RECHARGE RATES AND PROCESSES IN THE UPPER CROCODILE CATCHMENT

by:

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Submitted in partial fulfilment of the requirements

for the degree, of

Master of Science, Hydrogeology

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DECLARATION

I declare that: ***RECHARGE RATES AND PROCESSES IN THE UPPER CROCODILE CATCHMENT*** is my on work, that it has not 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 acknowledged by complete references.

Full name: Silindile Noluthando Zondi

Date: May 2017

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ABSTRACT

A study on groundwater recharge and processes controlling recharge was conducted in the Upper Crocodile catchment, located in the Johannesburg region. The catchment extends from the water divide south of Johannesburg, to the Hartbeespoort Dam in the North-West Province. The study area is predominantly underlain by the crystalline basement and meta-sedimentary rocks. The Upper Crocodile catchment is classified as a semi-arid region, receiving a mean annual rainfall of 699.3 mm/yr.

Groundwater recharge was quantitatively and qualitatively assessed using the water balance, baseflow separation, water table fluctuation and environmental isotope methods. The water balance and the baseflow separation methods resulted in recharge amounts of 4 and 5.8% of mean annual rainfall, respectively. The water table fluctuation method was only applied to the dolomitic aquifer and yielded a mean annual recharge estimate of 14% of the mean annual rainfall. Application of the isotopic shift method, which makes use of isotopically enriched water samples, resulted in a recharge amount of 10.19 to 23.90 mm/month obtained for the quartzites of the Witwatersrand Supergroup, south of the study area. Tritium was used to determine the residence time of stream water samples, collected during winter to represent baseflow. Additionally, it was used to understand the range of groundwater contribution to streams. The tritium values revealed that there are three types of water; i) relatively old water with lower tritium values, ii) intermediate tritium values indicating the possibility of mixing of older groundwater with more recent recharge and iii) high tritium values suggesting contamination from a local source/recent rainwater.

The results of groundwater recharge from the quantitative methods showed a temporal and spatial variability of recharge; this was attributed to the different processes that govern groundwater recharge. Climate appeared to have the most influence on potential groundwater recharge, with rainfall controlling the temporal variability of recharge while land cover, soil characteristics and geology influenced the spatial distribution of groundwater recharge.

Approximately 153 x 10^6 m^3/yr of wastewater was discharged into streamflow from wastewater treatment works as of 2008. The wastewater flow into streams overshadowed the baseflow contribution. The consequence of the presence of wastewater was reflected in the overestimation of groundwater recharge.

ACKNOWLEDGEMENTS

First and foremost, thank you to my supervisor Tamiru Abiye, for his patience, kindness and his continued support and encouragement. The knowledge he has imparted has been very beneficial. I am also grateful for all the isotope analyses he ran for me.

A big thank you goes to the GRECHLIM/PEER project for the financial support to complete this research.

My sincere gratitude to Aqeelah Davis, Khahliso Leketa and Khuliso Masindi for the motivation, the help and the knowledge they have shared with me.

Thank you to Mr Mike Butler for the efficiency of tritium analysis for my water samples.

I am also grateful to the Department of Water and Sanitation, Johannesburg Water, Rand Water and the South African Weather Services for the data they provided which enabled me to move forward with my thesis.

Finally, a big thank you to my friends and family for their continued support and motivation.

ABBREVIATIONS

$\delta^{18}O$	Oxygen – 18 isotopic composition
δD	Deuterium isotopic composition
AET	Actual evapotranspiration
BFS	Baseflow separation
BH	Boreholes
DWAF	Department of Water Affairs and Forestry
DACE	Department of Agriculture, Conservation and Environment
DARD	Department of Agriculture and Rural Development
LMWL	Local meteoric water line
MAR	Mean annual rainfall
mcm	million cubic meter
PET	Potential evapotranspiration
TU	Tritium Units
WB	Water balance
WTF	Water table fluctuation
WWTW	Wastewater treatment works

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1 INTRODUCTION

1.1 BACKGROUND

The quantification of groundwater recharge is something that is often overlooked especially in areas where fresh surface water is used as the main water supply. Because of the ease in which surface water can be obtained, groundwater is seldom prioritised. This is especially true in the case of groundwater abstractions where sustainable yields are often not enforced because of a lack of understanding of the importance of groundwater recharge rates, particularly by policy makers and those responsible for water management.

To realistically estimate groundwater recharge, one must have an understanding of the processes that govern recharge, the most common factors controlling recharge are climate, geology, soil characteristics and topography. These are responsible for the spatio-temporal variability of recharge (Lerner et al. 1990) in a catchment area.

The Johannesburg region, located in the Gauteng province, can be classified as having a semi-arid climate (Abiye 2011; Abiye et al. 2011), it is characterised by hot, warm summers and cold, dry winters (DWAF 2004; DARD 2011). Rainfall dominates during the summer but is often short lived and is defined by frontal rainfall (DWAF 2004). Recharge in semi-arid regions occurs episodically, most often during heavy rainfalls, in such regions recharge estimation studies can be challenging because mean annual evaporation tends to exceed the mean annual rainfall. Therefore, groundwater recharge is significantly less than rainfall (Beekman & Xu 2003; DWAF 2004; van Wyk 2010).

Part of the Upper Crocodile catchment is located within the Johannesburg region which is regarded as highly urbanised and as a result, has a continuously growing population and growing industries. Agriculture along with the increasing number of industries, mines and the population are heavily reliant on the fresh surface water thereby placing the available water resources under a great amount of stress. The reactionary results of increased population and industry are the pollution of the water source caused by the discharge of wastewater effluent, industrial waste, acid mine drainage (AMD), etc. The deterioration of the quality of surface water increases pressure on groundwater resources either through quality alteration or over extraction.

It must also be noted that groundwater forms an important source of water for the rural communities who don't have access to the municipal water (Dutt Tewari 2012; Abiye et al. 2015). Downstream commercial farmers are also highly dependent on groundwater from the Upper Crocodile catchment for irrigation purposes (DWAF 2008; Dutt Tewari 2012). The importance of groundwater is especially recognised during drought periods when surface water usually dries up because of a lack of rainfall and high evaporation and evapotranspiration (ET) rates. Groundwater is less susceptible to evaporation and ET and therefore makes it a more reliable water source. The above illustrates the importance of groundwater and subsequently the sustainable management of the water resource. The estimation of recharge rates can allow a sustainable yield to be determined which will ensure the longevity of the groundwater resource.

1.2 SOCIO – ECONOMY OF THE STUDY AREA

As mentioned above, the Upper Crocodile catchment is a highly urbanised area boasting developed industries and a successful mining sector which therefore attracts a large population to the urban areas. In 2000 the Upper Crocodile catchment had an urban population of approximately 2 million people and a rural population of about 200 thousand people (DWAF 2004). The rural population relies mostly on groundwater for domestic use and small-scale farming, on the other hand, the urban population's water demand is met by water transfers from the Vaal catchment (DWAF 2004). Rand water supplies water to areas in the Upper Crocodile catchment namely; the cities of Johannesburg, Centurion, Midrand and Pretoria to meet urban water requirements. Approximately 370 x 10^6 m^3/yr of water is transferred from the Vaal Dam, in the Upper Vaal catchment via the Crocodile River into the Upper Crocodile catchment for domestic, industrial and mining use. Nearly 75% of this water is returned into the catchment as wastewater. In addition to that, there are eight wastewater treatment plants located throughout the study area (Figure 1) that discharge a combined volume of approximately 274 x 10^6 m^3/yr to nearby rivers, contributing a significant amount to streamflow. The Rietvlei and the Hartbeespoort Dam are the unfortunate recipients of the wastewater discharge (DWAF 2004).

Rainfall is the primary recharge source, but when estimating recharge rates for the study area, one must be aware that rainfall is not the only source of recharge.

Thus, the water transfers from the Upper Vaal catchment and the effluent discharge must be taken into consideration when calculating recharge to ensure a more accurate representation of the recharge rates for the study area.

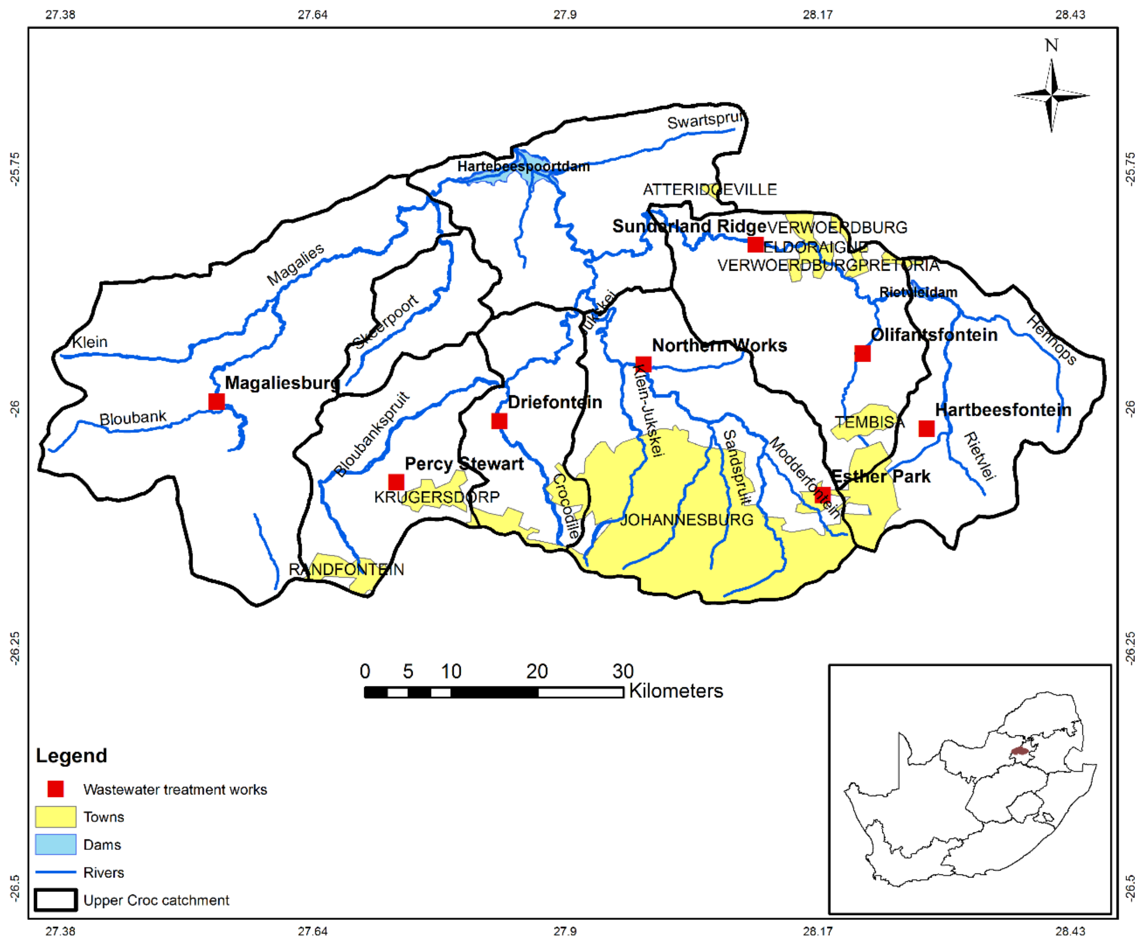

Figure 1: Location of the wastewater treatment plants in the Upper Crocodile catchment.

1.3 RESEARCH QUESTIONS

Are the results obtained from multiple groundwater recharge methods comparable? If not can the differences be explained?

Can the contribution of wastewater to groundwater recharge be quantified and what is the consequence of the presence of wastewater on groundwater recharge?

Which groundwater recharge processes must be considered for recharge rate studies and how do they relate to groundwater quantification?

Do the different aquifer types have any bearing on the spatial variability of recharge?

1.4 RESEARCH HYPOTHESIS

It is possible to obtain a reliable estimate of groundwater recharge by using several recharge estimation methods i.e. water balance, baseflow separation, water table fluctuation methods and environmental isotopes. Knowing more about recharge processes such as climate, geology, land cover and topography is fundamental in understanding the variability of groundwater recharge. The large volume of wastewater that is continuously released into streams will have an influence on groundwater recharge resulting in inflated recharge estimates.

1.5 AIMS AND OBJECTIVES

The study aims to employ different methods to estimate groundwater recharge rates and to identify and understand the primary recharge processes in the crystalline, karstic and meta-sedimentary aquifers of the Upper Crocodile catchment.

The primary objectives are:

- To apply the Water table fluctuation (WTF) method, baseflow separation (BFS) method, water balance method and environmental isotopes to estimate groundwater recharge in the Upper Crocodile catchment.
- To estimate the amount of recharge contributed by rainfall and any additional input sources such as effluent discharge.
- To quantify the spatial and temporal variations of recharge by estimating recharge rates for the different major aquifer types: Fractured crystalline aquifer, karstic aquifer and fractured meta-sedimentary aquifer.
- To identify and understand the processes that control groundwater recharge.

2 STUDY AREA

2.1 LOCATION

The Upper Crocodile catchment is located within the Gauteng and North-West Provinces and extends from the water divide in the south to Hartbeespoort Dam in the north (Figure 2). Geographically it can be found between the coordinates-25.678; 27.341 and -25.678; 28.472 degrees, covering an area of 4107 km^2. Hydrologically the study area is a part of the Crocodile West and Marico water management area (WMA), the study area falls within the A21 catchment which includes quaternary catchments A21A to A21H.

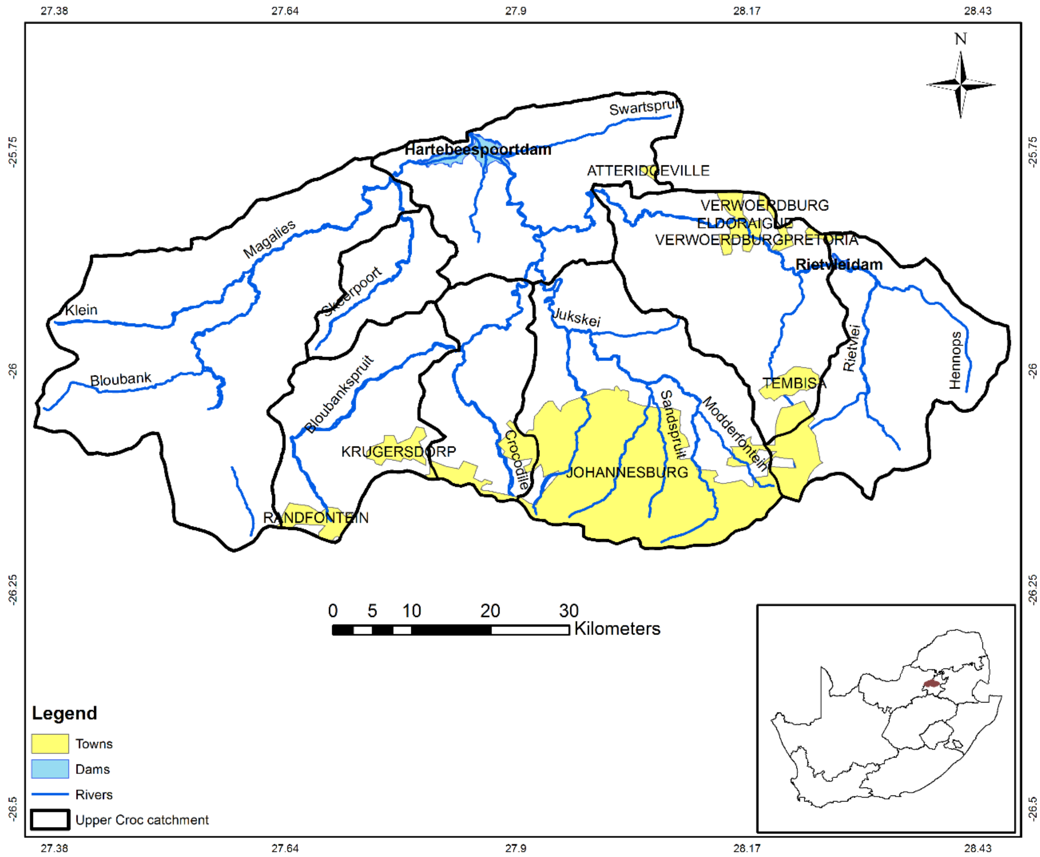

Figure 2: Study area of the Upper Crocodile catchment.

The elevation of the catchment decreases from south to north, with upstream areas having an elevation of approximately 1700 m.a.s.l. and the downstream areas with an elevation of approximately 1200 m.a.s.l. The study area is mostly situated in Johannesburg, a city that is highly urbanised and populated by many industries.

2.1.1 DRAINAGE

The catchment area is drained by the Crocodile River which flows from the south-west to the north, draining into Hartbeespoort Dam. The major tributaries feeding the Crocodile River in the catchment include the Hennops River in the east, Jukskei River, Rietspruit River and the Magaliesburg River in the west (Figure 2). All the main rivers in the catchment are perennial; they are sustained by a combination of runoff, baseflow and wastewater discharge.

Topographically the catchment is characterised by quartzite ridges in the northern and southern parts of the catchment, and the middle of the catchment is relatively low lying (land) dominated by dolomites and shales. The general topography dips towards the north hence surface drainage is from south to north (Figure 3).

