslope where the empirical equations are developed and calibrated. Generally, models developed with empirical equations to encapsulate the key parameters are site specific; however, we will show that our model was successfully extended to other hillslopes 39 km away from the site where it was calibrated and validated. Our task is to use the black box approach to find an empirical relationship between input variable (rainfall) and output variable (hillslope recharge).

3.3 Climatic induced phreatic levels

Since the direct driver of the recharge is the hydraulic gradient created by the transient phreatic level, we could try to find a relationship between rainfall and phreatic level at the foothill relative to stream water level along the foothill.

Therefore, in order to develop the hillslope recharge model, we take the following steps.

1. Assume key parameters found to be responsible for moving water to the foothill interface are encapsulated in a black box.
2. Find an empirical relationship to relate rainfall to hillslope recharge.

3.4 Instrumentation to collect data for model development

Paratiho

As part of another research objective to understand differences between rainfall recharge under pasture and forest hillslopes, in early summer of 2003 the Paratiho hillslope was instrumented with water level recorders and TDR soil moisture sensors and a rain gauge. Hourly data were recorded until June 2007. Although the topography and geology of the Paratiho hillslope was identical to those selected to model hillslope recharge to the shallow aquifer under Motueka river valley, it was 39 km away from the nearest recharge hillslope study site (Fig. 7).

Figure 7 Paratiho hillslope site is about 39 km away from the Korere where the model was calibrated. The transient water-level rainfall model which was developed for Korere hillslopes was successfully applied at Paratiho.

Figure 8 Installation of TDR and water-level recorders at Paratiho to measure rainfall recharge on pastoral hillslopes (August 2003).

Capacitance probes were installed along the slope from the edge of the stream to record transient phreatic levels (Figs 8 & 9). We will be using data from Paratiho to validate the empirical model developed to estimate transient phreatic levels for hillslopes 39 km away.

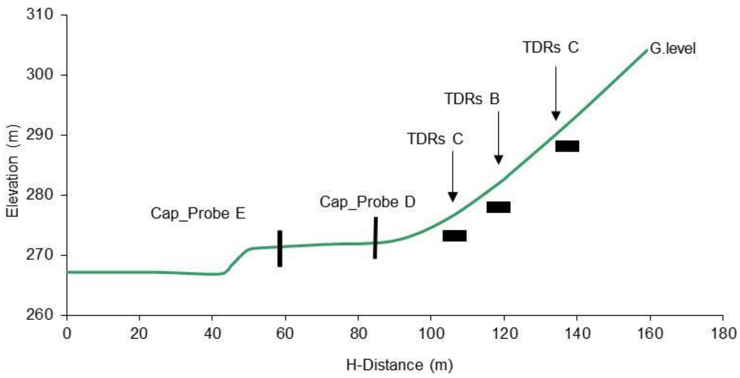

Figure 9 Instrumentation along the slope at Korere hillslope. TDR soil moisture sensors were installed to capture the wetting front dynamics.

Korere

Korere is our major hillslope site to measure rainfall and transient phreatic level to develop and calibrate the empirical model. Phreatic levels near the foothill area of the Korere site reach the ground surface both in winter rain events and during relatively larger summer rain events (see Fig. 10 B). Stream water level at the Korere site temporarily disappears in the summer, only to resurface during small rain events. Although the seasonal stream water levels at the Korere foothill vary, the stream at the Paratiho foothill flowed thoroughout the year.

Water level recorders, TDR soil moisture sensors and rain gauges were installed in the foothill area of the Korere hillslope (Fig. 10) in August 2006 and data were recorded until late August 2007. Instruments were configured to record hourly readings of water level and rainfall. The purpose of the instrumentation was to measure transient phreatic level and rainfall to develop the “Black Box” model.

Figure 10 Instrumenting Korere Hillslope with gently planar slopes showing (clockwise from top left) the Carson, Cemetery, Hyatt and Korere localities. Overland flow and live streams are common in large rain events (John Payne).

3.5 Data collection for model validation

For the purpose of model validation, four more hillslope sites, Cemetery (B), Forest(C), Hyatt (E) and Carson (J) (see Fig. 7) were instrumented with water level recorders to record water levels from Sept 2009 to June 2013. Daily rainfall was recorded only at Carson (J).

Figure 11 Capacitance type water level recorders were installed in PVC cased bore holes. All four sites were installed in a single day with the help of a post driver.

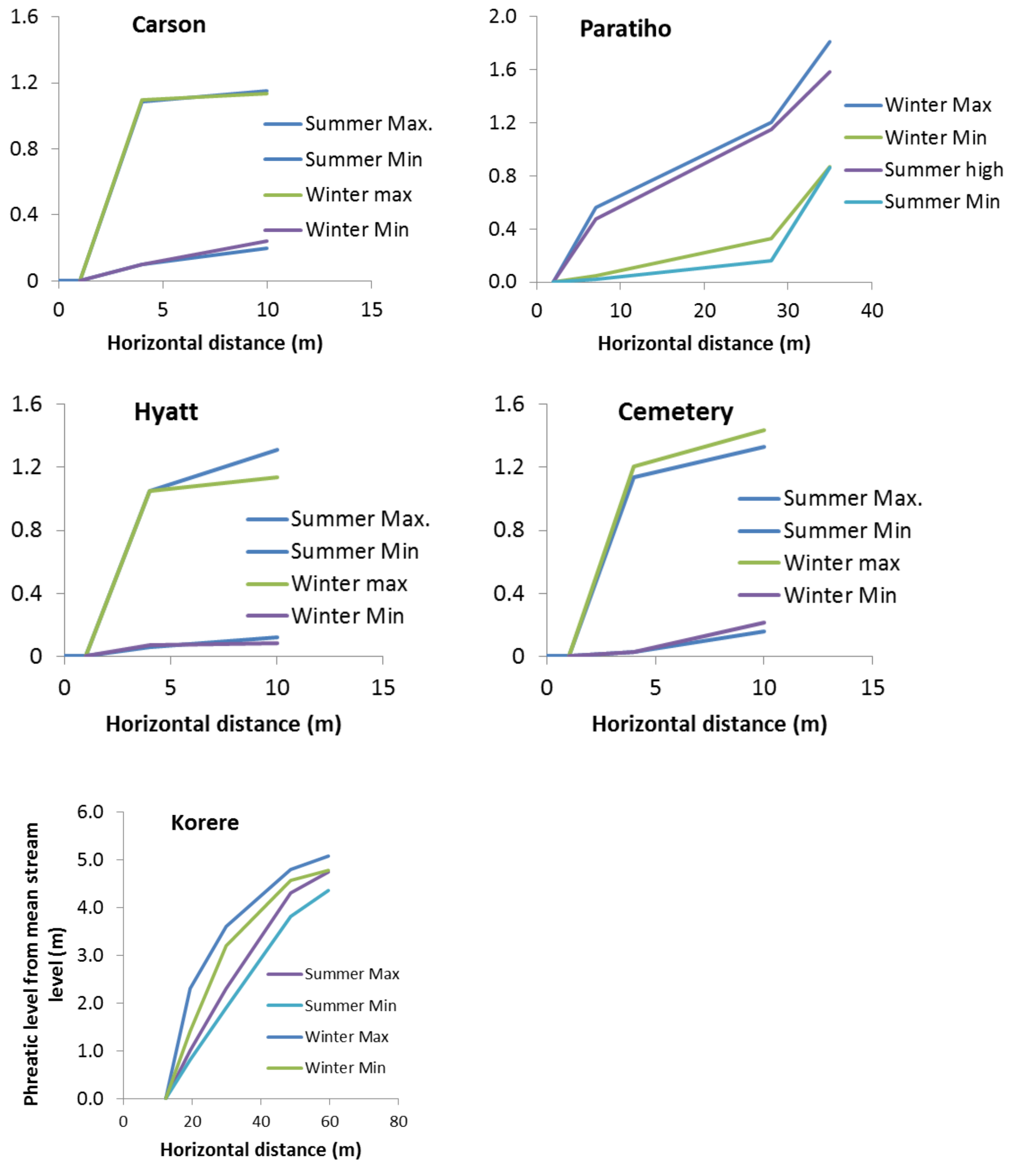

Figure 12 Maximum and minimum seasonal phreatic levels recorded relative to mean stream water level. Phreatic levels at the foot of the Korere hillslope did not show seasonal variation compared to other sites.

3.6 Model Concept: Estimating the transient phreatic level

In order to estimate a transient phreatic level to simulate the well hydrograph at the foothill, our plan is first to estimate the initial phreatic level for a given rain event and then estimate the transient recession limb of the phreatic level (Fig. 13). Flow net analysis (Eq-5) is then used to estimate the transient recharge.

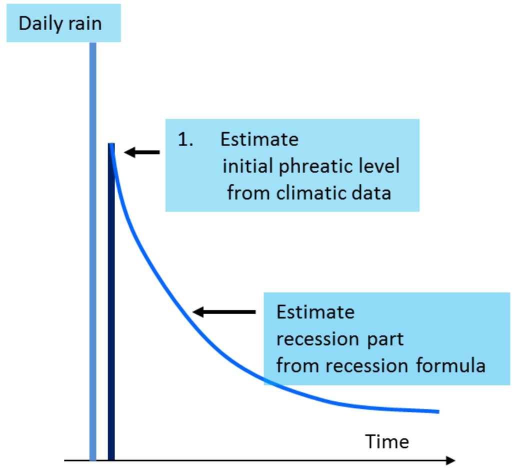

Figure 13 Concept of the estimating transient phreatic level. 1. Estimate the phreatic level due to daily rain. 2. Estimate the transient recession limb using the empirical master recession curve for the hillslope.

