A CONCEPTUAL UNDERSTANDING OF GROUNDWATER RECHARGE PROCESSES AND SURFACE-WATER/ GROUNDWATER INTERACTIONS IN THE KRUGER NATIONAL PARK

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A thesis submitted in fulfilment of the requirements for the degree of Master in the Faculty of Natural Sciences, Department of Earth Sciences, University of the Western Cape, Bellville, South Africa.

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Abstract

Key words: Recharge processes, Kruger National Park, Surface-water/ groundwater interactions, CRD, stable isotopes

In the Kruger National Park (KNP) which is the flagship conservation area in South Africa, the impact on groundwater should be kept to a minimum as groundwater plays a vital role in sustaining ecosystem functioning and sustaining baseflow to streams and rivers. For this reason groundwater has been recognized as one of the environmental indicators that need to be monitored.

The KNP has adopted a Strategic Adaptive Management (SAM) approach with clear ecosystem management goals. The achievement of these goals is evaluated by using environmental indicators. These indicators are evaluated against thresholds of potential concern (TPC). TPCs are a set of boundaries that together define the spatiotemporal conditions for which the KNP ecosystem is managed. TPCs are essentially upper and lower limits along a continuum of change in selected environmental indicators.

Historically, groundwater recharge and surface water interaction with rivers has tended to be overlooked in the KNP. This study proposes a conceptual model of

groundwater recharge processes in the KNP, defining when and how groundwater recharge occurs.

Two methods were used, the Cumulative Rainfall Departure (CRD) and stable isotopes of ^{2}H and ^{18}O. An adapted version of the CRD which incorporates a long and short term memory of the system was used to identify possible recharge processes. Further, using the CRD method a reliable reconstruction of the long term groundwater level trends are simulated using monthly rainfall totals with reference to the average rainfall over the entire time series 1936-2009.

The stable isotope of ^{2}H and ^{18}O samples from cumulative rainfall samplers, surface-water (streams and rivers) and groundwater from boreholes were collected monthly for approximately one year (May 2010 to July 2011). The isotope composition of the groundwater was used to establish whether recharge was immediate or delayed. Additionally, the isotopic composition of surface-water from rivers and streams were compared to that of groundwater to identify surface-water interactions.

Groundwater recharge in KNP occurs during the rainy summer months (December to March) and very little to none during the dry winter season (April to September). Recharge takes place during rainfall sequences 100mm or more. The stable isotope records collected from cumulative rainfall, groundwater and surface water (streams and rivers) indicate that groundwater experiences evaporation prior to infiltration. As the KNP experiences high evaporation rates, insignificant rainfall sequences contribute little or zero to recharge. The CRD analysis of groundwater level fluctuations shows that recharge to the aquifers respond to dry and wet cycles that last for 6 to 14 years. The KNP experienced several periods of below-average rainfall and hence no significant recharge took place to the basement aquifers. During a normal rainy season the water levels rise somewhat then starts receding again. It is only during major rainfall events that may occur every 100yrs to 200yrs causing the aquifers to fully recharge. This was perfectly illustrated by the high groundwater levels after the 2000 major rainfall event that recharged the aquifers fully. During below average rainfall years the overall water level trend is drastically declining. The system experiences higher natural losses than gains due to outflow of groundwater to streams and rivers.

The KNP is divided down the center by two geological formations, granites along the west and basalts along the east. The combination of the CRD model and the stable isotopic analysis suggest that the dominant recharge processes that occur in the southern region of the KNP are direct recharge via piston flow and indirect recharge via preferred pathways particularly streams and rivers. Along the eastern half of the KNP on the Basalts and Rhyolite direct recharge via piston flow are dominant. Groundwater is not recharged via small streams and rivers (Sweni and Mnondozi Rivers) as it was found that at these particular sites these rivers are detached and do not interact with groundwater. Along the western granitic areas the dominant recharge process are indirect recharge. Recharge takes place via preferred pathways particularly streams and rivers. It was found that ephemeral rivers (Nwatsisonto River) act as sinks for groundwater recharge and influent-effluent conditions are experienced along seasonal rivers (Mbyamiti River). The large perennial Sabie and its tributary the Sand River are consistently fed by groundwater, above all maintaining base flow during the dry season. These rivers act as basin sinks receiving groundwater discharge all year round.

Using the stable isotope composition of rainfall, surface-water and groundwater to act as a natural tracer, in combination with the CRD method proved invaluable to confirm the plausible recharge processes. The study provided a conceptual understanding of the groundwater system in the KNP forming the foundation to developing acceptable limits (TPCs) of the groundwater levels in the KNP. The model will serve as a guide for the recharge processes and for deciding on the location and time frames for data collection to ultimately set TPCs for groundwater in the KNP to sustainably manage the resource.

DECLARATION

I declare that **A CONCEPTUAL UNDERSTANDING OF GROUNDWATER RECHARGE PROCESSES AND SURFACE-WATER / GROUNDWATER INTERACTIONS IN THE KRUGER NATIONAL PARK** is my own work, that it has been submitted for any degree or examination in any other university, and that all the sources I have used or quoted have been indicated and acknowledge by complete references.

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Date: 15th May 2012

Signed:

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CHAPTER 1
INTRODUCTION

1.1. Background

Gaining a conceptual understanding of groundwater recharge processes and mechanisms is necessary for the sustainable utilisation of groundwater resources with particular reference to abstraction management and water allocation. To achieve sustainable utilisation, a better understanding of groundwater recharge processes and groundwater/surface-water interactions is needed.

In the Kruger National Park (KNP), which is the flagship conservation area in South Africa, the impact on groundwater should be kept to a minimum as it plays a vital role in sustaining Groundwater Dependent Ecosystems' (GDEs) functioning (Colvin *et al*., 2011; Colvin *et al*., 2008; February *et al*., 2007) and sustaining baseflow to streams and rivers. Groundwater is used to supply wildlife with water provisioning especially during periods of drought. A number of tourist camps and facilities situated in the park also consume large amounts of water. For these reasons, groundwater has been recognized as one of the environmental indicators that need to be monitored.

The KNP has adopted a Strategic Adaptive Management (SAM) approach with clear ecosystem management goals (Biggs *et al*., 2003). The achievement of these goals is evaluated by using environmental indicators. These indicators are evaluated against thresholds of potential concern (TPCs). TPCs are a set of boundaries that together define the spatiotemporal conditions for which the KNP ecosystem is managed. TPCs are essentially upper and lower limits along a continuum of change in selected environmental indicators. When the upper or lower TPC levels are reached, or when modeling predicts that they will soon be reached, an assessment is prompted in order to determine the cause of the extent of change. This assessment provides the basis for deciding whether management action is needed to reverse the change or whether the TPC should be recalibrated in light of new knowledge and/or understanding (Biggs *et al*., 2003).

Historically, groundwater recharge and surface- water interaction has tended to be overlooked in the KNP. The objective of this study is to gain an initial understanding (conceptual model) of the groundwater system within the southern region of the KNP, in particular determining groundwater recharge processes and surface-water interactions. A conceptual model of these processes and interactions would inform monitoring and management decisions related to the management of groundwater resources. The overall management of groundwater will therefore take into account the natural interaction/movement/exchange between groundwater and surface-water, including baseflow of rivers in dry periods and recharge of aquifers during wet periods. This research will provide the foundation for future groundwater research in KNP by providing a baseline to support management decisions particularly with respect to the first steps towards defining a TPC.

1.2. Aims and Objectives

The following main aims have been identified for this study

- To develop a conceptual model of groundwater recharge processes and surface-water/groundwater interactions in the southern region of the KNP.
- To provide guidance for developing 'Thresholds of Potential Concern' (TPCs) in order to assist the KNP in making sound management decisions concerning groundwater.

To address these objectives the following key questions will be addressed

- What are the dominant groundwater recharge processes?
- Are there any interactions between surface-water and groundwater, with particular reference to streams and rivers?
- What are the seasonal and long term trends of groundwater level behaviour in correlation with rainfall?

1.3. Scope and outline of the thesis

Chapter 1 provides the background and outlines the objectives and scope of the thesis.

Chapter 2 provides the theoretical background upon which this study is based. It overviews the types of groundwater recharge and their associated processes and recharge estimation methods used particularly in arid to semi arid areas. Methods that were used in this particular study are discussed in detail. These include the Cumulative Rainfall Departure (CRD) method and the theory and application of stable isotopes ^{2}H and ^{18}O.

Chapter 3 presents an overview of the KNP region in terms of the location, drainage patterns, climate, rainfall, topography, geology, soils, vegetation and geo-hydrology. Descriptions of the selected study sites within KNP are provided.

Chapter 4 gives an account of the application used to gain insight into potential recharge processes and the short- term and long- term groundwater level trends in correlation with rainfall using the CRD method.

Chapter 5 provides an account of using a tracer approach of stable isotopes ^{2}H and ^{18}O to evaluate groundwater recharge processes and to determine if there is bi-directional movement between groundwater and surface-water, particularly that of streams and rivers.

Chapter 6: Finally, a synthesis of the main findings from this study is given with some recommendations for future research work.

CHAPTER 2

Literature review of groundwater recharge and surface-water groundwater interactions related to the research questions

2.1. Introduction

This chapter aims to provide a definition of groundwater recharge, an overview of groundwater recharge processes and its interaction with surface-water. Further, the chapter introduces the basic concepts and methodologies in groundwater recharge and surface-water interaction studies and provides a detailed discussion of the methods used in this study.

2.2. The hydrological cycle

The hydrological cycle which consists of many varied and interacting processes involving rainfall, surface-water and groundwater are illustrated in Figure 2.1. Water evaporates from the oceans and land surfaces to become water vapour that is carried over the earth by atmospheric circulation. The water vapour condenses and precipitates on the land and oceans. The precipitated water vapour may be intercepted by vegetation, become overland flow over the land surface, infiltrate into the ground, flow through the soil as subsurface flow, or discharge as surface runoff (Todd and Mays, 2005).

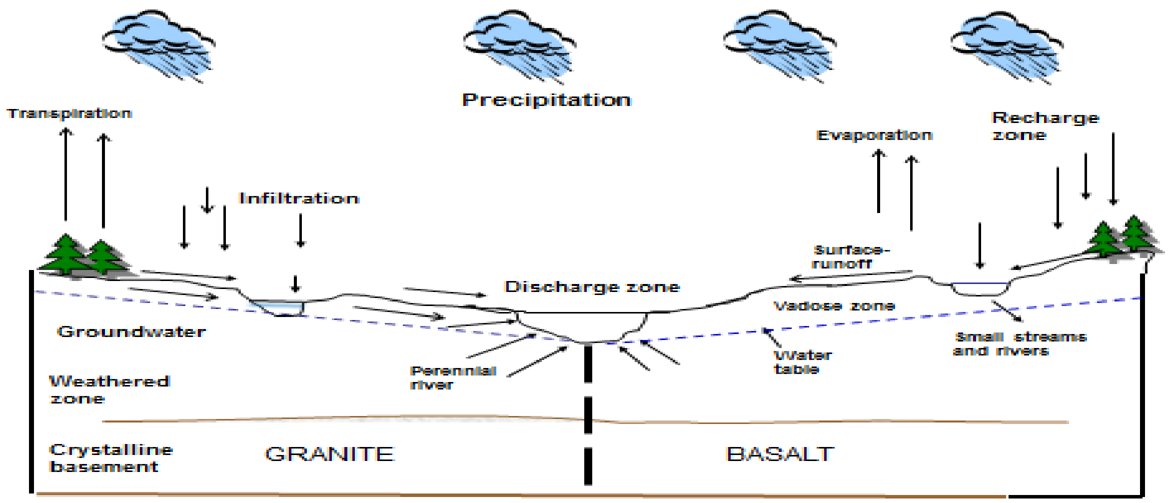


Figure 2.1 A conceptual representation of the hydrological cycle in the KNP (adapted from Todd and Mays, 2005).

Groundwater flow is but one part of this complex dynamic hydrologic cycle. The aquifers and confining beds that underlie any area comprise the ground-water system of the area (Heath, 1983). Hydraulically, this system serves two functions: it stores water to the extent of its porosity, and it transmits water (Heath, 1983). Groundwater is not static. It is part of a dynamic flow system where water enters groundwater systems in recharge areas and moves through them, as dictated by hydraulic gradients and hydraulic conductivities to discharge areas (Heath, 1983). It moves into (vertical direction) and through (horizontal direction) aquifers from areas of high water-level elevation to areas of low water-level elevation (Schwartz and Zang, 2002). The groundwater response or water level rise at a given point in an aquifer depends on the distance from the recharge area and the rate at which groundwater moves through the landscape which is dependent on the transmissivity, hydraulic conductivity and storativity of the aquifer system (Kirchner, 2003). Figure 2.2 shows schematically the high and immediate response of the water levels near the recharge area and the delayed lesser response farther away (Kirchner, 2003). Figure 2.3 illustrates that ground-water flow paths vary greatly in length, depth, and travel time from points of recharge to points of discharge in the groundwater system (Winter *et al*., 1998).

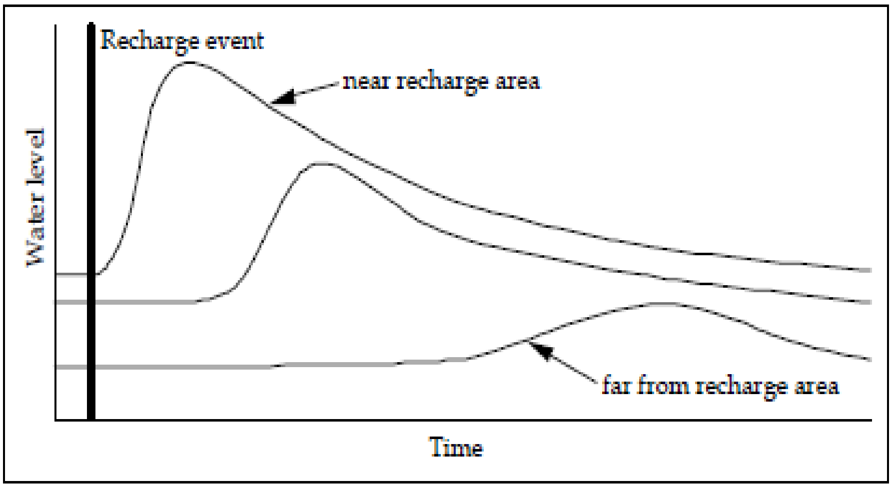


Figure 2.2 Shows schematically the high and immediate response of the groundwater levels near the recharge area and the delayed lesser response farther away (Kirchner, 2003).

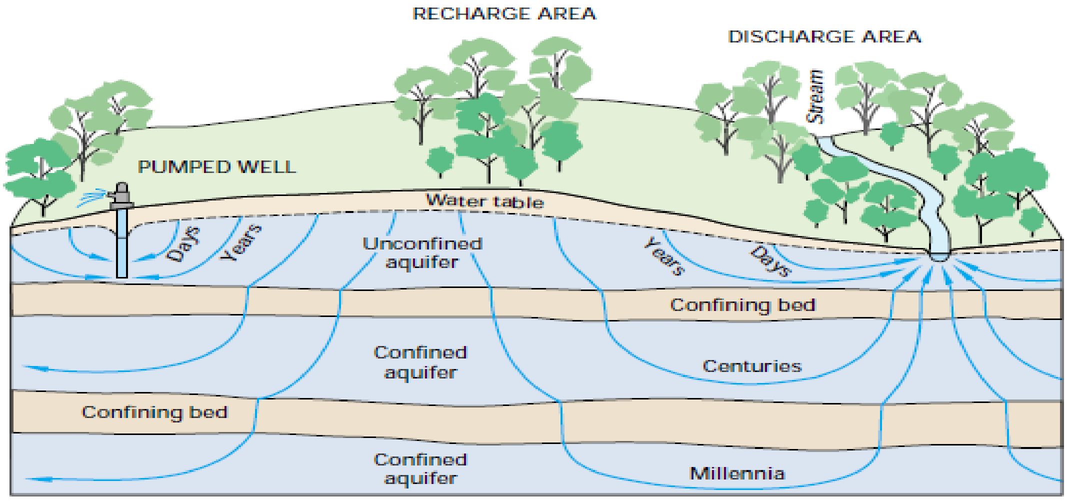


Figure 2.3 Illustrates that ground-water flow paths vary greatly in length, depth, and travel time from points of recharge to points of discharge in the groundwater system (Winter *et al*., 1998).

2.3. Definition and processes of groundwater recharge

When characterizing groundwater recharge, a distinction between potential and actual recharge needs to be made. Potential recharge is soil-water that percolates below the root zone, whereas actual recharge is soil-water that reaches the aquifer. Most potential recharge water will be stored in the vadose zone at negative pressures (suctions) and is not available for exploitation. Conversely, actual recharge is the amount of water that in fact reaches the water table, and can be exploited (Sophocleous, 2003). Therefore, for the purpose of this study groundwater recharge will be defined in very simple terms, as the portion of rainfall which reaches the water table thereby forming an addition to the groundwater reservoir. This is irrespective of whether it follows a preferential flow path via fractures, or drains through a column of soil, or infiltrates from free water in river channels or local surface depressions (Bredenkamp *et al*., 1995).

Natural sources of possible groundwater recharge in KNP include recharge from precipitation and river channels (including perennial, seasonal, and ephemeral flows). Recharge processes as classified by Lerner et al. (1990), De Vries and Simmers (2002); Healy and Scanlon (2010) and Scanlon et al. (2002) are summarised in Figure 2.4 which illustrates the process and three types of rainfall recharge to groundwater which is determined by factors which occur in almost the entire hydrological cycle. According to Sun (2005) these factors are interrelated, complex and natural phenomena, which are governed by the natural laws of conservation of energy, mass and momentum.

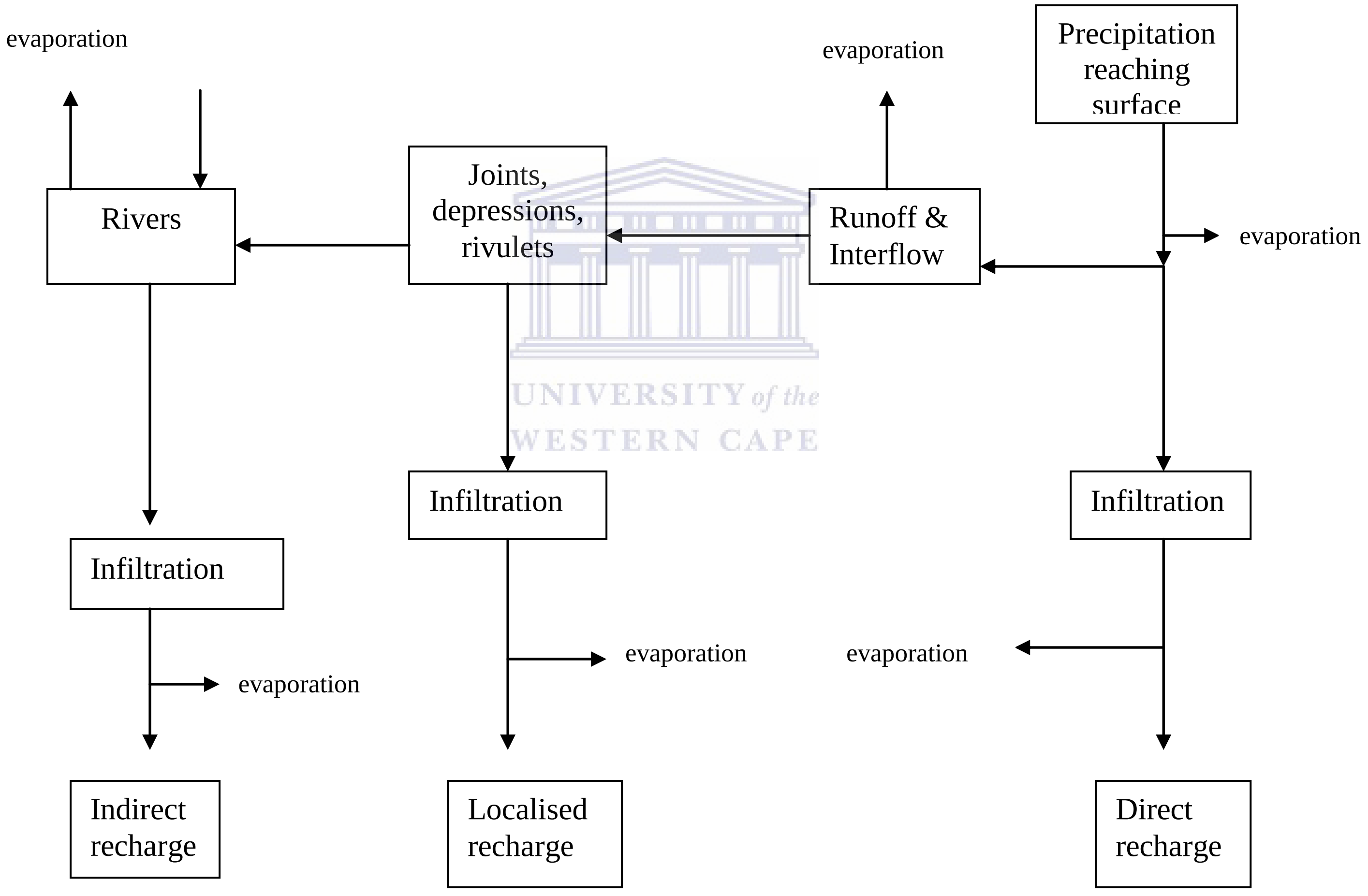


Figure 2.4 Illustrates the three types of recharge. Direct recharge refers to replenishment of the ground water reservoir from precipitation after subtracting interception, runoff and transpiration. Localised recharge results from the ponding of surface water in the absence of well-defined channels of flow whilst indirect recharge refers to percolation to the water table from surface-water courses (Lerner et al., 1997).

Direct recharge is water added to the groundwater reservoir in excess of soil moisture deficits and evapotranspiration by direct vertical percolation through the unsaturated zone. This normally occurs as piston-type flow or translatory flow, a process whereby precipitation that is stored in the unsaturated zone are displaced downwards by the next infiltration or percolation event without disturbing the moisture distribution. Indirect recharge is the movement of water from surface-water bodies, such as streams and rivers, to an underlying aquifer. Localised recharge is defined as concentrated recharge from small depressions, joints or cracks.

2.3.1. Groundwater level response

As water moves from the surface through the unsaturated zone, a proportion will be lost to evaporation, some will be taken up by plants (evapotranspiration) and some will remain within the unsaturated zone (de Vries, 2002). Thus, according to Xu and Beekman (2003), only part of infiltrated rainfall breaks through the zone where evapotranspiration occurs and percolates to the groundwater table i.e. recharge. The arrival time of this water at the water table is delayed due to 3-dimensional spreading of moisture. The duration of the recharge event is therefore prolonged with increasing thickness of the unsaturated zone (Figure 2.5). For that reason, water that reaches the water table may not necessarily result from a single rainfall event, but may represent a series of preceding rainfall events. Depending on the characteristics of the aquifer, that portion of rainfall that reaches the water table may cause a rise in water table levels and subsequently result in an increase in discharge down-gradient (Xu and Beekman, 2003). These processes determine a rainfall threshold above which groundwater recharge can effectively occur. Rainfall amounts below the recharge threshold will make little or no contribution to recharge.