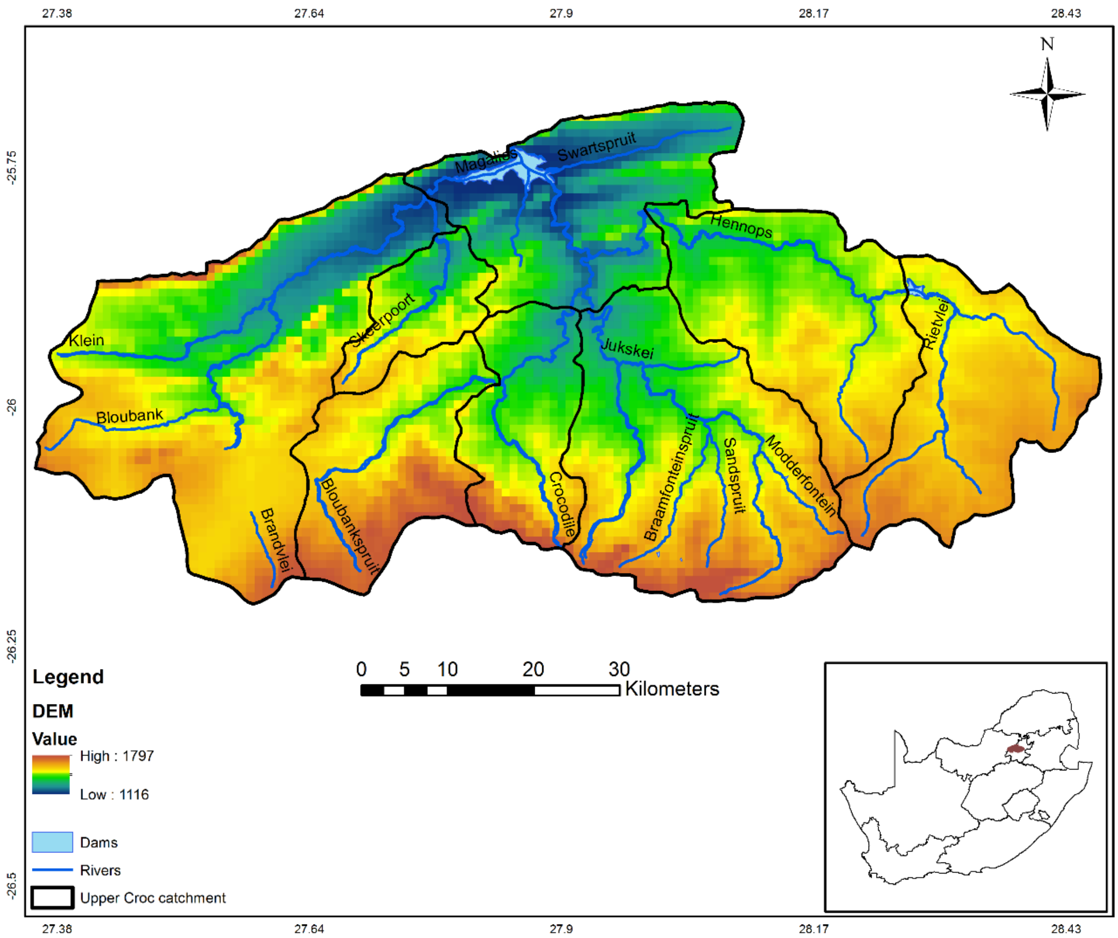

Figure 3: Digital elevation model of the Upper Crocodile catchment.

Significant surface water bodies within the study area include; the Rietvlei Dam, located in quaternary catchment A21A and Hartbeespoort Dam, located in quaternary catchment A21H. Hartbeespoort Dam is an important Dam as it is used as a water source for the downstream community (Figure 2).

2.2 CLIMATE

The Upper Crocodile catchment is classified as having a semi-arid environment. It is dominated by two seasons winter and summer, summer occurs between October and March and coincides with the wet season and winter occurs between April and September coinciding with the dry season (DWAF 2004; DACE 2004) (Figure 4).

The highest rainfall occurs during the summer months of October to March while winter months, April to September receive limited rainfall the reason for this is that the climate (rainfall) is mostly controlled by the Southern Hemisphere climate systems. In summer the intertropical convergence zone (ITCZ) migrates to its southern most point where it generates zones of low-pressure cells which are associated with frontal rainfall and localised thunderstorms (DWAF 2004; DARD 2011). In the winter months, the air circulation is dominated by the Kalahari High Pressure (HP) cell of the Subtropical HP system which brings with it cold dry air from the polar regions (DACE 2004, Abiye 2016). Figure 4 shows the mean monthly distribution of rainfall, peak rainfall occurs in January, and the lowest amount of rainfall falls in July. The mean annual rainfall for the years 1969 – 2014 is 699.3 mm.

According to the Köppen Classification the study area has a continental climate characterised by hot summer temperatures and dry winter temperatures. During summer, January has the hottest temperatures and July has the coldest temperatures (Figure 4). The mean annual temperature is recorded as 19.2 ^{o}C. Figure 4 shows a comparison of temperature, rainfall and PET; PET follows the distribution pattern of both temperature and rainfall with high PET values in summer and low PET values in winter. Often though PET exceeds rainfall, a rather common occurrence in semi-arid regions (Beekman and Xu 2003; DWAF 2004).

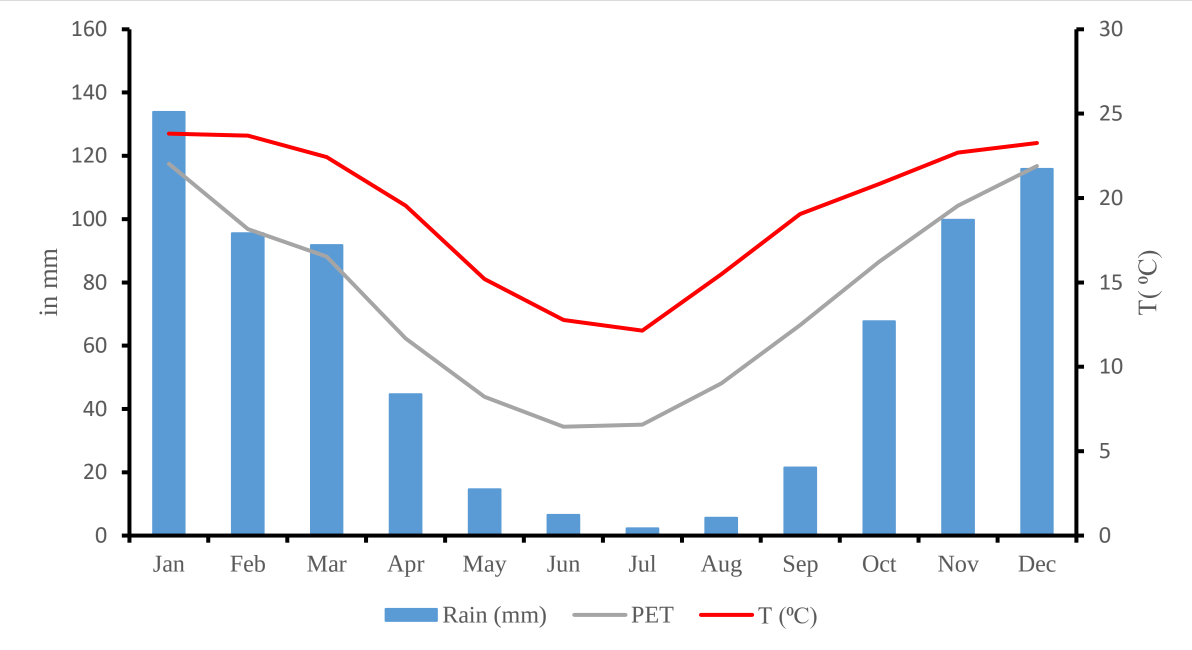

Figure 4: Mean monthly rainfall, PET and temperature.

2.3 GEOLOGY OVERVIEW

The catchment area is underlain by rocks that range from the Archaean to the Palaeo-Proterozoic era. These include Archaean crystalline basement rocks which are a combination of granitic, gneissic and granodiorite rocks these rocks have been eroded, weathered and tectonically altered (McCarthy & Rubidge 2005; Abiye 2011). The basement rocks are overlain by the Witwatersrand Supergroup, characterised by arenaceous and argillaceous sedimentary rocks which can be separated into two groups, the West Rand and the Central Rand Group (McCarthy & Rubidge 2005).

The Witwatersrand basin conformably overlies the sedimentary and volcanic rocks of the Dominion Group and unconformably overlies the basement rocks, where the Dominion Group is absent (McCarthy & Rubidge 2005). The Transvaal Supergroup, the youngest in the study area, consists of varying rock types that include sedimentary rocks, volcanic rocks, chemical rocks (banded iron formation (BIF)) and carbonate rocks (limestones and dolomites) (McCarthy & Rubidge 2005). All these rock types form a dome shape that extends across the catchment (Figure 5). The study area, in the past, has undergone structural deformation and as a result is plagued with shear zones and lateral strike-slip faults which cut across the different lithological units (Abiye 2011; Abiye et al. 2011).

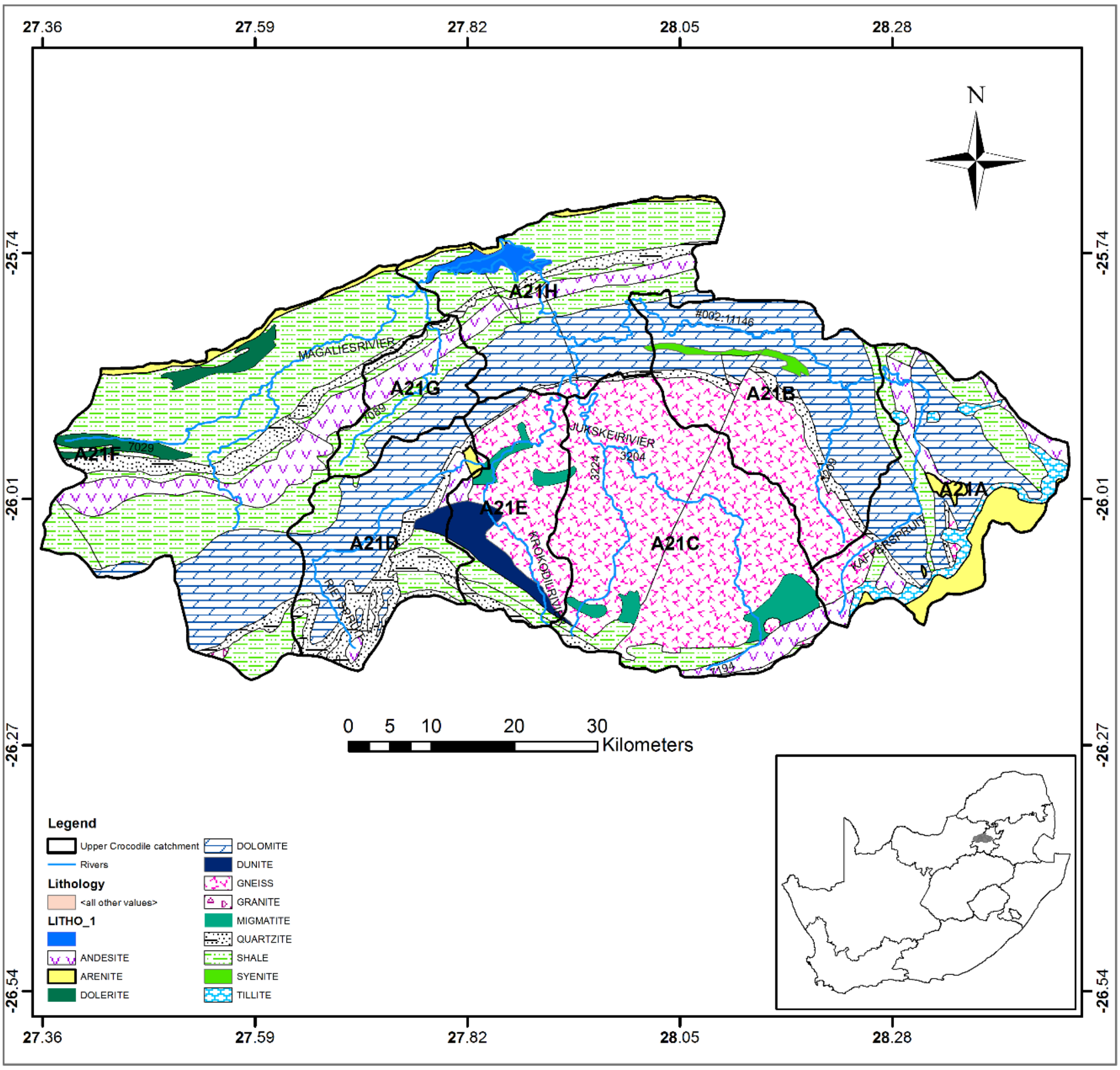

Figure 5: Surface geology of the Upper Crocodile catchment.

2.3.1 CRYSTALLINE BASEMENT

The crystalline basement comprises of the oldest rock types in the sequence, and they are all Archaean in age. The crystalline basement consists of greenstone remnants, the oldest rock type of the basement, gneissic, granitic and migmatite rock types, these rocks are unconformably overlain by the Witwatersrand Supergroup (McCarthy & Rubidge 2005; Abiye 2011). Geological maps (Figure 5) show a circular feature of the crystalline basement being surrounded by the rocks of the Witwatersrand, Ventersdorp and Transvaal Supergroups, this is famously known as the Johannesburg Dome.

2.3.2 WITWATERSRAND SUPERGROUP

The rocks of the Witwatersrand Supergroup were deposited during the Precambrian era, in an extensional geological setting. The Witwatersrand Supergroup is separated into two groups; the older West Rand Group and the younger Central Rand Group, these are further divided into five subgroups. The Central Rand Group consists of two subgroups namely the Johannesburg and the Turffontein Subgroups, which consist mostly of quartzites and shale. The West Rand Group is made up of three subgroups, the Hospital Hill, Government and Jeppestown Subgroups which are characterised by quartzites, conglomerates and shales. The subgroups are characterised by a variety of rock types such as interbedded quartzite and shale units, fluvial conglomerates, shales, quartzites, diamictites and andesitic lavas (McCarthy & Rubidge 2005). The shales and the quartzites of the West Rand and Central Rand Groups outcrop on the surface and can be seen south of the study area (Figure 5).

2.3.3 TRANSVAAL SUPERGROUP

The Transvaal Supergroup was first deposited approximately 2650 MA ago. Thermal subsidence of the basin due to rifting formed a shallow continental shelf, where the Transvaal Supergroup was deposited (McCarthy & Rubidge 2005). The Transvaal Supergroup can be divided into the Black Reef Formation, The Chuniespoort Group and the Pretoria Group (McCarthy & Rubidge 2005).

The Black Reef Formation was the first to be deposited, it conformably overlies the Ventersdorp Supergroup, and predominantly consists of quartzites and conglomerates. The Chuniespoort Group is categorised into three formations the Malmani Subgroup, the Penge and the Duitschland Formation. The Malmani Subgroup consists of various chert poor and chert-rich dolomite formations. The Penge Formation consists of the BIF and metamorphosed sedimentary rocks. The Duitschland Formation unconformably overlies the Penge Formation and is made up of carbonaceous rock types, diamictites and lavas.

The Pretoria Group is deposited unconformably on the Malmani Subgroup; it is subdivided into ten formations which consist predominantly of quartzites and shales with subordinate carbonate rock types, conglomerates diamictites and interbedded volcanic units (McCarthy & Rubidge 2005).

The Malmani dolomites and the Pretoria Group quartzites and shales extend north-east to south-west across the study area. The dolomites are characterised by a gently undulating topography (Abiye 2011; Abiye et al. 2011), while the Pretoria Group quartzites form ridges because of their high resistance to erosion and the shales form valleys because of the softer geological material which is more prone to erosion.

2.4 AQUIFER TYPES

Aquifers can be defined as geological units that are capable of transmitting substantial amounts of groundwater to wells and springs and are also capable of storing groundwater (Fetter 2001). Aquifers can be commonly described as being unconfined, confined or semi- confined (Fetter 2001). Confined aquifers are bounded by impermeable layers, at the top and bottom of the aquifer, referred to as aquicludes and are isolated from nearby aquifers. Unconfined aquifers, on the other hand, are overlain by permeable layers and are bounded by the water table. Semi-confined aquifers are also known as leaky aquifers and are confined by low permeable layers referred to as aquitards; these layers can allow for either recharge or discharge to occur.

The South African Department of Water Affairs has classified aquifers into four different classes: Intergranular (Class A), Fractured (Class B), Karst (Class C) and Intergranular and fractured (Class D) aquifers. Briefly, Class A aquifers are characterised by the unconsolidated and semi-consolidated material or material that has been exposed to weathering and has become partly consolidated; groundwater is transmitted through the intergranular spaces.

Class B aquifers are associated mostly with crystalline rocks and formations that have been subjected to lithification. In fractured aquifers, groundwater is transported by the fractures and flow occurs mainly in the weathered horizon.

Class C aquifers are characterised by carbonate rocks; they form as a result of rock dissolution caused by the interaction of carbonic acid (rainwater) with carbonate rocks. Groundwater in karst aquifers is transmitted by connected fractures, cavities and conduit systems present in the aquifer.

Lastly Class D, a combination of an intergranular and fractured aquifer. In this setting, the fractures act as a mode of transport for the groundwater while the matrix acts a storage facility. Class D aquifers are the most common in South Africa (Dutt Tewari 2012).

2.5 HYDROGEOLOGY

There are two types of aquifers in the study area, fractured aquifers which form shallow, low-yielding aquifers and karstic aquifers which are characteristic of deep, high-yielding aquifers (Figure 6). The latter are considered as a crucial source of water in South Africa due to their high yields (Dutt Tawari 2012).

The Malmani dolomites are classified as karst aquifers because they have a high storage capacity and are highly permeable as a result of karstification (Leskiewicz 1986). A process whereby weak carbonic acid, originating from rainwater dissolves carbonate rocks along zones of weakness such as joints, fractures, faults, etc. resulting in the formation of cavities, sinkholes and caves (Kafri et al. 1986; Leskiewicz 1986). Karstification is controlled by several geological factors such as lithology, stratigraphy and structural deformation (Kafri et al. 1986; Leskiewicz 1986). Karst aquifers in the study area are characterised by a gently undulating topography, extending across the catchment from east to west (Abiye 2011).