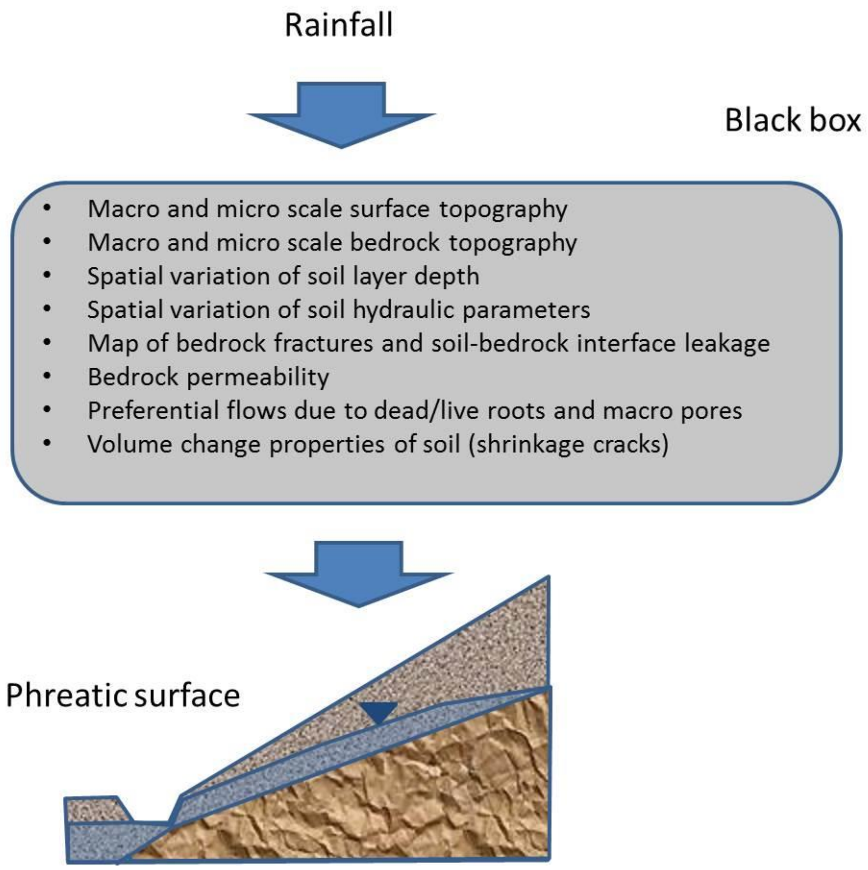

Figure 14 Our visualisation of the black box empirical approach to estimate phreatic level from rainfall. All key factors of the previous research are encapsulated in the black box.

Our first attempt to build a relationship between rainfall and phreatic level failed as the correlation between the phreatic level and rainfall was found to be very poor (Fig. 15). It is impossible to develop a meaningful empirical model relating rainfall to hillslope recharge using such a poor correlation. Correlations between rainfall and phreatic levels were poor for both Paratiho and Korere hillslope sites, suggesting many factors other than rainfall influence the development of phreatic levels.

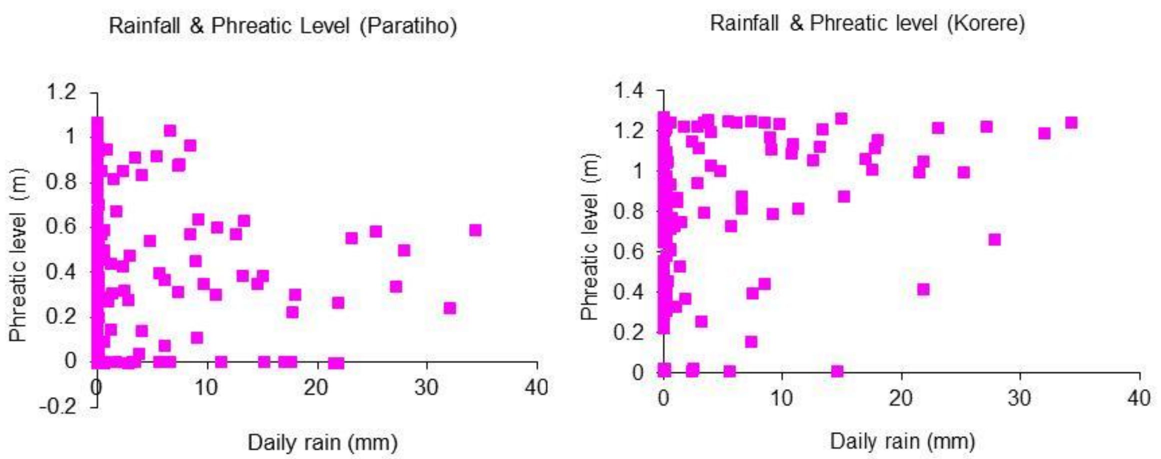

Figure 15 Poor correlation between phreatic level and rainfall was found for both Paratho and Korere hillsope sites. Not all rain events exceed the soil water deficit to contribute to the ground water.

3.7 Correlation between phreatic level and drainage

Main reasons for poor correlation between rainfall and phreatic levels could be due to the complex hydrological process described by the key parameters encapsulated in the black box. It is well known that phreatic levels do not respond to all rainfall events as precipitation does not exceed the soil water deficit under all rain events. There is a reasonable chance that phreatic levels could respond to drainage as water actually enters the subsoil to increase subsurface moisture content. Our next attempt was to see whether there was any correlation between drainage and phreatic levels.

3.8 Landcare Research irrigation scheduling model

The Landcare Research irrigation scheduling model developed by Andrew Fenemor and Tim Davie is a simple daily water balance model (Eq-4) that we used to calculate drainage and estimate irrigation water needs. If we could find an acceptable correlation between drainage and phreatic levels then we could also develop a relationship between rainfall and phreatic level. Use of drainage to relate rainfall to phreatic level is more meaningful as the drainage incorporates many of the key parameters we encapsulated in the black box.

$$SD_n = SD_{n-1} + AET - R_e - I_r \qquad [4]$$

Where,

R_e	=	Effective rain (rain reaching the soil)
AET	=	Actual evapotranspiration
SD_{n-1}	=	Soil water deficit at the start of the day
SD_n	=	Soil water deficit at the end of the day
I_r	=	Irrigation

1. Effective rain, R_e

Effective rain is estimated by allowing a fraction of rain water to drain through macropores if the daily rain total exceeds a predefined threshold rainfall value, which is taken as 2 mm. If the daily rain total is larger than the threshold value, then 10% of the total daily rain is assumed to be drained through macropores (Fig. 16).

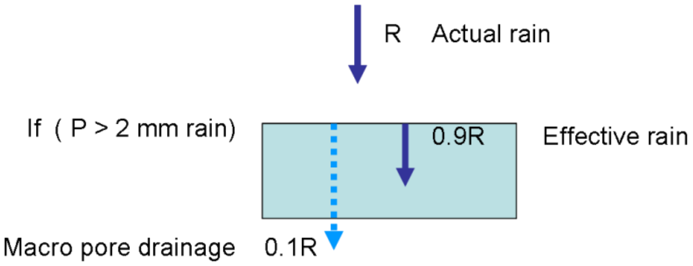

Figure 16 Estimation of effective rainfall after interception. Macropore drainage can be adjusted to site specific soil properties.

2. Actual evapotranspiration, AET

Actual evapotranspiration is driven by the soil water deficit (SD) as shown in Fig. 17. According to Figure 17, actual evapotranspiration (AET) reaches potential evapotranspiration (PET) if the SD < 0 and AET = 0 if the SD > readily available water capacity of the soil (RAW). For RAW < SD > 0, AET varies between 0 and PET according to the function shown in Figure 17. AET is taken as equal to PET if the soil is irrigated regardless of the level of the soil water deficit.

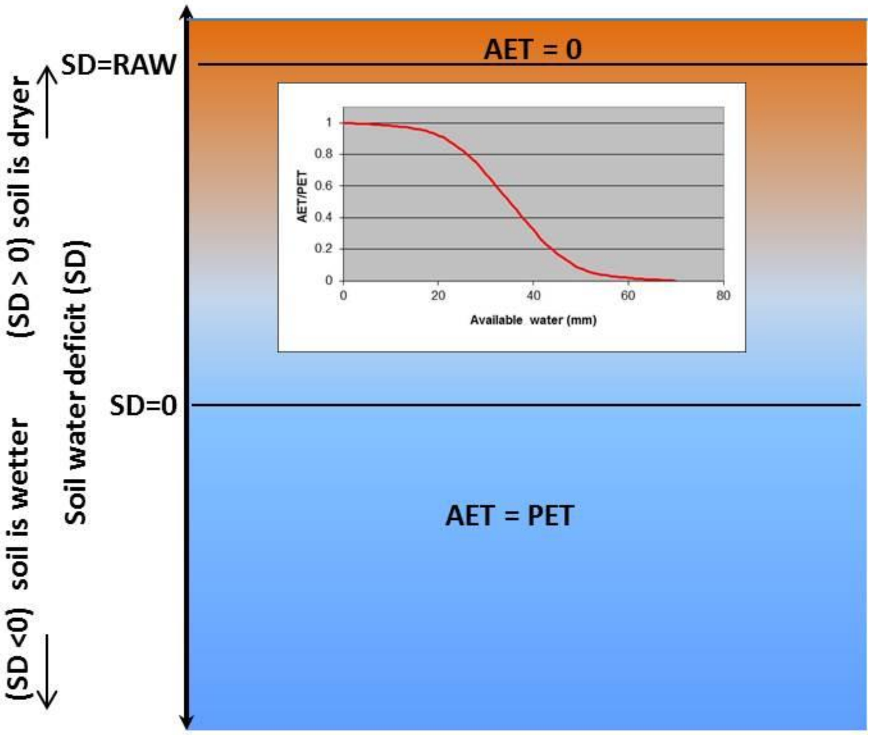

Figure 17 Estimation of actual evapotranspiration (AET) according to the soil water deficit (SD) for a given soil texture.