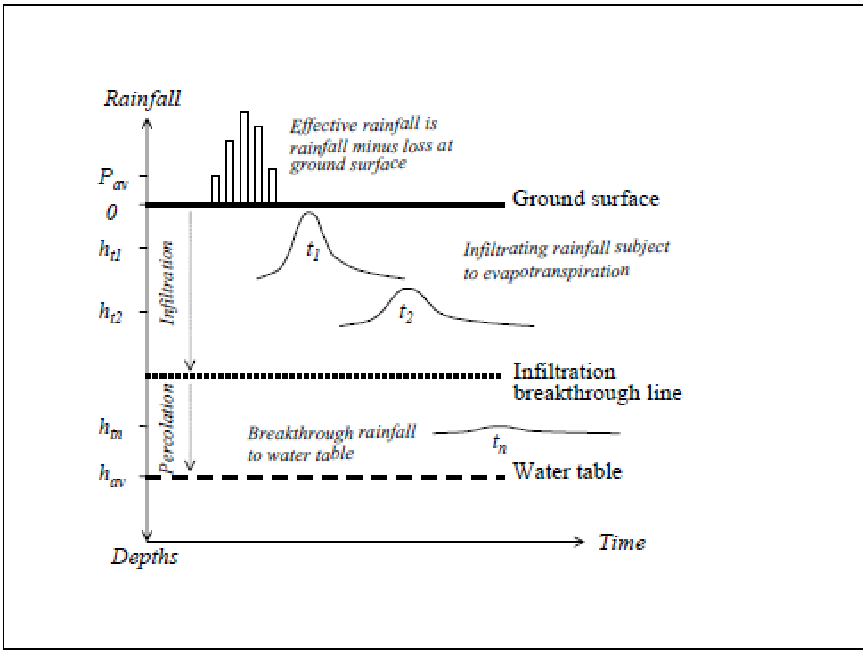


Figure 2.5 Schematic diagram of the rainfall infiltration process (Xu and Beekman, 2003).

The groundwater level response to rainfall recharge is a function of the depth to the water level, structure and texture of the unsaturated zone and the characteristics of the rainfall. The following types of responses can be distinguished (Xu and Beekman, 2003):

- **Rapid response:** within hours, days or a month of intensive rainfall. In this case, recharge normally occurs via preferential flow paths.
- **Intermediate response:** over a time span of a year or two. In this case, recharge normally occurs through direct and indirect flow paths,
- **Slow response:** taking longer than the above. In this case, recharge normally occurs through direct flow paths.

2.4. Groundwater/surface-water interaction

A conceptual model of recharge processes needs to consider the surface-water and groundwater flow systems and how they are linked. In order to prescribe surface-water/groundwater interaction namely recharge of groundwater by surface-water and discharge of groundwater to surface water. Lerner (2003), Vegter *et al*. (2003) and Xu *et al*. (2002) classified stream types according to stream flow characteristics, which are usually divided into three classes, namely ephemeral, seasonal and perennial which are schematically represented in Figure 2.6. They further build on this classification to propose a classification that considers the characteristics of the groundwater/surface water interaction in that stream or river. The classification is based on the stream or river's relation to the piezometric head in the aquifer system with which it is in hydraulic connection and indicates whether water is moving into or out of the aquifer (Figure 2.7).

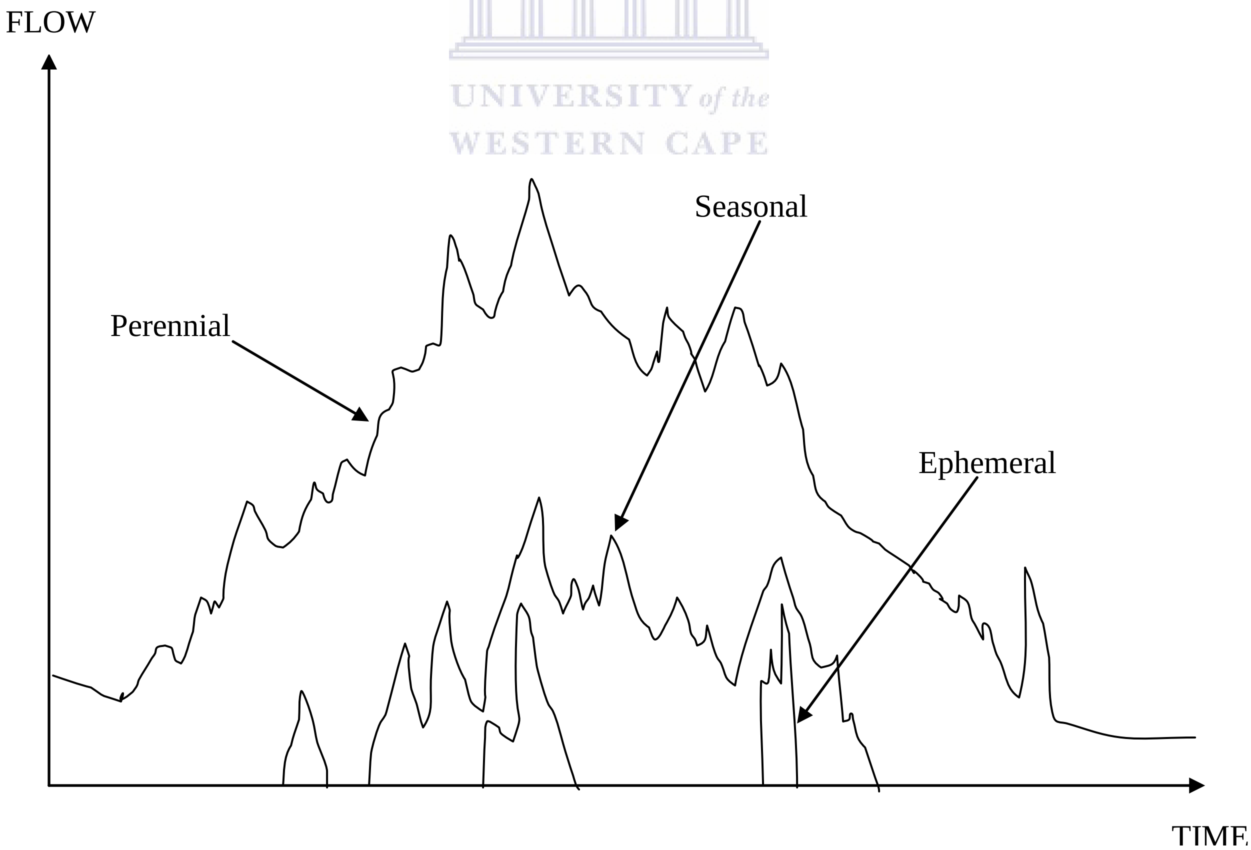


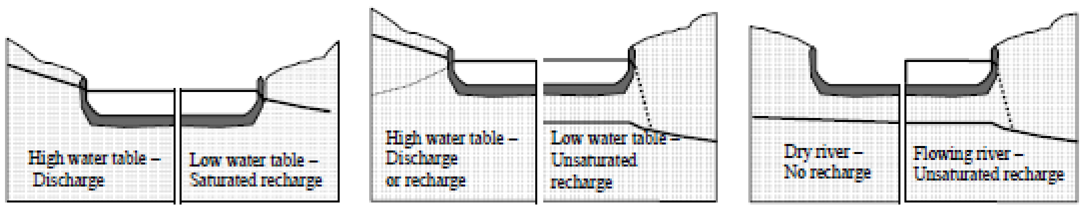
Figure 2.6 Classification of rivers by flow characteristics (Lerner, 2003).

2.4.1. Classification of streams and rivers according to vertical position

A. The piezometric surface slopes laterally down towards the stream (always connected). This typically is a perennial river. Groundwater reaches and emerges into the stream at all times. The piezometric surface at the stream is permanently above the stream stage and the material of the streambed is porous or fractured. These streams usually act as catchment drains.

B. The piezometric surface fluctuates alternately above and below the stream stage (intermittent). This is typically a seasonal stream. The stream material is underlain or bordered by alluvial deposits and/or porous decomposed rock. This stream is alternately influent and effluent.

C. The piezometric surface is at all times below the streambed level (detached). This is typically ephemeral stream conditions. Depending on the streambed material and piezometric surface, these streams can act as a sink and recharge groundwater if the stream bed material is pervious and the piezometric surface slope downward away from the stream. They contribute very little or no recharge if the streambed is impervious i.e. a remote or detached stream.



(A) **(B)** **(C)**

Figure 2.7 Classification of rivers by vertical positioning to the groundwater table (A) always connected, (B) intermittent and (C) detached (Lerner, 2003).

2.4.2. Bank storage

During the flood period of a stream, groundwater levels are temporarily raised near the channel by inflow from the stream. The volume of water stored and released after the flood is referred to as bank storage (Todd and Mays, 2005). As long as the rise in stage does not overlap the stream banks, most of the volume of stream water that enters the stream banks return to the stream in a period of days or weeks to later supplement stream flows. If the rise in stage is sufficient to overtop the banks and flood large areas of the land surface, widespread recharge to the water table can take place throughout the flooded area.

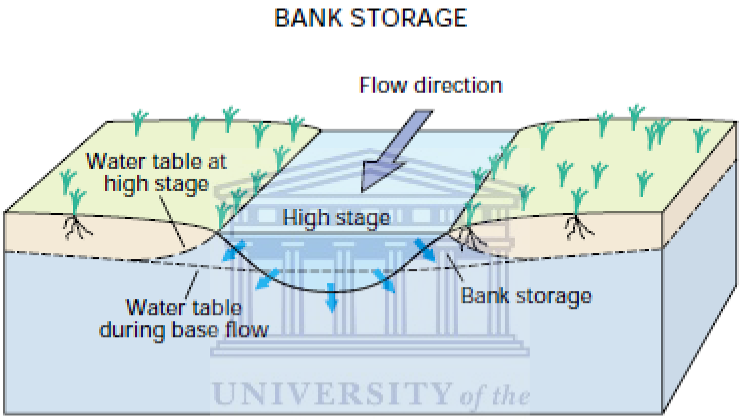


Figure 2.8 When stream levels rise higher than adjacent groundwater levels, stream water moves into the stream banks as bank storage (Winter et al., 1998).

2.5. Methods to estimate groundwater recharge

There are many review articles and publications that deal exclusively with recharge and the various estimation methods, Bredenkamp *et al*. (1995); Xu and Beekman (2003); Healy and Scanlon (2010) and Scanlon *et al.* (2002). The reliability of recharge estimates using different methods is variable. Methods based on surface-water and unsaturated-zone data provide estimates of potential recharge, whereas those based on groundwater data generally provide estimates of actual recharge. Bredenkamp *et al*. (1995) states that groundwater recharge estimation methods can be divided into physical and chemical methods. Physical methods attempt to estimate recharge from water balances calculated either from hydro-meteoric measurements, direct estimates of soil water fluxes based on soil physics or changes in the aquifers saturated volume based on water table fluctuations. Chemical methods are based on the distribution of natural tracers found in rainfall (commonly ^{2}H, ^{3}H, ^{14}C, $^{18}0$ and Cl) in the saturated and unsaturated zone.

There are various methods available for estimating recharge, each with its own limitations and applicability. Determining which of a wide variety of techniques is likely to provide reliable recharge estimates as a result of the complex processes associated with recharge as illustrated in Figure 2.4 is often difficult. Thus the selection and application of recharge estimation methods depends on the area of study and availability of data to develop an appropriate conceptual model to choose the most suitable and reliable estimation technique. Recharge estimation methods and their applications are well documented in the literature. A brief summary and comparison of the methods are listed in Table 2.1 adapted from Beekman and Xu (2003).

Bredenkamp *et al*. (1995) and Xu and Beekman (2003); provide an account of recharge estimation in semi-arid Southern Africa. They identified the following methods as the most promising for applications in semi-arid to arid environments:

- Cumulative rainfall departure (CRD);
- Water balance models;
- Chloride mass balance (CMB);
- EARTH model;
- Water table fluctuations;
- Groundwater models
- Saturated volume fluctuations (SVF)

Table 2.1: Recharge estimation methods applied in (semi)-arid Southern Africa (Beekman and Xu, 2003).

Zone	Approach	Method	Principle	Reliable	Ease	cost
Surface water	Physical	HS	Stream hydrograph separation: outflow, evapotranspiration And abstraction balances recharge	Med-high	low	low
		CWB	Recharge derived from difference in flow upstream and downstream accounting for evapotranspiration, in- and outflow and channel storage change	med	med	high
		WM	Numerical rainfall-runoff modeling; recharge estimated as a residual term	med	Med-high	high
Unsaturated	Physical	Lysimeter	Drainage proportional to moisture flux / recharge	med	high	high
		UFM	Unsaturated flow simulation e.g. by using numerical solutions to Richards equation	high	med	med
		ZFP	Soil moisture storage changes below ZFP (zero vertical hydraulic gradient) proportional to moisture flux / recharge	high	med	med
	Tracer	CMB	Chloride Mass Balance – Profiling: drainage inversely proportional to Cl in pore water	med	low	low
		Historical	Vertical distribution of tracer as a result of activities in the past	Med-high	Med-high	high
Saturated- Unsaturated	Physical	CRD	Water level response from recharge proportional to cumulative rainfall departure	Low-med	Low-med	med
		EARTH	Lumped distributed model simulating water level fluctuations by coupling climatic, soil moisture and groundwater level data	Low-med	med	Low
		WTF	Water level response proportional to recharge / discharge	med	low	Low
	Tracer	CMB	Amount of Cl into the system balanced by amount of Cl out of the system for negligible surface runoff / runoff	med	low	low
Saturated	Physical	GM	Recharge inversely derived from numerical modeling groundwater flow and calibrating on hydraulic heads / groundwater ages	Med-low	high	High
		SVF	Water balance over time based on averaged groundwater levels from monitoring boreholes	Med-low	Med-low	Med
		EV-SF	Water balance at catchment scale	Med-low	Med-low	Med-low
	Tracer	GD	Age gradient derived from tracers	high	med	high

HS: Hydrograph Separation- Baseflow
CWB: Channel Water Budget
WM: Watershed Modeling
UFM: Unsaturated flow modeling
ZFP: Zero Flux Plain
CMB: Chloride Mass Balance
CRD: Cumulative Rainfall Departure
EARTH: Extended model for aquifer Recharge and moisture Transport through Unsaturated hard rock
WFT: Water table Fluctuation
GM: Groundwater modeling
SVF: Saturated Volume Fluctuation
EV-SF Equal Volume- Spring flow
GD: Groundwater Dating

In this study an attempt is made to develop a conceptual model of groundwater recharge processes. These processes are complex and often require multiple approaches. The literature points to two methods that can be applied in the study area to meet the objectives of the study. These methods are the Cumulative Rainfall Departure (CRD) method and the use of stable isotopes ^{2}H and ^{18}O. These methods are discussed in detail in this chapter while the results of their application are discussed in chapter four.

2.5.1. Physical method: Water balance method

Water balance models have widely been used for estimating groundwater recharge (Manghi *et al*., 2009; Zagana *et al*., 2007). The idea of water balance is simple, and depends on the mass balance principle, comparing input and output into the area of study. The water balance methods can be used to estimate recharge as the residual of all the other fluxes such as precipitation, runoff, evapotranspiration and change in storage. The principle is that other fluxes can be measured or estimated more easily than recharge. The hydrological water balance for an area can be defined as:

RE (t) = R (t) - EVT (t) - RO (t) - Dq (t) **Equation 2.1**

Where RE (t) is recharge, R (t) is precipitation, EVT (t) is evapotranspiration, RO (t) is runoff and Dq (t) is change in volumetric water content, the term (t) designates that the terms are distributed through time.

The advantages of water-balance methods are that they use readily available data (rainfall, runoff, water levels), which are easy to apply and account for all water entering a system. Typical approaches include: catchment water balances, Saturated Volume Fluctuation (SVF) and the Cumulative Rainfall Departure (CRD) methods.

2.5.2. The Cumulative Rainfall Departure (CRD) method

Bredenkamp *et al*. (1995) demonstrated that the CRD series corresponds remarkably well with the fluctuations of the groundwater level. The rationale behind the departure method is that in any area, despite large annual variations in precipitation, equilibrium is established between the average annual precipitation and the hydrological response (run-off, recharge and losses from the system). Hence the equilibrium can be expressed as:

$$\mathbf{Rf_{av} = RO_{av} + RE_{av} + EVT_{av}} \qquad \textbf{Equation 2.2}$$

Where Rf_{av} = the average rainfall;

RO = the average run- off

RE = the average recharge; and

EVT = the average evapotranspiration

The CRD method is represented by the following equation:

$$\mathbf{{}_{av}{}^{1}CRD_{i} = Rf_{i} - kRf_{av} + {}_{av}{}^{1}CRD_{i-1}} \qquad \textbf{Equation 2.3}$$

Where:

CRD_i = accumulative rainfall departures from the mean at time i

Rf_i = rainfall at time i

k = 1 indicate natural conditions and k > 1 indicate that the aquifer system is being exploited

Rf_{av} = average rainfall

The CRD reflects the outcome of the natural balance of groundwater due to the combined effects of both recharge and losses from the system. It is commonly accepted that under natural conditions and in the absence of abstraction, a dynamic balance between recharge and drainage in an aquifer is established (Bredenkamp *et al*., 1995).

Bredenkamp *et al*. (1995) derived the following relationship between the CRD and the groundwater level response from the first principles;

$$\Delta h_i = a/S\,(Rf_i - k.\,Rf_{av}) \qquad \textbf{Equation 2.4}$$

Where Δh = the change in piezometric level for period i

a = the fraction of rainfall that constitutes recharge

S = the aquifer Storativity

Rf_i = rainfall for period i

Rf_{av} = average rainfall

k = 1 indicates that natural conditions apply

According to Bredenkamp *et al*. (1995), the derivation of the above equation is of great significance as it indicates that the change in water level will respond according to the cumulative rainfall departure from the mean, with a proportionality coefficient = a/S. This is in agreement with the concept illustrated by Grieske (1992) that the water level builds up over several years to reach an equilibrium condition representing the average status which is balanced by the average outflow or recession rate. The average rate of the water level recession is therefore a function of the average rainfall, assuming that the average rainfall is consistent for a specific region (Bredenkamp *et al*. 1995).

Equation 2.4 defines the correspondence between the water-level response and the rainfall departure:

1. the groundwater level will rise if $Rf > Rf_{av}$ and will decline when $Rf < Rf_{av}$
2. The losses from the system vary in proportion to average precipitation.

The following relationship can thus be used to determine the effective recharge according to the natural water balance applied to a selected time interval:

$RE = a\ (Rf - k.\ Rf_{av})$ **Equation 2.5**

Where k = 1 represents the natural situation and

k > 1 indicates that the system is being exploited

Equation 2.5 therefore indicates that:

- recharge occurs when the rainfall is more than the average for the selected time interval, and
- no recharge (i.e. natural losses) will occur when the rainfall is less than the average.
- the effect of pumpage (Δk) is incorporated inside the brackets i.e. its influence is reduced depending on the value of (a) which is normally less than 1.

Bredenkamp *et al*. (1995) further demonstrated that an improved correlation is attained between the groundwater level fluctuations if the cumulative rainfall departures are calculated relative to the moving average rainfall over the preceding 8 to 10 years. In terms of the CRD approach this implies that as the average rainfall changes, so the total water balance is adjusted. The aquifer responds differently during prolonged wet and dry rainfall cycles and the system adapts to the prevailing conditions. Many aquifers indicate a delayed (time lag) recharge response, which corresponds to the earlier rainfall (Bredenkamp *et al*., 1995). Recharge in a specific month depends on the average rainfall of more than one month preceding the current month. Bredenkamp *et al*. (1995) introduced a long and short term “memory” in the system of preceding rainfall conditions. The short-term memory accounts for the time lag in groundwater response to rainfall and can incorporate carry-over of recharge from year to year. The long-term memory represents the period over which the long-term reference rainfall is calculated.

The adapted CRD is expressed as:

$$ {}_{n}^{m}CRD_{\mathrm{i}} = \left[\frac{1}{m} \sum_{j=i-(m-1)}^{i} Rf_{\mathrm{j}}\right] - \left[k \times \frac{1}{n} \sum_{j=i-(n-1)}^{i} Rf_{\mathrm{j}}\right] + (\mathrm{CRD}_{\mathrm{i-1}}) $$

Equation2.6

Where

m = the number of months denoting the short term memory carry- over e.g. 1, 6, 12, 18, 24 months rainfall average values.

n = the number of months for which the long-term reference rainfall is calculated (long-term memory) e.g. 1, 3, 5, 7, 10 years rainfall average preceding a specific month.

Bredenkamp *et al*. (1995) concluded that the CRD method has proved to be a valuable method to simulate the groundwater level fluctuations of an aquifer where natural conditions still prevail (to large extent suited to the KNP). By using the long and short term memory of the system potential recharge processes can be identified as the water level response to rainfall events can be simulated. Particularly the short term memory can provide insights into the efficiency of recharge thereby suggesting the type of processes dominant for a particular aquifer system. In addition, the CRD applies even if an aquifer is exploited, as the effect of pumpage can be added to the rainfall reference value from which the rainfall departures are calculated.

2.5.2. Chemical method: Natural stable isotopes ^{2}H and ^{18}O

The use of stable isotopes, ^{2}H and ^{18}O, is based on the fractionation of water during phase changes. For example, $H_2{}^{16}O$ has a higher vapour pressure and diffusivity than does $H_2{}^{18}O$ leading to a higher ^{18}O content in water after evaporation. Likewise, $^{2}H_2O$ and $^{1}H_2O$ also fractionate, but to a greater extent than water with different oxygen isotopes due to a larger relative mass difference (Hunt *et al*., 2005). The different degrees of fractionation due to different processes results in the wide array of uses of ^{2}H and ^{18}O to investigate hydrologic systems (Levy, 2009). Hence, stable isotopes of oxygen and hydrogen have increasingly been used to investigate groundwater/surface-water interactions. The technique is also used to trace the provenance of groundwater and is very effective in identifying processes that alter the isotopic composition of rainfall prior to infiltration such as evaporation or mixing from different rainfall events (Clark and Fritz, 1997). Further, water from different sources such as precipitation, rivers and groundwater have different isotopic signatures, and those differences can be used to gain insight into the interactions between these components (Aggarwal *et al*., 2007). Because they are part of the water molecule, stable isotopes ^{18}O and ^{2}H are ideal conservative tracers of water movement (Hunt *et al*., 2005) as long as there are no phase changes along a flow path. These isotopes have therefore been used to locate and confirm groundwater discharge locations (Oxtobee and Novakowski, 2002), quantify groundwater discharge to surface water (Space *et al*., 1991) and distinguish the sources of groundwater recharge (Blasch and Bryson, 2007).

The δ Notation

Stable isotopic compositions are normally reported as delta (δ) values in parts per thousand (denoted as ‰ or permil). The ratio defined by the δ notation are calculated by:

δ = (R sample / R standard - 1) X 1000 Equation 2.6

Where "R" is the ratio of the heavy to light isotope, a positive delta value means that the isotopic ratio of the sample is higher than that of the standard (enriched in the heavy isotope) while a negative value means that the isotope ratio of the sample is lower that that of the standard (depleted in the heavy isotope). The isotopic compositions of δ^2H and $\delta^{18}O$ values are normally reported relative to an international reference standard, the SMOW standard (Standard Mean Ocean Water), or the equivalent VSMOW (Vienna-SMOW) standard. In practice, each laboratory has its own standard or set of standards that has been calibrated relative to the international standard. During the measurement, the isotopic ratio of the sample is compared to that of the laboratory standard and the result is recalculated to the VSMOW scale.

The Global Meteoric Water line

Craig (1961), observed that there is a relationship between the abundance of ^{18}O and ^{2}H relative to VSMOW, represented by the equation:

$\delta\,^2H = 8\delta O + 10‰$ Equation 2.7

Equation 2.7 describes what is known as the Global Meteoric Water Line (GMWL). The GMWL provides a reference by which local differences in water can be compared and thus assist is interpreting the provenance of water. If water samples have been subjected to evaporation the δ^2H and $\delta^{18}O$ relationship no longer plot on this line. This is due to the non-equilibrium kinetic effects that occur during evaporation which causes the fractionation of H and O isotopes to occur in different proportions (Kendall and McDonnell, 1998). As a consequence of fractionation processes, waters often develop unique isotopic compositions "fingerprints" (ratios of heavy to light isotopes) that may be indicative of their source or of the processes that formed them (Kendall and McDonnell, 1998).

Fundamental to the use of isotopes, is the need to understand the processes in which fractionation occurs, and then at what stages these occur within the hydrological cycle. As water moves through the hydrological cycle it is subjected to various conditions in which fractionation of the isotopes take place. The process starts with the evaporation of water from the ocean surface which has $\delta^{18}O$ and $\delta^{2}H$ values close to that of VSMOW. The evaporated water is typically isotopically lighter, or more depleted, than the water left behind.