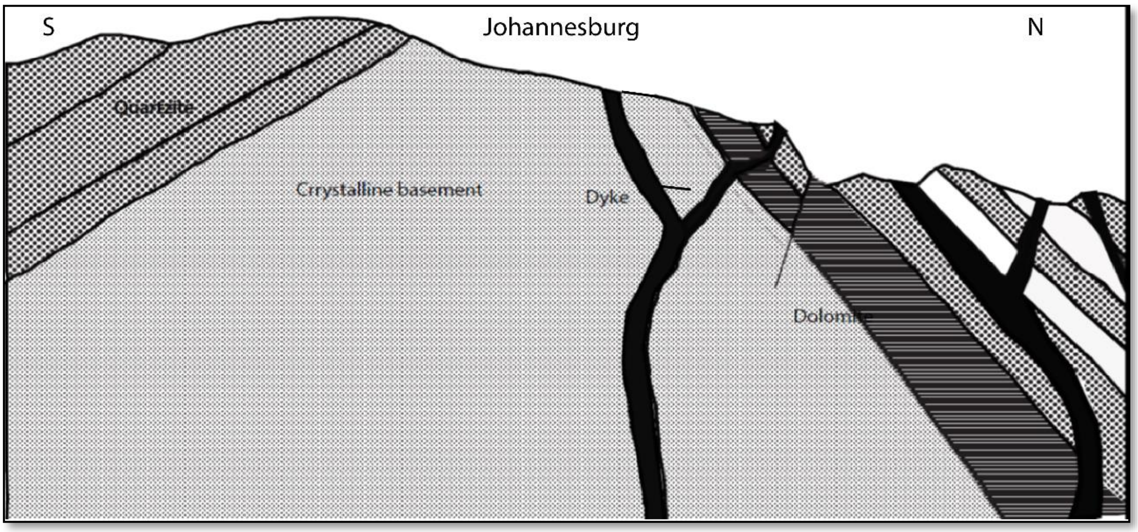

Figure 6: Hydrogeology cross section of the study area. Not to scale.

In karstic aquifers, groundwater occurrence is greatest where there is a vast network of connected cavities and conduits this is usually limited to a depth of 40 m.b.g.l (Abiye 2011), below this depth the aquifer is generally matrix dominated. Runoff in dolomitic terrains tend to be low and as a result, recharge is usually high. Chert-rich dolomite formations are generally more productive than the chert poor dolomites this is because of the soluble nature of the chert.

The chert-rich dolomites exhibit more fractures and joints along which groundwater can occur (Kuhn 1989; Leskiewicz 1986).

Other characteristic features of the karst aquifers in the catchment area are the compartmentalization of dolomites into isolated hydrogeological units as a consequence of dyke intrusions and structural deformation. Spring occurrences along geological contacts and zones of lineaments are common on the dolomitic terrain (Pietersen et al. 2011).

The Malmani dolomites are considered as moderate to highly productive aquifers with yields ranging from 15 l/s to 124 l/s (Abiye et al. 2011; Abiye et al. 2015), with the highest yields associated with the chert-rich dolomite formations (Kuhn 1989). The dolomites in the study area have a variable water table depth, with the water table being less than 1 m.b.g.l in some places to being as deep as 90 m.b.g.l.

Fractured aquifers can be found on the basement rocks and the quartzites and shales of the Witwatersrand Supergroup and the Pretoria Group. These fractured aquifers are commonly limited to the weathered horizon and the fractured media (eds. Wright & Burgess 1992; Abiye 2011).

Because these rock types have low primary and secondary porosity groundwater productivity is usually low. Spring occurrence is common especially between lithological contacts and along faults.

The quartzites of the Witwatersrand Supergroup form a ridge south of the catchment which also acts as a water divide between the Upper Vaal catchment and the Upper Crocodile catchment. The quartzites have been weathered and tectonically altered by a series of lateral strike-slip faults, shear zones and fractures. These structures are necessary for groundwater circulation as well as being important features that allow recharge to take place (eds. Wright & Burgess 1992).

The crystalline basement is made up of igneous rocks which virtually have no primary porosity. The amount of groundwater that occurs is controlled by the interconnectedness of fractures, joints and the distribution of fault zones as well as the thickness of the weathered profile (eds. Wright & Burgess 1992). Because the aquifers have low permeability runoff tends to be high in the vicinity of these aquifers, and subsequently, recharge will be low.

The fractured crystalline aquifers display low productivity with yields ranging from 0.01 to 0.98 l/s, and the fractured quartzitic aquifers have yields ranging between 1 and 14.6 l/s (Abiye et al. 2011; Abiye et al. 2015). The fractured aquifers are mostly used for domestic use, small scale localised farming and gardening.

The general direction of groundwater flow in the catchment area is from south to north; this is validated by the DEM (Figure 3) which shows that the area south of the catchment is at a higher elevation than the areas north of the catchment.

3 LITERATURE REVIEW

3.1 INTRODUCTION

South Africa is considered as a water poor country (Middleton & Bailey 2009), this applies to many other countries in the (semi) arid regions of Southern Africa this is mostly driven by the climatic conditions where annual evaporation greatly exceeds annual rainfall. Hence, it is important to have an understanding of groundwater recharge and the processes influencing recharge rates, to properly manage water resources. Quantifying recharge can be a challenging task because off all the complexities associated with it; these include the processes that govern recharge such as climate, geology, topography, soil characteristics and land cover (Gee & Hillel 1988). Most of these processes except rainfall are rarely included in recharge estimation methods making it difficult to determine the extent to which these processes affect groundwater recharge (Gee & Hillel 1988). Another problem with recharge estimation methods is choosing methods to quantify recharge because no one method has been identified that is capable of estimating recharge accurately, a number of methods should be used to obtain reliable recharge estimates (van Tonder & Xu 2001; Scanlon et al. 2002; Beekman & Xu 2003). Furthermore, all recharge estimation methods are associated with some degree of uncertainty therefore before using a particular method it is important to know the assumptions and the limitations of that method (Scanlon et al. 2002; Beekman & Xu 2003). Finally, it is hard to estimate recharge accurately if the time series data is incomplete. Complete data sets covering a long term period are uncommon in Southern Africa (Bredenkamp 1995; Adams et al. 2004; Sibanda et al. 2009; Lutz et al. 2015, Abiye 2016).

Over the past few decades, a lot of progress has been made in trying to understand groundwater recharge, especially in Southern Africa (Bredenkamp 1995; Beekman & Xu 2003). Some research has been conducted by Abiye (2011), Abiye et al. (2011), Abiye (2014), Abiye (2015) and Abiye (2016) in the Upper Crocodile catchment.

Most of this work investigated the groundwater – surface water interaction and the deterioration of groundwater and surface water quality as a result of mining with the exception of Abiye (2016) which is dedicated to recharge in Southern Africa. No detailed groundwater recharge studies have been conducted in the Upper Crocodile catchment. Hence this study aims to improve the knowledge and understanding of groundwater recharge by quantifying recharge and understanding the processes controlling it.

Recharge methods that have been selected for the study are; the water balance, baseflow separation and water table fluctuation method, these methods were chosen for their simplicity, data was easily obtainable, and they are cost effective. Environmental isotopes will also be used to assess recharge qualitatively. There are several methods which have been used in arid and semi-arid regions of Southern Africa to calculate recharge these will be explored below.

3.2 RECHARGE RATES IN ARID AND SEMI-ARID REGIONS

Numerous methods exist for analysing and estimating recharge, the use of these methods is dependent on the available or obtainable data, the areal extent of the study area and the timescale to which the methods will be applied (Hendrickx & Walker 1997; Scanlon et al. 2002; Adams et al. 2004). A detailed study on choosing the appropriate methods for groundwater recharge outlining the attributes of the different methods, the reliability of their results and their applicability in space and time has been conducted by Bredenkamp et al. (1995) and Scanlon et al. (2002).

Recharge methods can be grouped according to hydrologic zones namely surface water, unsaturated zone and saturated zone methods. Within these hydrologic zones, methods can further be classified as physical or chemical (tracer) methods (Table 1) (Bredenkamp et al. 1995; Scanlon et al. 2002; Beekman & Xu 2003; Healy 2010). Surface water and unsaturated zone methods calculate the potential recharge as there is no assurance that all the water infiltrated make it to the water table, water may be lost to processes taking place in the unsaturated zone (Scanlon et al. 2002; de Vries & Simmers 2002; Beekman & Xu 2003). Saturated zone methods calculate the actual recharge, defined as surface water that reaches the water table thus contributing to groundwater storage (Scanlon et al. 2002; de Vries & Simmers 2002; Beekman & Xu 2003).

Beekman & Xu (2003) came up with a list of promising recharge estimation methods for arid and semi-arid environments in Southern Africa the list includes; the chloride mass balance (CMB), cumulative rainfall departure (CRD), water table fluctuation (WTF), saturated volume fluctuation (SVF), groundwater modelling (GM) and the extended model for aquifer recharge and moisture transport through unsaturated hard rock (EARTH) methods. For a summary of all the methods applied in the arid and semi-arid regions in Southern Africa refer to Table 1.

3.3 CASE STUDIES

Many studies on the quantification of groundwater recharge have been carried out in the arid and semi-arid regions of Southern Africa over the decades. Bredenkamp et al. (1995), Beekman & Xu (2003) and Abiye (2016) have compiled a number of case studies conducted in Southern Africa giving a comparison of methods used in estimating recharge and their results. Below is a summary of case studies for recharge estimation in fractured and karst aquifers in the arid and semi-arid regions of Southern Africa.

Table 1: Common recharge estimation methods in (semi) - arid regions in Southern Africa. Source: Beekman & Xu (2003).

Zone	Approach	Method	Principle
Surface water	Physical	HS	Stream hydrograph separation: outflow, evapotranspiration and abstraction balances recharge
		CWB	Recharge derived from difference in flow upstream and downstream accounting for evapotranspiration, in- and outflow and channel storage change
		WM	Numerical rainfall-runoff modelling; recharge estimated as a residual term
Unsaturated	Physical	Lysimeter	Drainage proportional to moisture flux/ recharge
		UFM	Unsaturated flow simulation e.g. by using numerical solutions to Richards equation
		ZFP	Soil moisture storage changes below ZFP (zero vertical hydraulic gradient) proportional to moisture flux/ recharge
	Tracer	CMB	Chloride mass balance – Profiling: drainage inversely proportional to Cl in pore water
		Historical	Vertical distribution of tracer as a result of activities in the past (^3H)
Unsaturated-Saturated	Physical	CRD	Water level response from recharge proportional to cumulative rainfall departure
		EARTH	Lumped distributed model simulating water level fluctuations by coupling climatic, soil moisture and groundwater level data
		WTF	Water level response proportional to recharge/ discharge

	Tracer	CMB	Amount of Cl into the system balanced by amount of Cl out of the system for negligible surface runoff/ runon
Saturated	Physical	GM	Recharge inversely derived from numerical modelling groundwater flow and calibrating on hydraulic heads/ groundwater ages
		SVF	Water balance over time-based on average groundwater levels from monitoring boreholes
		EV-SF	Water balance at catchment scale
	Tracer	GD	Age gradient derived from tracers, inversely proportional to recharge; Recharge unconfined aquifer based on vertical age gradient (^{3}H, CFCs, ^{3}H/^{3}He); Recharge confined aquifer based on horizontal age gradient (^{14}C)
HS: Hydrograph Separation – Baseflow, **CWB**: Channel Water Budget, **WM**: Watershed Modelling, **UFM**: Unsaturated Flow Modelling, **ZFP**: Zero Flux Plane, **CMB**: Chloride Mass Balance, **CRD**: Cumulative Rainfall Departure, **EARTH**: Extended model for Aquifer Recharge and Moisture Transport through Unsaturated Hardrock, **WTF**: Water Table Fluctuation, **GM**: Groundwater Modelling, **SVF**: Saturated Volume Fluctuation, **EV-SF**: Equal Volume – Spring Flow, **GD**: Groundwater Dating			

3.3.1 FRACTURED AQUIFERS

de Vries & von Hoyer (1988) conducted a groundwater recharge study in Eastern Botswana. The geology of the area is similar to the geology of the Upper Crocodile catchment, it consists of an Archaean gneissic complex and Precambrian sandstones, quartzites, shales and dolomites. There are two types of aquifers; fractured aquifers characteristic of fractured media and weathered zones that form small shallow aquifers and highly permeable karst aquifers that develop in the dolomites.

A water balance method was used to estimate recharge which gave a mean annual recharge of 4% of a mean annual rainfall of 550 mm for the entire study area.

Abiye (2016) estimated the recharge rates for a small catchment underlain by crystalline basement rocks in Johannesburg, South Africa. The study used the WTF method which yielded recharge estimates of 98.9 mm/yr (14% of the annual rainfall) and the BFS method which gave an estimate of 189.1 mm/yr (27% of annual rainfall). Abiye (2016) explained that the BFS method had overestimated recharge as a result of the large inflow volume of wastewater into streams.

Sibanda et al. (2009) conducted a study on the Nyamandhlovu aquifer, a sandstone aquifer, located in Matebeland, Zimbabwe, which is classified as a semi-arid region. The mean annual rainfall of the area was given as 555 mm/yr. Methods used include the CMB, WTF, darcian flownet, ^{14}C and GM which gave recharge estimates ranging between 19 – 26 mm/yr, 2 – 50 mm/yr, 16 – 28 mm/yr, 22 – 25 mm/yr and 11 – 26 mm/yr, respectively. The study concluded that GM gave the best recharge estimates for aerial recharge.

A final recharge estimate of 15 -20 mm/yr based on GM was used for the study area which represented 2.7 – 3.6% of annual rainfall.

3.3.2 KARST AQUIFERS

Using the WTF method, Abiye (2016) estimated the mean recharge to be 118.2 mm/yr for a local dolostone aquifer in Johannesburg, South Africa representing 17% of the annual rainfall of 697 mm.

Bredenkamp (1988) used a rainfall – recharge method to estimate recharge for different dolomitic compartments of the Malmani dolomites. For the Steenkoppies compartment, which is part of the dolomites west of the study area, recharge was estimated as 15 % of a mean annual rainfall of 630 mm. Recharge for the Pretoria/Rietvlei compartment, located east of the study area, was estimated as 17 % of a mean annual rainfall of 682 mm.

Table 2 shows the recharge rates for the West Rand and East Rand dolomites that were calculated as an attempt to use the dolomitic aquifers as water supply during the drought of the 1980s. This information was documented in geohydrological reports, GH3866, GH3316, GH3440 and GH3501, authored by Bredenkamp (1993), Leskiewicz (1984), Bredenkamp et al. (1986) and Kuhn (1989), respectively.

Table 2: A summary of recharge rates for the Malmani dolomites.

METHOD	MAP (mm)	R% of MAP	Location	Report #
SVF	700	24*	West Rand dolomites	GH3866
BFS	725	12.5	East Rand dolomites	GH3316
SPRING FLOW	630	13.9	West Rand dolomites	GH3440
WATER BALANCE	639	10.3	East Rand dolomites	GH3501

*High recharge rates because of leakage from neighbouring compartments

3.4 GROUNDWATER RECHARGE PROCESSES

Natural groundwater recharge can be defined as the downward movement of surface water, originating from precipitation to the groundwater storage irrespective of recharge mechanisms (Lerner et al. 1990; Hendrickx & Walker 1997). Natural groundwater recharge is considered as the primary method for aquifer replenishment (Bredenkamp et al. 1995). Recharge can be defined by three principle mechanisms namely (Lerner 1990; Hendrickx & Walker 1997):

Direct recharge occurs when surface water is added to the aquifer via infiltration of the soil matrix through the unsaturated zone after evapotranspiration and runoff have been accounted for.

Indirect recharge occurs when precipitation accumulates in surface water bodies, such as streams and lakes before infiltrating the unsaturated zone and joining the aquifer.

And Localised recharge results from the localised ponding of surface water which subsequently infiltrates into the unsaturated zone.

The spatial and temporal variability of recharge is dependent on several factors Lerner et al. (1990), which have been identified as climate, geology, topography and land cover. Sewage will also be included as an anthropogenic recharge process as it plays a significant role in contributing towards recharge in the study area.

The spatio-temporal variability of precipitation is one of the major contributors to the spatial and temporal variability of groundwater recharge. The occurrence of precipitation alone is not enough to guarantee recharge, but rather recharge is dependent on the intensity, amount and duration of the precipitation.

Topography is responsible for driving both surface water and groundwater, which is the case in unconfined aquifers where the water table is likely to follow the surface topography. Areas with steep topography are prone to facilitating mountain front recharge, but they can also promote runoff depending on geology, soil characteristics, slope angle, rainfall intensity and duration (Winter et al. 1998). In the case of runoff occurrence, it eventually reaches the streams where indirect recharge can occur. Besides controlling the spatial distribution of recharge topography can also dictate where and how much precipitation occurs known as the orographic effect.

Land cover is also an important factor that controls the spatial variability of groundwater recharge. Take for instance catchments that are highly vegetated; recharge tends to be lower in vegetated areas because precipitation will be intercepted by plants. Furthermore, ET will be greater because of transpiration whereby plant roots take up the available soil moisture thus decreasing precipitation that could have been potential recharge. The soil texture, thickness and the moisture content are important for groundwater recharge. Favourable conditions for recharge to occur are thin soils with a low clay content, high permeability and high soil moisture content. Urban development plays a role in the amount of recharge that can occur as roads and paving create impermeable surfaces that can inhibit infiltration, promoting runoff and subsequently decreasing direct recharge (Lerner 1990).