3. Soil water deficit at the start and end of the day, SD_{n-1}, SD_n

Soil water deficit is the algebraic sum of irrigation, I_r, actual evapotranspiration, AET and effective rainfall, R_e as given by Eq-4.

4. Irrigation, I_r

Irrigation is applied if the soil water deficit exceeds a pre-defined trigger value which is taken as 50% of the RAW of the soil. 5 mm of water is allowed to stay in the soil if the irrigation exceeds the soil water deficit by 5 mm, and the excess water is allowed to drain.

5. Drainage

Drainage takes place only if the irrigation plus rainfall exceeds the soil moisture deficit.

Drainage = -0.1R + (SD-I_r) +5 if (SD-I_r) < -5 mm

Drainage = –0.1R if (SD-I_r) > –5 mm

Where 0.1 R is the macropore drainage if the rainfall exceeds the 2mm threshold value. Note that the sign of the drainage is negative.

Small variations in phreatic levels for larger rain events suggest the generation of overland flow or the lower antecedent conditions keeps the drainage relatively unchanged. The main black box model is now split into two black boxes. The new model concept is given by the flow chart in Figure 18 and Figure 19.

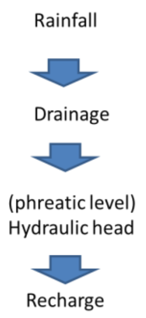

Figure 18 Flowchart of the empirical hillslope recharge model development. First, a relationship is built to estimate drainage from daily rainfall and then the second relationship is built to estimate phreatic levels using drainage. Recharge is assumed to be directly proportional to the transient hydraulic gradient of phreatic levels.

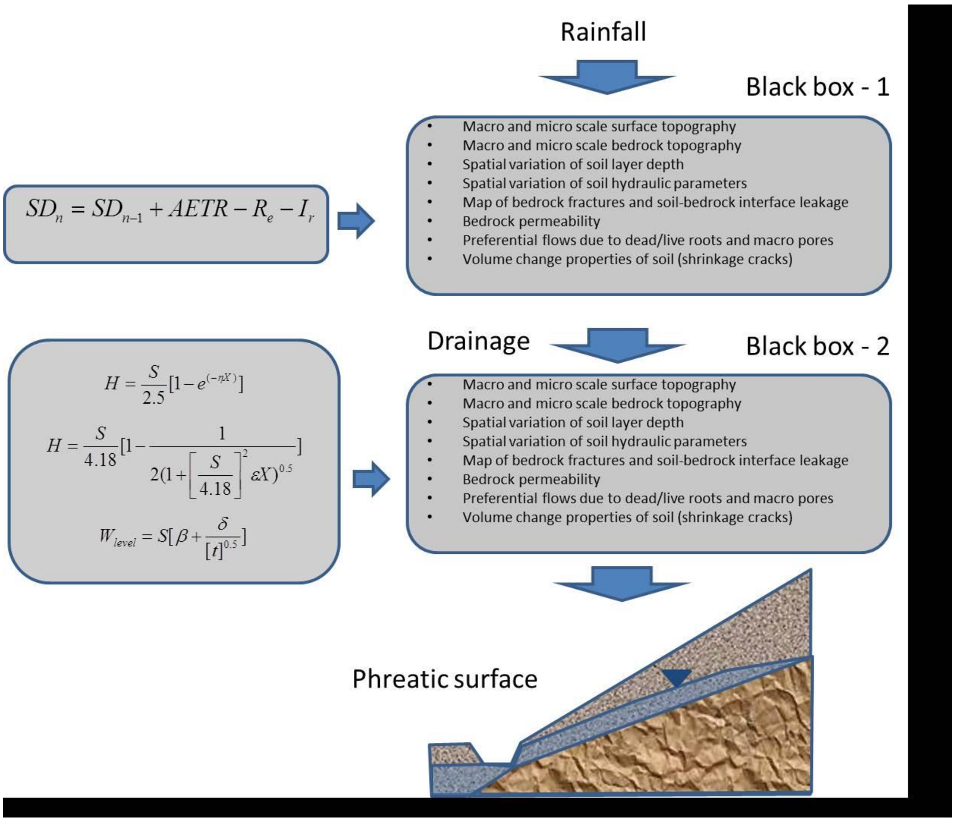

Figure 19 Black box model has been split into two parts: 1) drainage is estimated from rainfall; 2) phreatic level is estimated from drainage.

3.9 Model (empirical) equations

Our black box model to predict the phreatic level is now developed in two parts (Fig. 19):

1. Find model equation to predict the initial phreatic level for a given rain event.
2. Find a recession formula to estimate the transient limb of the phreatic level.

Measured data at the Korere hill slope site in the 2006/2007 summer and winter seasons were used to find empirical model equations and relevant parameters. Measured data at the other 5 sites were then used to validate the model.

Figure 20 Saturated hydraulic conductivities were measured at Korere (L) and Paratiho (R) sites using double ring infiltrometers.

3.10 Estimating the initial phreatic level

The irrigation scheduling model with the irrigation component switched off provides the daily drainage from daily rainfall. Table Curve 3D® from Systat Software was used to obtain the most suitable relationship (with the highest correlation coefficient) between the drainage and phreatic level for the 2006-2007 winter and summer. Relationships between the phreatic level and drainage were found to be site independent but they showed slight seasonal change. Seasonal phreatic levels for summer and winter are best estimated using two different empirical formulae. Therefore two different best fit relationships were developed to estimate the summer and winter phreatic levels given by Eqs-6 & 7.

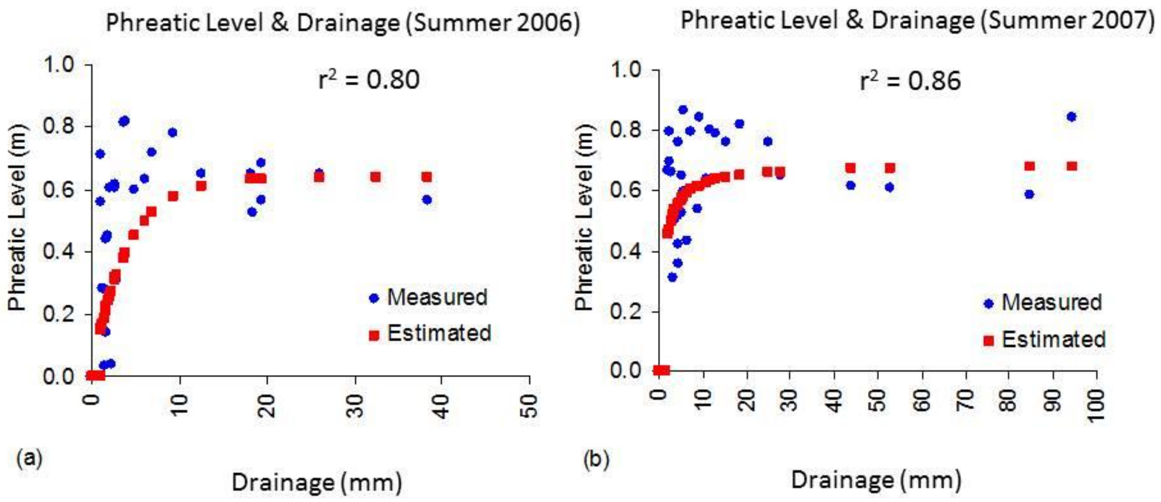

Figure 21 Correlation between all drainage events and phreatic level found to be reasonable for Korere.

For a known daily total drainage (X - mm), mean daily phreatic level (H - m) can be estimated for summer rain events using Eq-5 (Fig. 21a).

$$H = \frac{S}{2.5}[1 - e^{(-\eta X)}] \qquad [5]$$

where, S = scaling factor, and η = 0.2557

For a known daily total drainage (X - mm), mean daily phreatic level (H- m) can be estimated for winter rain events using Eq-6 (Fig. 21b). Drainage is measured in mm and the phreatic level is measured in metres.

$$H = \frac{S}{4.18}\left[1 - \frac{1}{2\left(1 + \left[\frac{S}{4.18}\right]^2 \varepsilon X\right)^{0.5}}\right] \qquad [6]$$

where, S= scaling factor, and ε = 1.006

3.11 Estimating the master recession curve

Once the initial peak phreatic level for a rain event is estimated, transient phreatic level can be calculated using a master recession curve. Master recession formulae are usually designed by using a linear or power type relationship between the phreatic level elevation and phreatic level decline rate for a given site. This was developed assuming the drainage is mainly controlled by the key parameters encapsulated in the black box which are site specific and independent of the seasonal variation. Drainage patterns of the measured recession data are shown in Figure 22.