The actual $\delta^{18}O$ values of precipitation reaching the ground depend on several factors (Mook, 2006; Clark and Fritz, 1997) and whether these factors will have an effect on the precipitation in the KNP environment are described below.

1. Latitude: Colder temperatures decrease the likelihood for heavier water to condense and fall as precipitation. Therefore, the higher the latitude, the generally more depleted the precipitation is in $\delta^{18}O$. This will not have an effect on the isotopic ratios in this study.
2. Proximity to the ocean: Air masses tend to drop their heavier precipitation first. Therefore, precipitation becomes more depleted in $\delta^{18}O$ as air masses move inland. This is likely to have an effect as different rainfall systems originating in the Indian Ocean, off the coast of Mozambique, and Atlantic Ocean off the Southern Cape coast, causes rainfall over the KNP. These systems will thus produce different isotopic ratios as a result of the distance traveled over land.
3. Altitude: Colder temperatures at higher altitudes generally result in precipitation that is more depleted in $\delta^{18}O$. This will not have an effect as the KNP falls into a narrow range of altitude ranges.
4. Season: Colder temperatures in winter result in precipitation that is relatively depleted in $\delta^{18}O$. This will have an effect as the temperature particularly during the hot summer months could result in rainfall experiencing evaporation before it reaches the ground.

5. Amount: large rainfall events tend to be depleted in $\delta^{18}O$ relative to small rains. This will have an effect on the isotopic signature as the intensity and duration of rainfall events will result in different isotopic ratios.

Due to all the effects mentioned above, local precipitation will tend to have a somewhat different relationship between ^{18}O and ^{2}H. It is therefore important in any local field study to sample local precipitation over time to establish the Local Meteoric Water Line (LMWL) against which one can compare isotopic values from water in other parts of the hydrologic cycle (Kendall and McDonnell, 1998). Therefore, the $\delta^{2}H$ and $\delta^{18}O$ values of groundwater and its relationship to the MWL can be used to determine if recharge is delayed or immediate and identify possible process that altered the isotopic composition of the precipitation prior to recharging the groundwater.

As rainfall infiltrates through the soil and sediment or preferential pathways to the water table, a number of processes can alter the isotopic composition of groundwater. In an area where the climatic conditions are conducive to significant evaporation, the ^{18}O in soil water tends to be isotopically heavier than the precipitation due to the preferential evaporation of the lighter isotopes (Abbott, 1997). Since the evaporation rates in the KNP are high during the dry winter months and particularly during the hot summer months, soil evaporation is likely to be an important process for changing the isotopic ratios of infiltrating water. Another means by which the isotopic compositions are altered is due to mixing of infiltrating water from different recharge events and flow paths. The path that water takes to the water table is dependent on the weathering thickness arrangement of soil particles, soil composition and preferential flow paths that influence the rate at which water can travel. In complex deposition and differential settings, the permeability of soil may vary significantly over small areas. For this reason water entering the soil may not continue at a uniform rate along uniform flow paths to the water table. This causes the mixing of isotopic signatures which originates from different rainfall events (Abbott, 1997).

Groundwater data points tending to plot below the meteoric water line indicate samples originating from surface water bodies (ponding in local surface depressions and streams) that have undergone evaporation, enriched in stable isotopes (Kendall and McDonnell, 1998). In the case of groundwater this is indicative of diffuse and indirect recharge processes where groundwater recharge occurred as a result of shallow soil layers, rivers, streams and local surface depressions that have undergone evaporation. Data points close to or tending to plot on the meteoric water line, indicate that samples are depleted in δ^2H and $\delta\ ^{18}$O (Kendall and McDonnell, 1998). In the case of groundwater this will be indicative of localised recharge processes from precipitation commonly via preferred pathways (fractures and cracks) thereby undergoing very little evaporation. Groundwater recharged in this way can therefore have an isotopic signature that is very similar to that of local precipitation during the periods of the largest precipitation (Mook, 2006).

Surface water, on the other hand, immediately undergoes evaporation leading to an enrichment of ^{18}O and therefore ^{18}O values that are higher than those of local precipitation. This effect is dependent on temperature and therefore greatest in summer (Mook, 2006). As evaporation occurs from a surface-water body, the surface water becomes enriched in both ^{18}O and ^{2}H. However, due to non-equilibrium evaporation (evaporation rate greater than the condensation rate), the rate of enrichment is more for ^{18}O than for ^{2}H relative to the GMWL. This result is a slope < 8 for the line relating δ^2H and δ^{18}O (the slope is usually between 1 and 5). The point of intersection of the evaporation line with MWL usually is taken as the isotopic composition of unaltered rainfall. The further along the evaporation line the data points lie, the greater the evaporation (Schwartz and Zang, 2002). Therefore, plots of δ^2H versus δ^{18}O in sampled water can reveal much about the evaporative processes that water has undergone.

2.6. Conclusion

To develop a conceptual model of groundwater recharge processes and surface-water interactions in the KNP environment the CRD method and the use of stable isotopes were identified as potential methods. The adapted CRD method using the long-term and particularly the short- term memory of the system can present insight into the recharge efficiency of the system thereby indicating what recharge processes are dominant. It can further be used to simulate groundwater level trends over the past 50 odd years. The use of stable isotopes ^{2}H and ^{18}O will provide confirmation into plausible recharge processes by indicating whether recharge is immediate or delayed. Additionally, it will act as a natural tracer to qualitatively identify interactions between surface and groundwater.

CHAPTER 3
STUDY AREA

3.1: Introduction

The physical characteristics of the KNP environment will play an important role in groundwater recharge and surface-water/groundwater interactions. Hence, understanding the characteristics of the climate, geomorphology and aquifers would improve the understanding of groundwater recharge processes and its interaction with surface-water systems. In this chapter the location, drainage patterns, climate, topography, geology, soils, vegetation and hydrogeology of the KNP are discussed and what this means for recharge processes.

3.2: Geographic Location and Drainage Patterns

The Kruger National Park (KNP) is situated in the Mpumalanga and Limpopo Province along the north-eastern boundary of the Republic of South Africa (RSA). It is an elongated park of about 2 million ha measuring approximately 350 km north to south and 65 km on an average east to west (Venter, 1990). The KNP is part of the northeastern South African Lowveld, the low-lying area between the footslopes of the Drakensberg Great Escarpment to the west and the extensive Mozambique coastal plain to the east (Figure 3.1). It is drained by two major rivers systems- the Limpopo system which forms the northern boundary of the park and the Crocodile system which forms the southern boundary (Du Toit, 1998). The park interior is drained by five large naturally perennial rivers (i.e. Luvuvhu, Letaba, Olifants, Shingwedzi and Sabie Rivers), which flow west to east across the park before flowing into Mozambique and the Indian Ocean (Venter, 1990). Important seasonal and ephemeral rivers which originate in the park are the Phugwane and Mphongolo Rivers in the north, Tsende, Timbavati, Nwaswitsontso, Ripape and Sweni Rivers in the central region, and the Mbyamiti (seasonal), and Nwaswitshaka and Mnondozi Rivers in the south (Figure 3.1). These rivers predominantly carry water following heavy rainfall during summer months. The Sand River which is a tributary of the Sabie River flows all year round (observations post 2000).

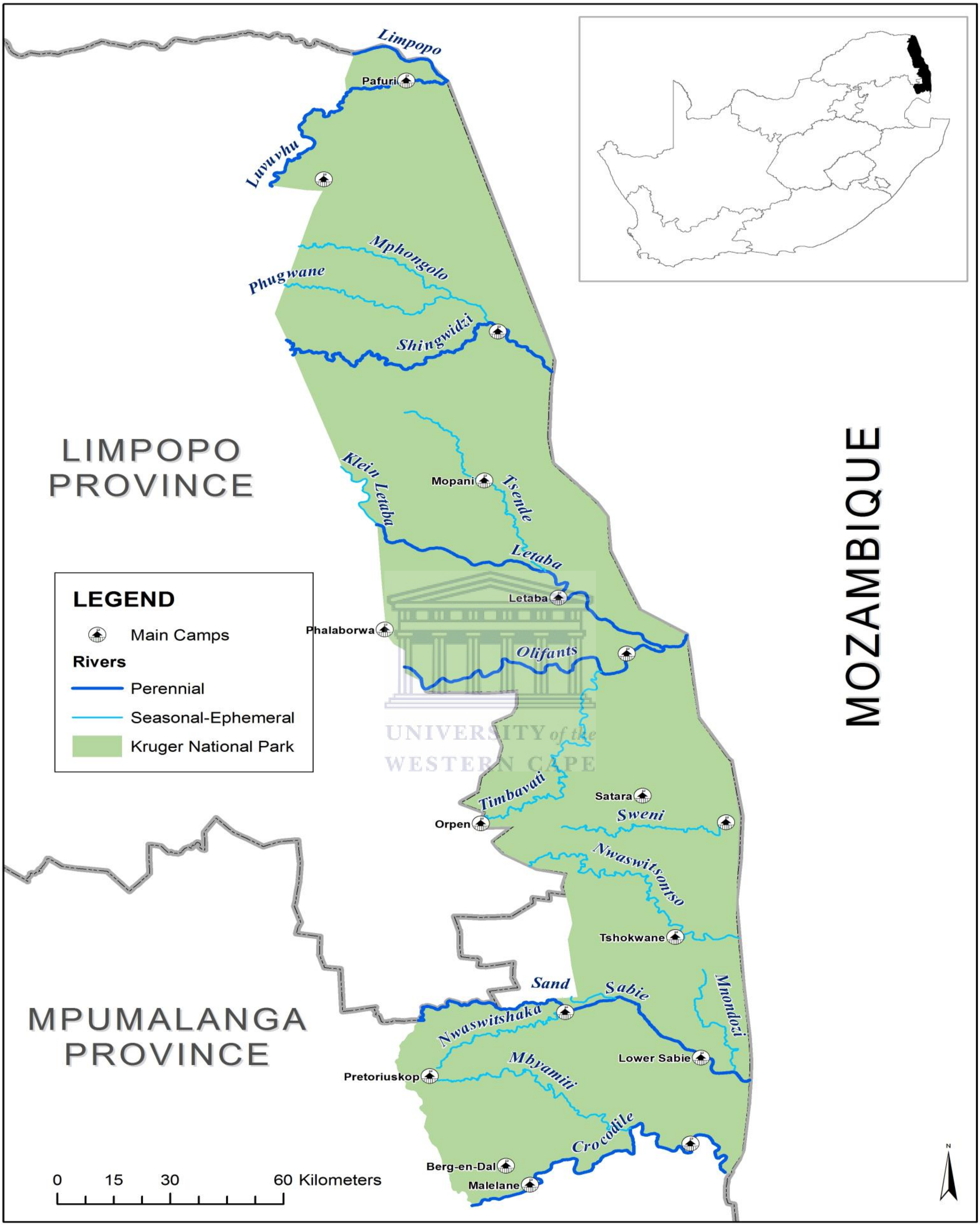


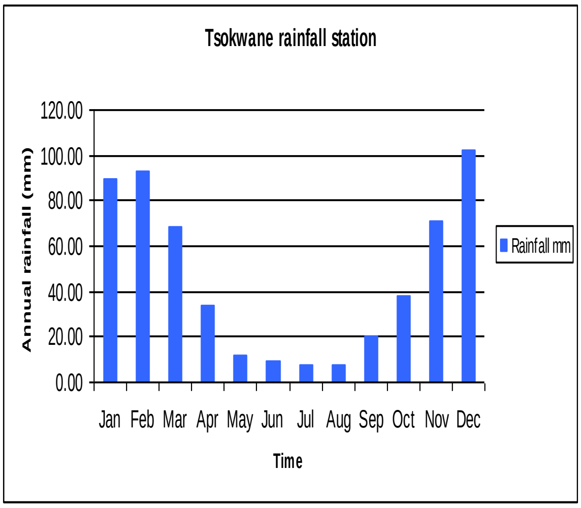
Figure 3.1 The location and rivers of the KNP.

3.3. Topography

The topography of the KNP largely reflects differences in resistance to weathering of the underlying rocks and the intensity of dissection in areas that flank major rivers (Schutte, 1986; Venter, 1990). The most outstanding topographic feature is the Lebombo Mountains along the eastern border with Mozambique from the Shingwedzi River southwards. Within the borders of the KNP the Lebombo reach a maximum height of 497m above sea level. The Malelane Mountains in the south-western part of the KNP average about 800m above sea level. The remainder of the park is a gently undulating landscape between 200m and 400m above sea level with a gentle gradient to the east (Schutte 1986; Venter 1990).

3.4. Climate

The KNP falls within two climatic zones, the south and central portions fall into the Lowveld bushveld zone (potential evaporation of 6mm/day in October), and the north in the northern arid bushveld zone (potential evaporation of 7mm/day in October (Venter *et al*., 2003). These are influenced by anti-cyclonic systems that move over southern Africa from a west to east in a semi-rhythmical way (Venter, 1990; Gertenbach, 1980; Venter *et al*., 2003). With its sub-tropical climate the KNP experiences warm to hot summers and mild dry winters. There is a slight spatial trend in temperature from cooler in the south to hotter in the north (Gertenbach, 1980). Temperatures may reach 44°C in the summer and seldom fall below freezing point in winter (Venter, 1990). The mean annual rainfall (MAP) of the KNP decreases from south to north and from west to east. The MAP of the park varies from approximately 740mm at Pretoriuskop in the south to about 440mm at Pafuri in the north. The entire park experiences a MAP of 530mm. Most of this rain occurs in the form of thunder storms between November and March with a peak in January and February. Between June and August very little rain occurs (Gertenbach, 1980; Venter, 1990). The storms are typically of short duration lasting just minutes or hours. As a result, the rainfall intensity is often high leading to flash floods in the ephemeral drainage lines (Venter *et al*., 2003).



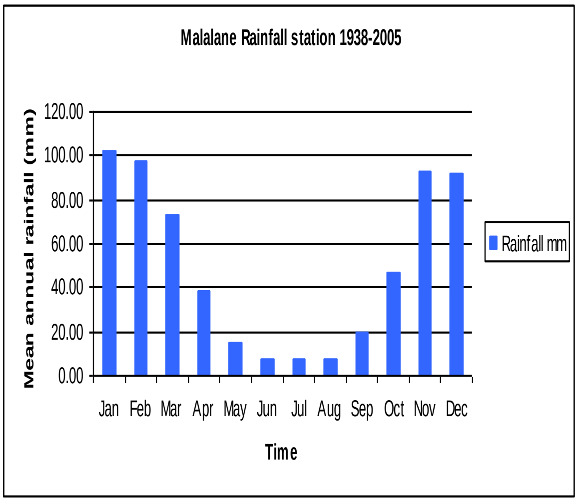


Figure 3.2 Rainfall patterns in the KNP, most rain occurs in the form of thunder storms between November and March with a peak in January and February. Between June and August very little rain occurs.

The long term rainfall of the KNP oscillates through periods of above and below average rainfall. Rainfall cycles last approximately 10 years each (Gertenbach, 1980; Venter, 1990).

In 2000 abnormally high rainfall events which caused a one in two hundred year flood directly related to tropical cyclone Eline that originated in the Indian Ocean and moved through the Mozambique Channel before swinging inland over the study area. The extreme rainfall was concentrated in two periods, from the 5 to 10 February and 22 to 25 February 2000 which resulted in disastrous flooding, loss of hundreds of lives and severe damage to infrastructure. The combination of the two systems and high levels of earlier soil moisture from an already wet December resulted in the excessive flooding (Smithers *et al*., 2001).

3.5. Geology

A diverse assemblage of igneous, sedimentary and metamorphic rocks, as well as unconsolidated sediments deposited over a time span of more than 300 million years (Ma) occur within the borders of the KNP. The most important litho-stratigraghic units that are present in the KNP include, the Basement Complex which consists of ancient granitoid rocks of Swazian age (>3090Ma), sedimentary and volcanic rock of the Soutpansberg Group and the volcanic rock of the Karoo Supergroup (Venter, 1990). The distribution and characteristics of these different rock formations are summarised in (Figure 3.3 and Table 3.1). The strike of the lithology is north-south, so that the geological succession changes from west-east. Granitic rocks in the west and the basaltic rocks in the east underlie the majority of KNP. A thin north-south strip of sedimentary rocks separates the granitic and basaltic rock formations (Venter *et al*., 2003). For the purpose of the study I will only describe the geologic formations that are relevant to the specific study areas. A detailed description of the geology of the KNP are given in Bristow and Venter (1986); Schutte (1986); Venter (1990); and Barton (1986), and Walraven (1986).

3.5.1. The Basement Complex

Orpen Gneiss:

The area to the east and south-east of Orpen consist of fine- grained grey to dark grey homogeneous gneiss without ultramafic xenoliths. The gneiss is cut by a few fine-to coarse-grained pegmatite veins. The Orpen Gneiss consists of quartz, plagioclase, biotite and minor amounts of microcline and sphene (Schutte, 1986).

Nelspruit Granite Suite

The Nelspruit Granite Suite complex forms a huge batholitic outcrop in the southern part of the KNP. It consists of banded gneiss, gneiss, migmatite and porphyritic granite. The banded gneisses grade into unbanded gneiss and migmatite (Schutte, 1986).

3.5.2. The Lebombo Group

The Lebombo Group embraces the volcanic rocks at the top of the Karoo and represented here by the Letaba and Jozini Formations.

Letaba Formation:

The Letaba Formation consists of extrusive mafic volcanic rocks which rest conformably on the Clarens Formation. The basalts are considered to represent flood basalts that extrude along fissures associated with the fragmentation of Gondwanaland. Outcrops are poor and are mostly covered by a black turf soil. Some good outcrops do occur along the main rivers (Schutte, 1986).

Jozini Formation:

The Jozini Formation is situated above the Letaba Formation and comprises of rhyolite to dacite lava flows that formed the Lebombo Mountains. Inter-fingering of basalt and rhyolite flows occurs along the contact between the Letaba and Jozini Formations. The base of the Jozini is taken where the rhyolite flows become dominant (Schutte, 1986). The rhyolite reaches a maximum thickness of approximately 2.5 km in the most southern part of the KNP (Bristow and Venter, 1986).

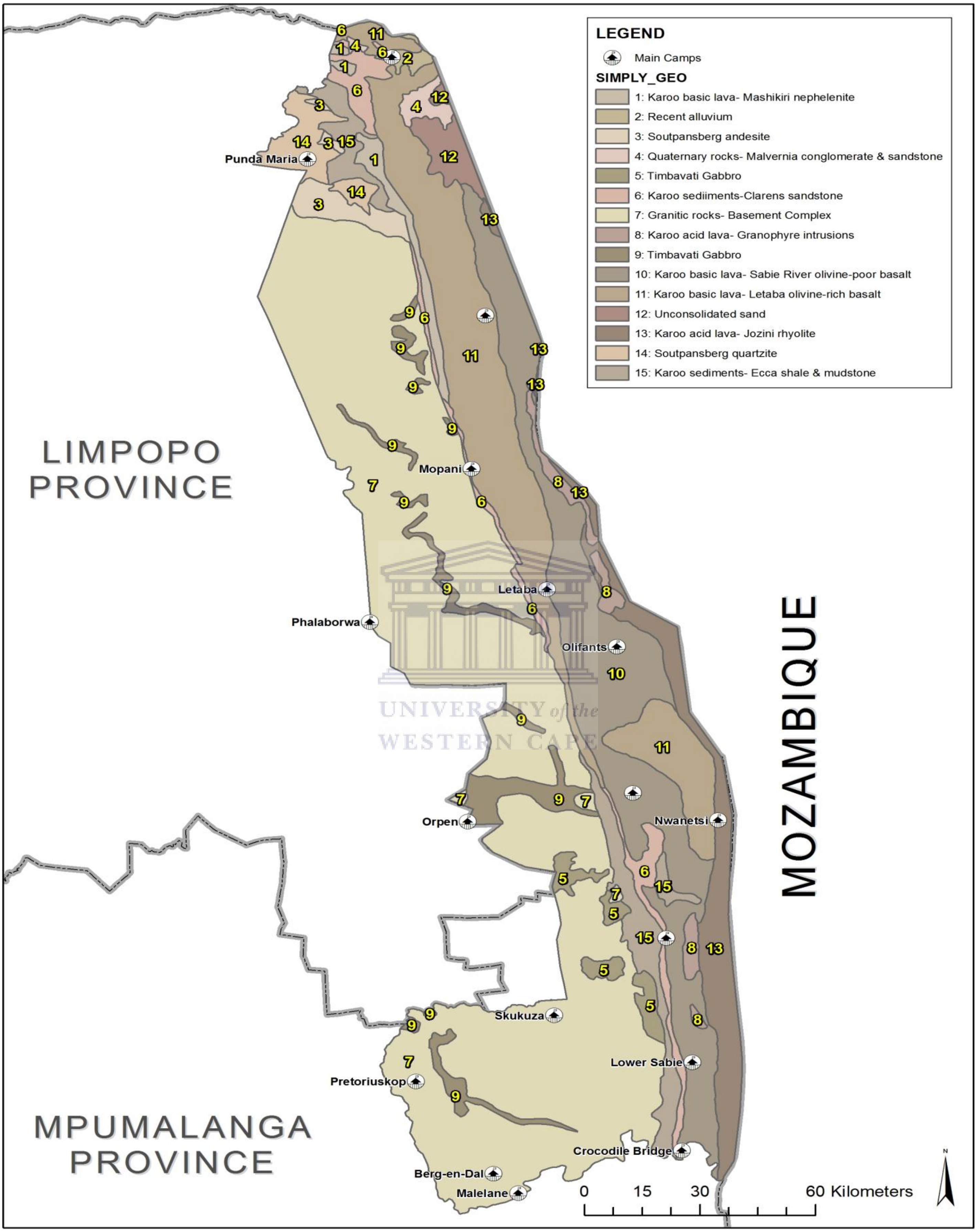


Figure 3.3 A simplified geological map of the KNP (Bristow and Venter, 1986).

Age (Ma)	Major units	Major sub-units	Dominant rock types
<130	Recent deposits	-	Alluvium, sand
		Malvernia formation	Fossiliferous conglomerate and sandstone
<175	Karoo sequence	Tshokwane granophyre	Granophyre
~175		Jozini rhyolite formation	Rhyolite, dacite
~190		Sabie river basalt formation	Olivine-poor basalt
~200		Letaba basalt formation	Olivine-rich basalt
~200		Mashikiri nephelinite formation	Nepheline lavas
~200-300		Clarens sandstone formation	Fine-grained sandstone
		Ecca group	Shale, mudstone, grit, conglomerate, coal
~1000	Timbavati gabbro		Gabbro, quartz gabbro, olivine gabbro
~1800	Soutpansberg group	Nzhelele formation	Quartzitic sandstone, shale, basalt
		Wyllies Poort quartzite formation	Quartzite, sandstone
		Fundudzi formation	Sandstone, quartzite
		Sibasa basalt formation	Basalt
~2050	Basement complex	Phalaborwa igneous complex	Syenite
~2200		Tsheri pegmatite	Muscovite-bearing pegmatite
~2650		Baderukwe granite	Granodiorite, granite
~3200		Nelspruit granite suite	Granite, gneiss, migmatite
~3500		Orpen gneiss	Gneiss
>3500		Makhutswi gneiss	Gneiss, migmatite, amphibolite
		Goudplaats gneiss	Gneiss, migmatite, amphibolite
		Murchison sequence	Amphibolite, schist
		Barberton sequence	Schist, amphibolite

Table 3.1 A summary of the stratigraghy of the major rock formations in the KNP (Venter, 1990).