The type of geology will have an influence on the amount of recharge that occurs. The structural features, type of aquifer, aquifer materials and hydrogeological parameters of an aquifer will determine the extent to which recharge occurs. Fractured aquifers where flow is facilitated only by fractures and the weathered zone(s) will experience less recharge than karst aquifers where flow occurs through karst structures. Thus, the permeability of the aquifer will be the most important factor determining the amount and rate of recharge that occurs (Healy 2010).

Wastewater treatment plants discharge wastewater into nearby streams, one of the implications of this is that wastewater can become artificial recharge especially in places with highly permeable aquifers. The presence of wastewater in streamflow will result in unrealistic recharge estimates, especially if using the baseflow separation method (Abiye 2016). The addition of wastewater in a catchment will act to inflate the amount of natural groundwater recharge, adding another dimension to the complexities of recharge estimation.

3.5 METHODS

3.5.1 WATER BALANCE METHOD

The water balance method is governed by the law of mass conservation, for water storage systems, where the method takes into account the water flowing into and out of the aquifer. The water mass balance equation is written as Inputs – Outputs = ΔS where the input is rainfall and outputs are evapotranspiration and surface runoff. The components of a water balance are best described by simple hydrological processes occurring in the unsaturated zone (Figure 7). From Figure 7 the water balance method equation can be written as:

$$R = P - ET - RO - \Delta S \tag{1}$$

Where P is rainfall in mm, ET is evapotranspiration in mm, RO is runoff in mm and ΔS is change in storage. The water balance method is used in an attempt to measure all the fluxes of the water balance (i.e. P, ET, RO and ΔS) to estimate potential recharge. Potential recharge is estimated from the residual of the water balance fluxes (Gee & Hillel 1988; Bredenkamp et al. 1995). The method assumes that the occurrence of recharge is direct. The reliability and accuracy of the recharge estimates are dependent on the accuracy of the other components. If the errors or the level of uncertainty is high for the other components, these will be carried forward to the recharge estimates (Gee & Hillel 1988; Bredenkamp et al. 1995).

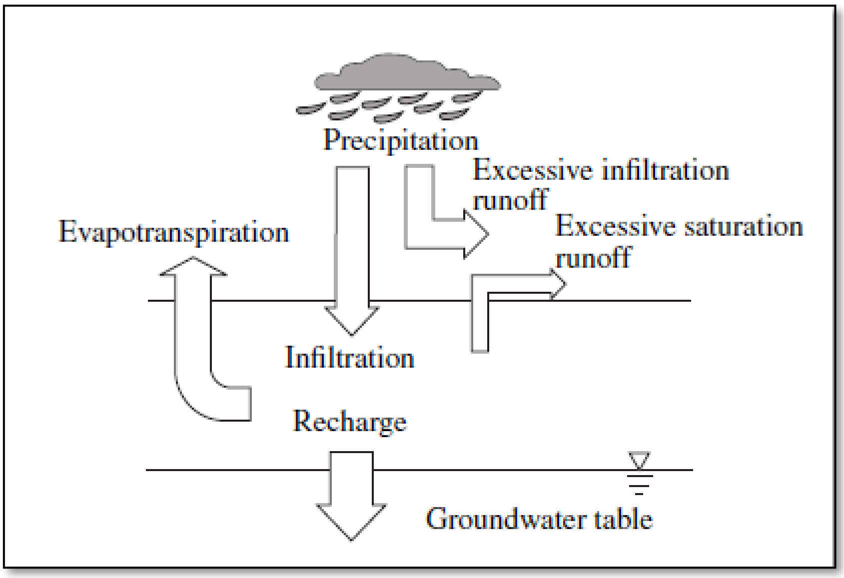

Figure 7: Water balance components. (Source: Hsin-Fu et al. (2007)).

McCabe & Markstrom (2007) applied the Thornthwaite –Monthly Water Balance model to estimate the components of the water balance. The TMWB model requires monthly total rainfall (in mm) and mean monthly temperature (in ^{o}C) as input parameters to estimate the rest of the components. The latitude of the monitoring station must be known as it is needed for day length correction which is included in the PET computation, 30^{o} S was the latitude used for the calculations. In addition, the approximate field capacity should be known as it is needed to compute the soil moisture storage. The field capacity of the soil is obtained by multiplying the water holding capacity (%) by the rooting depth (in m). In the study area, the soil type is predominantly sandy loam, and the dominating vegetation type is grassland (DWAF 2004). Using the Thornthwaite & Mather (1957) tables the field capacity was determined to be 150 mm assuming a water holding capacity of 15% and a rooting depth of 1.00 m (Table 3).

A brief description of all the components estimated by the TMWB model is given below:

Snow Storage is controlled by the temperature threshold value. If the temperature falls below the temperature threshold value, then all rainfall is considered to be snow, which accumulates as snow storage.

Runoff is separated into two components, direct runoff (DRO) and runoff generation. DRO represents runoff, that occurs as a consequence of impervious surfaces or oversaturation of soil, it is calculated as DRO = Rain – drofrac, where drofrac is 5% of rainfall. Runoff generation is runoff generated from surplus, a factor of 0.5 is used to determine the portion of surplus that becomes runoff.

Table 3: Suggested field capacity values based on the tables of Thornthwaite & Mather (1957). Source: Bakundukize et al. (2011).

Vegetation	Soil texture	Water holding capacity (%)	Rooting depth (m)	Field capacity (mm)
Shallow rooted crops (spinach, peas, beans, beets, carrots, etc.)	Fine sand	10	0.50	50
	Fine sandy loam	15	0.50	75
	Silt loam	20	0.62	125
	Clay loam	25	0.40	100
	Clay	30	0.25	75
Moderately rooted crops (corn, cereals, cotton, tobacco)	Fine sand	10	0.75	75
	Fine sandy loam	15	1.00	150
	Silt loam	20	1.00	200
	Clay loam	25	0.80	200
	Clay	30	0.50	150
Deep rooted crops (alfalfa, pasture, grass, shrubs)	Fine sand	10	1.00	100
	Fine sandy loam	15	1.00	150
	Silt loam	20	1.25	250
	Clay loam	25	1.00	250
	Clay	30	0.67	200
Orchards	Fine sand	10	1.50	150
	Fine sandy loam	15	1.67	250
	Silt loam	20	1.50	300
	Clay loam	25	1.00	250
	Clay	30	0.67	200
Mature forest	Fine sand	10	2.50	250
	Fine sandy loam	15	2.00	300
	Silt loam	20	2.00	400
	Clay loam	25	1.60	400
	Clay	30	1.17	350

Potential Evapotranspiration (PET) is defined as the amount of evaporation that would occur if there was an unlimited amount of surface water. It is calculated using the Hamon method: $PET_{Hamon} = \ 13.97 \times d \times D2 \times Wt$, where $Wt = \frac{4.95\ x\ e^{0.062T}}{100}$,d is the number of days in a month, D is the mean monthly hours of daylight in 12 hours, Wt is the saturated water vapour in grams per cubic meters and T is the mean monthly temperature in degrees Celsius.

Soil moisture storage (ST) the soil moisture storage represents that amount of moisture (water) stored in the soil. The soil moisture can vary between the maximum soil moisture storage which is the equivalent to field capacity or the minimum of soil moisture storage corresponding to the wilting point. The amount of available soil moisture is controlled by P – PET, if P-PET >0, then the P – PET value is added to the preceding soil moisture value if the soil moisture reaches field capacity the excess water goes towards runoff and recharge. If after adding P-PET to soil moisture and the soil moisture storage has not reached field capacity then the Thornthwaite – Mather tables have to be consulted to calculate the change in storage.

Actual Evapotranspiration (AET) is the actual amount of evaporation that occurs when water is limited. If P – PET >0 then AET = PET. If P – PET<0 then AET = P + ST.

A Deficit occurs when P –PET<0 indicating that there is no excess water, it is calculated as PET – AET.

A Surplus occurs when there is an excess of water which only occurs when the soil moisture storage is at field capacity, it is calculated as (P-PET) – ST.

A screenshot of the graphical user interface along with output display can be seen in Figure 8 and Figure 9.

Date	PET	P	P-PET	Soil Moisture	AET	PET-AET	Snow Storage	Surplus	ROtotal
Jan-1960	18.0	44.1	23.9	173.9	18.0	0.0	0.0	0.0	14.9
Feb-1960	18.2	62.9	41.6	200.0	18.2	0.0	0.0	15.5	17.2
Mar-1960	28.5	32.8	2.7	200.0	28.5	0.0	0.0	2.7	10.0
Apr-1960	70.1	48.5	-24.1	175.9	70.1	-0.0	0.0	0.0	6.6
May-1960	98.0	150.6	45.1	200.0	98.0	0.0	0.0	21.0	20.1
Jun-1960	147.5	74.7	-76.6	123.4	147.5	0.0	0.0	0.0	10.0
Jul-1960	156.7	151.8	-12.5	115.7	151.9	4.8	0.0	0.0	10.7
Aug-1960	142.1	88.1	-58.4	81.9	117.5	24.6	0.0	0.0	6.0
Sep-1960	96.2	60.6	-38.6	66.1	73.4	22.8	0.0	0.0	3.8
Oct-1960	56.8	115.6	53.0	119.1	56.8	0.0	0.0	0.0	6.2
Nov-1960	29.6	13.4	-16.8	109.1	22.8	6.8	0.0	0.0	0.9
Dec-1960	17.6	84.5	62.6	171.7	17.6	0.0	0.0	0.0	4.3
Jan-1961	17.5	10.0	-8.0	164.9	16.4	1.1	0.0	0.0	0.6
Feb-1961	24.0	43.5	17.4	182.3	24.0	0.0	0.0	0.0	2.2
Mar-1961	41.9	90.0	43.6	200.0	41.9	0.0	0.0	25.9	17.4
Apr-1961	60.7	27.0	-35.1	164.9	60.7	0.0	0.0	0.0	7.8
May-1961	100.5	123.5	16.9	181.8	100.5	0.0	0.0	0.0	9.4
Jun-1961	131.7	111.6	-25.6	158.5	129.3	2.3	0.0	0.0	7.2
Jul-1961	155.5	133.5	-28.7	135.8	149.6	5.9	0.0	0.0	7.5
Aug-1961	132.1	84.4	-51.9	100.6	115.4	16.7	0.0	0.0	4.6
Sep-1961	83.5	140.8	50.3	150.8	83.5	0.0	0.0	0.0	7.2
Oct-1961	53.6	66.5	9.5	160.3	53.5	0.0	0.0	0.0	3.4

Save

Figure 8: An example of the output screen for the TMWB model (Source: U.S. Geological Survey (USGS))

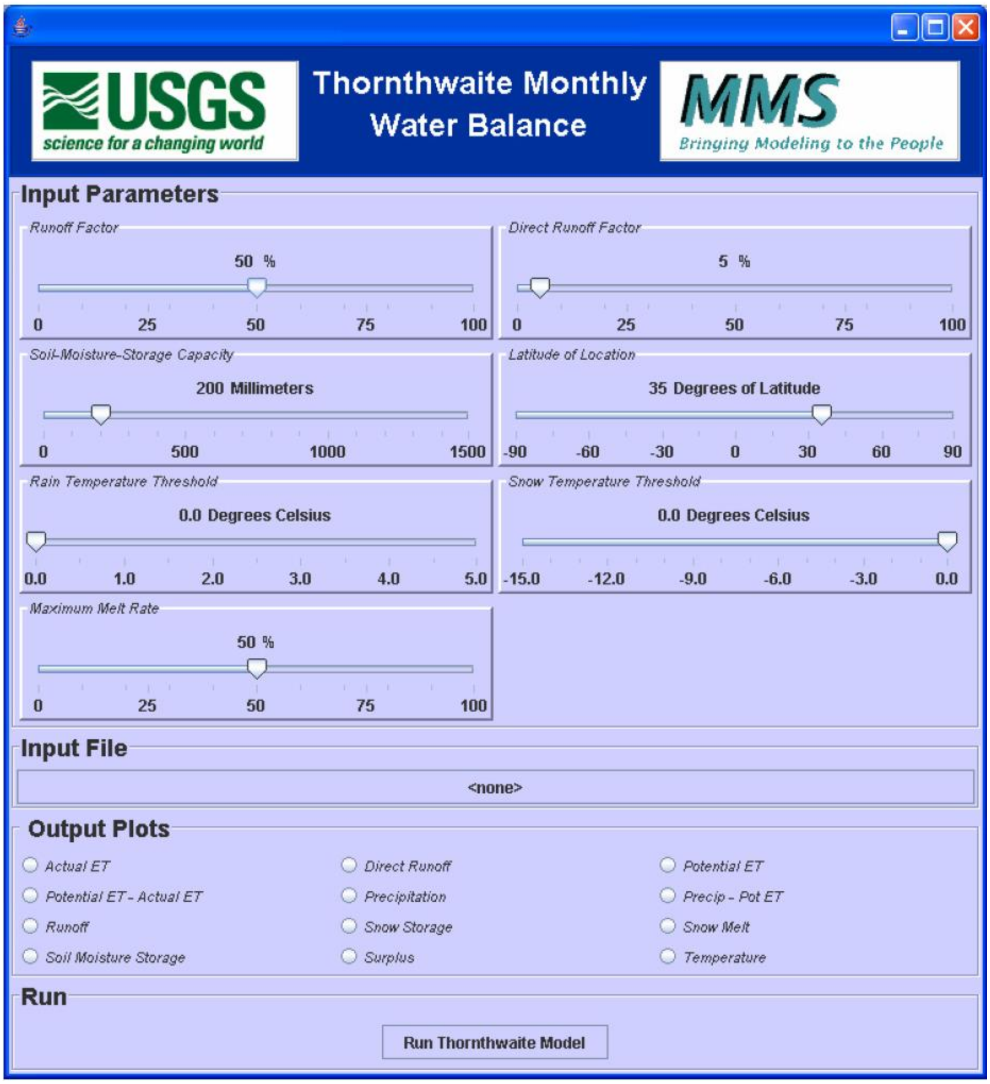

Figure 9: Graphic user interface for the TMWB model (Source: USGS).

3.5.2 BASEFLOW SEPARATION METHOD

To fully appreciate the BFS method, an understanding of groundwater – surface water interaction (GSI) must first be acquired. GSI can be loosely defined as the continuous water exchange between surface water and groundwater; this interaction is controlled by the geomorphology, soil characteristics, geology and climate (Winter et al. 1998; Sophocleous 2002). To illustrate the hydrological connection between groundwater and surface water the physical processes of GSI are explained.

The GSI can be viewed as a two-way process; streams can either lose water to groundwater, referred to as a losing stream or streams can gain water via groundwater inflow, known as a gaining stream (Figure 10) (Winter et al. 1998; Sophocleous 2002).

In a gaining stream, the water table intersects the stream channel allowing groundwater inflow, the rate of flow depends on the slope of the water table and the aquifer properties (Sophocleous 2002). In a losing stream, the water table is lower than the stream channel thus the stream loses its water to groundwater, the rate of stream loss depends on the properties of the underlying alluvium. It is possible for a stream to change from a gaining stream to a losing stream and vice versa along the course of its flow (Winter et al. 1998; Sophocleous 2002). The BFS method can only be used on gaining streams.

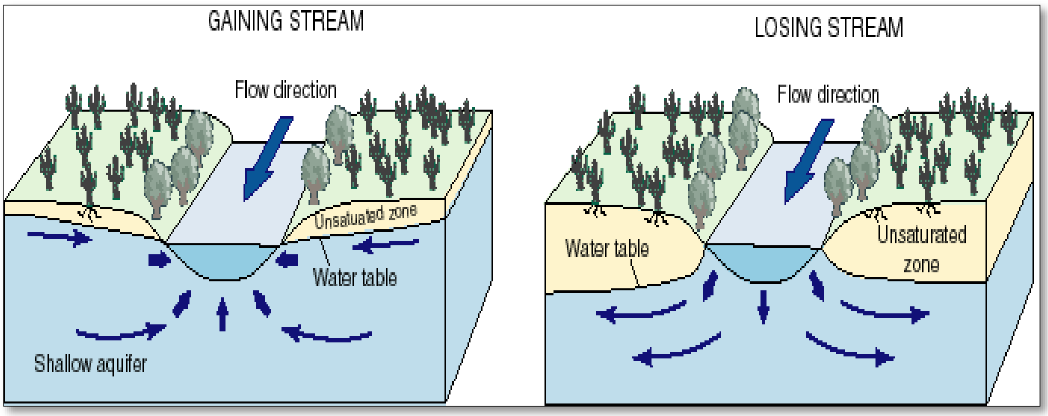

Figure 10: Groundwater - surface water interaction: gaining and losing stream. (Source: USGS).

The BFS method separates streamflow into a surface runoff component and a baseflow component, representing groundwater discharge, based on time series data of stream discharge. This method provides a way of estimating groundwater recharge, assuming that groundwater discharge is equal to groundwater recharge over a long-term period (Frôhlich et al. 1994; Bredenkamp et al. 1995; Wittenberg 2003; Healy 2010); this is valid assuming that transmission and evapotranspiration losses are negligible.

The effect of a rainfall event on a hydrograph is defined by a peak, which corresponds to surface runoff (Figure 11). The assumption is; the peak of the hydrograph should closely resemble the peak of a precipitation event. During winter when rainfall is minimal the streams are sustained by the continuous discharge of groundwater, the baseflow (Frôhlich et al. 1994; Bredenkamp et al. 1995; Wittenberg 2003).

Timeplot, an Excel based program, was used for the baseflow separation method, it is based on the single parameter digital recursive filter method of Nathan & McMahan (1990). The principle behind the digital filter method is that a digital filter (mathematical operations) is used to partition high frequency waves from low frequency waves (Lyne & Hollick 1979). In the case of streamflow, surface runoff produces higher flows corresponding to high-frequency waves, and baseflow produces low flows corresponding to lower frequency waves. Thus streamflow is filtered based on the flow volume (Nathan & McMahan 1990; Eckhardt 2005).