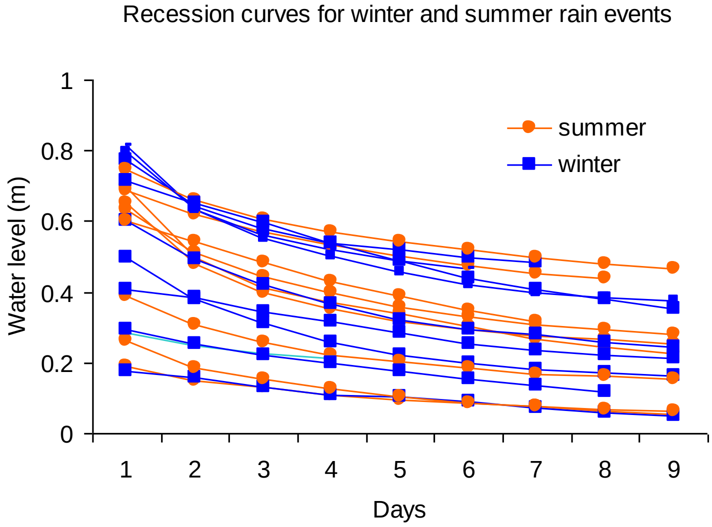

Figure 22 Recession curves at Korere for different seasons. A master recession formula was developed using a range of recession curve segments in Table Curve 3D®.

The empirical relationship given in Eq-7 was developed to predict the recession behaviour of the phreatic level following the initial phreatic levels given in Eqs-5&6.

$$W_{level} = S[\beta + \frac{\delta}{[t]^{0.5}}] \qquad [7]$$

where, β and δ are given by Eqs-8 and 9, which are found to be site and season independent. The site-dependent parameter S is related to the catchment area and the length of the hill slope to the interface boundary of the hillslope. β and δ parameters given in Eqs-8 & 9 reflect the initial phreatic level H at day = 0 found in Eqs-5 & 6. Time step t is taken as 1 day.

$$\beta = [-0.029 + 0.479H^2] \qquad [8]$$

$$\delta = [1.495 + 0.5248H] \qquad [9]$$

3.12 Model Calibration

Scaling factor

Extending the empirical model equations to predict the phreatic water levels at other sites requires a scaling factor. Most preferably, the scaling factor S should represent the geometry of the catchment recharge area and interface boundary between the foothill and shallow aquifer. The relationship between the scaling factor and the catchment /boundary length was developed using the data available at all 6 sites (e.g. Fig. 23). The catchment boundary and hill slope boundary interface for each hill slope site were manually plotted on ARC-View GIS maps to estimate the scaling factor given in Table 1 and Figure 24. The effect of scaling factor on the estimated phreatic level is shown in Figure 25. Saturated hydraulic conductivities were measured at Korere and Paratiho sites at up to 1.5 m depth. Saturated hydraulic conductivities were used in Eq-3 to estimate recharge to the aquifer.

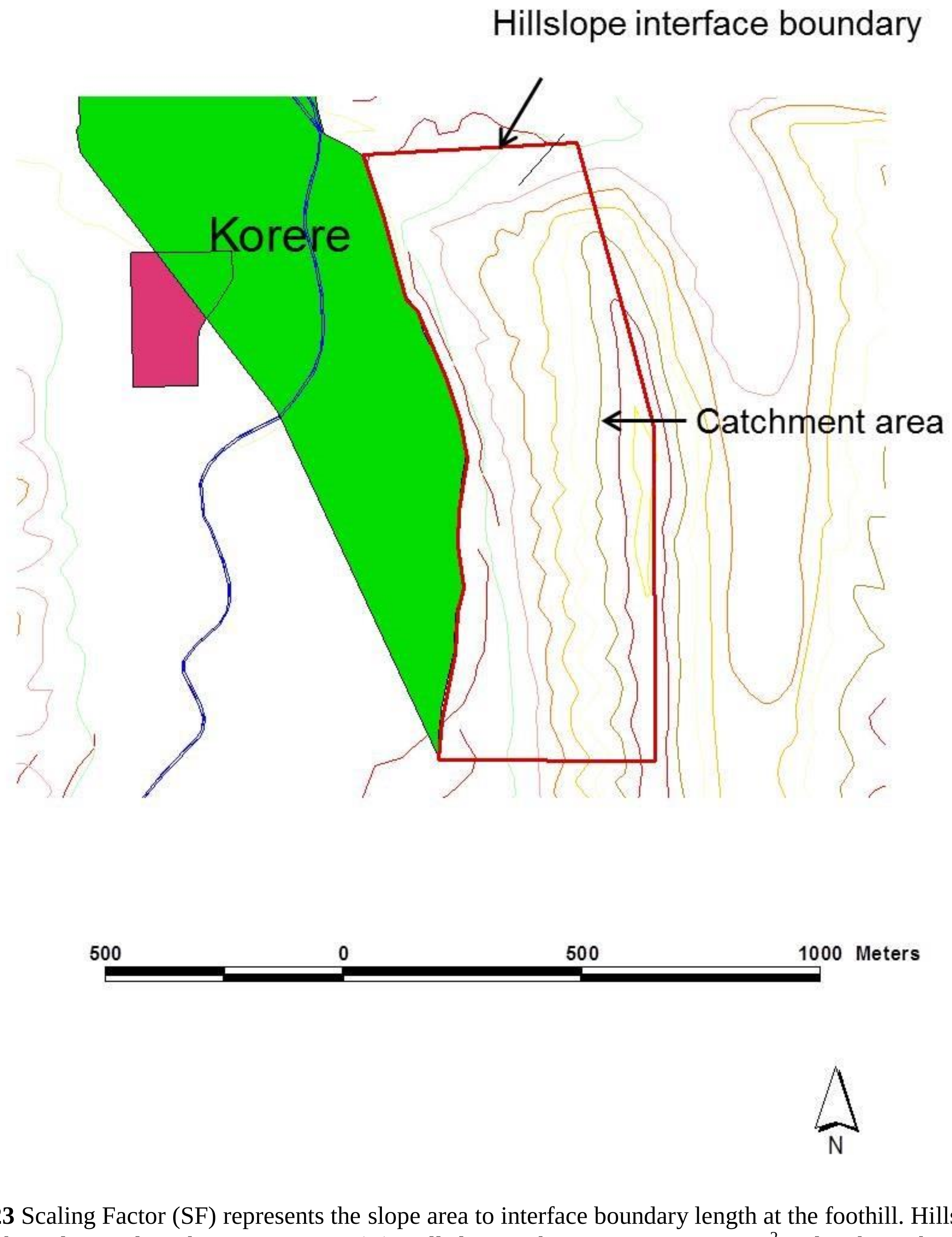

Figure 23 Scaling Factor (SF) represents the slope area to interface boundary length at the foothill. Hillslope recharge boundary and catchment at Korere (G). Hill slope recharge area = 524156 m^2 and recharge boundary length = 1273 m.

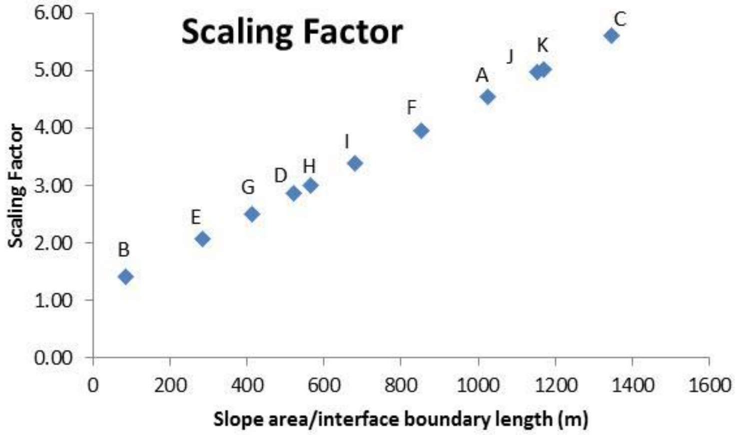

Figure 24 Scaling factor for selected hillslopes.

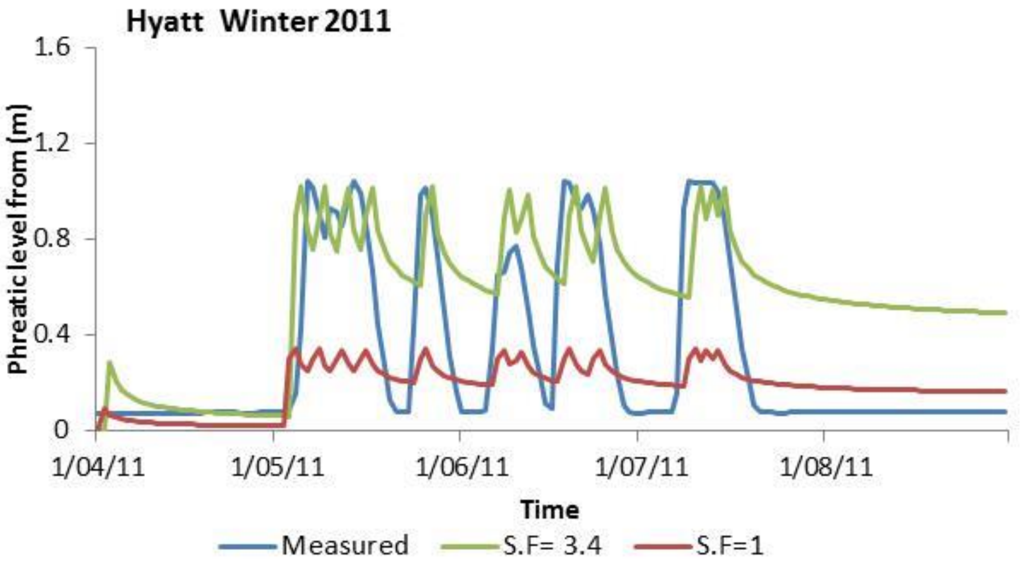

Figure 25 Effect of scaling factor on phreatic level. SF = direct representation of the slope area/interface boundary length.