3.6. Soils

There is strong correlation between geology and soils of the KNP (Venter, 1986; Venter, 1990; Venter *et al*., 2003). Soil profiles generally become shallower as rainfall decreases towards the north. This is particularly noted for the coarse- grained soils derived from the granitic materials, where soil depths decrease from approximately 150 cm in the Pretoriuskop area (rainfall 750 mm/yr) to 30 cm north of Phalaborwa (rainfall 350 mm/yr). On a broad scale the areas of the park that are underlain by the granitic/gneiss of the basement complex are charcterised by distinctive hillslope sequences (catena). From crest to valley bottom the soils usually occur in the following pattern: along the crest and midslopes sandy-hydromorphic (coarse grained) soils, duplex soils along the foot slopes and complex alluvial soils are found along the valley bottoms. The Karoo sequence (basalt) which is a predominantly flat landscape (low undulation) produces soils that have high clay content with olivine-rich clay soils in the northern plains and olivine-poor soils in the southern plains. Alluvial soils occur along most of the drainage lines in the KNP, the extent of which increases as the size of drainage lines increase. Older river terraces and gravels also occur along the major rivers. The most extensive alluvial deposits are found along the Limpopo and Luvuvhu rivers in the northern KNP (Venter, 1986; Venter, 1990; Venter *et al*., 2003). The distribution of soils across the KNP is provided in Figure 3.4.

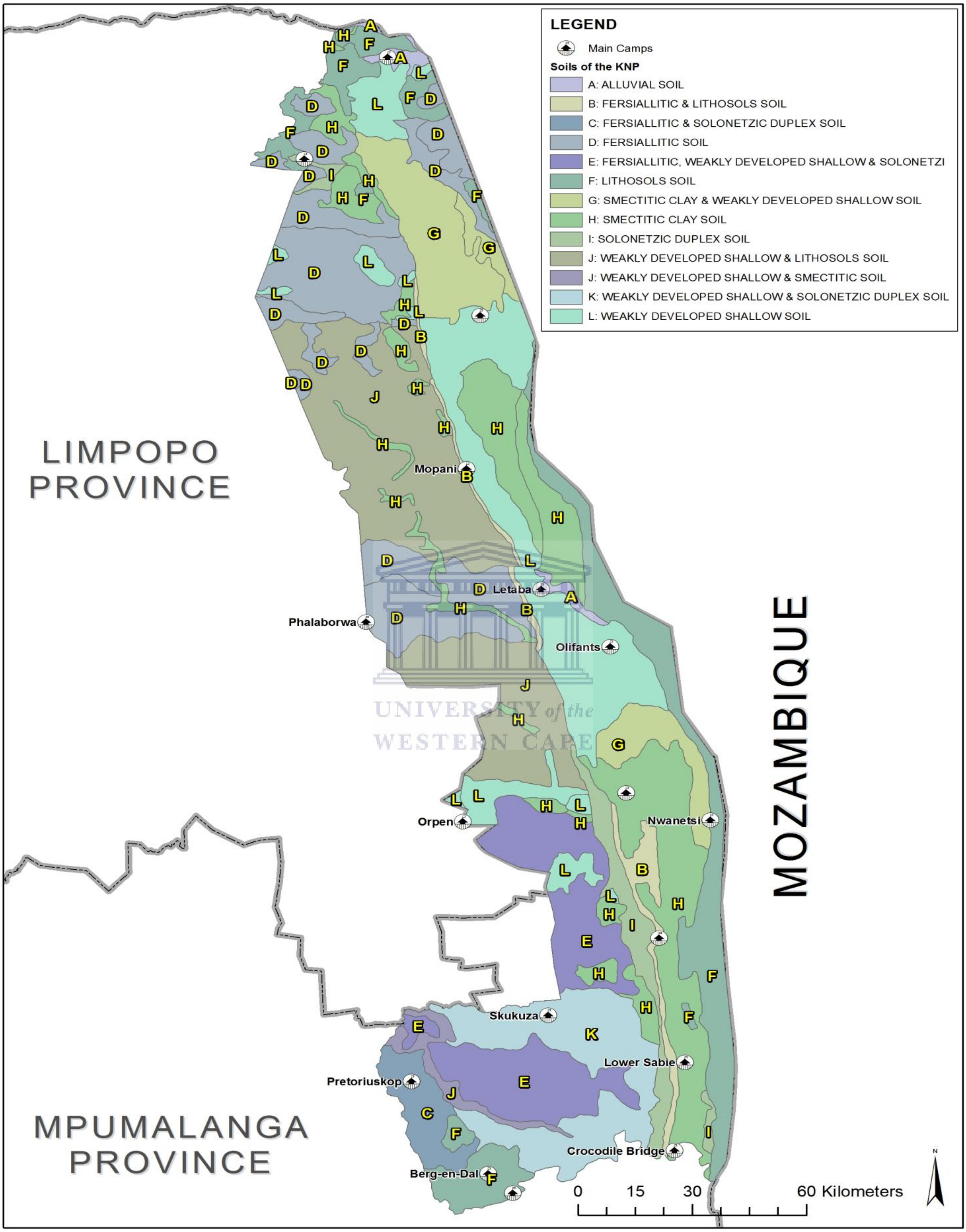


Figure 3.4 Soil distribution map of the KNP (Venter, 1990).

3.7. Vegetation

The vegetation of the KNP is classified as semi-arid to arid wooded savanna (Gertenbach, 1986; Venter, 1990). The vegetation of the park is divided between two main ecological types. The broad-leaved savannas that occupy approximately 75% of the park of which 50% are *Colophospermum mopane* shrubveld and remaining 25% are made up fine leaved savanna (Venter et al. 2003). The vegetation in the south-western part of the KNP which is underlain by granitoid rocks is characterised by relatively dense woodland. The dominant tree species in this area are *Combretum apiculatum., C. zeyheri Sond., C. collinum Fresen. Subsp. Suluense.* The area underlain by basalt south of the Olifants River consists mainly of *Acacia nigrescens / Sclerocarya birrea.* North of the Olifants River the basalts are characterised by riparian thickets which occur along all the major drainage lines. These thickets are usually denser in the granitic areas compared to the basaltic areas (Venter and Gertenbach, 1986). The distribution of vegetation across the KNP is provided in Figure 3.5.

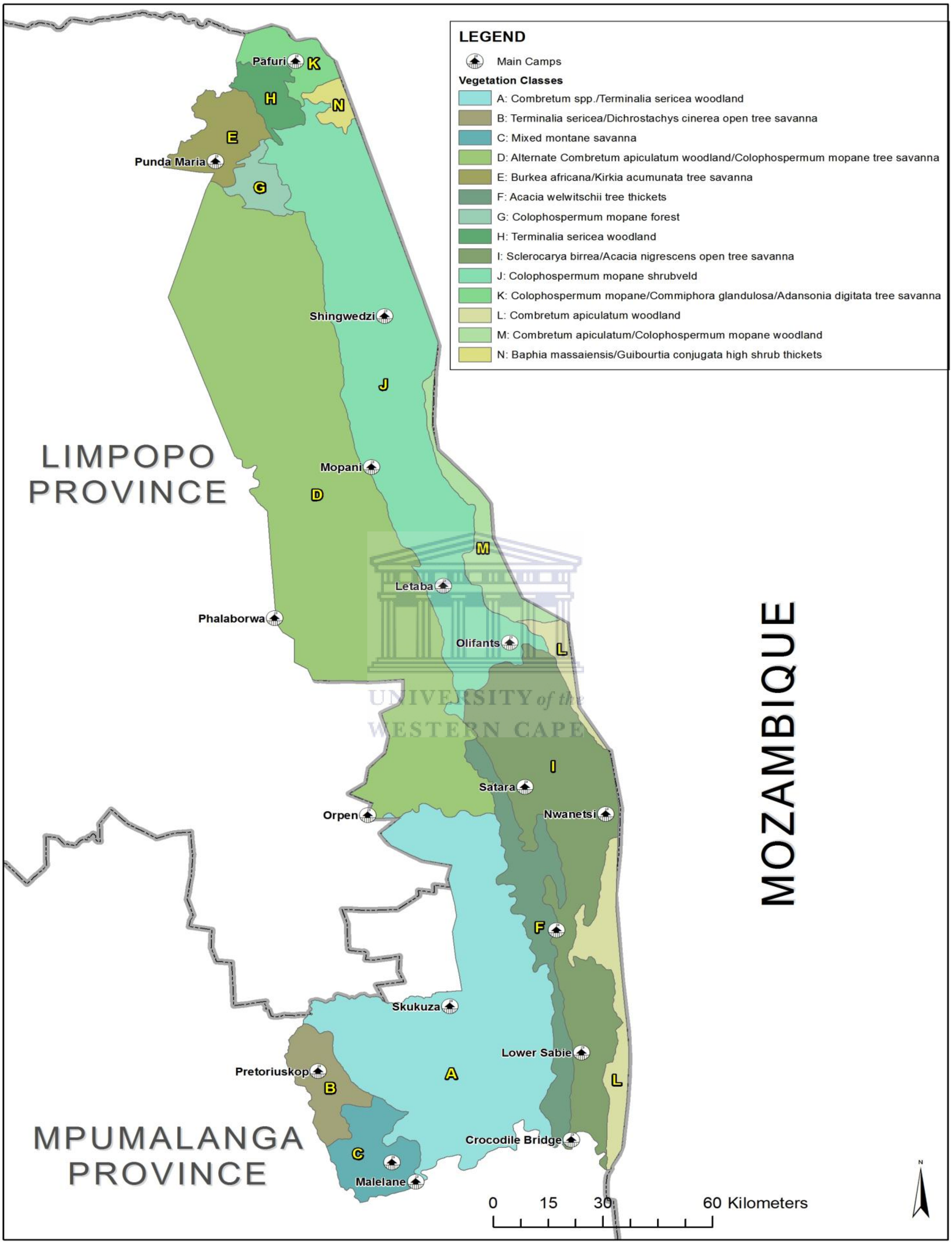


Figure 3.5 Vegetation distribution map of the KNP (Venter, 1990).

3.8. Occurrence of groundwater in the KNP

There are three types of aquifers namely: crystalline aquifers, consolidated aquifers and the unconsolidated aquifers. These aquifers are derived from crystalline, sedimentary/meta-sedimentary and unconsolidated sedimentary rock formations respectively. An aquifer is characterized by two main intrinsic properties namely porosity and permeability, which affect the manner of storage and transmission of water. Hard rock aquifers possess secondary or double porosity and secondary permeability whereas unconsolidated aquifers possess primary porosity and primary permeability characteristics (*Adams et al*., 2004).

The aquifers of the KNP mainly occur in crystalline hard rock lithologies. Generally the term crystalline basement refers to igneous and/or metamorphic rocks such as, granites, gneisses, meta-quartzite and basalts. Crystalline basement aquifers are classified as two-layered systems comprised of fractured bedrock overlain by a weathered/regolith zone. These rocks are characterised by very low primary porosity and almost all water movement and storage in these rocks takes place along fractures, faults, weathered zones and other secondary features that enhance the aquifer potential. While interconnected fractures may have high permeabilities, they have very little storage potential (Witthuser *et al*., 2011). In these aquifers it is not the rocks themselves that transmit the groundwater but the fractures and fissures that form the conductive openings through the impervious rock matrix. Crystalline Basement aquifers are depended on the presence of higher porosity material such as weathered regolith or adjacent alluvium to store groundwater (Witthuser *et al*., 2011). The weathered/regolith zone acts as a reservoir that slowly feeds water down towards fractures in the bedrock. Fractures exposed at the surface act as preferential flow paths (Wright, 1992). The classical model of a Basement aquifer is a zone of high porosity but low permeability clay-rich weathered material overlying a fractured zone of low porosity but high permeability (Witthuser *et al*., 2011).

Natural groundwater quality in most Basement environments is generally good, with low salinities and neutral to slightly acid pH values being common. Elevated salinities are typically encountered in areas of low recharge, discharge and/or prolonged residence times in the subsurface. The complex geology and discontinuous nature of many Basement areas means that natural water quality can change over relatively short distances, laterally and with depth (Witthuser *et al*., 2011).

In the crystalline Basement aquifers of the KNP groundwater is mainly stored in water filled joints, fractures, faults, zones of weathering/regolith and structures such as dyke intrusions. The aquifer systems mostly form anisotropic and heterogeneous groundwater bodies as the geometry and arrangement of fractures, joints and faults are influenced by various tectonic features (Du Toit, 1998). Groundwater flow and storage in these aquifers are inherently very complex. The following types of aquifers can be expected in the KNP.

- Composite aquifers comprised of a variable thickness of regolith overlying bedrock where the upper part is frequently fractured.
- Deep fractured aquifers which are composed mainly of crystalline material (igneous and metamorphic rocks) charcterised by an intact and relatively unweathered matrix with a complex arrangement of interconnected fracture systems.
- And, alluvial aquifers, where alluvial material overlies or replaces the weathered overburden and creates distinct intergranular aquifer types which can be found along river systems such as the Crocodile, Limpopo, Sabie and Sand Rivers (Fischer *et al*., 2008)

Du Toit (1998) evaluated some 1011 boreholes in KNP of which 41% were drilled in the Letaba Formation (basalt) and 37% drilled in the Basement Complex (granites). The depths to groundwater levels varied from 1.2m to 40m. The boreholes are primarily drilled in the weathered/regolith zone between 30m-55m below the surface. The aquifers were found to be low yielding at 1.5l/s to 3l/s and rarely exceeded 5l/s. Du Toit (2009) generated a water level contour map using a water level correlation with topography together with a Digital Elevation Model (DEM) to interpolate water levels to a KNP scale. The groundwater flows from a regional perspective from west to east through the Park, closely following the surface water drainage regions. Further assessment of the generated groundwater piezometric map shows groundwater drainage to be more complicated and localized in the southern and to a lesser extent the northern part of the Park (Du Toit, 2009).

3.9. Study sites

The boreholes that are used in this study are existing unused boreholes in the KNP. The locations of these boreholes are by no means designed for sustainable water supply and particularly not for scientific observation. The study sites that were chosen were selected based on the continuity of the groundwater level data in boreholes and availability of rainfall time series data from the nearest rainfall gauging station. Boreholes were selected to be situated on the two main geologic formations (basalts and granites) and relatively close to streams and rivers that flow along these formations. This was an important criterion to draw comparisons on dominant recharge processes. Access to these sites had to be relatively easy particularly during the wet season and traveling distance between the sites had to be limited as far as possible, to ensure logistical feasibility.

Five sites were selected in the southern region of the KNP (Figure 3.6). Three of these Rietpan, Mafutsu and Sweni-hide are located on the southern basalts and two, Jock Msanimond, on the granites. The study sites are representative of 2 main geological units, namely the Karoo Sequence consisting of the Letaba formation (basalts), the intrusive Jozini formation (rhyolite) and the Basement complex which consists of the Nelspruit granite suite and Orpen Gneiss (granites). Each area has a different rainfall average varying from MAP 650mm in Malalane to MAP 540mm in Tshokwane. The boreholes are located relatively close to the Sweni, Nwaswitsontso, Mbyamiti and Mnondozi Rivers. Samples from the Sabie and Sand Rivers were additionally included, representing two perennial rivers in the region.

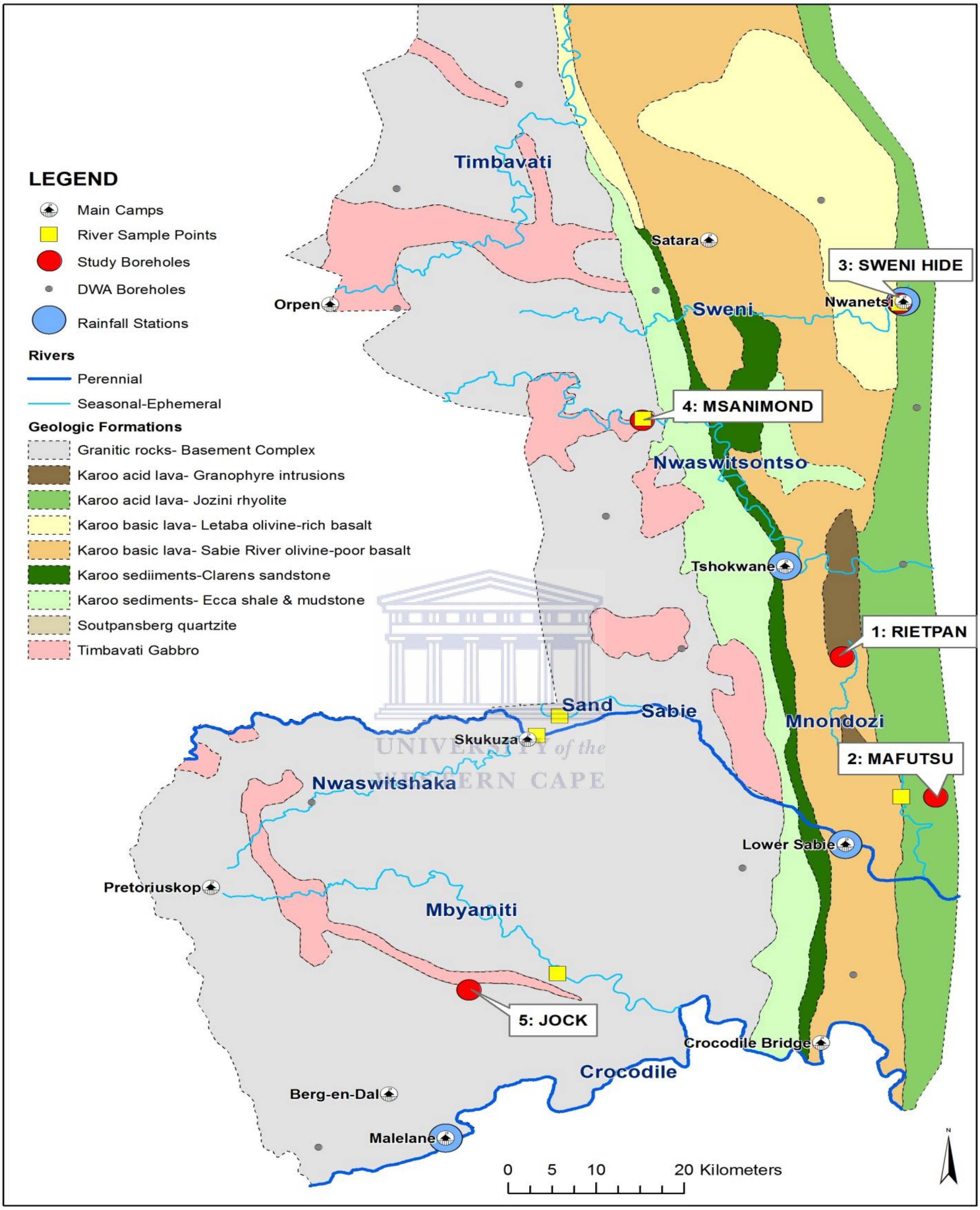


Figure 3.6 Map of southern KNP showing the location of boreholes (red dots) and rainfall stations (blue dots), and the surface-water (yellow squares) sampled in this study.

3.9.1. Description of specific boreholes (Bh) used

1) Bh-Rietpan is located at E31.91421 S24.8969 on the basalts (Karoo Sequence). The area is underlain by the Letaba Formation which is characterised by mafic and ultra mafic extrusive rocks, basalts and andesite. The area is characterised by extensive flat plains with moderately deep to shallow, red and brown, structured clayey soils (olivine-poor). The vegetation is open *Sclerocarya birrea/Acacia nigrescens* tree savanna (Figure 3.5). The nearest rainfall station is located at the Tshokwane picnic site receiving MAP 545.7 mm. The borehole is drilled to a depth of 50 m in the weathered zone. The groundwater is moderately saline with an electrical conductivity (EC) of 2.91 mS/cm. The moderate salinity is indicative that no significant recharge recently took place.

2) Bh- Mafutsu is located at E32.00947 S25.0643 on the rhyolites in south-eastern KNP, relatively close to Lower Sabie rest camp which experiences an MAP of 562.82mm. The area is characterised by undulating low mountains and hills with rocky outcrops. The soils are mainly talus with very shallow to moderately deep brown eutrophic apedal loam and clay. The vegetation is broad leaved deciduous bushveld with no Mopane (Figure 3.5). The borehole is drilled to a depth of 50 m in the weathered zone. The borehole is low yielding as the borehole is pumped dry within 5 to10 minutes of pumping at a constant rate of 0.25l/s during sampling. The groundwater has a low salinity EC 1.295 mS/cm and pH 7.34 which indicates that groundwater recharge is experienced frequently. The borehole is located close to ephemeral streams and the Mnondozi River (3.53 km away) which flows during the wet season and maintains pools along certain reaches of the river.

3) Bh- Sweni-hide is located at E31.97269 S24.4736 also on the rhyolites close to the Nwanetsi picnic site which receives a MAP 510 mm. The area is characterised by flat plains with broad leaved deciduous bushveld with no Mopane (Figure 3.5). The borehole is drilled on a rhyolitic rocky outcrop with no soil layer close to the borehole but rather an exposed rocky surface that is not representative of the surrounding landscape. Like Mafutsu the borehole is low yielding where, and was pumped dry within 10 to 15 minutes at 0.25 l/s during sampling. The seasonal Sweni River, which maintains pools during the dry season, flows approximately 50 meters away from the borehole. The elevated EC levels of 3.47 mS/cm could be a result of surface water/ groundwater interaction as during the dry winter months standing pools of water in the river are evaporated (Leyland *et al*., 2008).

4) Bh-Jock is located at E31.92509 S25.2775 in the south western KNP on the Basement Complex which in the southern part of the park consists of granitic rocks, gneiss, migmatite, amphibolites and schist (Schutte, 1986). The area, underlain by the Nelspruit Granite Suite, is characterised by undulating plains with the occurrence of numerous granitic koppies and rocky outcrops scattered around the area. The nearest rainfall gauging station is at the Malalane rest camp which experiences a MAP of 600 mm. The surrounding vegetation is broad-leaf to fine-leaf bushveld, typically acacias (Figure 3.5). The borehole is drilled to a depth of 25 m in the weathered zone. The soils are shallow, medium to coarse sand to loam over weathered bedrock. The groundwater is fresh with an of EC 1.377 mS/cm indicative of low mineralization which suggest that frequent recharge occurs. The borehole is located close to small ephemeral streams and the seasonal Mbyamiti River which is 9.36 km away.

5) Bh-Msanimond is located at E31.70972 S24.6138. The area is underlain by the Orpen Gneiss which consists of fine-grained grey to dark grey homogeneous gneiss without ultramafic xenoliths. The Orpen Gneiss consists of quartz, plagioclase, biotite and minor amounts of microcline and sphene (Schutte, 1986). The area is characterised by slightly undulating plains with very few rocky outcrops scattered around. The soils are shallow to moderately deep eutrophic red and yellow coarse sand and loamy sand with duplex sodic clays. The vegetation is broad-leaf to fine-leaf bushveld, typically acacias 9Figure 3.5). The nearest rainfall gauging station is at the Tshokwane picnic site receiving MAP 545.7 mm. The borehole is drilled to depth to 37 m in the weathered zone. Msanimond has highest EC values at 5.41mS/cm which is indicative of enhanced water-rock interactions, the high mineralization suggests infrequent recharge. The high salinities associated with this groundwater indicate that insignificant recharge was experienced under the present climatic conditions. The borehole is located close to the ephemeral Nwaswitsontso River (240 m away) which only flows after high rainfall events.

3.1. Implications for recharge

The seasonal nature of the rainfall in the KNP means that groundwater recharge will only occur in the wet season, November to April, with a discharge period in the dry season between May and September. The year-to-year and long-term trends in rainfall, as well as frequency, duration and intensity of individual storm events will affect recharge processes.

The permeability of the surface and subsurface materials can greatly affect recharge processes. Healy and Scanlon (2010), describe that recharge is more likely to occur in areas that have coarse-grained, high permeability soils as opposed to areas with fine-grained low-permeability soils. Coarse-grained soils have relatively high permeability and are capable of transmitting water rapidly. The presence of these soils promotes recharge because water can quickly infiltrate and drain through the root zone before being extracted by plant roots. Finer-grained sediments are less permeable, but are capable of storing greater quantities of water. Therefore, in areas of finer-grained sediments there would be decreased infiltration, enhanced surface runoff, increased plant extraction of water from the unsaturated zone, and decreased recharge relative to coarse-grained sediments.