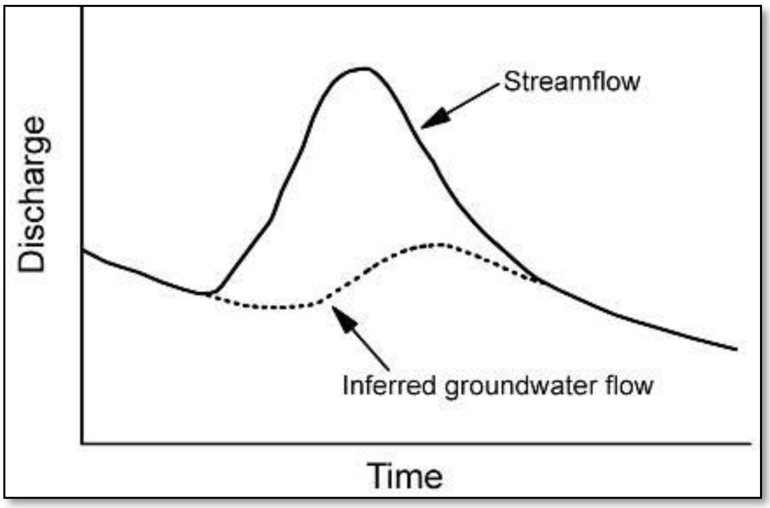

Figure 11: A streamflow hydrograph showing the baseflow and runoff components.

The digital filter method is preferred for this study as it is better suited for analysing long-term streamflow data, although digital filters are not based on physical processes the methods are objective and easily repeatable (Nathan & McMahan 1990) making it favourable for baseflow comparisons and large data sets. The computation of baseflow using the Nathan and MacMahon (1990) method is:

$$R_{k+1} = \alpha R_k + \frac{(1+\alpha)}{2}(Q_{k+1} - Q_k) \quad (2)$$

Where R_k is runoff in m^3/s, Q_k is streamflow in m^3/s and α is the baseflow filter parameter.

Once the baseflow has been computed, wastewater discharge must be subtracted from the baseflow estimate to determine the amount of recharge contributed by precipitation relative to effluent discharge.

The advantage of using this method is that it gives a spatial rather than a point estimate of recharge additionally the data is readily available therefore making it cost effective and efficient.

3.5.3 WATER TABLE FLUCTUATION METHOD

The WTF method assumes that the rise in water table depth is caused by the direct result of a precipitation event, provided natural conditions prevail (Healy & Cook 2002; Scanlon et al. 2002; Healy 2010; Shi et al. 2015). This method should only be applied to shallow unconfined aquifers as the groundwater fluctuations are better displayed, the WTF method necessitates there be a groundwater level change (Healy & Cook 2002). The data requirements are time series data of groundwater levels and the specific yield of the aquifer. Recharge computed using the WTF method is given as:

$$R = S_y * \frac{\Delta h}{\Delta t} \qquad (3)$$

Where S_y is specific yield, Δh is the change in water table height in m and Δt is the change in time in years. A graphical approach was used to calculate Δh whereby Δh is the difference between the peak of groundwater level rises and the lowest point of the extrapolated antecedent recession curve (Figure 12).

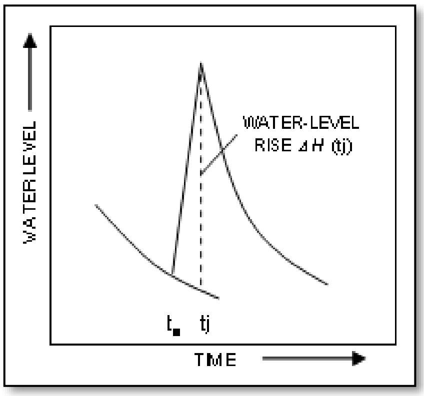

Figure 12: Determination of water level rises in boreholes.

The antecedent recession curve is the path the water table (hydrograph) would have followed in the event had there not been any rainfall. Equation 3 is only applied to water table rises as they signify recharge. This method is popular among hydrogeologists as it is simple to use, data can be obtained easily, and it doesn't take into account the flow mechanisms of recharge (Healy & Cook 2002).

Specific yield is defined as the volume of water that drains under the influence of gravity in an aquifer (Meinzer 1923). The specific yield formula is given as:

$$S_y = \emptyset - S_r \qquad (4)$$

Where $\emptyset$ is soil moisture, and S_r is specific retention. Literature values for specific yield were used for the WTF method because there was no pumping test data from the study area that was made available. The specific yield values used (Table 4) were obtained from the Far West Rand (FWR) dolomites of the Malmani Subgroup, in Carletonville, Johannesburg. Enslin & Kriel (1968) estimated the specific yield values for the FWR dolomites by carrying out a water balance study for the FWR dolomite compartment, which was being dewatered. The study allowed the authors, Enslin & Kriel (1968), to observe the relationship between specific yield and depth, revealing that specific yield decreased with increasing depth. Specific yield varied from 9.1 to 1.3 % with an increase in depth from 61 to 146 m.b.g.l (see Table 4). This method of calculating specific yield is superior because it considers the heterogeneous nature of the aquifer as the water table is lowered. Whereas other methods such as pumping tests, firstly assume a homogeneous aquifer and secondly only obtain a single specific yield value through calculations (Enslin & Kriel 1968). Based on the discovery of Enslin and Kriel (1968), that specific yield changes with depth it is clear that a single specific yield value is not representative of the entire compartment.

Table 4: Specific yield values from the FWR dolomites.

(Source: Enslin & Kriel (1968)).

DEPTH (MBGL)	Specific yield (%)
61	9.1
76	5.5
107	2.6
126	2
146	1.3

3.5.4 ENVIRONMENTAL ISOTOPES

3.5.4.1 STABLE ISOTOPES

The stable isotopes of oxygen and hydrogen are important tracers, especially in hydrological studies as they are useful in tracing the movement of groundwater, locating groundwater sources and they provide insight into hydrological processes (Clark & Fritz 1997). The use of O and H isotopes is based on their abundance in water; this guarantees that their composition remains the same unless there is a significant amount of evaporation occurring or mixing of meteoric waters with different isotopic compositions (Clark & Fritz 1997).

Kinetic and equilibrium processes are responsible for isotopic fractionation, which results in the isotopic variation of vapour and rain (Dansgaard 1964; Clark & Fritz 1997; Hoefs 2009). Kinetic fractionation is a process that partitions stable isotopes from each other based on their mass during unidirectional processes. Evaporation, driven by kinetic isotopic fractionation in the hydrological cycle, is a process that fractionates the lighter water molecule isotopes from the heavier isotopes. Therefore, the vapour phase and consequently the vapour mass will be enriched in the lighter isotopes (reflecting a vapour mass that is isotopically depleted) and the remaining water will be enriched in the heavier isotopes. The rate of kinetic fractionation is strongly influenced by temperature and humidity. Under conditions of low humidity and high temperature kinetic evaporation is favoured, at low temperatures when humidity nears 100% equilibrium fractionation between water and vapour dominates and evaporation is minimised. (Dansgaard 1964; Clark & Fritz 1997).

The first systematic study of rainwater was carried out by Craig (1961) who recognised a correlation between δD and $\delta^{18}O$ of rainwater worldwide, which corresponded to a line of best fit defined by the equation:

$$\delta D = 8 * \delta^{18}O + 10 ‰ \qquad (5)$$

Equation 5 is known as the global meteoric water line (GMWL) it is constructed from the δD and $\delta^{18}O$ averages of local meteoric water lines (LMWL) worldwide. The LMWL is dependent on climate, geographic location and source region of evaporation to form clouds (Craig 1961; Clark & Fritz 1997; Hoefs 2009).

The distribution of O and H isotopes in meteoric water is controlled by several factors such as (Dansgaard 1964):

The rainout effect is responsible for the progressive depletion of isotopic ratios, as the vapour mass follows a trajectory from its moisture source to higher latitudes and altitudes. The rainout process is driven by decreasing temperature.

Temperature effect which is related to the progressive depletion of the rainfall's isotopic composition with decreasing temperatures. It is also responsible for driving the rainout process.

Altitude (latitude) effect which is responsible for the progressive depletion of isotopic signatures with increasing altitude (latitude). The altitude effect is temperature dependent because condensation occurs as a result of decreasing temperature with increasing altitude.

And seasonal effect which is caused by the shift in temperature with changing seasons resulting in the seasonal variation of rainfall isotopes.

The isotope effects of Dansgaard (1964) reveal that the oxygen isotopic evolution of precipitation has a strong temperature dependency (Dansgaard 1964; Clark & Fritz 1997; Hoefs 2009).

Rainfall within the Johannesburg region is currently being collected by Professor Tamiru Abiye, of the Hydrogeology programme at the University of Witwatersrand, and being analysed for $\delta^{18}O$ and δD with the expectation of constructing a LMWL for the Johannesburg region.

Deuterium excess (d-excess) is generally defined by the equation:

$$\text{d-excess} = \delta D - 8 * \delta^{18}O \text{ ‰} \qquad (6)$$

It measures the ratio of δD and $\delta^{18}O$ in water and shows the deviation of a single sample from the GMWL (Dansgaard 1964; Froehlich et al. 2002). For this study d-excess will be calculated based on the constructed LMWL for Johannesburg, South Africa.

D-excess can be used to trace climatic processes at the source region i.e. oceanic or continental source, and determine the moisture source of rainfall, with high d-excess values being characteristic of local moisture sources whereas low d-excess values are associated with a regional circulation (Dansgaard 1964; Froehlich et al. 2002; Hoefs 2009).

The variability of d-excess is primarily caused by the source region of a vapour mass, sub-cloud fractionation processes (evaporation and condensation), relative humidity and temperature. (Merlivat & Jouzel 1979; Froehlich et al. 2002).

Isotopic ratios are reported in δ notation:

$$\delta = \left(\frac{R_{sample}}{R_{standard}} - 1\right) * 1000 \text{ per mil (‰) , where} \qquad (7)$$

$$R = \frac{D}{H} \; or \; \frac{O^{18}}{O^{16}}$$

The Allison et al. (1984) isotopic shift method will be applied to estimate recharge; this method is based on the relationship between isotopic enrichment and recharge. The soil water is isotopically enriched as a result of evaporation, the incoming rainwater then mixes with the soil water to eventually recharge the groundwater storage. This process is reflected by an isotopic concentration profile showing a combination of evaporated soil water and rainwater. The method requires stable isotopic compositions of rainfall and soil water, in this case, spring water, to compute for recharge which is given as:

$$R = \left(\frac{22}{\delta D\ shift}\right)^2 \text{ or } \quad R = \left(\frac{3}{\delta 18O\ shift}\right)^2 \qquad (8)$$

Where δD and δ^{18}O are deuterium and oxygen isotopic composition in ‰, respectively.

3.5.4.2 RADIOISOTOPES

Tritium is the radiogenic isotope of hydrogen; it has a half-life of 12.43 years. The tritium concentration in water can be expressed as a ratio of one tritium atom to 10^8 hydrogen atoms which is defined as 1 tritium unit (TU).

Tritium is used as a dating tool in groundwater studies to provide a means for determining the residence time of groundwater (Clark & Fritz 1997; Healy 2010).

The residence time of groundwater is important because it gives an indication of how long the groundwater has been in circulation for and whether the groundwater was a part of a deep or shallow circulation (Beekman & Xu 2003). Additionally, tritium can also be used to understand the range of groundwater contribution to surface water bodies (Michel 1992; Clark & Fritz 1997). The residence time can be calculated by comparing tritium from groundwater to tritium from rainwater. Residence time is calculated as:

$$t = t_{1/2} In\, (\frac{A_o}{A_{obs}})/In2 \qquad (9)$$

Where A_o is the presumed initial activity in TU and A_{obs} is the observed activity in TU. Based on the tritium values, groundwater can be classified as one of six water types (Clark & Fritz, 1997). Water with:

1. < 0.8 TU sub-modern groundwater recharged prior to 1952
2. 0.8 – 4 TU mixture of sub-modern and recent recharge
3. 5 – 15 TU modern recharge (<5 to 10 years)
4. 15 – 30 TU some bomb tritium present
5. > 30 TU considerable component of recharge from 1960 or 1970s
6. > 50 TU dominantly 1960s recharge

In this study tritium values are not expected to exceed the input function value which is currently at 5.6 TU, any values above this will be an indication of additional tritium input source(s).

4 METHODOLOGY

4.1 DESK WORK

The desktop study involved a review of material relevant to the research; this included a literature review on recharge methods used for recharge rates estimation, recharge processes and their effect on groundwater recharge and previous studies of recharge rates in Southern Africa. It also involved sourcing hydrogeological data and wastewater discharge from various institutions.

4.2 DATA COLLECTION

Long-term discharge measurements and borehole water level data (time series data) were gathered from the national database of the Department of Water and Sanitation (DWS) and the National Groundwater Archive (NGA), respectively. Existing hydrometeorological data for weather stations found within and around the catchment area were obtained from the DWS and the South African Weather Services (SAWS), this included temperature, rainfall and evaporation data. Files used for the Geographical Information System (GIS) software included shapefiles obtained from RQIS database of the DWS and the Water Research Commission (WRC). DEM tiff files were obtained from the USGS National Elevation Dataset, and land cover tiff files were obtained from the South African National Biodiversity Institute (SANBI) website. Wastewater discharge and water transfer volumes were sourced from groundwater assessment reports by the DWA and Rand Water, respectively.

4.3 FIELD WORK

Field work undertaken included:

a) Rainfall sample collection by Professor Tamiru Abiye in the Johannesburg area.
b) Water sample collection at different stream locations around the catchment area in June. The water samples were collected in 1L bottles for tritium.

c) Monthly sampling of spring water at Alberts Farm, Northcliff, Johannesburg. The water samples collected were used for the environmental isotope methods.

4.4 LAB WORK

The lab work portion involved the analysis of the water samples which were conducted in the Hydrogeology Lab, at the University of the Witwatersrand (South Africa), except the tritium water samples which were analysed at iThemba Labs, Johannesburg. The water samples were analysed for tritium and the stable isotopic composition of $\delta^{18}O$ and δD.

4.5 MEASURING EQUIPMENT

The stable isotopes of $\delta^{18}O$ and δD were analysed by using the Liquid Water Isotope Analyzer-model 45-EP at the University of the Witwatersrand (South Africa). The instrument contains the laser analysis system and an internal computer, Liquid autosampler, a small membrane vacuum pump, and a room air intake line that passes air through a Drierite column for moisture removal. A Hamilton microliter syringe was used to inject 0.75 µL of the sample through a PTFE septum in the autosampler. The injection port of the autosampler is heated to 46°C to help vaporise the sample under vacuum immediately upon injection. The vapour then travels down the transfer line into the pre-evacuated mirrored chamber for analysis. A 1.5-mL aliquot of a sample (filtered if it is cloudy or contains sediment) or standard is pipetted into a 2 –mL autosampler glass vial and closed with PTFE septum caps. Five standards were used in the analysis. The laser machine is capable of providing accurate results with a precision of approximately 1 ‰ for δD and 0.2 ‰ for $\delta^{18}O$ in liquid water samples of up to at least 1000 mg/L dissolved salt concentration.

Tritium analysis required the water samples to be distilled and subsequently enriched by electrolysis. The electrolysis cells consist of two concentric metal tubes, which are insulated from each other. The outer anode, which is also the container, is of stainless steel. The inner cathode is of mild steel with a special surface coating. Approximately 500 ml of the water sample, having first been distilled and containing sodium hydroxide, is introduced into the cell. A direct current of approximately 10–20 ampere (A) is then passed through the cell, which is cooled because of the heat generation.

After several days, the electrolyte volume is reduced to approximately 20 ml. The volume reduction of approximately 25 times produces a corresponding tritium enrichment factor of approximately 20. Samples of standard known tritium concentration (spikes) are run in one cell of each batch to check on the enrichment attained. For liquid scintillation counting samples are prepared by directly distilling the enriched water sample from the now highly concentrated electrolyte. 10 ml of the distilled water sample is mixed with 11 ml Ultima Gold and placed in a vial in the analyser and counted 2 to 3 cycles of 4 hours. Detection limits are 0.2 TU for enriched samples.

5 RESULTS AND DISCUSSION

The study period for all the methods was based on a hydrological year starting from October to September, with the exception of the environmental isotopes.

Within the Upper Crocodile catchment not all quaternary catchments had meteorological stations. Therefore, for the methods that required information such as temperature and or rainfall the catchments without the meteorological stations used the data from neighbouring catchments that had stations. The quaternary catchments with no temperature or rainfall data had to have similar elevations and be in proximity with the catchment whose data it was sharing. Only catchments A21C and A21F had both temperature and rainfall data while catchments A21A, A21B and A21H only had rainfall data.

5.1 WATER BALANCE METHOD

Potential recharge for the entire catchment was estimated using a TMWB model by McCabe & Markstrom (2007) for the hydrological years of 1995 to 2004. The TMWB model requires meteorological data such as rainfall and temperature as input parameters to estimate the rest of the water balance components.