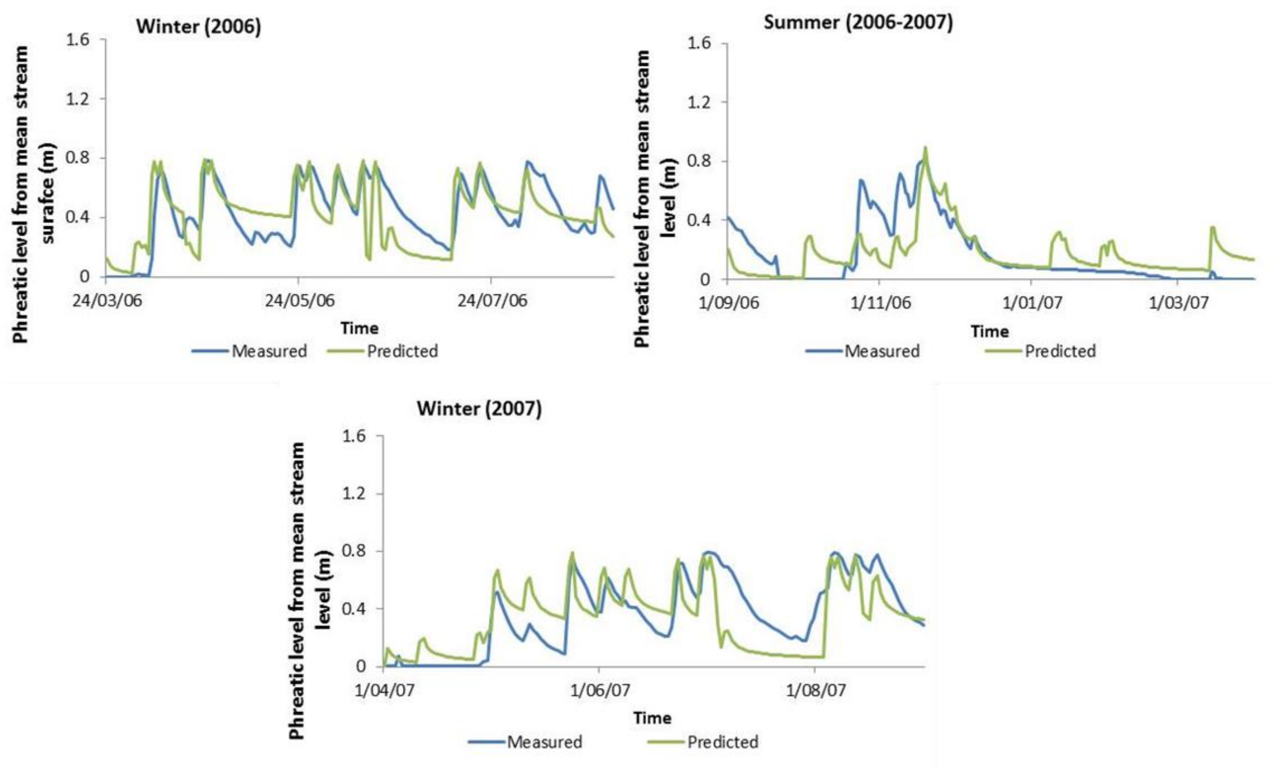

Figure 26 Model calibration for Korere hillslope. Summer and winter seasons use two different empirical formulae to estimate the peaks of the phreatic level.

Table 1 Hillslope geometry and soil hydraulic characteristics

Hillslope site	Hillslope	Boundary	Recharge slope area/ boundary length (m)	Scaling Factor (S)	Interface thickness				
	Catchment	Length			(m)		Permeability		Ksat
	Area (m2)	(m)				Zone	Indicator	Soil series	m/d
A	1408951	1377	1023.2033	4.54	3	2	M/R	hill soil	0.03
B - Cemetry	217174	2537	85.6	1.42	3	1	M	hill soil	0.04
C - Forestry	3978673	2960	1344.1	5.61	3	1	M/S	hill soil	0.02
D	693606	1334	519.9	2.86	3	2	M/R	hill soil	0.03
E	840032	2960	283.8	2.08	3	3	M/S	hill soil	0.03
F - Carson	1191970	1400	851.4	3.97	3	3	M/S	Silt loam	0.03
G (korere)	524156	1273	411.7	2.50	3	3	M/R	hill soil	0.04
H	1154175	2044	564.7	3.01	3	3	M/S	Silt loam	0.03
I - Hyatt	1514601	2225	680.7	3.40	3	3	M/S	hill soil	0.03
J	1813748	1574	1152.3	4.97	3	3	M/S	Silt loam	0.02
K	1595653	1364	1169.8	5.03	3	1	M	hill soil	0.04

4 Results & Discussion

Comparisons of predicted and measured phreatic levels at Carson (J), Hyatt(E), and Cemetery(B) sites are shown in Figure 27. Measured and estimated phreatic levels for both summer and winter seasons matched better at the Carson site than at the Hyatt and Cemetery sites. The main reason for poor agreement between measured and estimated phreatic levels at the Cemetery site was the use of rainfall data measured at Carson to predict phreatic levels at Cemetery. However, as seen in Figure 27, the model was effective at predicting the peak phreatic levels but poor in predicting the recession part of the transient phreatic levels that reflects the base flow component. Missing peaks of measured phreatic levels at Cemetery, which is 12 km from Carson, are due to local rainfall not being well represented by rainfalls at Carson.

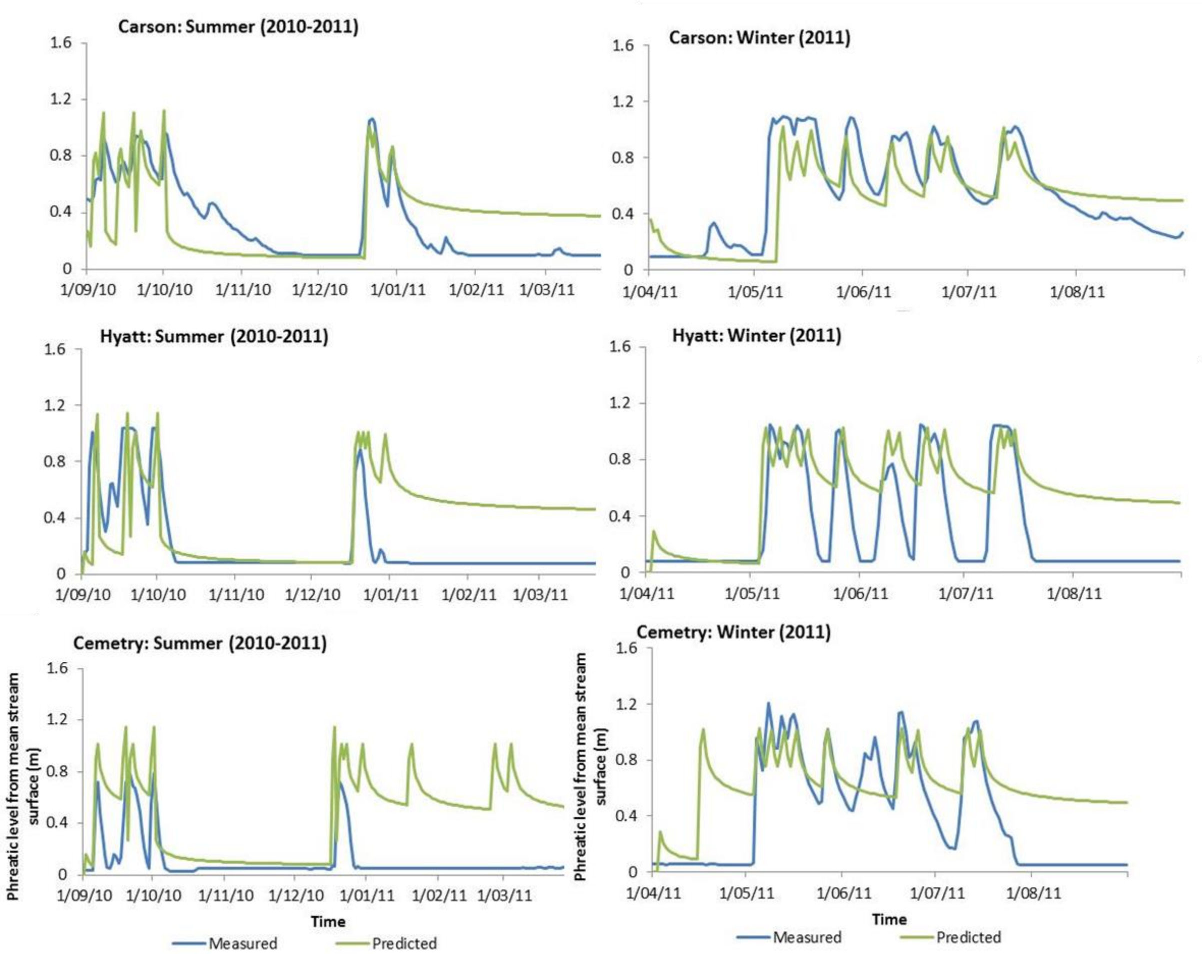

Figure 27 Application of the model to predict phreatic levels to 3 other hillslope sites where phreatic levels were measured. Rainfall measured at Carson was used for other sites. No phreatic levels were measured at forestry hillslope.