The land surface topography plays an important role for both direct and indirect recharge. Steep slopes tend to have low infiltration and high runoff rates. These conditions are favorable to indirect and localised recharge processes through small depressions, cracks and fractures. Flat land surfaces that have poor drainage are more conducive to direct recharge (Healy and Scanlon, 2010). The flow mechanism through the unsaturated zone is thus likely to be via piston flow as precipitation that is stored in the unsaturated zone is displaced downward by the next infiltration event as a result of the poor drainage and runoff.

Hypothesis

Along the basalts, which are characterised by flat plains with fine-grained low permeable clayey soils, direct recharge via piston flow is expected to be the dominant process. Along the granites, characterised by steep to moderate slopes and rocky outcrops with coarse-grained highly permeable soils, exposed cracks and fractures, indirect and localized recharge via preferential pathways are expected to be dominant. In the granites the local surface hydrology particularly perennial, season and ephemeral streams is expected to play a vital role in contributing to indirect recharge as preferred pathways.

CHAPTER 4

PHYSICAL METHOD: CUMALATIVE RAINFALL DEPARTURE (CRD)

4.1. Introduction

Long term periods of drought and the migration of wildlife from the KNP resulted in artificial water provision for wildlife in the park. This resulted in the construction of dams and the drilling of boreholes, a present estimate of 1000-2000 boreholes has been drilled across the park (Du Toit, 1998). Very few water level measurements have been taken since the first borehole was drilled in the mid-1920s (Du Toit, 1998). There is huge knowledge gap on borehole information such as frequent historical water table depths, lithological logs, screen depths, water strikes, blow yields and aquifer characteristics. Regular water level measurements were introduced for the first time in 2001 when measurements were taken twice a year (at the end of the wet and dry seasons respectively) at a selected number of boreholes across the Park (Du Toit *et al.* 2009). Presently groundwater levels are measured at unused boreholes across the KNP using Solinst level loggers. The loggers were installed in 2007 by The Department of Water Affairs (DWA) and are programmed to take water level readings that are corrected to atmospheric pressure from barometer-loggers installed congruently at hourly intervals. The data are downloaded twice a year, before and after the wet season. Rainfall data have been collected in the park since the 1920s. Presently rainfall is measured on a daily basis by field rangers in their respective sections using standard rain gauges or in some cases weather stations. Rainfall data is consolidated centrally at Skukuza on a monthly basis.

To develop a conceptual model of the dominant groundwater recharge processes in the southern region of the KNP the Cumulative Rainfall Departure (CRD) method as described in Chapter 2s used. The short-term and long- term memory of the system was incorporated to infer potential recharge processes. The results are discussed based on the two main geologic formations in the KNP namely the southern basalts and the southern granites.

4.2. Methodology

The water level data was sourced from (DWA) Limpopo Geo-hydrology Directorate and rainfall from KNP data bases. The water level measurements taken by the park (section rangers) from 2001 to 2007 together with the data logger data collected by DWA from 2007 till May 2009 were used to determine the correlation between the CRD and observed water level response from the five selected boreholes and rainfall gauging stations. The water level data from these boreholes was processed using monthly water level measurements that were extracted from the data sets taken on the 30th of every month except for February which was taken on the 28th. Long term monthly rainfall data obtained from the nearest rainfall gauging stations (based on the continuity of the data) were utilized to calculate the annual and monthly rainfall rates which is plotted as a bar graph beneath the borehole hydrograph. As described in chapter 2, the CRD was derived from first principles using equation 2.3 (page). The short and long term memory of the system was incorporated using equation 2.5 (page).

For the CRD simulation 1, 2, and 3,6,12 month average rainfall values for the short-term memory and for the long-term reference value the average rainfall for 3, 5, and 10 years preceding a specific month was used. A correlation coefficient between the measured water level and all the simulated CRD values was calculated. The CRD simulation with the highest correlation coefficient to the measured water level response was used.

4.3. Results

4.3.1: Long term trends in groundwater levels

Analysis of the rainfall data from two stations (Tsokwane and Lower Sabie) from 1990 to 2009 shows an oscillation between wet and dry cycles with a 6-10 year return interval (Figure 4.1). The values were determined as the annual rainfall over the 20 year period minus the MAP over the same period 545.7 mm and 562.82 mm respectively. The results show an oscillation of 6-10 years between wet and dry cycles. These wet and dry cycles have a significant impact on groundwater recharge as the annual rainfall between 1996-2000 are above the average with a peak in 2000 (flood event) causing a substantial potential for recharge, thereafter in 2002-2010 the rainfall is below average resulting in the reduction in recharge potential.

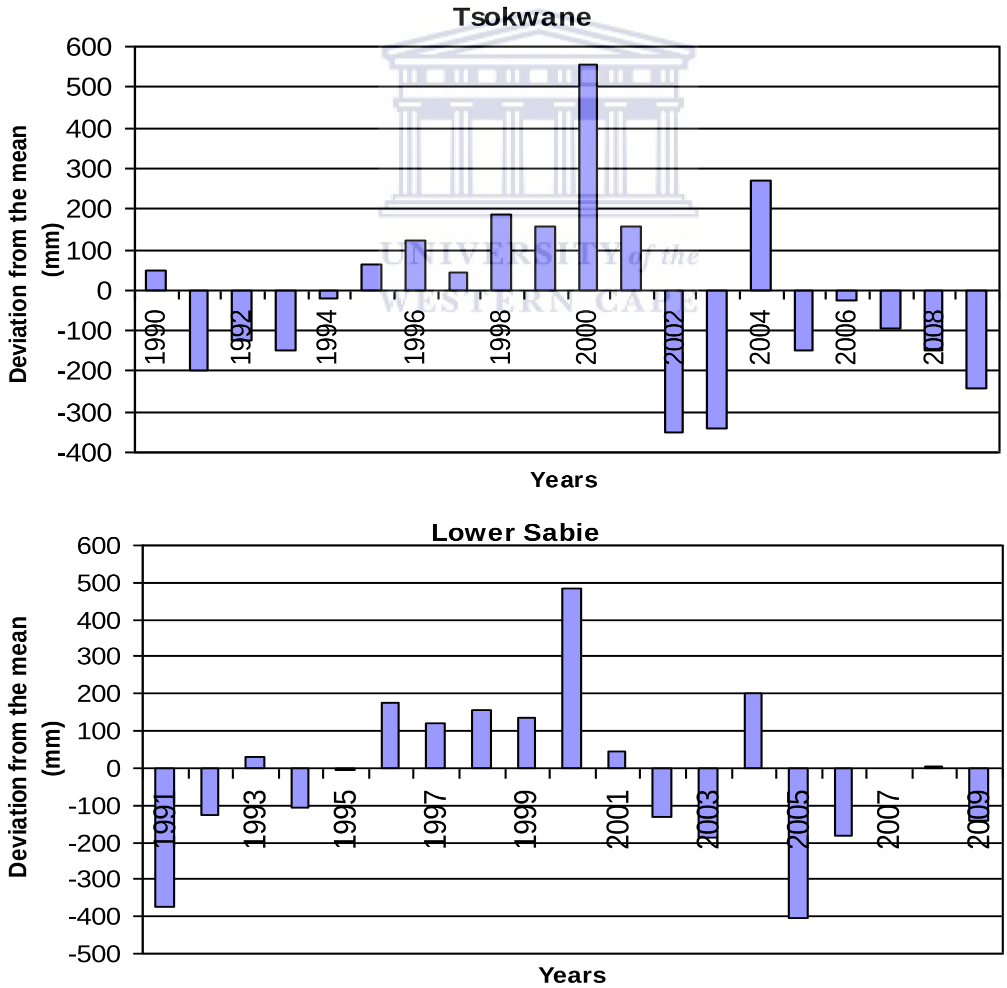


Figure 4.1 Annual rainfall deviations from the mean rainfall over an 18-year period for Tsokwane and Lower Sabie area.

Using the CRD derived from first principles, a reliable reconstruction of the full series of water level fluctuations could be extrapolated from rainfall data (Figure 4.2). Long-term groundwater level trends are simulated with the CRD model using monthly rainfall totals referenced to the average rainfall over the entire time series 1936-2009. The results of the model shows that two prominent low recharge periods occurred during the severe droughts of 1964 – 1976 and 1990- 1994 (Figure 4.2). There are three prominent high recharge periods which occurred in 1955-1961, in 1980 with a peak in 1988 and then in 1996 to 2000 where the park experienced a 1 in 200 year flood. Indications are that the KNP is currently moving into another low recharge period which started around 2001/2002.

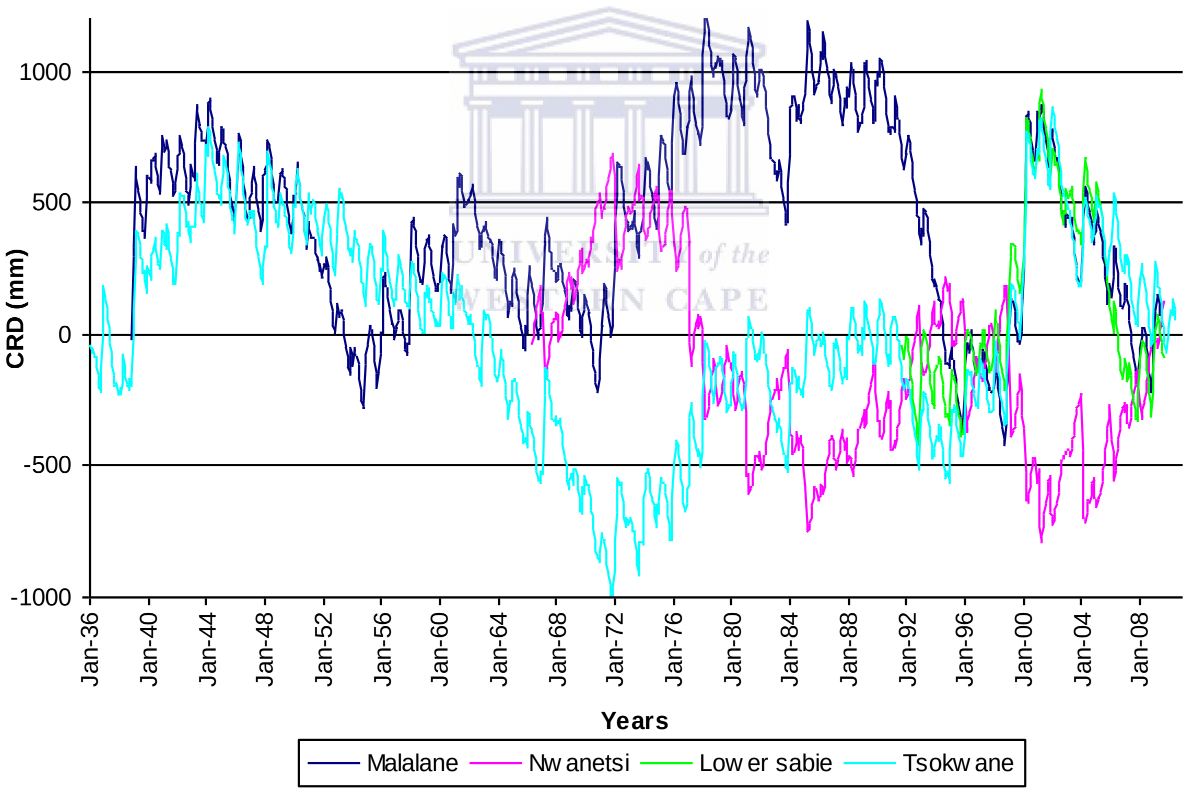
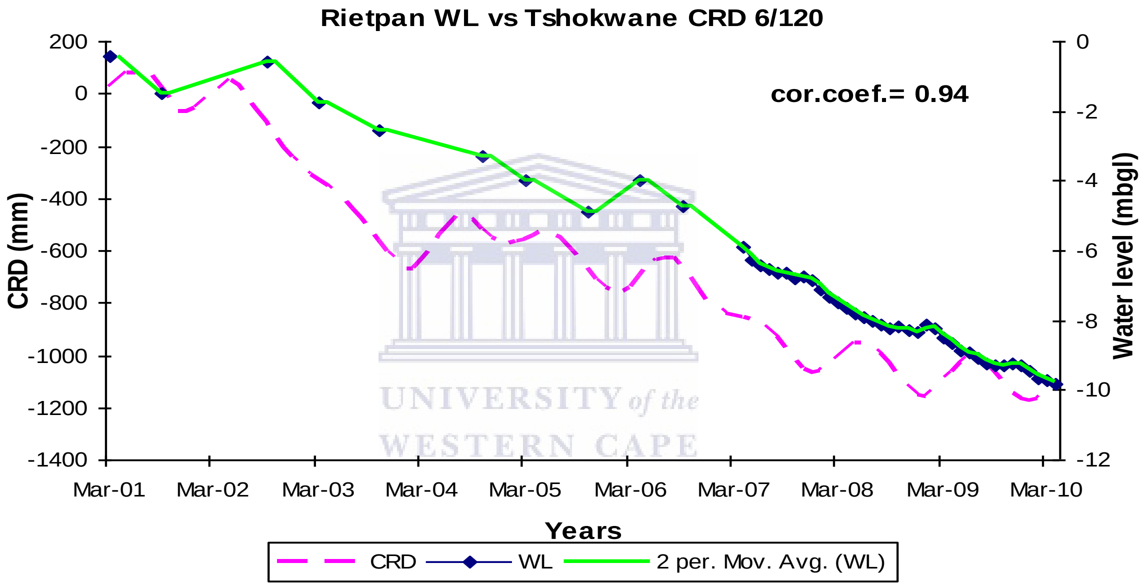


Figure 4.2 Shows the simulated water level behaviour as derived from CRD first principles reflected by the cumulative rainfall departure of four rainfall stations from 1936 to 2009 for four areas in the KNP.

4.3.2. Southern Basalts

On the southern basalts at Rietpan, Mafutsu and Sweni-hide the best simulation model was achieved using a CRD 6/120 for Rietpan (R^2 = 0.94, P = 0.01) a CRD 12/120 for Mafutsu (R^2 = 0.84, P = 0.01) and CRD 1/36 for Sweni-hide (R^2 = 0.56, P = 0.01). A CRD 6/120 is the average rainfall over 6 months relative to a 120 month average rainfall, CRD 12/120 is the average rainfall over 12 months relative to a 120 month average and CRD 1/36 averages 1 month's rainfall relative to a 36 month average rainfall.

4.3.2.1. Bh- Rietpan



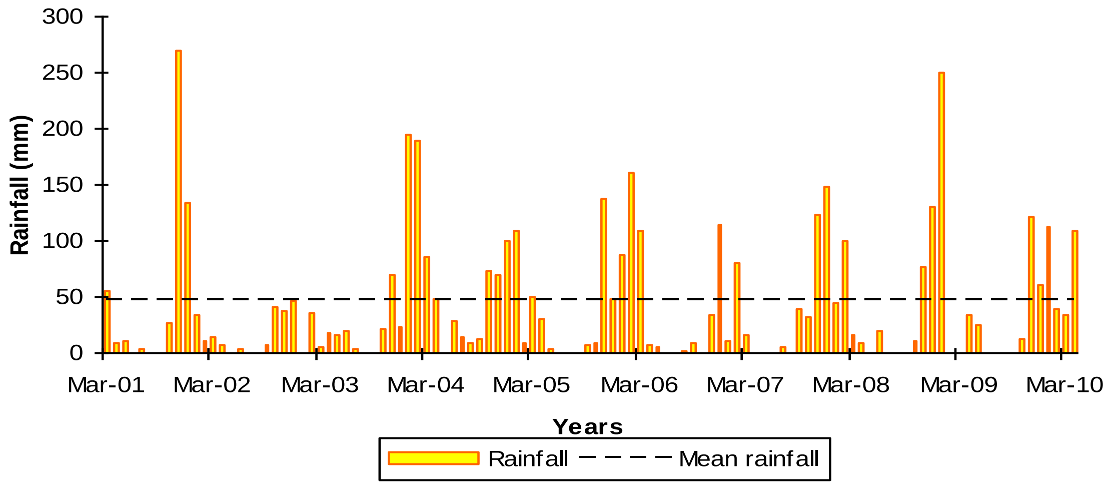


Figure 4.3 Observed borehole data from Rietpan and CRD simulated water levels as derived from the monthly rainfall for Tshokwane situated 14.11 km away.

The water levels progressively declined by 9.39 m for Rietpan and 6.35 m for Mafutsu over the last decade. The aquifer recharged fully during the 2000 flood but gradually declined over the subsequent years. The simulated CRD response shows there is a short-term memory of cumulative preceding rainfall events over 6 months and its impact on groundwater levels. The groundwater level response of the system is intermediate. The 120 month long term memory of the system and gentle fluctuations observed during a recharge event is indicative of the size of the aquifer system and distance of the borehole from the recharge area (Kirchner, 2003). Thus the borehole water level response represents that of a discharge area.

4.3.2.2. Bh- Mafutsu

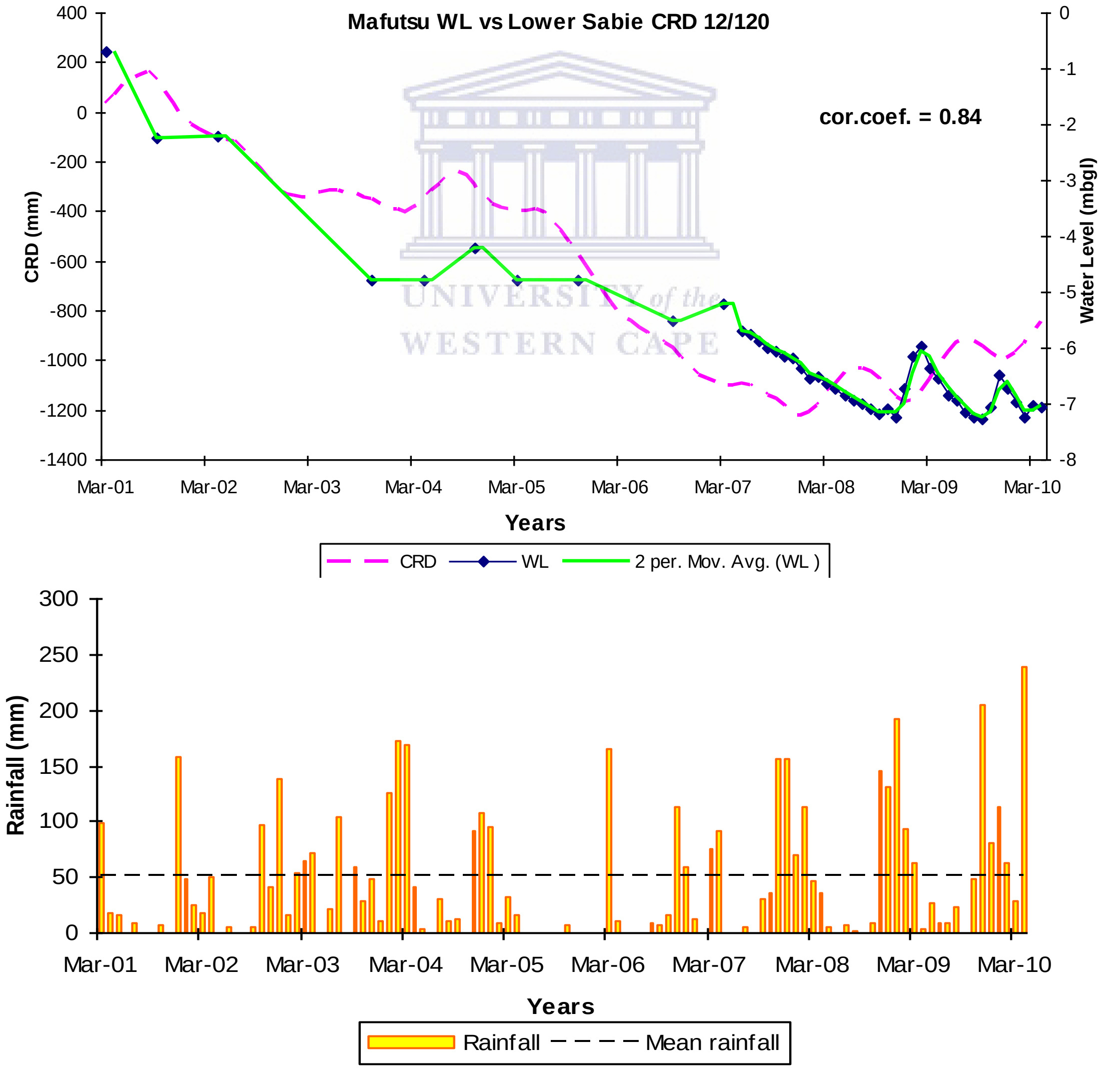


Figure 4.4 Observed borehole data from Mafutsu and CRD simulated water levels as derived from the monthly rainfall for Lower Sabie situated 11.38 km away.

At Mafutsu it is only during rainfall years well below average that there is a decline in water levels. The water levels response and simulated CRD suggest that the water level in the borehole responds to moderate term (yearly) rainfall averages. The intensity of the rainfall also determines the response of the water levels. In 2006 rainfall year one significant rainfall event took place in March (165mm) the water level responded a year later, and then quickly declined (Figure 4.4).

4.3.2.3. Bh- Sweni- Hide

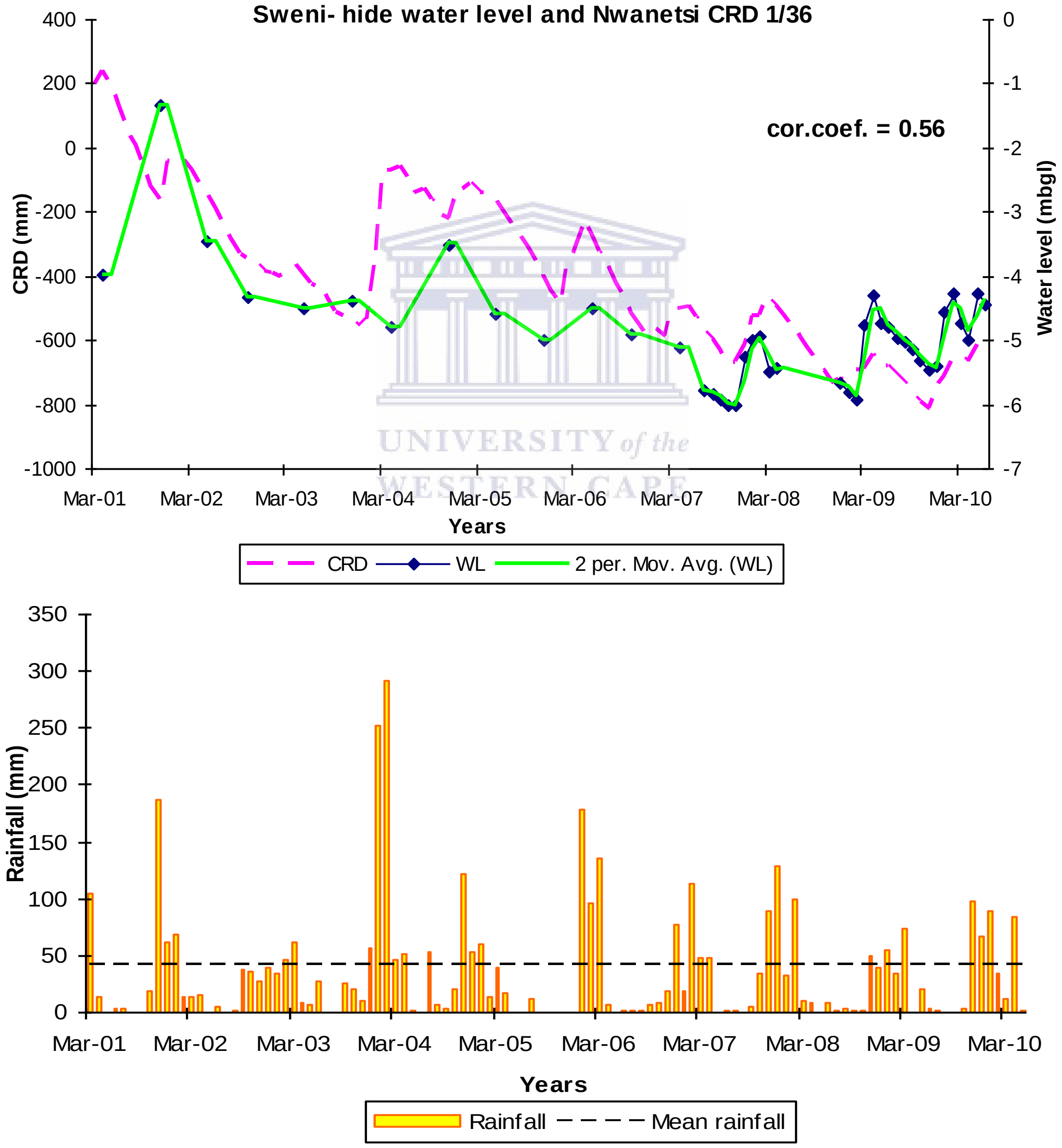


Figure 4.5 Observed borehole data from Sweni-hide and CRD simulated water levels as derived from the monthly rainfall for Nwanetsi situated 25.84 km away.