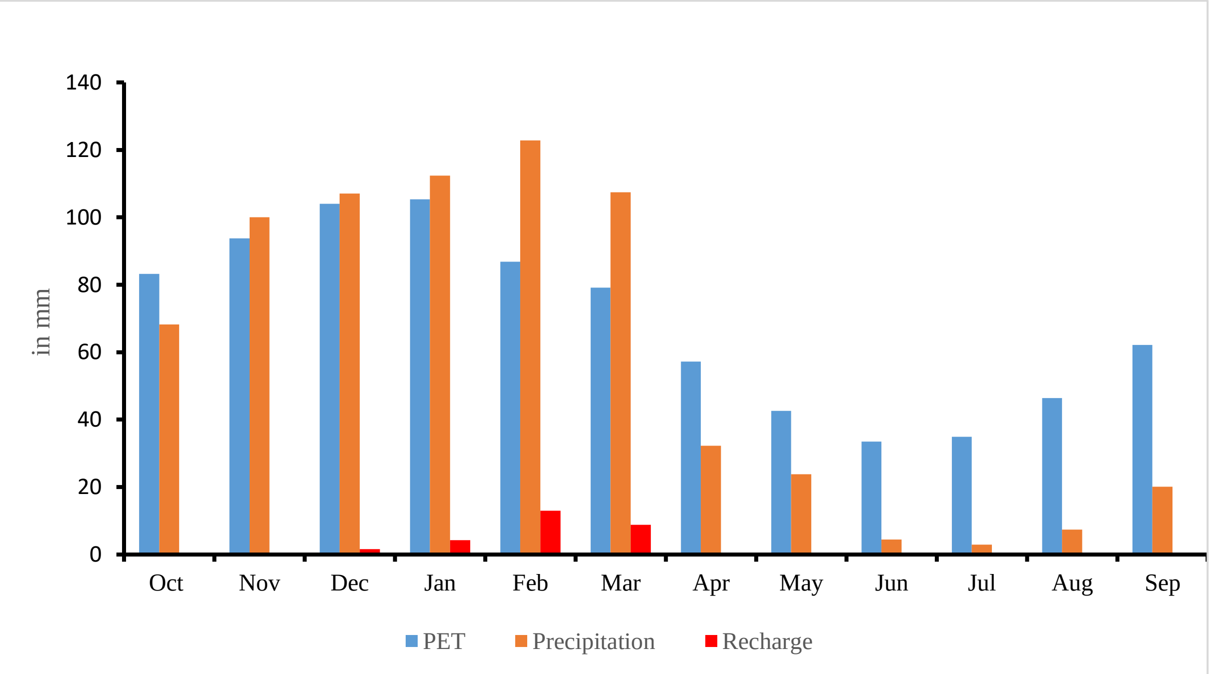

Figure 13: Mean monthly water balance components.

Figure 13 illustrates the mean monthly distribution of rainfall and PET for the study area. From the graph, it is evident that that two distinct seasons dominate the catchment area, a wet season which extends from October to March, with peak rainfall occurring in February and a dry season extending from April to September with the least amount of rainfall falling in July. A similar pattern can be observed with PET, where high PET estimates coincide with the hot summer months and the low PET estimates, the cold winter months. During the summer months when rainfall exceeds PET there is a surplus of water meaning there is potential for recharge to occur given that the soil moisture is at field capacity, determined to be 150 mm, and runoff has been accounted for. In the dry winter months where PET is high and far exceeds rainfall there is a water deficit. Available water stored in the soil will be taken up by evapotranspiration and as a result, soil moisture will be below field capacity and thus recharge is unlikely, assuming recharge does not occur via preferential flow.

Figure 13 also shows a distribution of mean monthly recharge for the entire catchment area throughout the hydrological period of 1995 to 2004. The distribution pattern of recharge loosely reflects that of the mean monthly rainfall for the wet season. Recharge only occurs during the wet summer months between November and March with peak recharge occurring in February. No recharge is observed for the dry months as during this time PET is greater than rainfall resulting in a water deficit. The above is in agreement with what was mentioned above where the potential of recharge occurrence is dependent on rainfall exceeding PET and soil moisture being at field capacity. A delay between recharge and the onset of rainfall can be seen in Figure 13, rainfall occurs from October, but the recharge response is only seen from December, furthermore in November there is a water surplus, but no recharge occurs. A possible explanation for this is that during the onset of the rainy season the rainfall is still replenishing the soil moisture that was lost during the dry season, inferring that a threshold value (field capacity) must first be met before recharge can occur (Bakundukize et al. 2011). The monthly recharge rates of the study area vary between a minimum and maximum value of 0 and 13 mm, respectively (Table 5).

The role of land cover on recharge estimates should also be noted, from Figure 14 it can be seen that the catchment area has been heavily modified by urban development. The consequence of this is increased runoff due to a lack of pervious surfaces; this will impact the amount of recharge that can occur as little to no infiltration can occur.

Table 5: Mean monthly components of the water balance for the hydrological period of 1995 – 2004.

MONTHS	PET (mm)	Rainfall (mm)	P-PET (mm)	Recharge (mm)	R% of rain
OCT	83.2	68.2	-18.4	0	0
NOV	93.8	100.1	1.2	0.4	0
DEC	104	107	-2.4	1.6	1
JAN	105.4	112.3	1.3	4.3	4
FEB	86.9	122.8	29.8	13	11
MAR	79.1	107.4	22.9	8.9	8
APR	57.2	32.2	-26.6	0	0
MAY	42.6	23.8	-20	0	0
JUN	33.5	4.5	-29.2	0	0
JUL	34.9	2.9	-32.1	0	0
AUG	46.4	7.4	-39.4	0	0
SEP	62.1	20.1	-43	0	0
ANNUAL	829.1	708.7	-155.8	28.2	4
MAX	105.4	122.8	29.8	13	11
MIN	33.5	2.9	-43	0	0

Figure 15 shows the annual average rainfall, PET and recharge computed using the TMWB model for the hydrological period of 1995 to 2004. Mean annual rainfall and PET values for the entire catchment area for the duration of the study are given as 709 mm and 829 mm, respectively (Table 6). Annual rainfall varies between a minimum of 372 mm for the year 2003 and a maximum of 1072mm for the year 1996, whereas the annual PET minimum and maximum are given as 792 mm for the year 1996 and 887mm for the year 2003, respectively (Table 6).

Recharge is calculated as the difference between rainfall and the rest of the water balance components (i.e. ET, change in storage and runoff). Figure 15 illustrates the variable nature of recharge, clearly, the amount of recharge that occurs annually is predominantly controlled by rainfall and PET.

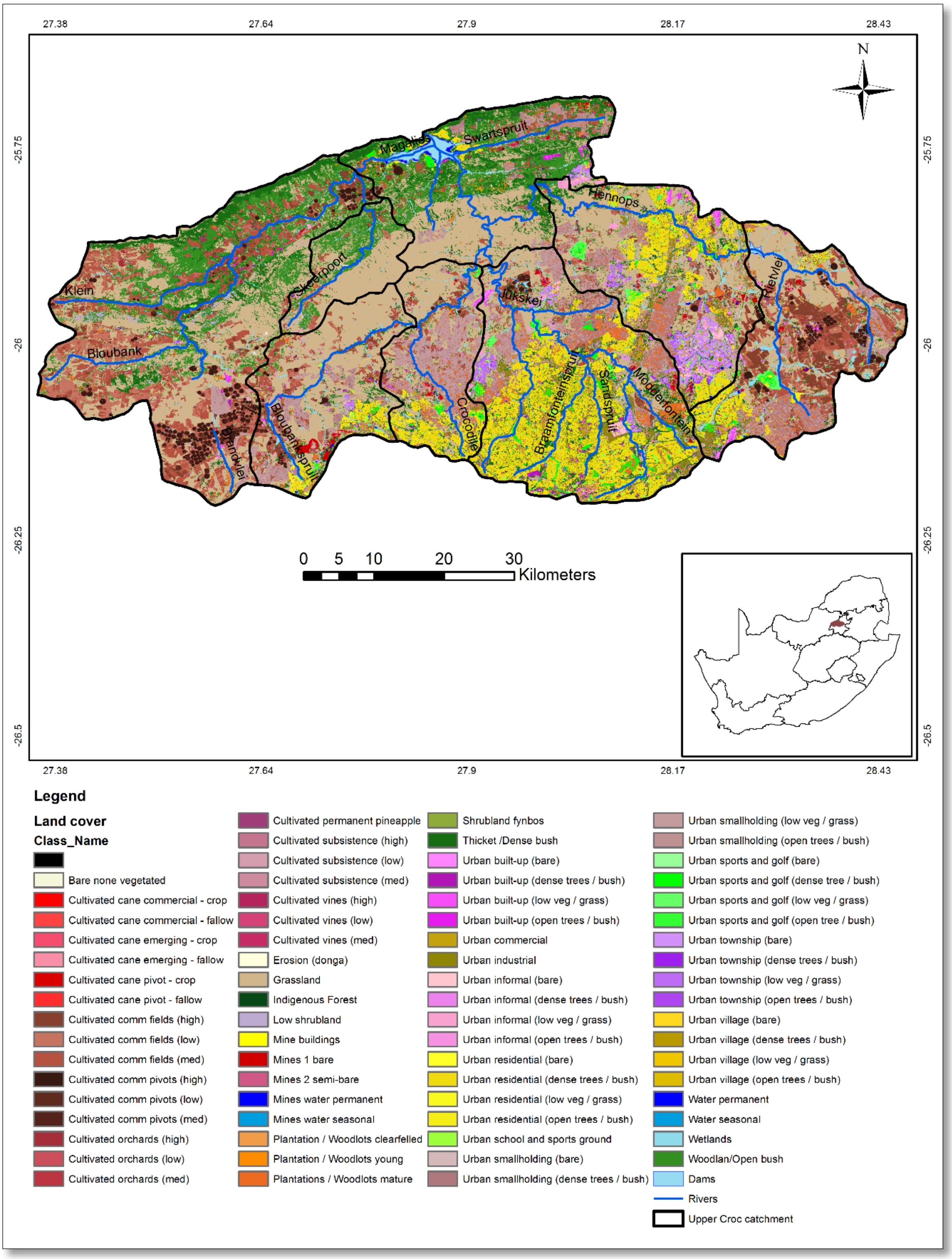

Figure 14: Land cover of the Upper Crocodile catchment.

For the years 1996, 1997 and 2000 when annual rainfall was greater than PET the amount of recharge was substantially higher than for the years when PET was greater than rainfall. A reason for recharge being substantially higher than normal can be attributed to annual rainfall exceeding the mean annual rainfall. And because the amount of rainfall was higher than average rainfall water surplus would be higher than average, translating to soil moisture being at field capacity for longer periods hence higher recharge rates. The years 2002 and 2003 have the lowest recorded rainfall and the highest recorded PET values, such observations point to drought conditions. Because of the water deficit, soil moisture would have been below field capacity, and the low amounts of rainfall would not have been enough to replenish the soil moisture hence no recharge occurred for that period.

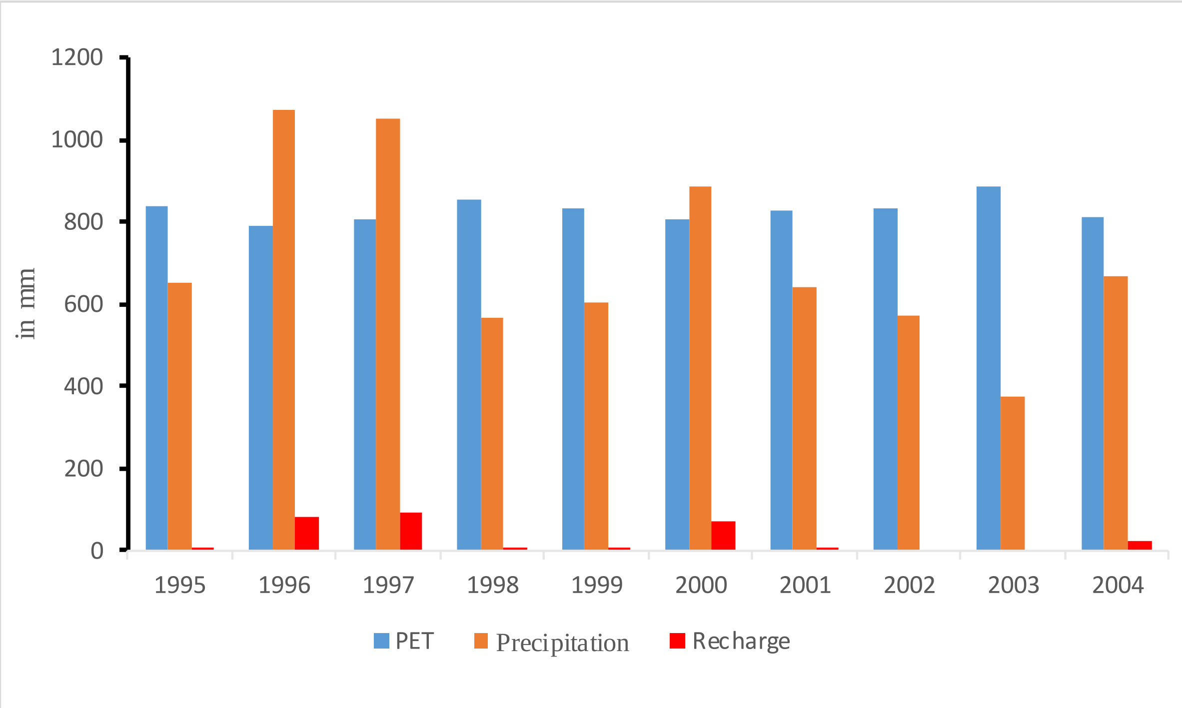

Figure 15: Mean annual water balance components.

Recharge can still occur even if PET is greater than rainfall, it is termed episodic recharge, and it is induced by high-intensity rainfall events (van Wyk 2010; Abiye 2016), for a single rainfall event rainfall can exceed PET on a single day. In such a case, it is likely that recharge will occur via preferential flow rather than diffusive flow, that way recharge can occur without the soil moisture being at field capacity.

The annual recharge rates of the study area vary between a minimum and maximum value of 0 and 90.5 mm/yr, respectively (Table 6), with the mean annual recharge rate calculated as 28.2 mm/yr representing 4% of the mean annual rainfall of 708.7 mm.

This value is in agreement with a study conducted by de Vries and von Hoyer, (1988) who also observed a recharge of 4% for a water balance study in a similar geological setting.

Figure 15 also draws attention to that fact that recharge is not necessarily a yearly occurrence and is rather sporadic in nature, which is not uncommon in semi-arid regions because of the spatio-temporal variability of meteorological (especially PET and rainfall) conditions as well as the hydrogeological environments (van Wyk 2010).

Table 6: Mean annual components of the water balance for the hydrological year of 1995 – 2004.

YEAR	PET (mm)	Rainfall (mm)	P-PET (mm)	Recharge (mm)	R% of MAP
1995	837.6	654.8	-215.5	5.7	1
1996	791.5	1072.5	227.4	79.5	7
1997	804.8	1051.9	194.5	90.5	9
1998	855.7	566.3	-317.7	7	1
1999	831.5	605.1	-256.7	0.7	0
2000	806.5	885.3	34.5	71.1	8
2001	826.5	640	-218.5	2.8	0
2002	835.8	572	-292.5	0	0
2003	886.7	372.3	-533	0	0
2004	814.4	667.2	-180.6	24.6	4

5.2 BASEFLOW SEPARATION METHOD

Baseflow was estimated from daily streamflow data during 1998/10-2003/09 (Appendix A). Streamflow data was obtained from the hydrology database of the DWS, which has numerous monitoring stations throughout the catchment, monitoring stations used for this study area can be found in Table 7.

Mean annual baseflow estimates were calculated using Timeplot, an Excel based program that separates baseflow from streamflow by filtering high flows from low flows, for quaternary catchments A21A – A21G. The filter parameter used for all the baseflow calculations was 0.995; this was found to be the best filter value for rivers in South Africa (Smakhtin & Watkins 1997).

Quaternary catchment A21H has been excluded from the baseflow calculations as baseflow estimates would not be reflective of the natural conditions of the catchment. The streamflow discharge monitoring station for catchment A21H is located downstream of the Hartbeespoort Dam, and streamflow discharge is heavily influenced by man's control of the dam.

A reason for calculating baseflow for each quaternary catchment instead of the whole catchment was to see the spatial and temporal variability of recharge throughout the catchment area.

Within the study area, there are eight wastewater treatment plants (Figure 1) located in proximity to rivers that discharge treated wastewater into the nearby streams. Thus, to account for the additional input into the streams the volume of wastewater discharge was subtracted from the baseflow estimate to give a more reliable estimate of baseflow. If wastewater discharge is not factored into baseflow calculations, it results in the overestimation of baseflow (Abiye 2016).

Table 7: Location of the stream discharge monitoring stations.

STATION NO	Catchment	Place	Latitude	Longitude	Drainage Area km2
A2H090	A21A	Hennops River @ Van Riebeeck Nat Res	-25.88555	28.30277	451
A2H014	A21B	Hennops River @ Skurweberg	-25.79828	27.98539	527
A2H044	A21C	Jukskei River @ Vlakfontein	-25.89550	27.93481	761
A2H049	A21D	Bloubank Spruit@Riet Spruit @ Zwartkop	-25.97681	27.83639	372
A2H045	A21E	Krokodil River @ Vlakfontein	-25.89275	27.91483	290
A2H013	A21F	Magalies River @ Scheerpoort	-25.77703	27.76117	1001
A2H034	A21G	Skeerpoort River @ Scheerpoort	-25.82492	27.77181	160

Figure 16 shows long term daily rainfall and the apparent baseflow; the effluent discharge has not been accounted for, for the period 1998/10 to 2003/09 it illustrates that rainfall has a seasonal influence, with the majority of rainfall falling between October and March. The mean annual rainfall for the five-year period was 603 mm/yr, for the entire study area. The lowest rainfall values were recorded for the year 2003 with a MAR of 372 mm/yr and the highest for 2000 with a value of 1223 mm/yr. A comparison of the apparent baseflow and rainfall shows that baseflow closely resembles the distribution pattern of rainfall, suggesting that rainfall has an influence on baseflow. It should be noted that the apparent baseflow remains relatively high even during the dry months of April to September, this is clearly observed in quaternary catchments A21A, A21D, A21G and A21H especially from the year 2001. The substantial baseflow volume in winter can be attributed to the presence of wastewater. Wastewater is discharged continuously throughout the year thus while baseflow fluctuates seasonally wastewater remains constant hence in winter the apparent baseflow appears to be higher than expected.