4.1 Extending the model outside Korere hillslope site

As shown in Figure 28, model prediction with onsite rainfall records at Paratiho hillslope, which is 39 km from the Korere hillslopes, was surprisingly successful. This suggests that to a certain extent both sites share the same key parameters encapsulated to reflect the hillslope hydrological process at Korere hillslope.

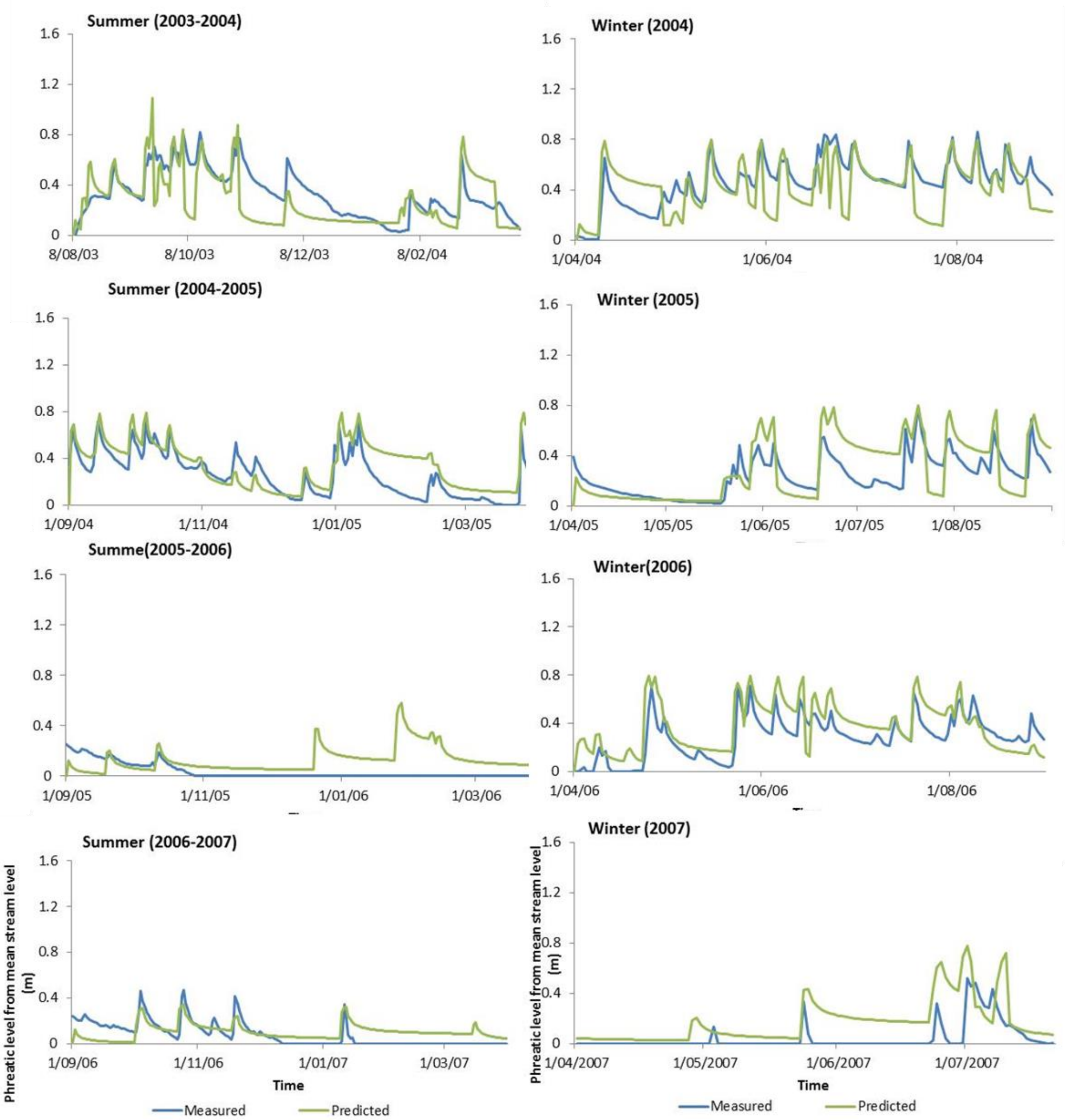

Figure 28 Successful application of the model to estimate phreatic levels at Paratiho site which is 39 km away from Korere hillslope.

Figure 29 (a,b,c,d & e) show the estimated and measured phreatic levels at the foothills for all 5 sites for the entire monitoring time period. Figure 30 shows the potential hillslope recharge estimated from Eq-3 to the shallow unconfined aquifer of the Motueka river valley.

Estimation of the recharge is determined by the foothill aquifer interface geometry and saturated hydraulic conductivities described earlier and in Eq-3. The model is better at predicting peak phreatic levels, and therefore peak recharge, than the phreatic recessions which need further work.

Phreatic levels were not present during the monitoring period at the forestry(C) site.

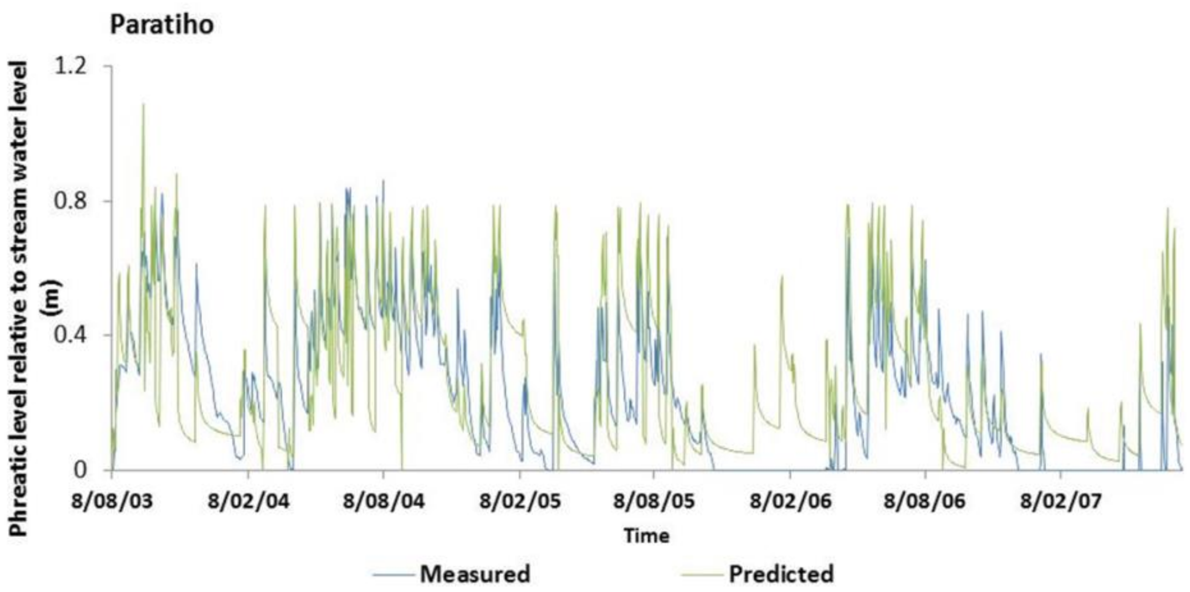

Figure 29(a) Phreatic levels measured and predicted at Paratiho site.

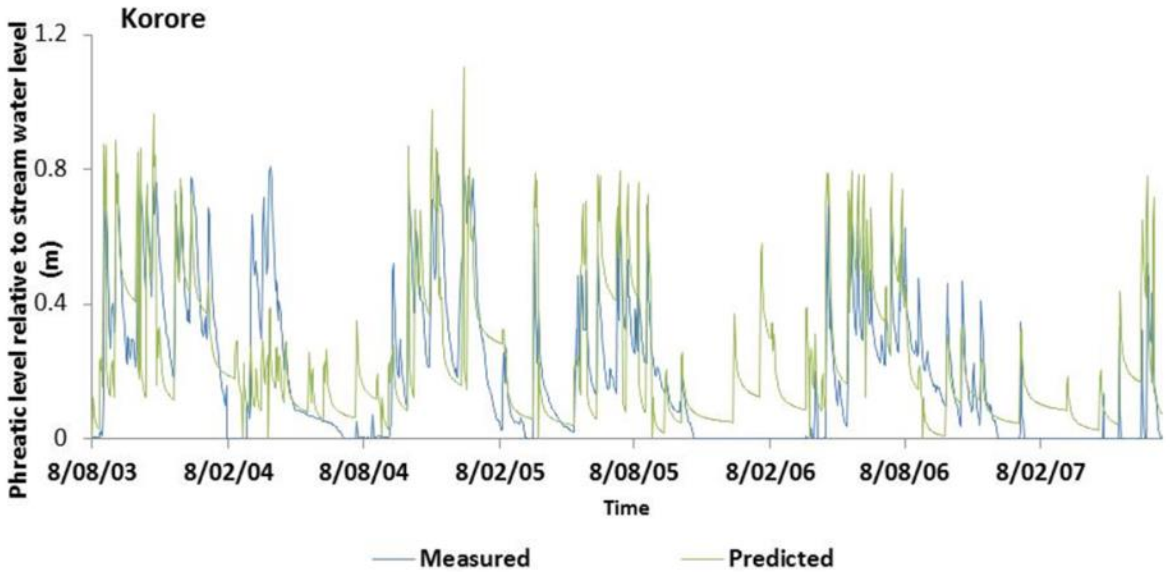

Figure 29(b) Phreatic levels measured and predicted at Korere site.