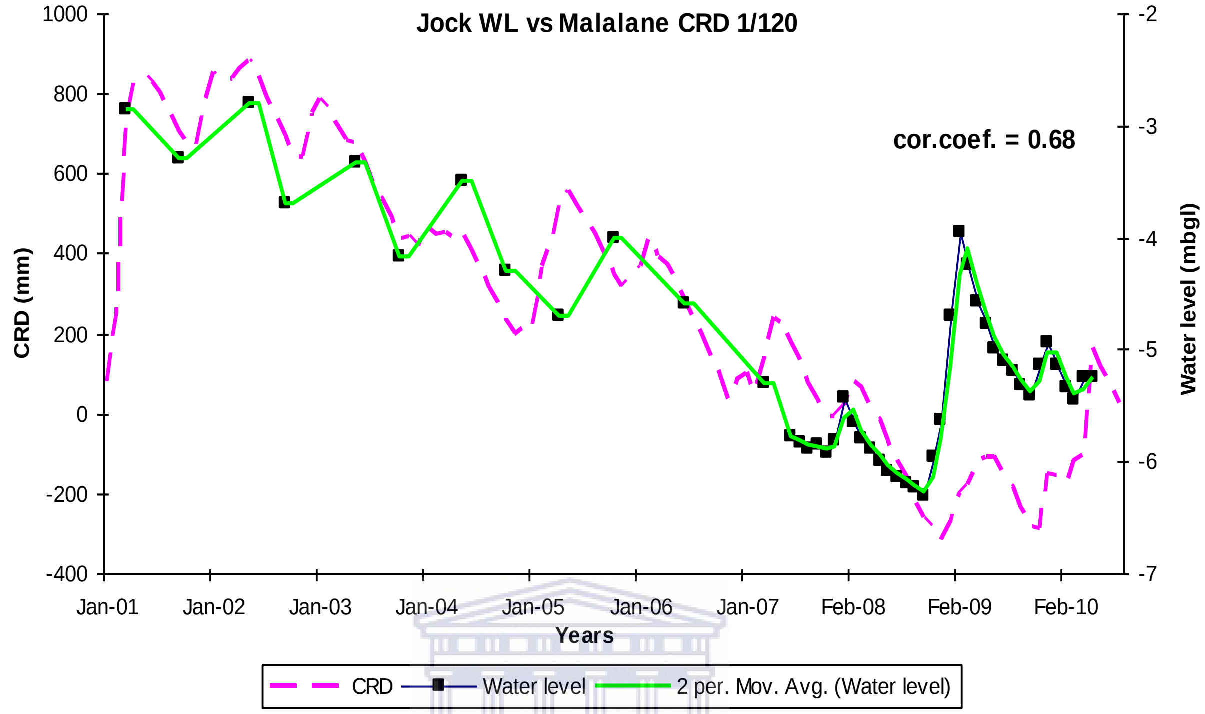
At Sweni-hide the flood event caused the water level to steeply rise by 3 m then quickly declined. Over the past decade the groundwater level trends display seasonal fluctuations in response to local rainfall over time. This response is typically indicative of a local scale groundwater flow system. Groundwater levels rise after rainfall events particularly during the wet season January- March, then decline in the dry winter months. The groundwater level responds to a 1 month cumulative series of preceding rainfall events. The water level responds rapidly to rainfall events.

4.3.3: Southern granites:

On the southern granites at Jock and Msanimond, the best simulation model was achieved using a CRD 1/120 for Jock (R^2 = 0.68, P = 0.01) a CRD 12/120 for Msanimond (R^2 = 0.57, P = 0.01). A CRD 1/120 is the average rainfall over 1 month relative to a 120 month average rainfall, CRD 12/120 averages the rainfall over 12 months relative to a 120 month average rainfall.

The groundwater level fluctuations trends that are observed at Jock display season fluctuations to local rainfall over time (Figure 4.6). This response is typically indicative of a local scale groundwater flow system. Groundwater levels rise after rainfall events particularly during the wet season January to March, then decline in the dry winter months. In Figure 4.6 the measured water level is plotted against the simulated CRD response. Analysis of the CRD illustrates the groundwater level responds to a 1 month cumulative series of preceding rainfall events. The aquifer experiences a rapid response to rainfall event.

4.3.3.1. Bh- Jock



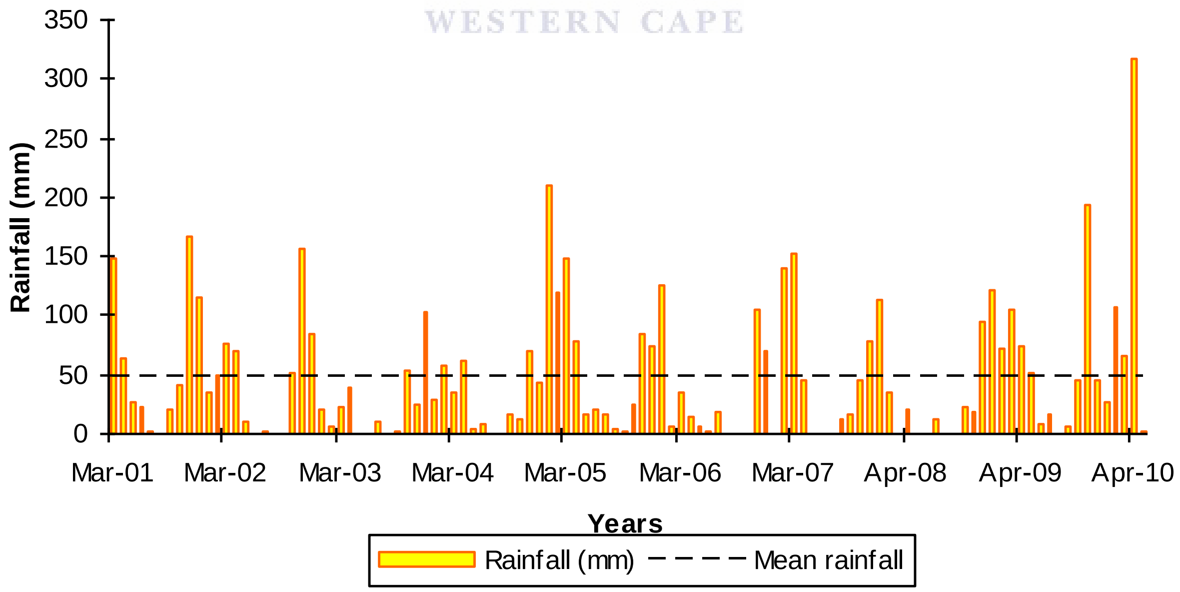
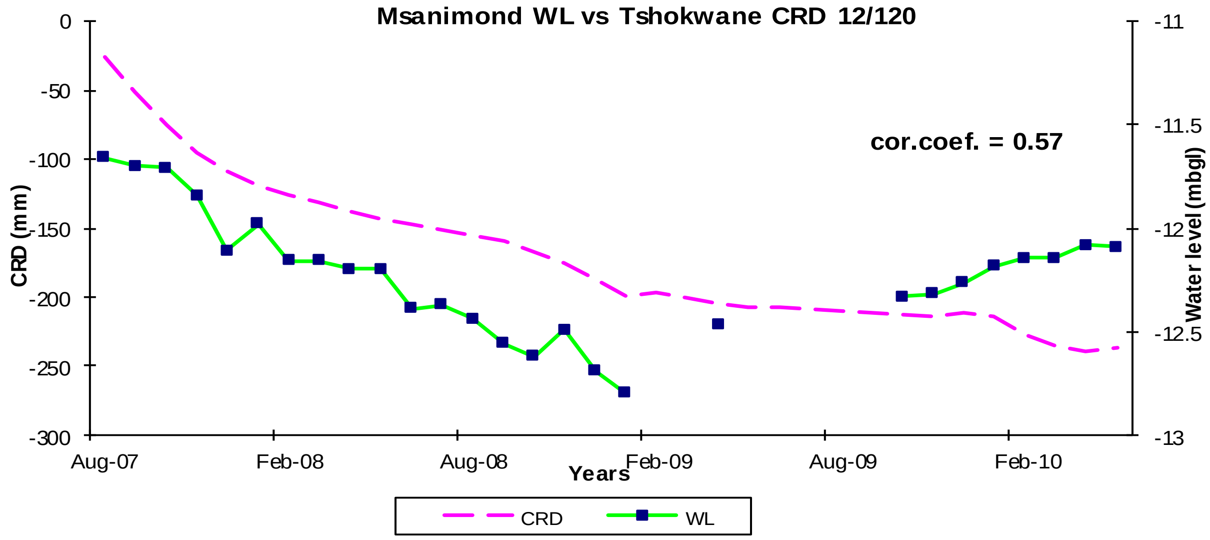


Figure 4.6 Observed borehole data from Jock and CRD simulated water levels as derived from the monthly rainfall for Malalane situated 19.48 km away.

4.3.3.2. Bh- Msanimond



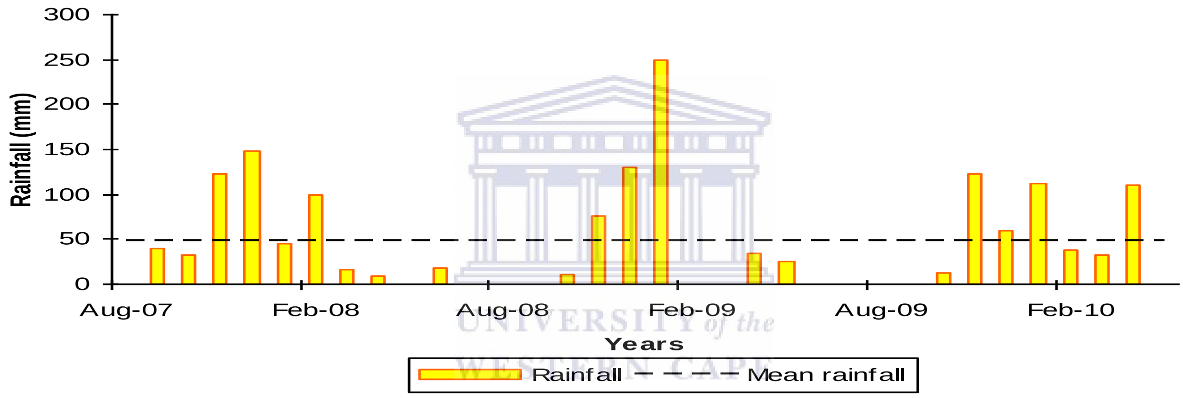


Figure 4.7 Observed borehole data from Msanimond and CRD simulated water levels as derived from the monthly rainfall for Tshokwane situated 23.64 km away.

The groundwater level trends that are observed at Msanimond are relatively stagnant. The measured water level is plotted against the simulated CRD response (Figure 4.7). Interpretation of the CRD illustrates the groundwater level responds to a cumulative 12 month series of preceding rainfall events. The aquifer experiences an intermediate to long term response to rainfall events. The borehole is located in close proximity to the Nwaswitsontso River which is a large ephemeral stream that may be providing bank storage recharge to the system during periods of flow. The response time of the water level to rainfall could be the result the unsaturated thickness and low saturated and unsaturated hydraulic conductivities of the aquifer. Further, the minimal fluctuation in the water level response could indicate that the borehole is situated far from the recharge area and is actually in a discharge area. The system did not respond to seasonal fluctuations. Although the data series is not sufficiently long, the system seems to respond to long term wet and dry cycles.

4.4. Discussion

The results from the CRD simulation and the measured water level response indicate that groundwater recharge is driven by periods of above-normal rainfall. Recharge occurs during the rainy summer months (December to March) and very little to none during the dry winter season. There are low and high recharge periods that vary in duration from 6 up to 14 years. These periods are characterised by seasonal fluctuations i.e. there are gains and losses during consecutive hydrological years, but because the total recharge added to the system annually is lower or higher than the natural losses depending on the type of period, the water level will have either a declining or a rising trend.

The best simulations along the basalts between the measured water levels and simulated CRD were obtained using a CRD 12/120 and 6/120. The short term memory in this case proves to be cumulative preceding rainfall events over 6 to 12 months. This implies that the water level response is intermediate which means that recharge occurs via direct and localised flow paths. Here piston flow and surface runoff ponds in small depressions are dominant processes as ponding was observed at the sites during the raining season.

Where a weak correlation coefficient between the simulated CRD response and the measure water level was obtained as at Sweni-hide (R^2 = 0.56) the distance the nearest rainfall station was in excess of 25 km which misrepresents the spatial distribution of rainfall. The frequency of measured water levels from 2001-2007 which were only taken twice a year compared to monthly rainfall data further affected the result. However, the measured water level response closely follows the trend of the CRD. The short term memory of

1and 12 months obtained implies that the aquifers respond within 1 and 12 month to cumulative preceding rainfall events. The 1 month short term memory obtained on the southern granites at Jock suggest that the water level responds rapidly to rainfall events, which in this case shows that localised recharge will occur via preferred pathways such as through fractures, cracks and joints and indirect recharge via surface water channels from small streams and the Mbyamiti River. At Msanimond the water level response is intermediate indicating that recharge could be taking place indirectly via surface water channels through the Nwaswitsonto River.

Recharge does not only occur in a vertical direction but also in horizontal direction. The water level rise at a given point in an aquifer therefore also depends on the distance the borehole is from the recharge area. The very low transmisivities and storativity inherent in Basement aquifers means that the lower part of the aquifer is only recharged several months after rainfall events. As a result the water level response observed at Rietpan, Mafutsu and Msanimond suggest that these boreholes are situated in a discharge zone and that Jock and Sweni-hide are potential recharge areas.

The 120 month long term memory of the system and the water levels response of the boreholes which all show a declining trend suggest that the piezometric response over the southern region of the KNP are comparable. This implies that the average rainfall for the different areas over periods of approximately 10 years fluctuate in a similar pattern. This confirms that there is a high level of interdependency on the regional long term rainfall patterns.

Surface- water / groundwater interactions

Recharge processes along the basalts are dominated by direct recharge via piston flow and along the granites recharge via preferred pathways are dominant. I therefore propose that particularly along the granites there is a strong link between surface-water and groundwater as the streams and rivers in the area are acting as preferred pathways for recharge. Conversely along the basalts where direct recharge via piston flow is dominant, surface- water interactions with streams and rivers will possibly not play a significant role in recharge. In the southern region of the KNP the overall declining water level trends over the past decade where natural losses are greater than natural gains groundwater could be discharging into the perennial rivers such as the Sabie and Sand Rivers, thereby maintaining the baseflow component of these rivers.

CHAPTER 5

CHEMICAL METHOD: STABLE ISOTOPES ^{2}H AND ^{18}O

5.1. Introduction

Stable isotopic analysis of ^{2}H and ^{18}O of precipitation, surface water and groundwater can play an important role in identifying recharge processes. Therefore, the isotope approach was used to test whether the predicted recharge processes inferred from the CRD model are plausible. The isotope composition of the groundwater is used to establish whether recharge was immediate or delayed. Additionally, the isotopic composition of surface-water from rivers and streams were compared to that of groundwater to identify surface-water interactions particularly if they act as preferred path ways for recharge.

5.2. Methods

At each site groundwater, surface-water and rainfall samples (n=176) were collected on a monthly basis during the dry season (May to August 2010), the wet summer season (January to April 2011) and winter (May to July 2011).

5.2.1. Rainfall

Cumulative monthly rainfall samples were collected via permanent rainfall collectors installed on-site at a close proximity (10 m- 50 m) to the boreholes. The rainfall collectors are standard plastic 100 mm rain gauges that are filled with at least 30 mm of paraffin oil to prevent evaporation. There were indications that evaporation occurred prior to the collection of the water samples. Water samples collected during the dry season (June- August 2011) at a few collection stations showed unusually high $\delta^{18}O$ and $\delta^{2}H$ values, I suspect this resulted from the small amount of rain collected for the period (2 mm) that could not be separated successfully from the oil layer in the rainfall collector. Additionally during the rainy season when the rainfall exceeded 100 mm the collectors overflowed, resulting in the samples evaporating. These samples were considered non- representative samples.

5.2.2. Surface-water

Surface water samples were collected from seasonal and ephemeral rivers (Mbyamiti (Bh- Jock), Sweni (Bh-Sweni-hide), Mnondozi (Bh-Mafutsu), and Nwaswitsontso (Bh-Msanimond)) and perennial rivers (Sabie and Sand). Water samples collected from the associated rivers and streams were sampled directly (grab samples) about 10-15 cm below the water surface.

5.2.3. Groundwater

Groundwater samples were collected from the selected boreholes by purging the aquifer using a submersible pump powered by a small generator. The water samples were collected directly from the outlet pipe/tap after the electrical conductivity (EC) stabilized or 3 bore volumes were abstracted to ensure that water representative of the aquifer are sampled. The EC was measured in-situ using a hand help multi- probe. All the water samples that were collected were stored in insulated bottles and followed sampling protocol described by Weaver *et al*., (2002).

5.2.4. Analysis

All the samples collected (n=176) were analysed for $\delta^{18}O$ and $\delta^{2}H$ using a ThermoFinnigan DeltaXP mass-spectrometer at the University of Cape Town (UCT). At UCT the samples are prepared off line, for $\delta^{2}H$ the analysis is done on hydrogen obtained in glass “break seal” tubes through high temperature reduction of water (3 µL) on specially prepared zinc. The $\delta^{18}O$ analyses are done on dioxide that has equilibrated with the water sample at a constant temperature (25 ºC). Between 1 and 3ml of water is added to reactions vessels, and then evacuated by pumping through capillary tubes connected to the mass spectrometer (Clark and Fritz, 1997). The measured values are reference against UCT in-house standards CTMP and the secondary internal standard DLMICE calibrated to VSMOW.

5.3. Results

5.3.1. Rainfall

A total of 45 rainfall samples were collected throughout the sampling period. These samples were used to develop a Local Meteoric Water Line (LMWL). The isotopic composition of the groundwater and surface-water is compared to the LMWL to establish whether these waters experienced evaporation. The equation for the best fit line, (LMWL) was developed using a regression analysis with an equation $\delta^2H= 8.66\delta\ ^{18}O + 2.23$ ($R^2= 58.45$)

To test for seasonal effects the plot in Figure 5.1 shows that there are strong seasonal variations in $\delta^{18}O$ in rainfall as more depleted values are observed during the summer rainy season compared to that of rainfall occurring during the dry winter season. This could be the consequence of the amount effect where large rainfall events tend to be depleted in $\delta^{18}O$ relative to small rains. This will have an effect on the isotopic composition as the intensity and duration of rainfall events will result in different isotopic ratios (Kendall and McDonnell, 1998; Clark and Fritz, 1997). Figure 5.2 illustrates the relationship between the rainfall amount and $\delta^{18}O$ indicating a poor correlation $R^2 = 0.3$.

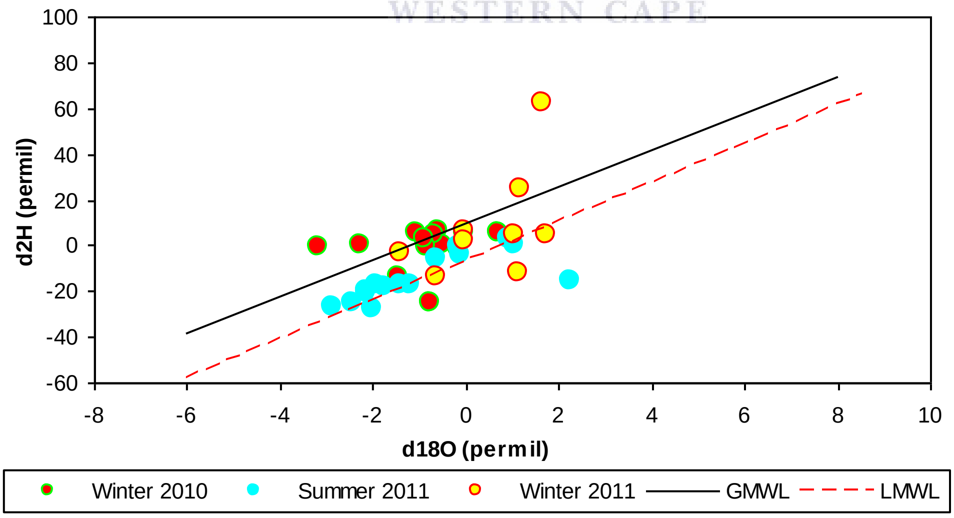


Figure 5.1 A scatter plot of d2H (permil) vs. d18O (permil) of the seasonal variations in rainfall for all collection stations.

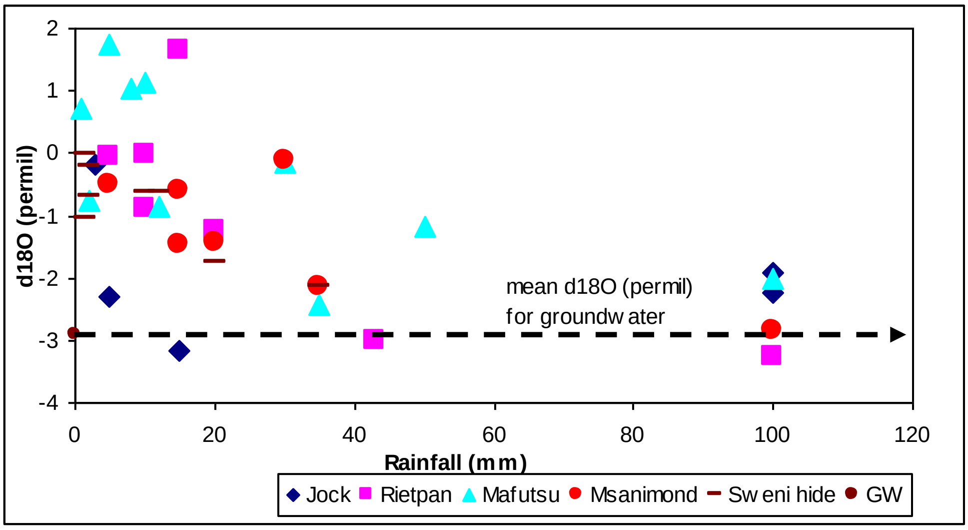


Figure 5.2 A scatter plot of d18O (permil) and rainfall amount. The dashed line represents the mean d18O (permil) of groundwater.

5.3.2. Groundwater isotopic composition

The isotopic composition of the groundwater in the study area will be used here to determine whether recharge is immediate or delayed. The δ^2H vs. $\delta^{18}O$ data for all groundwater sampled are plotted against the LMWL providing information on the secondary processes acting on the water as it travels from precipitation to groundwater (Figure 5.3). The majority of groundwater sampled cluster either close to or to the right of the local meteoric water line, indicating that these samples have undergone evaporation prior to infiltration (Kendall and McDonnell, 1998). Groundwater samples had the most negative $\delta^{18}O$and δ^2H values of all the samples collected, with a mean value of -2.91‰ and -31.83‰ respectively with lowest standard deviations between samples. Groundwater values showed some variation between seasons although no definitive trend is observed in Figure 5.4. The mean $\delta^{18}O$ compositions of all the samples taken differ slightly among the boreholes, ranging from -3.71‰ at Msanimond to -2.35‰ at Sweni hide. There is no relationship R^2 = 0.51 between the groundwater levels and $\delta^{18}O$ (Figure 5.5), additionally the low standard deviation (Stdev = 1.125) between groundwater samples allows for the assumption that groundwater over the entire area exhibits similar values. There is no relationship R^2= 0.25 with the electrical conductivity (EC) and $\delta^{18}O$ (Figure 5.6). This indicates that the source of the groundwater has undergone minimal evaporation and that evaporative concentration is not a significant process for these samples. The high salinity exhibited in a few boreholes is largely due to water rock interactions.