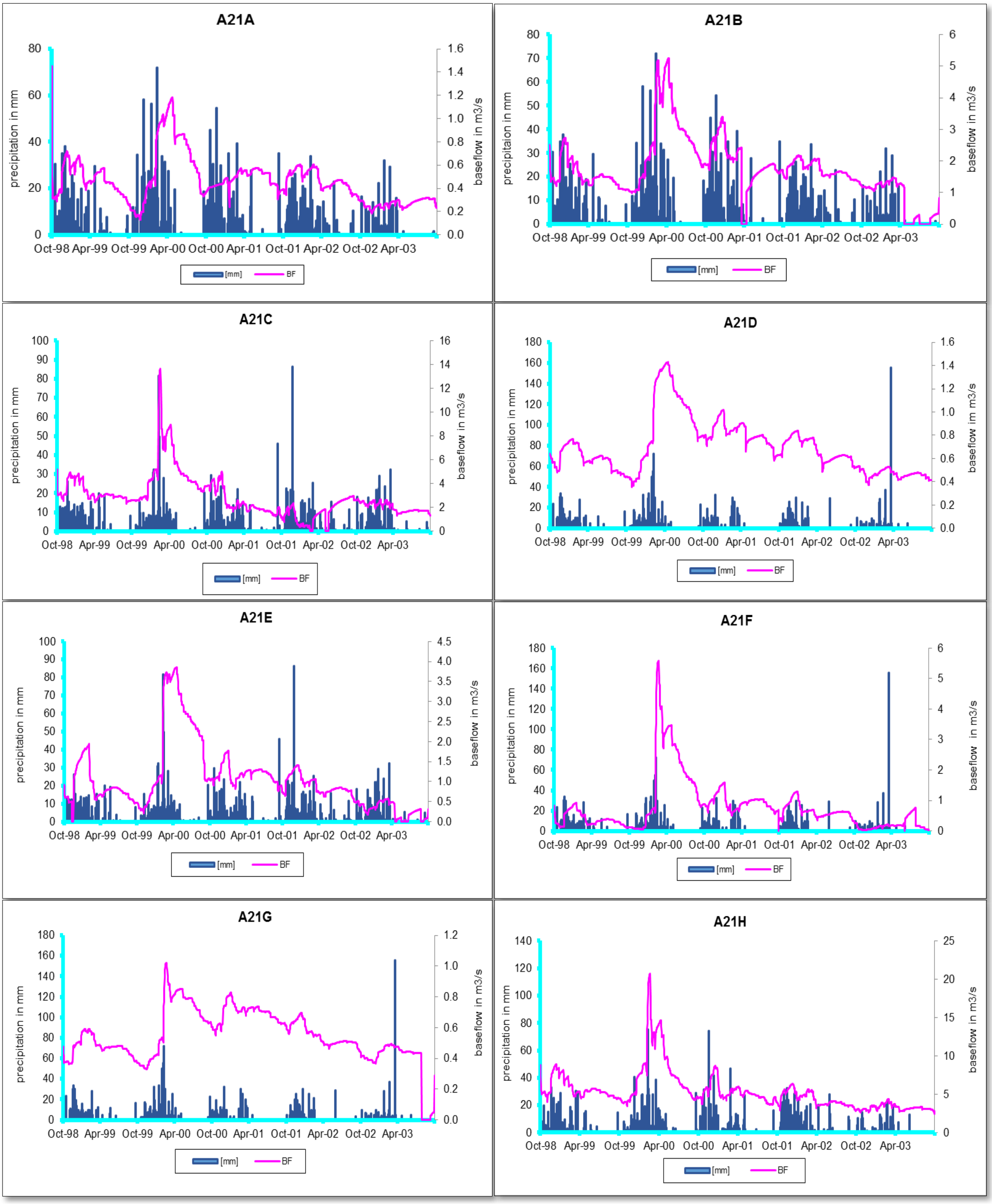

Figure 16: Plots of apparent baseflow and rainfall for quaternary catchments A21A - A21H.

While there is still baseflow in winter, the wastewater component dominates the flow; this is further justified by the apparent baseflow of 2003. It is mentioned above that 2003 had the lowest rainfall yet the apparent baseflow is not significantly different from the previous year(s), a plausible explanation is that the flow is predominantly wastewater.

Table 8 shows the original baseflow values obtained from Timeplot along with the naturalised baseflow estimates, which were calculated as the difference between baseflow and total wastewater discharge. Total wastewater discharge contributes more than 50% of total streamflow. Hence wastewater is bound to play a role in the estimation of baseflow. The mean annual baseflow estimates along with the baseflow percentage of rainfall for each quaternary catchment is summarised in Table 8. The mean annual baseflow estimates range between 6.7 and 108.41 mm/yr. It should be noted that the baseflow estimates given represent the minimum amount of recharge as it doesn't take into account losses incurred such as transmission and evapotranspiration losses (Risser et al. 2005).

Table 8: Summarised results from timeplot for the hydrological period of 1998 – 2003.

Catchment	Area (km2)	Runoff (mcm)	Total WWD (mcm)	BF without WWD (mcm)	MABF (mm)	MAR (mm)	BF % of MAR
A21A	451	6.4	12.8	3	6.7	616.8	1.1
A21B	527	29.1	39.4	19.2	36.5	616.8	5.9
A21C	761	72.1	80.4	24.9	32.32	655.2	4.9
A21D	372	4.6	9.13	12.9	34.76	542.1	6.4
A21E	290	17.5	14.6	20.8	71.77	655.2	11.0
A21F	1001	20.8	0.4	25.2	25.2	542.1	4.6
A21G	160	3.3	-	17.3	108.41	542.1	20.0
UCC	3593	153.8	156.7	123.2	34.29	603.1	5.8

BF = Baseflow WWD = wastewater discharge MABF = Mean annual baseflow

MAR = Mean annual rainfall mcm = million cubic meters

UCC = Upper Crocodile catchment

Figure 17 shows the annual baseflow for the duration of the study for each catchment. It can be seen that the amount of baseflow is controlled by rainfall to an extent, where the year with the highest amount of baseflow coincides with the wettest year, and the same applies for the driest year. Dolomites are known to be highly permeable and can accommodate large amounts of groundwater, hence, constant groundwater discharge takes place.

The differing baseflow estimates across the catchment can be attributed to the influence of the quaternary catchment characteristics such as geology, topography, land cover and the distribution of rainfall (Zhang et al. 2013). Catchments with similar geology (Figure 5), land cover (Figure 14) and physiography (Figure 3) may have different baseflow estimates (Queener & Stubblefield 2016) as a result of other recharge processes influencing groundwater recharge.

Assuming that baseflow is equal to recharge over a long-term period and wastewater discharge has been accounted for then catchment A21G has the highest recharge of 20% and catchment A21A has the lowest recharge of 1.1%.

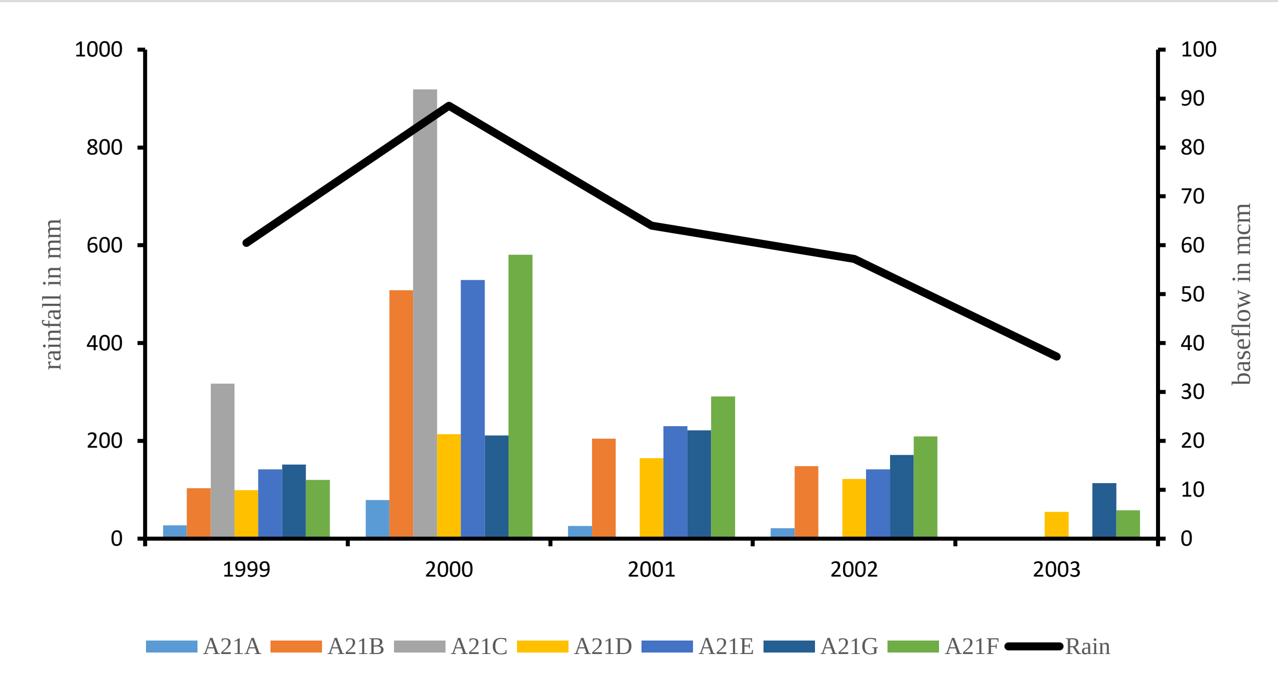

Figure 17: Mean annual baseflow of the quaternary catchments, A21A - A21G for the hydrological period of 1998 – 2003.

Quaternary catchment A21A has an unexpectedly low recharge rate of 1.1% for the five year period; this is surprising considering that a large area of it is underlain by dolomites (Figure 5) and the amount of rainfall it receives is substantial. An explanation for this could be that the wastewater discharge coming from Hartbeesfontein WWTW (Figure 1) has been overestimated and as a consequence baseflow (recharge) has been underestimated.

Catchment A21G is the smallest catchment (Figure 5) and yet has the largest recharge estimate of 20% and the lowest runoff volume of 3.3 mcm; this can be attributed to the underlying dolomites which are known to be highly permeable and can accommodate large groundwater storage. Therefore, the higher recharge in dolomitic terrains results in low surface runoff (Table 8) (Abiye et al. 2011). There is also a large spring, the Ngosi spring (Figure 18) located within the catchment with a discharge of approximately 100 l/s (Abiye 2015), the spring discharge makes its way to the closest stream and becomes a component of streamflow. Like baseflow, spring discharge is equal to recharge over a long-term period. Therefore, the presence of the spring must be responsible for the high recharge in catchment A21G. Based on the small surface area, of catchment A21G, and the large flows of the spring it can be inferred that the surface catchment of A21G differs from the groundwater catchment. To justify this, a study conducted by Abiye (2011) and Abiye et al. (2011) revealed that the spring has an O and H isotopic composition of -5.42 and -31.0‰, respectively and ^{3}H value of 0.6. The low tritium value suggests that the spring water is older than fifty years and therefore must have been in circulation for a very long time. The depleted isotopic composition of the spring water indicates it was a part of a deeper more regional circulation (Abiye 2011; Abiye et al. 2011).

Catchments with a similar geology and physiography such as A21C and A21E (Figure 3 & Figure 5) would be expected to have similar recharge values but that it not the case. The above could be due to hydrogeological differences such as the extent of the weathered zone or the degree and the connectedness of fractures, etc (Risser et al. 2005). A more plausible explanation is that baseflow in catchment A21E is inflated by the water transfers received from the Upper Vaal catchment. More than 370 x 10^6 m^3/yr of water is transferred to the Crocodile River, which flows through catchment A21E, this water is mostly for domestic and industrial use in the Johannesburg area (DWAF 2004).

Recharge for catchment A21C (4.9%) is not surprising considering that it is underlain by a crystalline basement consisting of granitic and gneissic rocks (Figure 5) which are known for their low porosity.

For the years 2001, 2002 and 2003 no recharge occurs for A21C, a few explanations can be given to account for this. Firstly, just like A21A, the wastewater discharge from the two WWTW (Figure 1) in quaternary catchment A21C may have been overestimated translating to baseflow being underestimated.

Figure 18: Ngosi spring issuing on dolomitic rocks in quaternary catchment A21G (Photo by Tamiru Abiye).

The effects of this would be exacerbated during the low rainfall period. Secondly, recharge in fractured crystalline rocks occurs mostly in the weathered zone meaning storage for recharge is limited. During periods of low rainfall like in 2002 and 2003, suggesting drought conditions, which is most likely accompanied by high ET rates, the little recharge stored in the weathered horizons of the basement rock will be subjected to evaporation, hence there is no recharge. In addition to that no baseflow from the underlying aquifers will be contributed to streamflow because the Jukskei River is underlain by low porosity crystalline rocks (Figure 5) thus there will be little to no surface water – groundwater interaction. It can thus be inferred that streamflow is mostly a combination of runoff (during the wet months) and wastewater discharge with baseflow having very little contribution in catchment A21C.

The mean annual baseflow estimate for the entire study area is given as 5.8% of rainfall. The recharge value of 5.8% is in close agreement with recharge values given for semi-arid regions in Southern Africa and for large scale catchments which are 1 – 5% of rainfall (Gieske 1992; Scanlon et al. 2006; Abiye 2016).

Figure 14 shows the land cover of the Upper Crocodile catchment, a large majority of the catchment area is highly urbanised, south of the catchment there are industrial and urban areas and west and east of the catchment the areas are mostly residential.

What this implies is that in the urbanised areas the surfaces will be impervious due to roads and pavements, such conditions will facilitate runoff and inhibit direct recharge as only a little, or no infiltration can occur.

From Table 8 and Figure 17, it can be deduced that recharge has a spatial and temporal variability across the catchment. It appears that the main processes influencing the occurrence of recharge include climate, sewage, geology and land cover.

A similar BFS study conducted by Abiye (2016) for a quaternary catchment located in the city of Johannesburg, which is predominantly underlain by crystalline basement rocks, yielded a baseflow estimate of 27.1% of rainfall. The study attributed the overestimation of the baseflow to the presence of wastewater discharge in the catchment area. The above highlights the importance of accounting for effluent discharge and to question the reliability of the baseflow separation method if additional input sources are not considered.

5.3 WATER TABLE FLUCTUATION METHOD

The WTF method was one of four methods used to estimate recharge for the study area; it requires groundwater level time series data and specific yield. The specific yield values were obtained from literature for a karst aquifer in the Malmani dolomite, south of the study area. During the dewatering of an aquifer, the authors observed a relationship between specific yield and depth where specific yield decreased with increasing depth, illustrating the complex nature of dolomitic aquifers. The specific yield values used for the computation of recharge for the study area are shown in Table 4. The locations of the boreholes are shown on Figure 19, the boreholes are distributed between catchments A21A, A21D and A21F, all located on the Malmani dolomities. Monthly groundwater levels from 43 different boreholes (BH) were used to quantify recharge for the hydrological period of 1991 – 1996 (Appendix B.1).

Groundwater level hydrographs were constructed for only six of the boreholes to represent the fluctuations (Figure 20). Figure 20 illustrates that there is variability in groundwater level fluctuations amongst the boreholes. The water table for BH’s 036353A, 36350 and 37772 shows the least amount of fluctuation with groundwater level changes of less than 3 m for the boreholes mentioned above. BH’s 36008 and 36349 have the highest fluctuations of groundwater levels with changes exceeding 20 m for BH 36349.

Distribution patterns of the rest of the boreholes in their respective catchments are similar variations of the ones seen in Figure 20 (Appendix B.2).

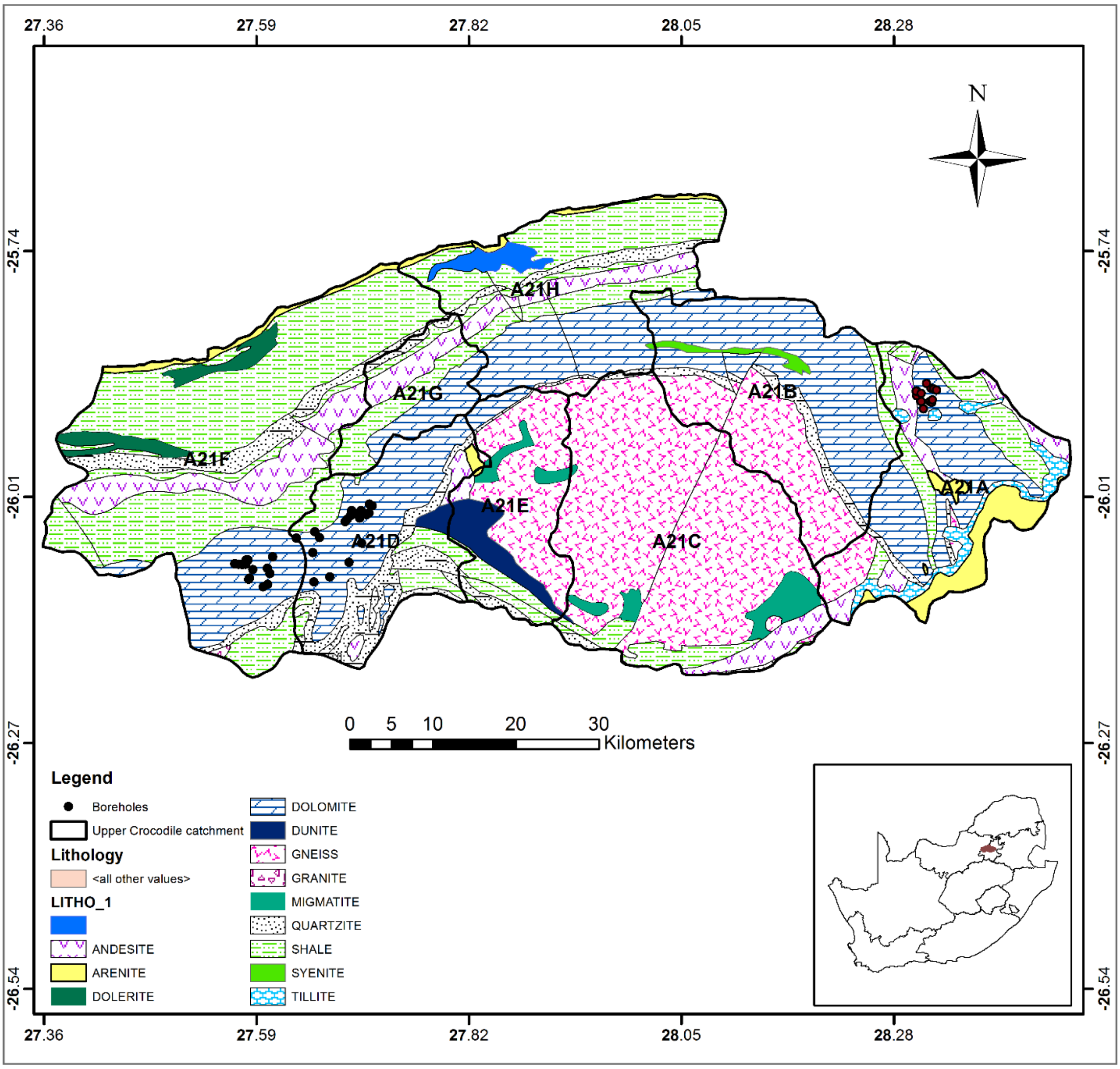

Figure 19: Location of the boreholes located on the Malmani dolomites.