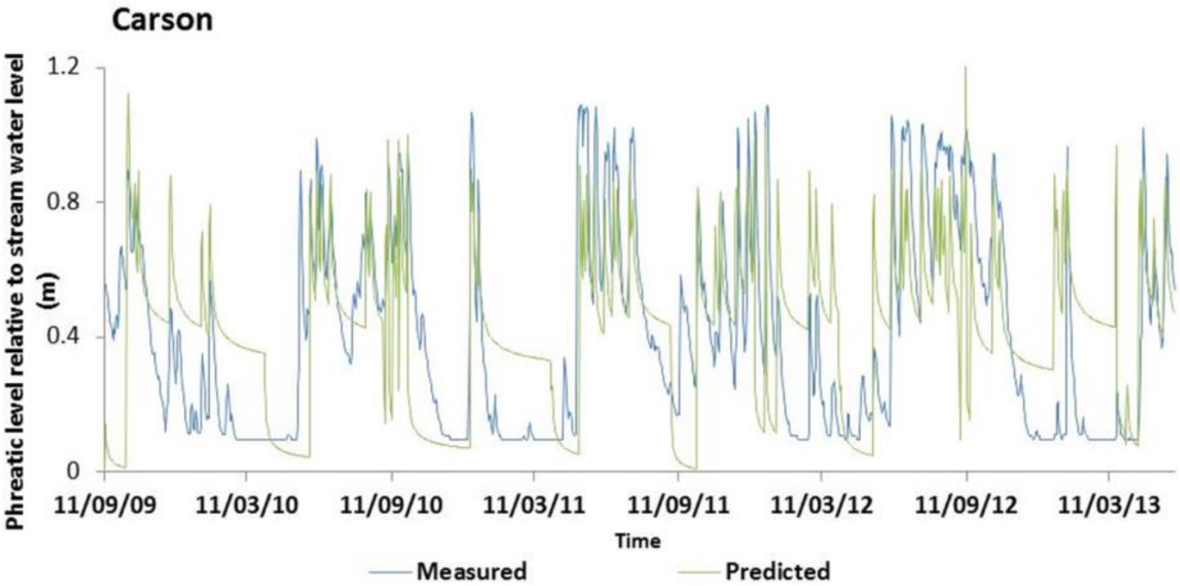

Figure 29(c) Phreatic levels measured and predicted at Carson site.

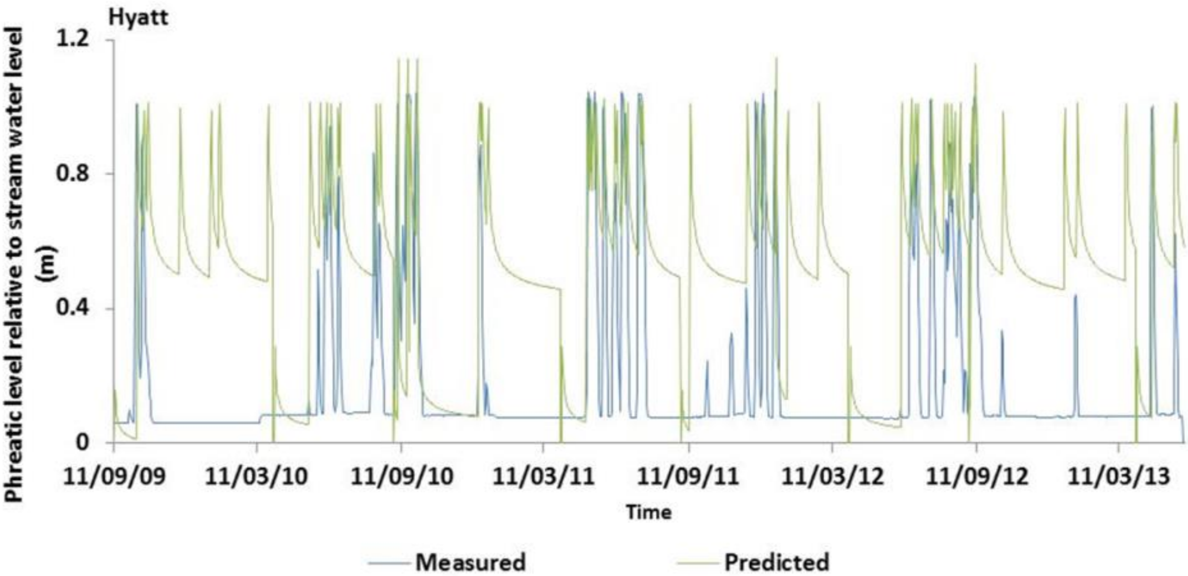

Figure 29(d) Phreatic levels measured and predicted at Hyatt site.

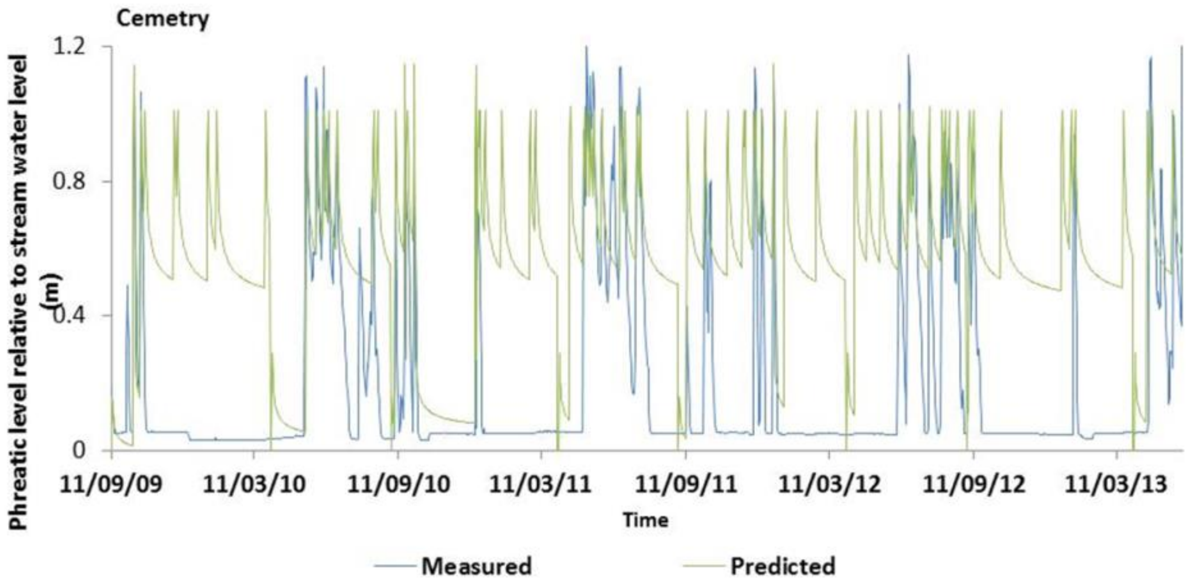

Figure 29(e) Phreatic levels measured and predicted at Cemetery site.

Summary of method for estimation of recharge

1. Obtain time series of daily rainfall record.

2. Estimate the scaling factor S by defining the recharge interface between hillslope and foothilland the area of the contributing hillslope catchment . We used Arc View software to generate contour maps of the hillslope area to estimate the catchment boundary. See Figure 23.

3. Time series rainfall data are then used to estimate the phreatic levels using Eq-4, Eq-5, and Eq-6 for summer and winter periods. Use the appropriate site-specific scaling factor. Now we have a time series of phreatic levels based on rainfall events.

4. Estimate the recession part of the hydrograph for each phreatic level (for each rain event) using Eq-7 to produce a time series of phreatic levels.

5. Estimate the appropriate number of potential drops and flow tubes by manually inspecting the foothill recharge interface and average phreatic levels for summer and winter seasons. These parameters must be selected to form approximate square shapes, as described in the theory section.

6. Use the above parameters and measured saturated hydraulic conductivities in Eq-3 to estimate recharge.

A sample calculation of recharge estimation for the Paratiho site on 30 May 2004 is given in Table 2 below. A total of 90 mm rain was recorded on this day at the Paratiho site. Estimated phreatic level from model Eq-6 is used in Eq-3 to estimate the recharge.

Table 2 Sample calculation of recharge estimation for the Paratiho site, 30 May 2004

Date	Rainfall (mm)	Phreatic level (m)	Stream w.level (m)	No of potential drops	No of flow tubes	Sat. H.Conductivity (m/day)	Recharge m^3/day/m
30-05-04	90	103.67	102.90	05	15	0.01	0.023

The number of potential drops and flow tubes must be decided by drawing a flow net at the recharge interface to form an approximate square shape. In this study the number of potential drops and flow tubes were decided by visual inspection of phreatic levels and the stream water level.