Analysis of the rainfall over the entire range of amounts indicates that the $\delta^{18}O$ and $\delta^{2}H$ values for cumulative rainfall events less than 35 mm tend towards the positive end (Figure 5.3). During cumulative rainfall sequences in excess of 35-50 mm and particularly during rainfall sequences 100 mm or more the isotopic values are more depleted and closely resemble groundwater values.

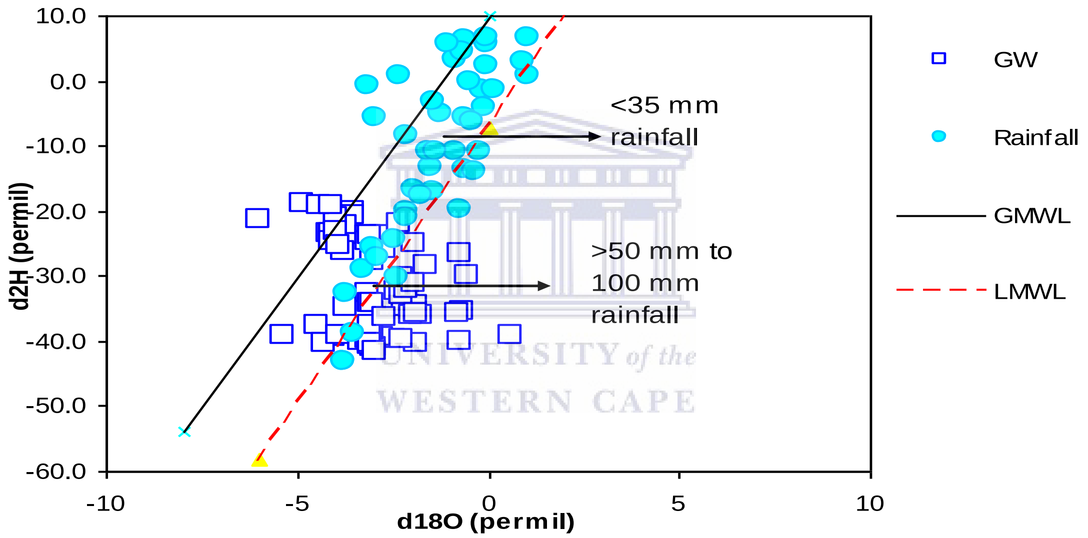


Figure 5.3 d2H (permil) vs. d18O (permil) scatter plot of all groundwater and all rainfall samples.

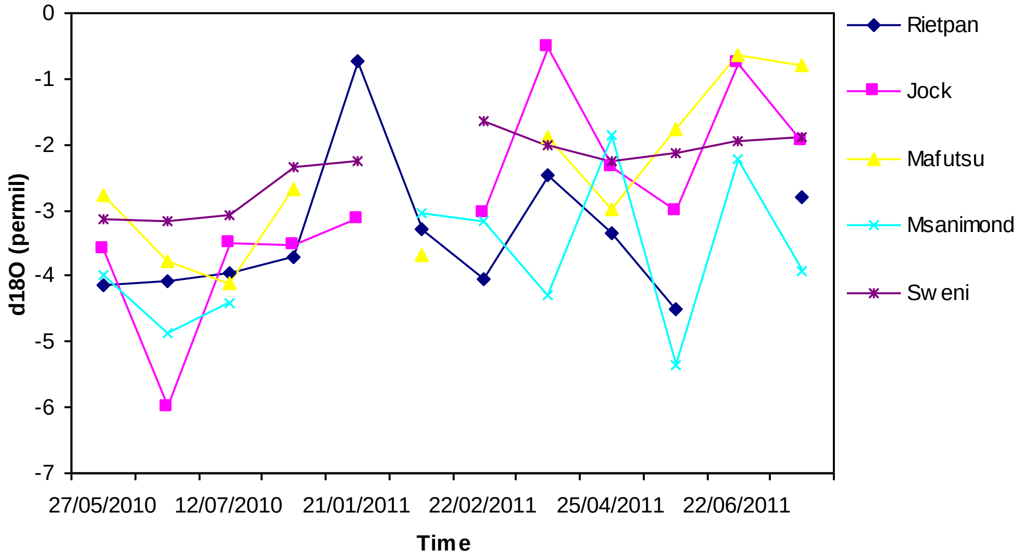


Figure 5.4 Shows the seasonal variability of d18O (permil) of groundwater.

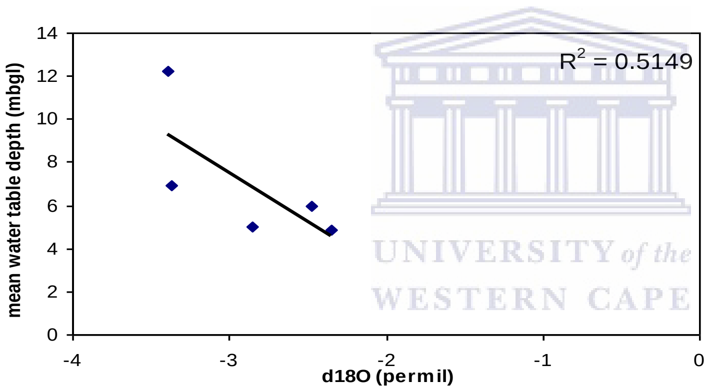


Figure 5.5 Groundwater levels (mbgl) vs. d18O (permil).

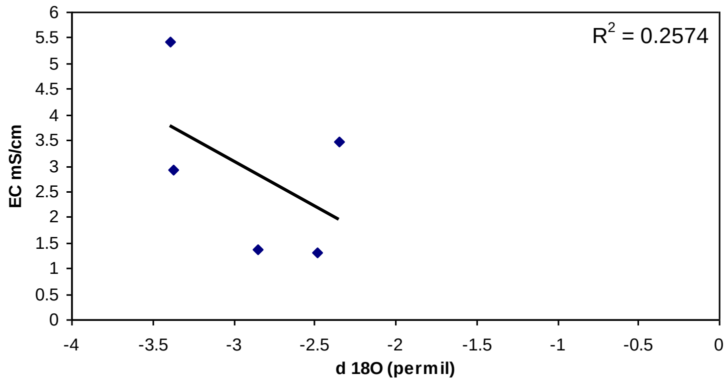


Figure 5.6 Electrical Conductivity (EC) vs. d18O (permil).

5.3.3. Surface water (rivers and streams)/ groundwater interactions

The isotopic values of the rivers and streams that were used in this study are compared to that of groundwater to qualitatively evaluate surface-water / groundwater interactions and the temporal variability of these interactions.

5.3.3.1. The perennial Sabie and Sand Rivers

The stable isotope data plotted in Figure 5.7 provides some interesting insights. Surface water values from the perennial rivers (Sabie and Sand) exhibit a distinct groundwater signature throughout the year. The average δ^2H and δ^{18}Ovalues of groundwater are -31.6‰ and -2.91‰ and surface-water -26.2‰ and -2.43‰ respectively, suggesting surface-water dependence on groundwater discharge.

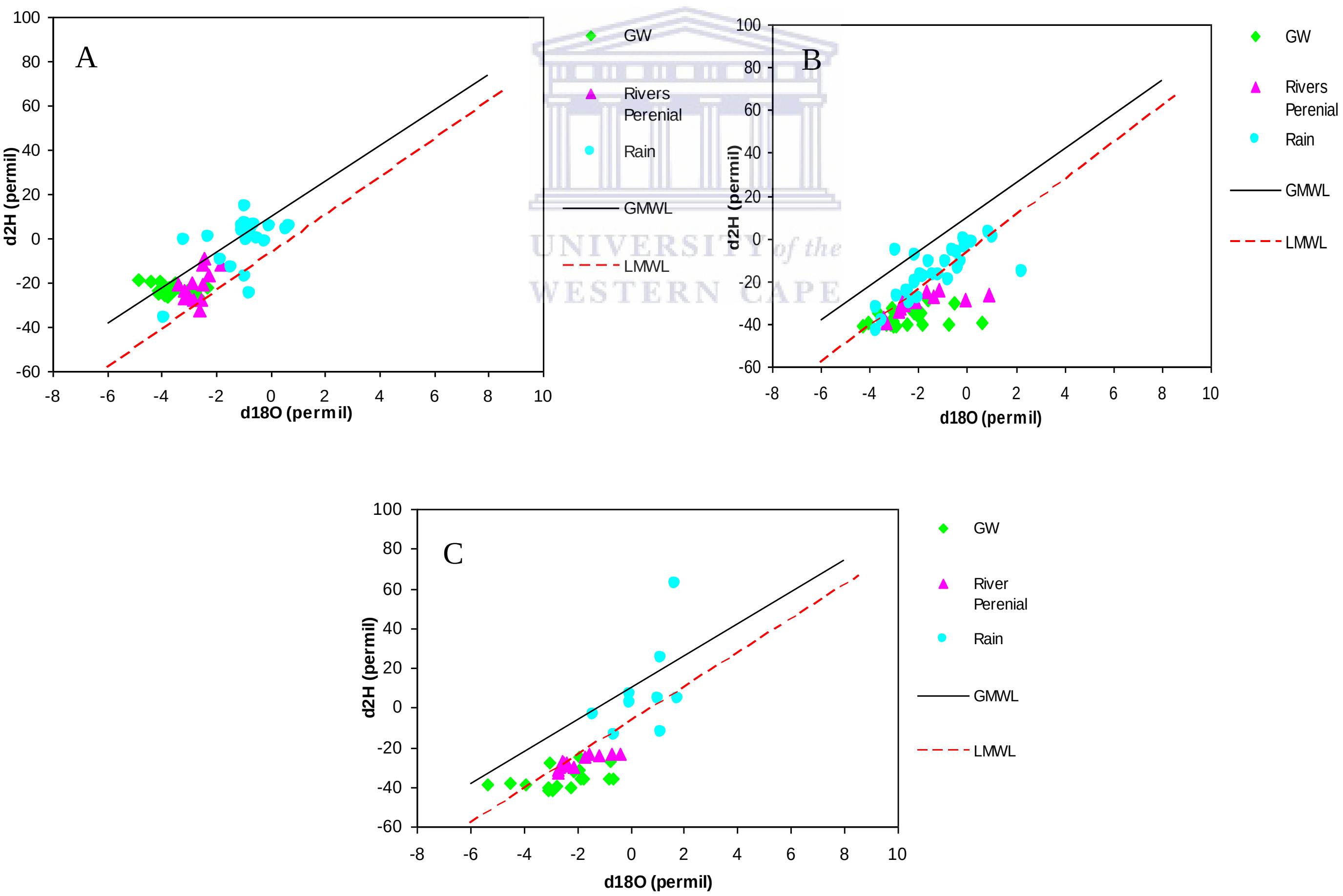
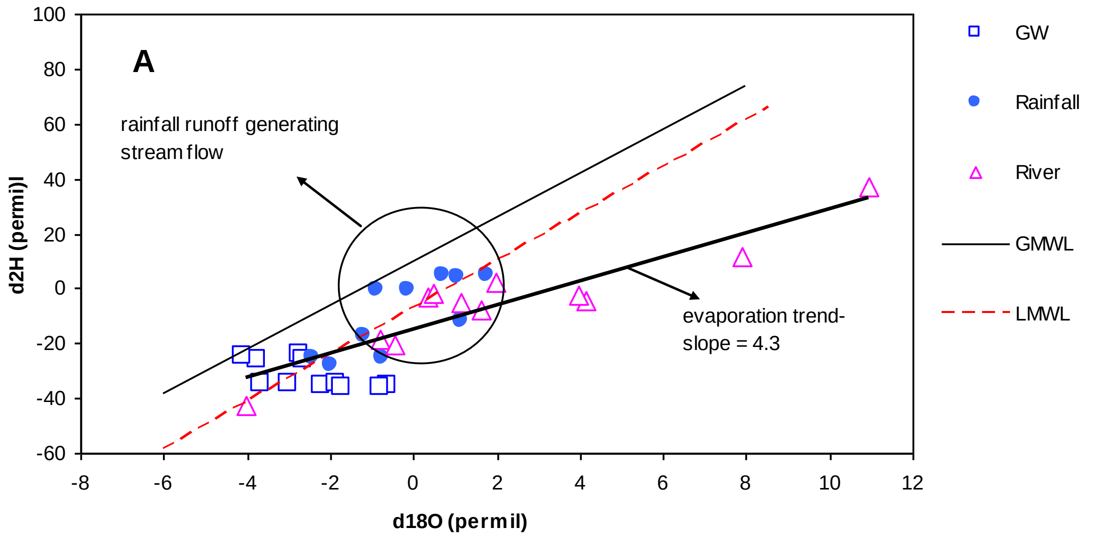


Figure 5.7 Illustrates d2H (permil) vs. d18O (permil) concentrations for the perennial Sabie and Sand River and all the groundwater and rainfall sampled during, A (Winter-May-August 2010), B (Summer Jan-April 2011) and C (winter -May-July 2011).

5.3.3.2. The seasonal and ephemeral rivers

Basalts



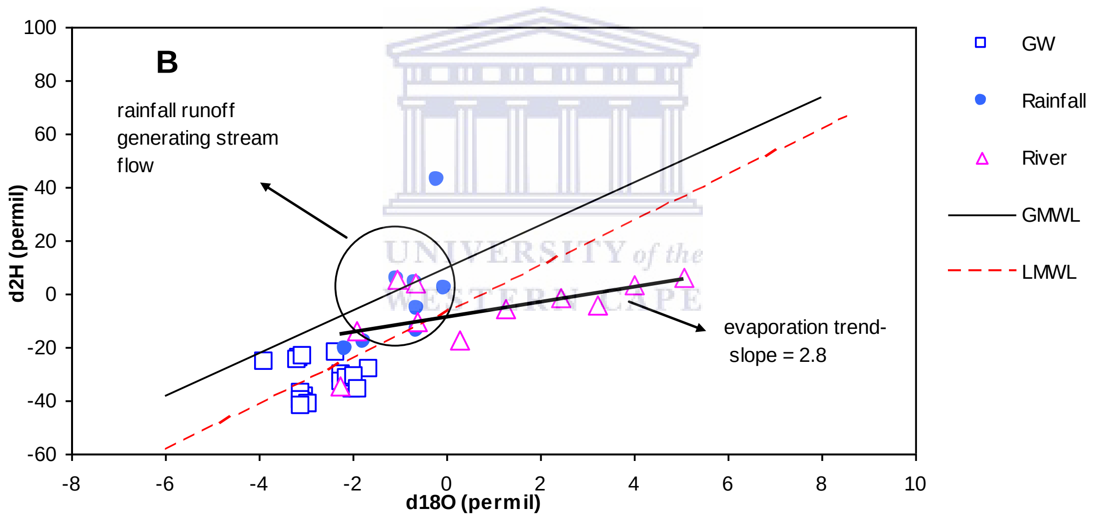
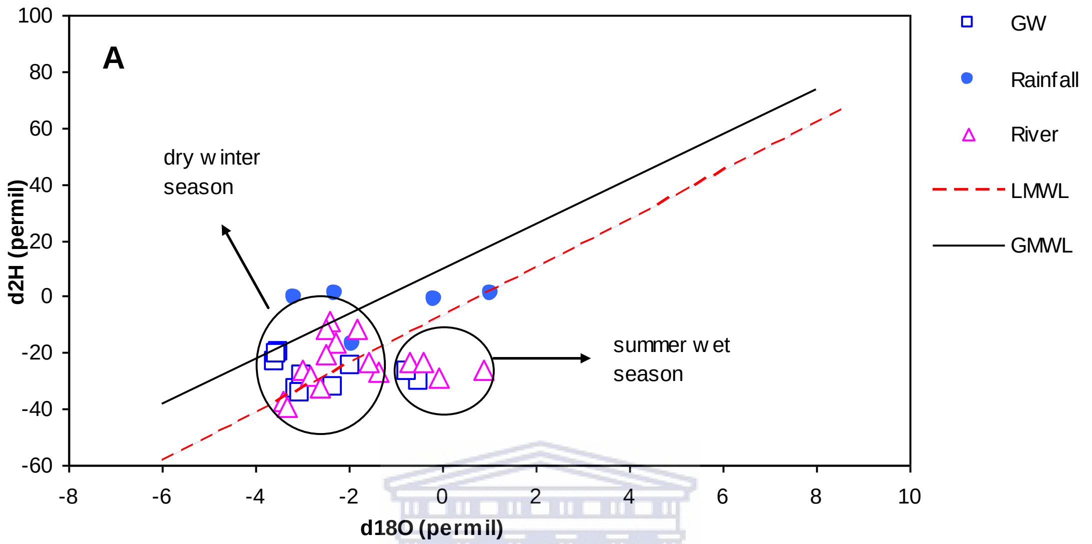


Figure 5.8 Shows a scatter plot of, (A) Bh-Sweni-hide, rainfall and Sweni River d2H (permil) versus d18O (permil) concentrations for all seasons. (B) Bh-Mufutsu, rainfall and Mnondozi River 2dH (permil) versus d18O (permil) concentrations for all seasons.

The isotopic composition in Figure 5.8 suggest that stream flow in the Mnondozi and Sweni River is generated by surface runoff as rainfall δ^2H and δ^{18}O values are similar to stream flow during the rainy season. However as the stream flow subside it causes standing pools of water along certain reaches of the rivers. These pools are progressively evaporated with a slope between 2 and 4, the distribution of the sample points are explained by the gradual evaporation of up to 100% of the total volume of water towards the end of the dry season. The water plotting closest to the LMWL is from the summer period. Those furthest away with highest isotope values are during the winter period.

Granites:



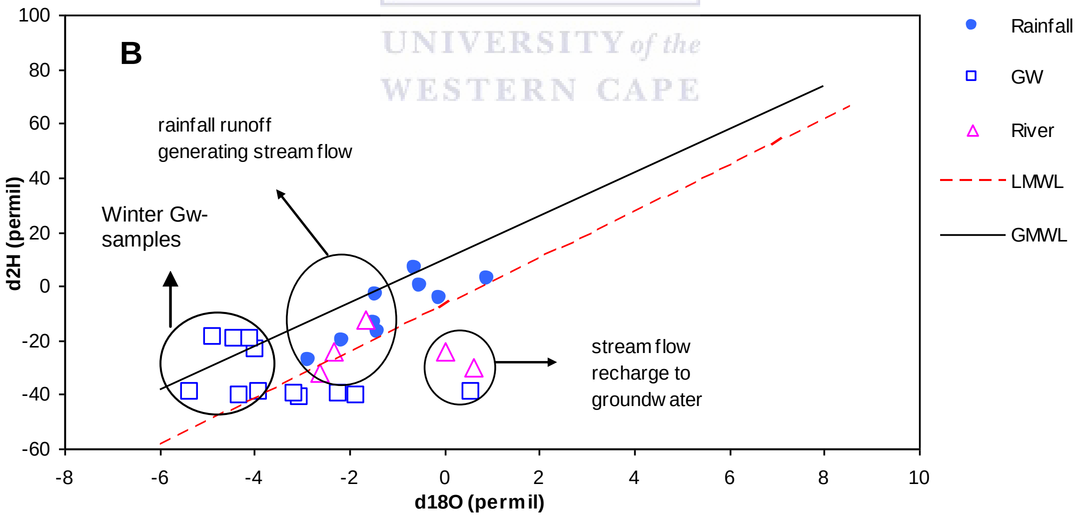


Figure 5.9 Shows a scatter plot of, (A) Bh-Jock, rainfall and Mbyamiti River 2dH (permil) versus d18O (permil) concentrations for all seasons. (B) Bh-Msanimond, rainfall and Nwatsisonto River d2H (permil) versus d18O (permil) concentrations for all seasons.

Figure 5.9 (A) shows at Bh- Jock that there are two distinct seasonal clusters of groundwater and surface-water, the mean isotopic values for surface-water in the Mbyamiti River during summer when the river experiences high flows are, δ^2H = -.31.69‰ and $\delta^{18}O$= -1.47‰; the groundwater values are, δ^2H= -32.32‰ and $\delta^{18}O$= -2.26‰. During the dry winter months when the river experiences low flow conditions the mean isotopic values for surface-water values are δ^2H= -20.5‰ and $\delta^{18}O$= -2.07‰, the groundwater values are, δ^2H= -24.0‰ and $\delta^{18}O$= -2.72‰. Referring to Figure 5.8 (B) at Bh- Msanimond during the wet summer season surface runoff from high rainfall events generates stream flow for 1 to 2 months in the ephemeral Nwaswitsontso River. The river water experiences evaporation as it plots to right of the LMWL. The groundwater exhibits the same evaporated signature shown by the clustering of the groundwater and river isotopic values during the rainy season.

5.4. Discussion

The isotopic data collected in this study can be used to characterise groundwater recharge. The isotopic composition of rainfall particularly cumulative amounts 50 mm to 100 mm or more that occur during the rainy summer months contribute to recharge. Since the evaporation rates in the KNP are high during the dry winter months and particularly during the hot summer months, the groundwater experiences evaporation prior to infiltration. This suggest that recharge to groundwater is delayed as evaporation most likely occurred from shallow soil profiles as a result piston flow and from surface-water bodies (ponding in local surface depression, streams and rivers) as indirect recharge.

The stream material of the seasonal Mbyamiti River and the ephemeral Nwaswitsontso River is underlain by alluvial deposits and weathered porous granite material (coarse river sand) and rock. As a result when the Mbyamiti River experiences peak flows during the wet summer period groundwater is recharged indirectly through the river bed. Conversely, during the dry winter months groundwater is discharged into the river maintaining low flows and pools. Therefore the Mbyamiti River is alternately influent and effluent (intermittent). The Nwaswitsontso River only flows after high rainfall events. As a result during the rainy season after significant rainfall events groundwater experiences indirect recharge through the river bed which acts as a preferred pathway (recharge sink).

Groundwater recharge via rivers and streams along the basalts are not a significant process as groundwater interactions with the Sweni and Mnondozi Rivers were not identified. The stream bed materials of these rivers are impervious with very low hydraulic conductivities due to the clayey soils commonly found on the basalts and rhyolite. Based on the observed $\delta^{2}H$ and $\delta^{18}O$ values of groundwater and stream water these rivers contribute very little to groundwater recharge. These rivers and streams are detached or remote from the groundwater system.

The perennial Sabie and Sand Rivers are connected to the groundwater system. During the dry winter periods when these rivers experience low flow conditions and throughout the wet summer season during high flow periods groundwater is consistently discharged into these rivers. Additionally, post 2000 flood during the last decade no significant recharge events took place. The groundwater levels are progressively declining and consequently discharging residual water into these rivers particularly maintaining baseflow during the dry winter months. The low hydraulic conductivity and transmissivity of the basement aquifers can explain the slow release over 10 years (Winter *et al*., 1998). Therefore, these rivers act as catchment drains.

Based on the results surface-water / groundwater interactions are important processes for recharge as the streams and rivers that flow through the study area provide important zones for recharge. Therefore the plausibility of the recharge processes as described in the analysis of the CRD model which suggests that infiltration to the water table via direct recharge (piston flow) along the basalts and indirect recharge via preferred flow paths (small depressions, streams and rivers) along the granites are confirmed.

CHAPTER 6

SYNTHESIS

6.1. Conceptual Model

A conceptual model consists of a set of assumptions that verbally and or visually describe the aquifer systems, based on field observations and data interpretation (Adams *et al*., 2004). In this study I developed a conceptual model of groundwater recharge processes in the KNP. The development of the conceptual model attempts to answer where, when and how groundwater recharge occurs.

Groundwater recharge in KNP occurs primarily during the rainy summer months (December to March) with very little to no recharge during the dry winter season. Recharge takes place during rainfall sequences of 100 mm or more. It is evident from the stable isotope records collected from rain, groundwater and surface-water that groundwater has experienced evaporation prior to infiltration (Figure 5.3). As the KNP experiences high evaporation rates (Venter *et al*., 2003), insignificant rainfall sequences contribute extremely little to groundwater recharge. The CRD analysis of groundwater level fluctuations shows (Figure 4.2) that recharge to the aquifers respond to dry and wet cycles that last for between 6 to 14 years.

During normal rainy seasons the groundwater levels rise somewhat, then start receding again. It is only during major rainfall events that are expected to occur every 100 to 200 years that the aquifers fully recharge. This was well illustrated with the high water levels after the 2000 (1: 200 year flood event) that recharged the aquifers fully. During below- average rainfall years the overall water level trends sharply decline (Figures 4.3 and 4.4) as the system experiences higher natural losses than gains due to outflow of groundwater to streams and rivers. This is particularly apparent in the perennial rivers where groundwater maintains baseflow.

Combination of the CRD model and the stable isotopic compositions used in this study suggest that the dominant recharge processes that occur in the southern region of the KNP are a combination of direct recharge via piston flow and indirect recharge via preferred pathways particularly streams and rivers. Figure 6.1 represents a simplified conceptual model of the dominant groundwater recharge processes and surface water groundwater interactions associated with the different geologic types in the southern region of the KNP.

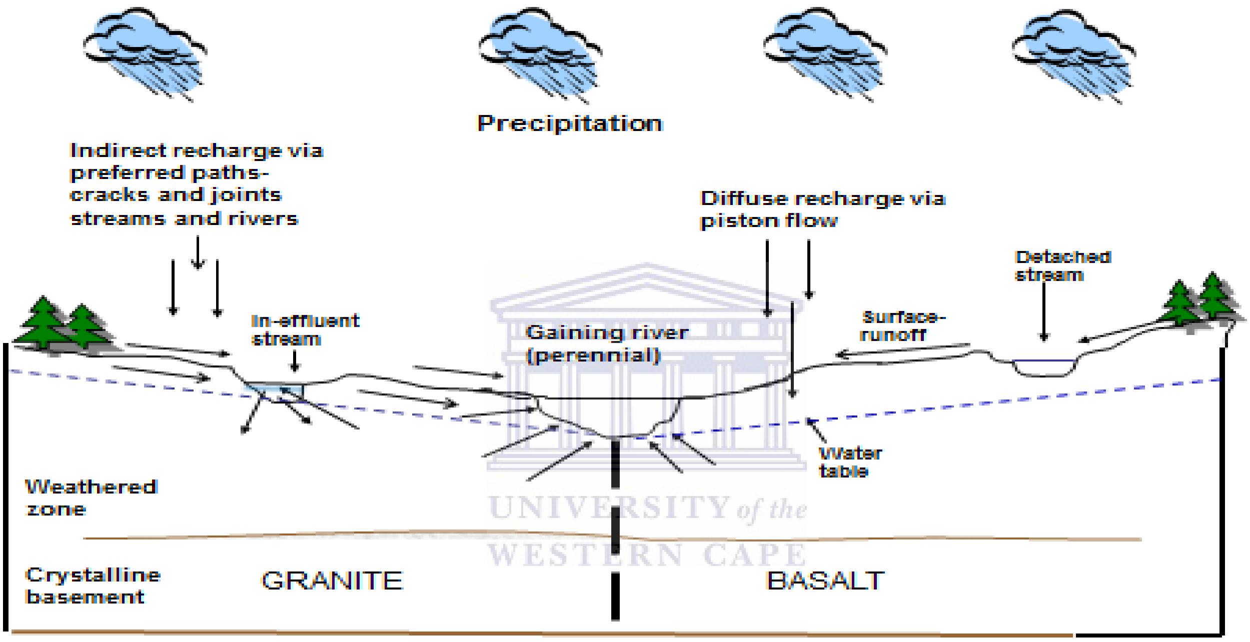


Figure 6.1 A conceptual model of the dominant recharge processes and surface water groundwater interactions associated with differing geology types in the southern region of the KNP.

Along the eastern KNP on the basalts and rhyolite, direct recharge via piston flow is dominant (Figure 6.1). Groundwater is not recharged via small streams and rivers such as the Sweni and Mnondozi Rivers as it was found that at these particular sites these rivers are detached and do not interact with groundwater. In contrast, along the western granitic areas the dominant recharge processes are indirect recharge (Figure 6.1). Recharge takes place via preferred pathways particularly streams and rivers. It was found that ephemeral rivers e.g. the Nwatsisonto River act as surface-water sinks for groundwater recharge and influent-effluent conditions are experienced along seasonal rivers e.g. the Mbyamiti River. The large perennial Sabie River and its tributary the Sand River are consistently fed by groundwater which maintains base flow during the dry season. These rivers act as basin outflows receiving groundwater discharge all year round.

It was found that the CRD model is a simple and a rapid method of predicting groundwater level fluctuations from rainfall. The method does not require all the environmental variables needed compared to physically based models. Groundwater level observations are required at each monitoring site over a reasonable period approximately 3 to 10 years. The method does incorporate cumulative preceding rainfall events contributing to recharge, making use of a short-term and long- term memory of the system. Further, the CRD model provided a good simulation of groundwater levels over the last 50 odd years and equally provided an estimate of the recharge efficiency and processes that was used to infer recharge processes that occur at the various study sites in the KNP. Use of the stable isotope composition of rainfall, surface-water and groundwater to act as a natural tracer in combination with the CRD method proved invaluable to confirm the plausible recharge processes.

6.2. Implications for management

Sustainable aquifer management can be described as the abstraction of less than or equal to that is replenished, with due consideration of the unavoidable losses (Kirchner, 2003). In the KNP there is increasing pressure on groundwater resources to accommodate the growing tourism demand (Biggs *et al*., 2003). Also, with the ever growing demand for water resources particularly outside the KNP and the potential impacts of climate change on future water resources there is a need for tools that allow water managers to make decisions on water, particularly the setting of TPCs for sustainable aquifer management.

The permeability and storativity in fractured aquifers varies with depth and it has been found that the water-bearing capacity of most fractured aquifers decrease considerably when the water level is lowered by more than 20 to 30 m (Kirchner, 2003). The aquifers of the KNP are predominantly basement fractured aquifers where water is mainly stored in joints, fractures and faults. Based on the groundwater levels fluctuations observed in the KNP the groundwater levels decline considerably during long periods of below-average rainfall. Abstraction of groundwater will cause an additional drawdown of the water table adding to the natural recession during these times in particular. It is therefore important for the management of KNP to set acceptable (TPCs) to strictly manage groundwater abstraction rates during extended (5 to 10 year) periods of below- average rainfall and between major rainfall events.

6.3. Recommendations

This conceptual model forms the foundation to developing acceptable TPCs of the groundwater levels in the KNP. The conceptual model can serve as a guide for the dominant recharge processes and for deciding on the location and time frames for data collection. The conceptual model can be revised and adjusted as additional data and analyses provide new insights into the groundwater system.

In order to inform sound management decisions, it is crucial to have a good understanding of the hydrological processes of the KNP. The main knowledge gaps on hydrological processes in KNP reside on understanding the dominating runoff generation processes, groundwater flow and recharge and groundwater / surface-water interactions. The following additional research and monitoring will assist in optimizing the conceptual groundwater model to ultimately set TPCs for groundwater in the KNP to sustainably manage the resource:

- Additional data should be acquired to optimize the conceptual model of the groundwater systems in the KNP using techniques evaluated by this baseline study

- Further isotope data collection of groundwater, surface-water and particularly rainfall is recommended to refine baseline values to establish long-term trends. In order to increase the efficiency, the rainfall collector design needs to be improved to guarantee that no evaporation takes place, and an expanded spatial distribution of collectors and long- term records is needed to develop a robust local meteoric water line (LMWL) for KNP.
- A broader study on the Sabie River catchment which is a highly diverse and biologically important perennial river using stable isotope and baseflow separation techniques will be useful to evaluate what contribution groundwater has on river flow.
- Vadose zone profiling and monitoring needs to be conducted to refine groundwater recharge processes and to evaluate runoff generation mechanisms such as, interflow and overland flow. (Hill slope processes).
- Measurements of chloride concentrations in groundwater and rainfall are recommended to try an attempt to quantify recharge rates using chloride mass balance method and the adapted CRD approach using rainfall and water level data to quantify recharge is further recommended.

CHAPTER 7

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APPENDICES

Appendix A

Borehole Name	X	Y	Geology formation	Lithologies	Topography	Soil Type	Vegetation
Sweni- Hide	31.97296	-24.47361	Jozini Formation	Acid/intermediate/alkaline extrusive rocks (rhyolite, felsite, quartporphyry)	Low mountain and hills	Rock outcrops and stony soils	Broad-leaved deciduous bushveld – no mopane
Msanimond	31.70972	-24.61379	Orpen Gneiss	Predominantly meta-arenaceous rocks (quartzite, gneiss, migmatite, granulite)	Slightly undulation plains	Sandy soils with duplex sodic clays	Broad and fined leaved bushveld Acacias
Rietpan	31.91421	-24.89687	Letaba Formation	Mafic/ultramafic extrusive rocks (basalt, andesite)	Flat plains	Red and dark clays	Fine-leaved trees – Acacia
Mafutsu	32.00947	-25.06430	Jozini Formation	Acid/intermediate/alkaline extrusive rocks (rhyolite, felsite, quartporphyry)	Low mountain and hills	Rock outcrops and stony soils	Broad-leaved deciduous bushveld – no mopane
Jock	31.53283	-25.29565	Nelspruit Granite Suite	Predominantly meta-arenaceous rocks (quartzite, gneiss, migmatite, granulite)	Slightly undulation plains	Sand soils with dupex sodic clays	Broad and fine leaved bushveld-acacias

Table 7.1 Provides a brief description of the environmental characteristics of the area of the selected borehole sites

Appendix B

Borehole name	Depth (mbgl)	WL - max-depth (mbgl)	WL-min-depth (mbgl)	WL-mean-depth (mbgl)	EC mS/cm	pH	Temperature °C
Sweni- hide	80	6.02	1.35	4.86	3.47	7.84	26
Msanimond	37	12.28	11.66	12.22	5.41	7.6	26.3
Rietpan	55	9.81	0.42	6.93	2.91	6.91	24.4
Mafutsu	50	7.26	0.71	6	1.295	7.34	25.9
Jock	25	6.30	2.85	5.02	1.377	7.78	24.6

Table 7.2 A summary of the water levels statistics of the boreholes used for the period 2001-2010. Also, the groundwater chemistry for the period the study was conducted (2010-2011).

Appendix C

Rainfall station	Period Years	MAR(mm)	MAR Min (mm)	MAR Max (mm)	Stdev
Nwanetsi	1967-2005	510	142	947	190.47
Tsokwane	1967-2005	545.7	100	1105	180.2
Lower Sabie	1991-2005	562.82	159.9	1044.4	230.4
Malalane	1939-2005	600	203	1353	191.2

Table 7.4 A table of the rainfall statistics used for the study.

Bh name	Rainfall station name	River name	Type of river	Distance rainfall station is from Bh (km)	Distance river sampling point is from Bh (km)
Sweni-hide	Nwanetsi	Sweni	Seasonal	25.84	0.05
Msanimond	Tshokwane	Nwaswitsontso	Ephemeral	23.64	0.24
Rietpan	Tshokwane			14.11	
Mafutsu	Lower Sabie	Mnondozi	Seasonal	11.38	3.53
Jock	Malalane	Mbyamiti	seasonal	19.48	9.36

Table 7.5 Illustrates the distance the rainfall stations and isotope sampling points for surface-water are relative to the boreholes.

Appendix D

	δ18 O	min	max	mean	Stdev	δD	min	max	mean	Stdev
GW		-5.3	0.58	-2.91	1.125		-41.95	-18.77	-31.83	15.4
Rainfall		-3.26	6.08	-0.60	1.74		-28.84	62.39	-4.38	12.55
Perennial rivers		-8.21	-1.19	-2.81	1.26		-34.02	-19.95	-27.79	12.9
Seasonal/ ephemeral rivers		-4.04	10.9	0.07	3.04		-42.91	37.28	-14.29	13.2

Table 7.6 A summary of the stable isotope statistics of all the samples collected over the period (n- 176).

Table 7.7 Stable isotope analysis for the groundwater samples.

Sample Name	Date	d18O (‰)	d2H (‰)
Jock- GW	27/05/2010	-3.58	-23.63
Jock- GW	12/07/2010	-3.50	-20.15
Jock- GW	16/08/2010	-3.53	-20.75
Jock Gw	18/01/2011	-3.12	-32.67
Jock Gw	21/02/2011	-3.06	-34.14
Jock Gw	22/03/2011	-0.52	-30.01
Jock Gw	29/04/2011	-2.35	-32.47
Jock Gw	28/05/2011	-3.01	-27.88
Jock Gw	24/06/2011	-0.76	-26.66
Jock Gw	22/07/2011	-1.95	-24.86
Rietpan-GW	27/05/2010	-4.13	-23.58
Rietpan-GW	17/08/2010	-3.71	-22.38
Rietpan-GW	14/07/2010	-3.96	-21.02
Rietpan-GW	7/07/2010	-4.06	-23.33
Rietpan GW	22/02/2011	-4.05	-39.37
Rietpan GW	24/03/2011	-2.45	-39.89
Rietpan GW	3/5/2011	-3.36	-39.96
Rietpan GW	25/05/2011	-4.50	-37.70
Rietpan GW	22/06/2011		-38.93
Rietpan GW	20/07/2011	-2.79	-39.57
Mafutsu-GW	28/05/2010	-2.76	-23.88
Mafutsu-GW	9/07/2010	-3.78	-26.12
Mafutsu-GW	14/07/2010	-4.12	-24.76
Mafutsu-GW	19/08/2010	-2.68	-25.95
Mafutsu GW	20/01/2011	-2.21	-34.91
Mafutsu GW	1/2/2011	-3.69	-34.71
Mafutsu GW	22/02/2011		-32.62

Mafutsu GW	24/03/2011	-1.88	-34.52
Mafutsu GW	3/5/2011	-3.00	-34.27
Mafutsu GW	25/05/2011	-1.75	-35.95
Mafutsu GW	22/06/2011	-0.64	-35.51
Mafutsu GW	20/07/2011	-0.80	-35.74
Msanimond-GW	27/05/2010	-3.98	-23.26
Msanimond-GW	5/07/2010	-4.88	-18.77
Msanimond-GW	13/07/2010	-4.42	-19.24
Msanimond-GW	18/08/2010	-4.10	-19.25
Msanimond GW	19/01/2011	0.58	-39.19
Msanimond GW	28/01/2011	-3.04	-41.07
Msanimond GW	23/02/2011	-3.17	-39.74
Msanimond GW	23/03/2011	-4.30	-40.50
Msanimond GW	28/04/2011	-1.86	-40.41
Msanimond GW	24/05/2011	-5.36	-38.97
Msanimond GW	23/06/2011	-2.23	-39.86
Msanimond GW	21/07/2011	-3.92	-39.07
Sweni hide GW	26/05/2010	-3.15	-23.83
Sweni hide GW	26/05/2010	-3.16	-24.69
Sweni hide GW	20/08/2010	-2.34	-22.07
Sweni hide GW	19/07/2010	-3.07	-23.69
Sweni hide GW	25/01/2011	-2.25	-30.31
Sweni hide GW	22/02/2011	-1.63	-28.48
Sweni hide GW	23/03/2011	-2.00	-36.10
Sweni hide GW	28/04/2011	-2.25	-33.02
Sweni hide GW	24/05/2011	-2.13	-31.94
Sweni hide GW	23/06/2011	-1.95	-31.22
Sweni hide GW	21/07/2011	-1.87	-35.89
Sweni hide GW	25/01/2011	-2.94	-41.10
Sweni hide GW	23/02/2011	-3.03	-38.60
Sweni hide GW	23/03/2011	-3.10	-37.52
Sweni hide GW	28/04/2011	-3.06	-40.58
Sweni hide GW	24/05/2011	-3.08	-40.31
Sweni hide GW	23/06/2011	-2.92	-41.44
Sweni hide GW	21/07/2011	-3.11	-41.95
Sweni hide GW	20/08/2010	-3.89	-25.43

Table 7.8 Stable isotope analysis of the rainfall samples.

NAME	DATE	d18O (‰)	d2H (‰)
Jock- Rain	29/06/2010	-3.16	-0.68
Jock-Rain	12/07/2010	-0.18	-1.28
Jock Rain	18/01/2011	2.25	-15.50
Jock Rain	21/02/2011	-1.91	-16.83
Jock Rain	29/04/2011	1.03	0.68
Rietpan-Rain	7/07/2010	-0.04	5.55
Rietpan-Rain	17/08/2010	-0.87	3.28
Rietpan rain	31/01/2011	-3.26	-28.84
Rietpan Rain	22/02/2011	-1.23	-5.09
Rietpan Rain	24/03/2011	-3.00	-25.66
Rietpan Rain	22/06/2011	1.66	62.39
Rietpan Rain	20/07/2011	-0.05	6.78
Mafutsu- Rain	6/07/2010	-0.87	-0.73
Mafutsu- Rain	14/07/2010	0.71	5.21
Mafutsu-rain	19/08/2010	-0.77	-25.12
Mafutsu Rain	1/2/2011	-2.00	-27.90
Mafutsu Rain	22/02/2011	-0.16	-0.59
Mafutsu Rain	24/03/2011	-1.19	-16.87
Mafutsu Rain	3/5/2011	-2.44	-25.07
Mafutsu Rain	25/05/2011	1.14	-12.12
Mafutsu Rain	22/06/2011	1.04	4.43
Mafutsu Rain	20/07/2011	1.74	4.61
Msanimond-Rain	30/06/2010	-1.47	-13.37
Msanimond-Rain	5/07/2010	-0.61	6.31
Msanimond-Rain	18/08/2010	-0.50	-0.04
Msanimond Rain	19/01/2011	0.90	2.84
Msanimond Rain	28/01/2011	-1.42	-16.87
Msanimond Rain	23/02/2011	-0.11	-4.25
Msanimond Rain	23/03/2011	-2.86	-27.03
Msanimond Rain	28/04/2011	-2.13	-20.02
Msanimond Rain	24/05/2011	-1.43	-3.21
Sweni-Rain	19/07/2010	-0.69	4.47
Sweni- Rain	20/08/2010	-1.06	5.60
Sweni Rain	25/01/2011	6.08	7.76
Sweni Rain	22/02/2011	-0.62	-5.51
Sweni Rain	23/03/2011	-2.15	-20.98
Sweni Rain	28/04/2011	-1.77	-17.67
Sweni Rain	24/05/2011	-0.62	-13.64
Sweni Rain	23/06/2011	-0.03	2.21
Skukuza Rain	19/04/2010	-0.92	5.39
Skukuza Rain	19/04/2010	-0.92	6.70
Skukuza Rain	7/04/2010	-3.90	-35.92
Skukuza Rain	20/04/2010	-0.94	-17.39
Skukuza Rain	7/05/2010	-1.84	-9.69
Skukuza Rain	30/05/2010	-1.05	3.13
Skukuza Rain	2/07/2010	-0.96	14.46
Skukuza Rain	2/07/2010	0.57	4.22
Skukuza Rain	28/10/2010	-0.35	-14.03

Skukuza Rain	31/10/2010	3.38	2.33
Skukuza Rain	17/11/2010	-0.25	-11.02
Skukuza Rain	19/11/2010	-0.87	-10.79
Skukuza Rain	24/11/2010	-3.74	-32.73
Skukuza Rain	15/01/2011	-2.37	-30.30
Skukuza Rain	23/01/2011	-3.52	-38.85
Skukuza Rain	15/02/2011	-0.43	-6.33
Skukuza Rain	15/02/2011	-2.94	-5.62
Skukuza Rain	27/01/2011	-2.46	-24.41
Skukuza Rain	28/01/2011	-3.75	-43.22
Skukuza Rain	23/02/2011	0.18	-1.44
Skukuza Rain	27/03/2011	-0.76	-19.64
Skukuza Rain	30/03/2011	-1.56	-10.93
Skukuza Rain	3/4/2011	-2.15	-8.32

Table 7.9 stable isotope analysis of the surface-water samples.

NAME	DATE	d18O (‰)	d2H (‰)
Biyamiti-River	27/05/2010	-2.42	-8.97
Biyamiti-River	29/06/2010	-2.49	-11.42
Biyamiti- River	12/07/2010	-2.82	-27.86
Biyamiti-River	12/07/2010	-2.30	-16.22
Biyamiti-River	12/07/2010	-1.85	-11.58
Biyamiti- River	16/08/2010	-2.51	-20.59
Biyamiti- River	18/01/2011	-3.42	-37.17
Biyamiti- River	18/01/2011	-3.34	-39.44
Biyamiti- River	21/02/2011	-1.37	-27.04
Biyamiti- River	22/03/2011	-0.10	-28.59
Biyamiti- River	29/04/2011	0.87	-26.23
Biyamiti- River	26/05/2011	-0.69	-23.43
Biyamiti- River	24/06/2011	-0.42	-23.46
Biyamiti-River	12/07/2011	-2.64	-32.23
Biyamiti- River	22/07/2011	-1.58	-23.55
Mnondozi- River	27/05/2010	0.33	-3.06
Mnondozi- River	14/07/2010	1.15	-5.21
Mnondozi- River	6/07/2010	0.49	-1.94
Mnondozi- River	19/08/2010	1.98	2.11
Mnondozi- River	20/01/2011	-4.04	-42.91
Mnondozi- River	1/2/2011	-0.79	-18.51
Mnondozi- River	1/2/2011	-0.44	-20.54
Mnondozi- River	22/02/2011	1.60	-7.68
Mnondozi- River	24/03/2011	4.13	-4.73
Mnondozi- River	3/5/2011	3.97	-2.49
Mnondozi- River	25/05/2011	7.89	11.79
Mnondozi- River	22/06/2011	10.94	37.28
Nwaswitsontso	19/01/2011	-2.62	-31.64
Nwaswitsontso	19/01/2011	-2.34	-23.76
Nwaswitsontso	28/01/2011	-1.66	-12.10
Nwaswitsontso	23/02/2011	0.02	-24.28
Nwaswitsontso	23/03/2011	0.60	-29.92

Sweni-River	26/05/2010	-1.93	-13.49
Sweni-River	8/07/2010	-0.62	-10.69
Sweni- River	13/07/2010	1.24	-5.86
Sweni-River	20/08/2010	2.43	-1.40
Sweni-Rain	19/07/2010	-0.69	4.47
Sweni- Rain	20/08/2010	-1.06	5.60
Sweni River	25/01/2011	0.26	-17.31
Sweni River	23/03/2011	-2.29	-34.30
Sweni River	28/04/2011	3.23	-4.12
Sweni River	24/05/2011	4.01	3.41
Sweni River	23/06/2011	5.08	6.11
Sabie River	5/07/2010	-3.17	-26.70
Sabie River	6/07/2010	-2.88	-19.95
Sabie River	20/08/2010	-3.41	-20.64
Sabie River	19/08/2010	-3.21	-21.80
Sabie River	26/01/2011	-2.76	-32.56
Sabie River	23/02/2011	-2.66	-28.48
Sabie River	23/03/2011	-2.61	-31.05
Sabie River	24/05/2011	-2.69	-30.07
Sabie River	23/06/2011	-2.56	-26.92
Sabie River	21/07/2011	-2.70	-32.63
SandRiver	5/07/2010	-3.16	-23.51
SandRiver	20/08/2010	-2.54	-27.30
Sand	23/02/2011	-2.58	-29.52
Sand	23/03/2011	-2.83	-34.02
Sand	24/04/2011	-2.73	-30.04
Sand	24/05/2011	-2.39	-27.89
Sand	23/06/2011	-2.73	-31.15
Sand	21/07/2011	-2.50	-27.89