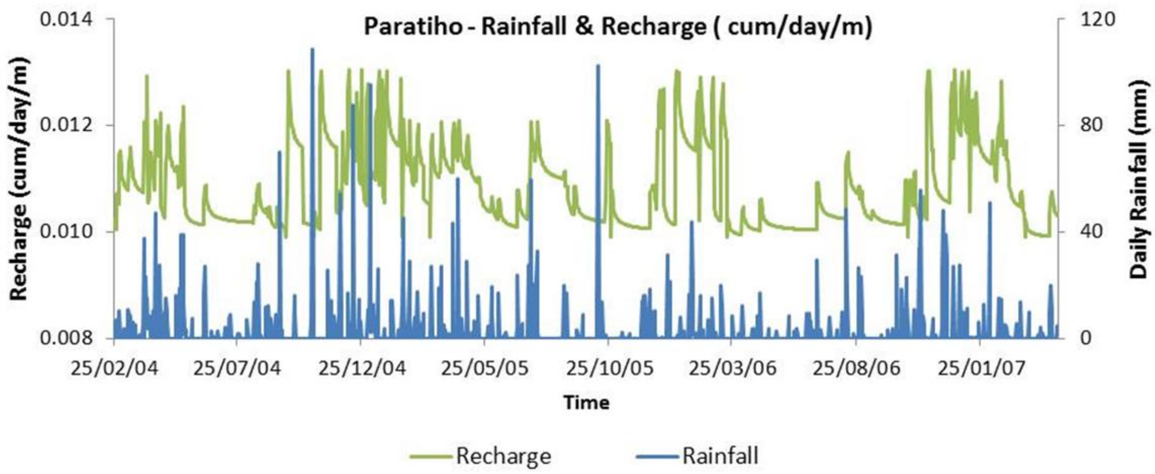

Figure 30(a) Estimated recharge at Paratiho.

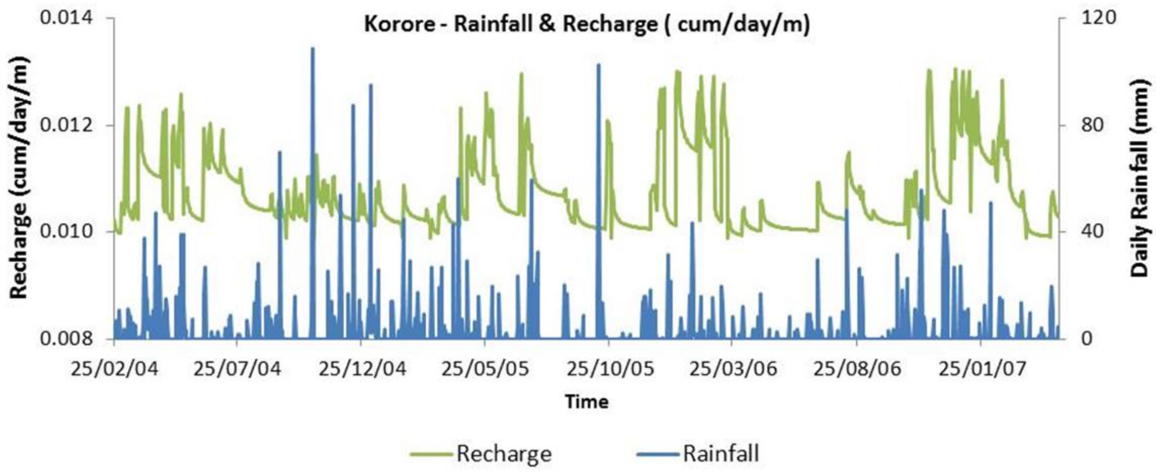

Figure 30(b) Estimated recharge at Korere.

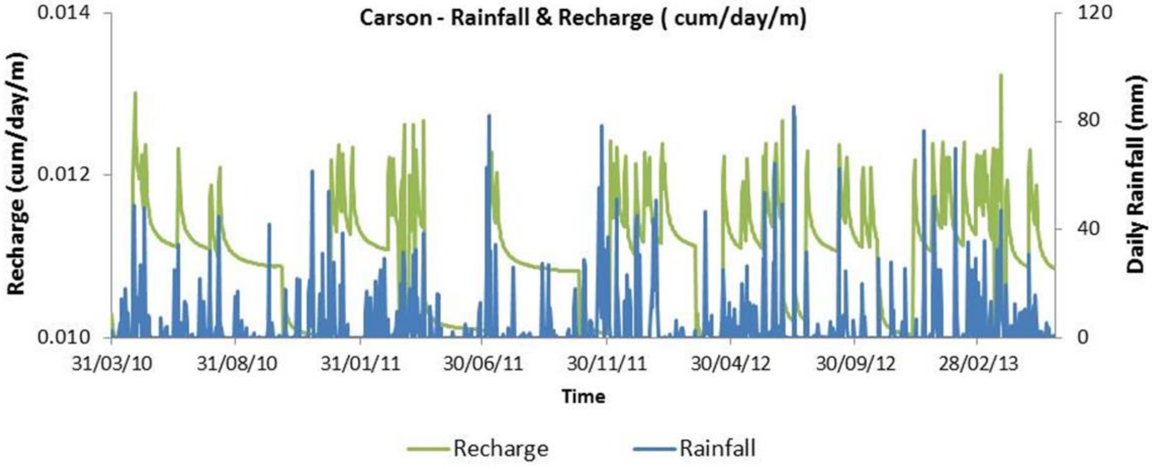

Figure 30(c) Estimated recharge at Carson.

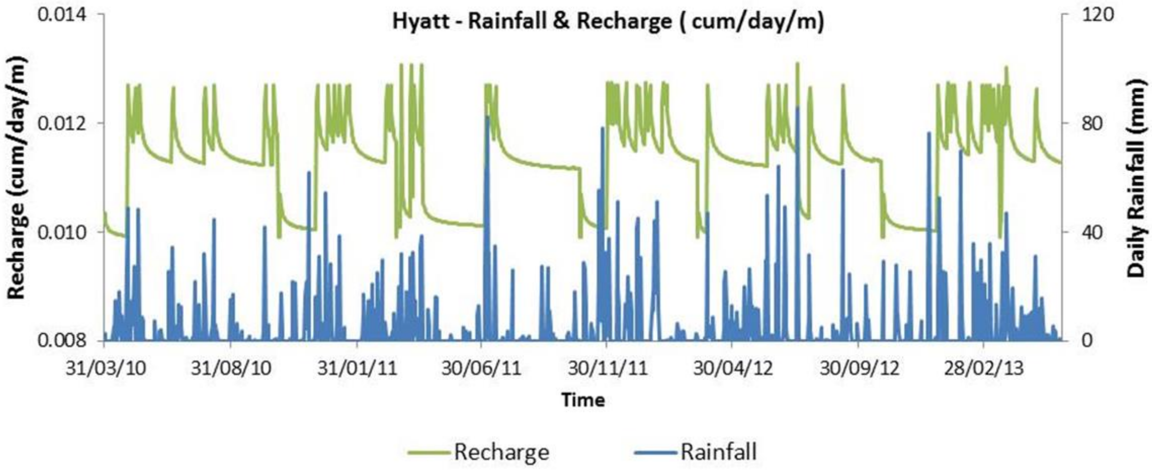

Figure 30(d) Estimated recharge at Hyatt.

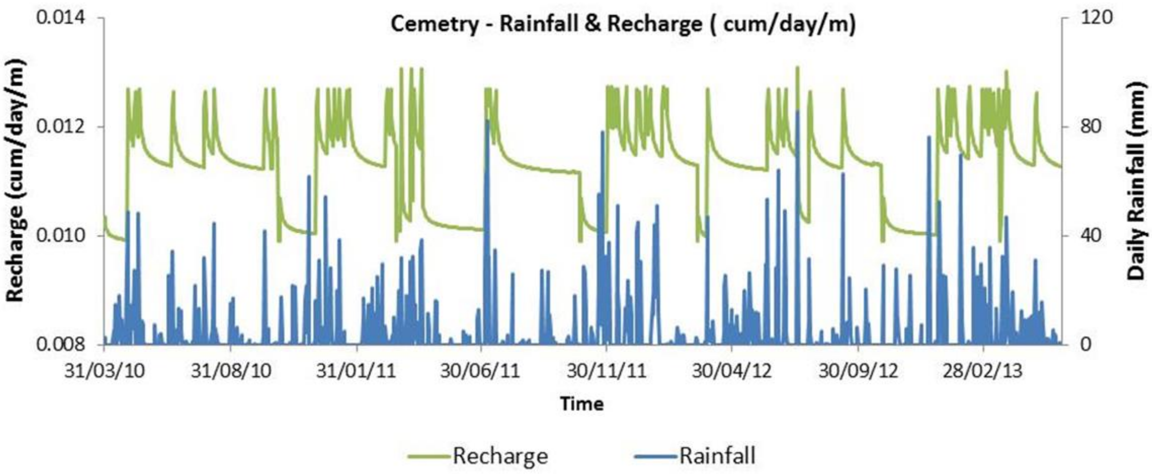

Figure 30(e) Estimated recharge at Cemetery.

5 Conclusion

A tool to estimate climatic driven phreatic levels at the foothill-shallow aquifer interface was developed using modelled relationships between rainfall and drainage, and then between drainage and phreatic level. Empirical relationships were developed in a black box using measured recession curves that reflect the actual hillslope process. The validity of the model was successfully tested by comparing estimated phreatic levels with measured values for four hillslope sites. The model was better at predicting peak recharge rates than recharge rates during flow recessions, because the phreatic recessions were not steep enough at some sites. More work is needed on this aspect of the model if it is to be used elsewhere.

Recharge to the shallow aquifer is assumed to be directly proportional to transient hydraulic gradient created by the phreatic levels at the foothill. The same methodology may be extended to build simple tools to estimate climate-induced stream flows, thus avoiding complex models that demand physical parameters that are difficult to obtain.

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