A comparison of groundwater level fluctuations and rainfall variability for the same period shows there is a relationship between the two. The relationship may not be pronounced for all the boreholes, especially for the BHs with the lowest fluctuations, but groundwater level responses can be seen after rainfall events suggesting recharge will have a small seasonal influence (Figure 20). During the wet season, some BHs respond rapidly to the incoming rainfall while the others show more of a gradual response.

The rate of response of groundwater levels from rainfall events could be controlled by recharge mechanisms, geology and land cover.

A rapid response of groundwater levels after a rainfall event can be attributed to recharge occurring via preferential flow where recharge occurs through connected sinkholes and fractures, and a thin soil cover. Delayed responses of groundwater levels after a rainfall event suggest that recharge is direct and occurs through less permeable geological cover. The soil thickness and the soil moisture could also be responsible for the delayed response; a relatively thick soil cover could delay water flow through the unsaturated zone. If rainfall falls on dry soil, the infiltrating rainwater will first replenish the soil moisture before reaching the groundwater storage. In winter groundwater level fluctuations can still be observed even though rainfall is at a minimum, suggesting water levels respond to individual rainfall events with recharge occurring via preferential flow or it could be a result of other sources of recharge. Overall there is a large variability in groundwater level changes, which will translate to a large variability of groundwater recharge throughout the dolomitic aquifers.

The recharge for each borehole was calculated by applying equation 3. For the recharge calculations, 38 BHs were used (Table 9), the others were excluded because they were receiving induced recharge, possibly from the nearby surface water bodies that had a much greater influence on groundwater levels in comparison to rainfall. Because the WTF method assumes recharge is from rainfall, the other five boreholes were not included. Table 9 shows the mean annual water level rises of the 38 BHs along with recharge and the recharge percentage of the rain; the mean annual water level rises range from 0.7 to 4.3 m.

Recharge has a spatial variation across the dolomites with a minimum mean annual recharge of 17.5 mm/yr recorded at BH 36352 and a maximum mean annual recharge of 270.3 mm/yr being recorded at BH 36356. The mean annual recharge estimate for the Malmani dolomites, in the study area, was calculated to be 99 mm/yr, representing 14% of a mean annual rainfall of 676.8 mm for the hydrological period 1991-1996. The obtained recharge values are similar to studies conducted by Abiye (2016), Bredenkamp (1988), Leskiewicz (1984) and Bredenkamp et al. (1986).

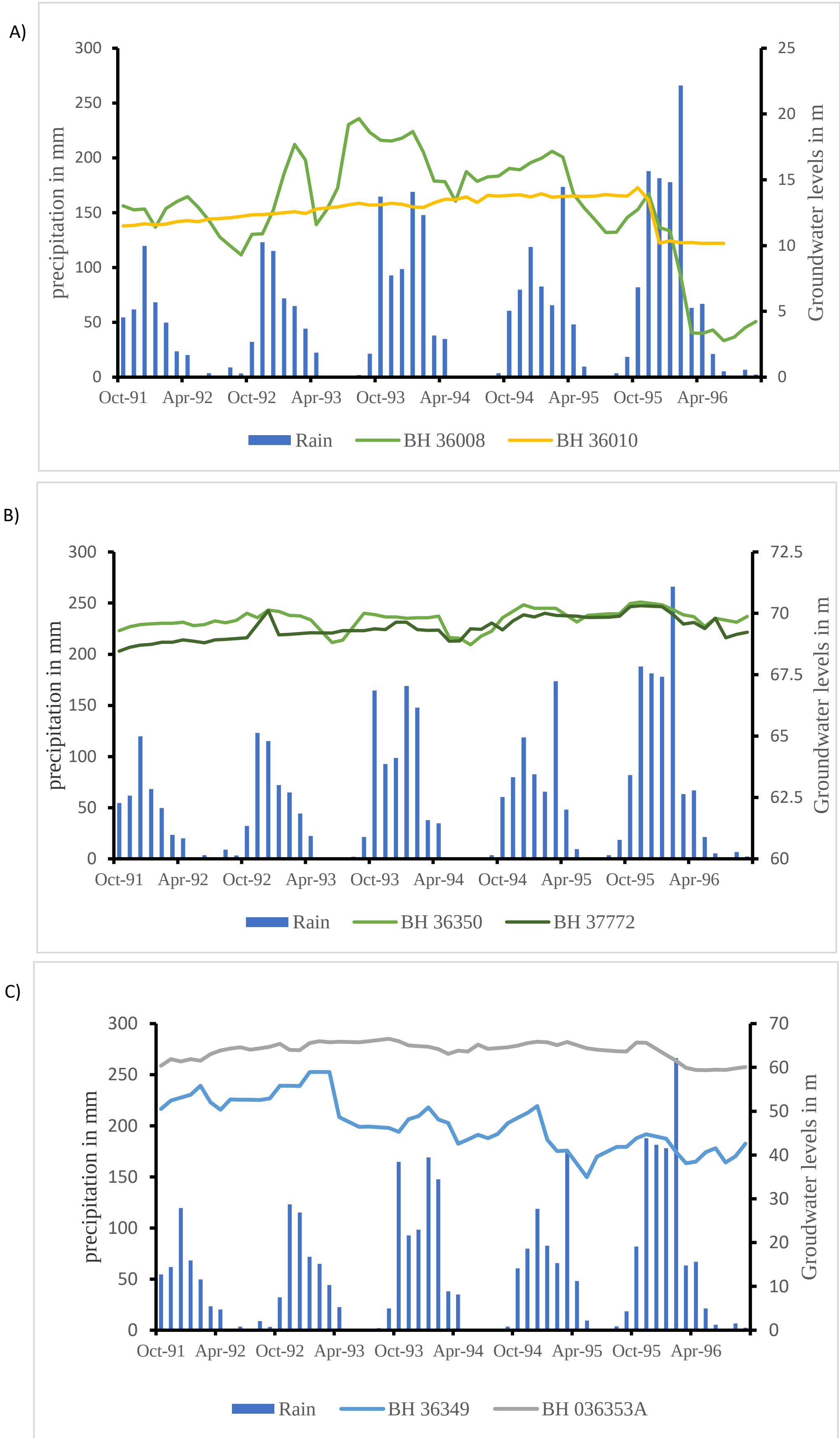

Figure 20: Groundwater level fluctuations and rainfall for quaternary catchments: A) A21A, B) A21D AND C) A21F.

Table 9: The water table fluctuation method results for the hydrological year 1992 – 1996.

BOREHOLES	Latitude	Longitude	Δh(m)	Sy	MAR(mm)	Recharge % of MAR
35727	-25.8937	28.30019	1.1	0.091	102.6	15.2
35730	-25.9131	28.30749	1.3	0.091	119.6	17.7
36003	-25.8987	28.30042	2	0.091	180.9	26.7
36007	-25.8906	28.31569	0.9	0.091	84.3	12.4
36010	-25.8917	28.31859	1	0.091	89.5	13.2
36020	-25.8939	28.29971	1.2	0.091	108.7	16.1
36051	-25.8964	28.30513	1.9	0.091	171.3	25.3
36056	-25.9053	28.30511	1	0.091	87	12.9
36059	-25.8857	28.31073	2.1	0.091	188.2	27.8
36063	-25.8929	28.32166	1.4	0.091	125.7	18.6
37794	-26.0179	27.71107	2.7	0.091	245.6	36.3
37788	-26.026	27.69096	1.1	0.091	99.6	14.7
37786	-26.0277	27.70514	2.1	0.091	192.5	28.4
36356	-26.0948	27.66543	3	0.091	270.3	39.9
36342	-26.0523	27.65417	2.6	0.091	232.8	34.4
36337	-26.0792	27.68627	0.7	0.091	62.7	9.3
36341	-26.046	27.64915	0.7	0.091	68.1	10.1
36334	-26.0584	27.69968	1.4	0.091	126.6	18.7
36320	-26.0163	27.70823	2.1	0.091	194.6	28.8
36601	-26.0764	27.57467	1	0.091	93.5	13.8
36602	-26.0766	27.57652	1.1	0.091	103.5	15.3
36603	-26.0771	27.5765	1.4	0.091	128.6	19

37775	-26.0813	27.57435	0.7	0.091	64.2	9.5
37779	-26.0871	27.58198	0.8	0.091	74.6	11
37792	-26.0325	27.6843	0.9	0.055	46.8	6.9
37789	-26.0288	27.68757	0.9	0.055	47.9	7.1
37785	-26.0257	27.7078	0.9	0.055	47.1	7
37784	-26.103	27.59814	2	0.055	108.4	16
37783	-26.1058	27.59323	1.1	0.055	58	8.6
036353A	-26.0687	27.64697	3.5	0.055	190	28.1
36325	-26.035	27.68205	0.8	0.055	46.1	6.8
36350	-26.0733	27.60337	1.1	0.055	60.4	8.9
36599	-26.0978	27.57742	0.8	0.055	46	6.8
37772	-26.0809	27.56259	1	0.055	55.7	8.2
37773	-26.0812	27.56661	1.5	0.055	82.7	12.2
37774	-26.0818	27.57058	0.7	0.055	36.7	5.4
37782	-26.0963	27.57831	1.7	0.055	91	13.4
36343	-26.053	27.62917	3.4	0.026	88	13
36324	-26.0313	27.69752	0.7	0.026	18.1	2.7
36322	-26.0231	27.69918	4.3	0.026	111.7	16.5
36321	-26.0231	27.68882	0.9	0.026	22.5	3.3
36352	-26.0913	27.60041	0.7	0.026	17.5	2.6
37781	-26.0855	27.59765	1.1	0.026	28.6	4.2

Annual rainfall varies throughout the study period, with a minimum of 413.5 mm for the year 1992 and a maximum of 1061.8 mm for the year 1996. Overall the annual rainfall appears to be increasing from 1992 to 1996.

The rainfall values for the years 1992 and 1993 were significantly lower than the MAR value of 676.8 mm, indicating possible drought conditions during that time. A look at Figure 21 shows the annual recharge plot against the annual rainfall for the period 1991-1996, from the graph the temporal and spatial variability of recharge is clear.

Apart from 1992, the total annual recharge reflects the same trend as rainfall amount, the exception being A21F where apart from 1992 recharge appears to decrease with increasing rainfall.

A possible explanation is the dolomites to the west of the area (A21F) are extensively used for irrigation to support the agricultural industry (DWAF 2008; Pietersen et al. 2011) therefore it is possible that the rate of abstraction of groundwater has exceeded recharge hence the annual recharge decline.

The variability in groundwater fluctuations amongst the boreholes can be a result of the fracture systems intercepting the boreholes and/or the complex nature of karst aquifers (Risser at al. 2005). The hydrogeological properties of the aquifer such as transmissivity, specific yield and hydraulic conductivity along with recharge mechanisms also play a role in the variability of water level fluctuations.

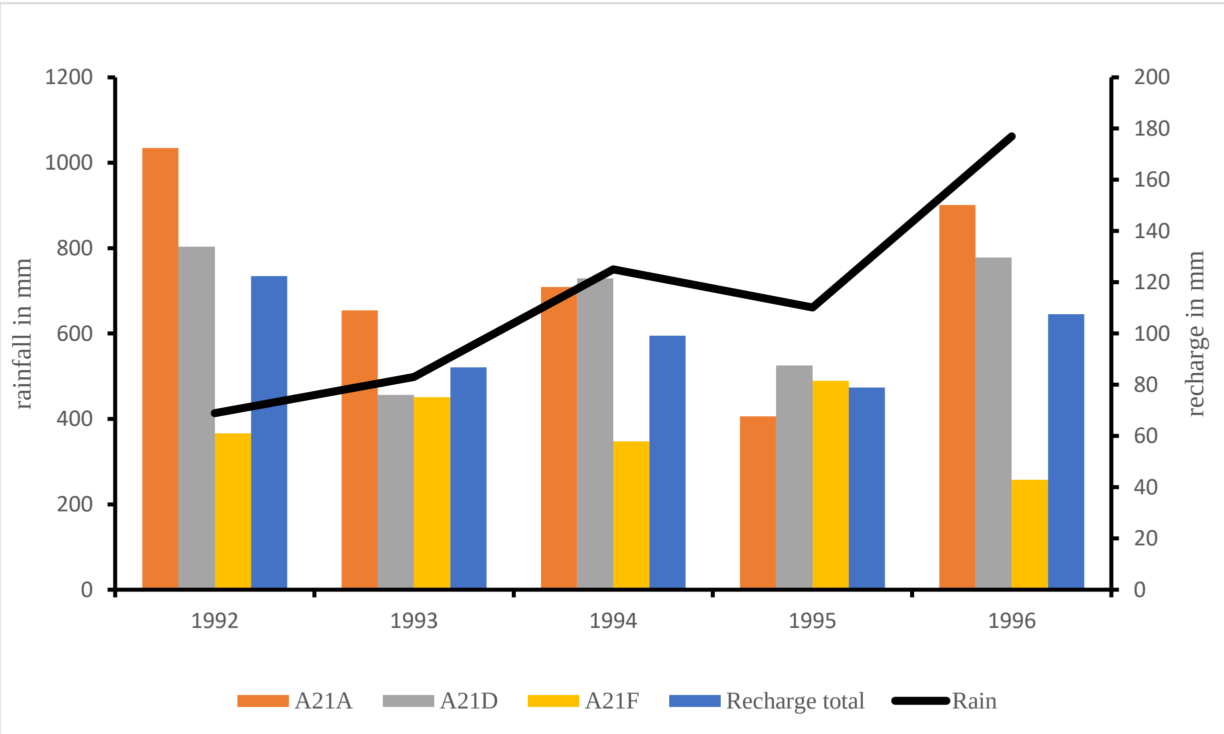

Figure 21: Annual groundwater recharge and rainfall of quaternary catchments A21A, A21D and A21F for the hydrological period of 1992 – 1996.

Recharge occurring through a connected network of fractures or sinkholes can have high transmissivities resulting in the rapid movement away from the water table thus groundwater level responses will be low (Risser et al. 2005; Somaratne 2014); this could explain why some of the boreholes show little response to rainfall.

A likely reason as to why most of the groundwater levels of the boreholes in catchment A21A show little response to rainfall can be attributed to spring water seepage through secondary structures, from the Rietvlei springs. Other possible causes of the larger groundwater level variability include the borehole proximity with respect to the nearby surface water bodies.

During the wet summer months, the surface water bodies receive rain which increases the surface water level, when this occurs the hydraulic head of the surface water will be greater than that of groundwater, and the surface water will recharge the groundwater. In winter this process is reversed, and groundwater will be discharged to surface water bodies. The spatio-temporal variability of rainfall is also responsible for groundwater level variability. Finally, large groundwater changes can be a result of missing data (Sibanda et al. 2009; Lutz et al. 2015).

The general cyclic nature of groundwater levels during summer and winter is because of groundwater recharge and discharge. If one considers the average groundwater level of the boreholes what can be seen is the overall groundwater level remains relatively consistent (Appendix B.1), with the exception of a few boreholes. The above could be due to the highly permeable nature of dolomitic aquifers or replenishment from a regional groundwater flow.

Figure 21 draws attention to the fact that recharge has a spatio-temporal variability across the dolomites, with catchments A21A, A21D and A21F exhibiting different recharge estimates in space and in time. The factors that control recharge include climate, hydrogeological characteristics of the aquifer, depth to water table, recharge mechanism, soil moisture and aquifer use.

5.4 ENVIRONMENTAL ISOTOPES

5.4.1 STABLE ISOTOPES

For the past three years (2013 – 2016) Professor Tamiru Abiye has been collecting rainfall in the Johannesburg region to analyse for the oxygen and deuterium isotopic composition. A LMWL was constructed for Johannesburg, using the collected data (Appendix C), it has a regression line of $\delta D = 6.4\ \delta^{18}O + 8.5$ ‰. The majority of the rainfall points cluster around the LMWL and the rest either fall above or below it. Samples falling below the LMWL are isotopically enriched indicating that rainfall was subjected to evaporation processes prior to recharge while the isotopically depleted samples falling above the line suggest that there was low humidity in the vapour during rainfall.

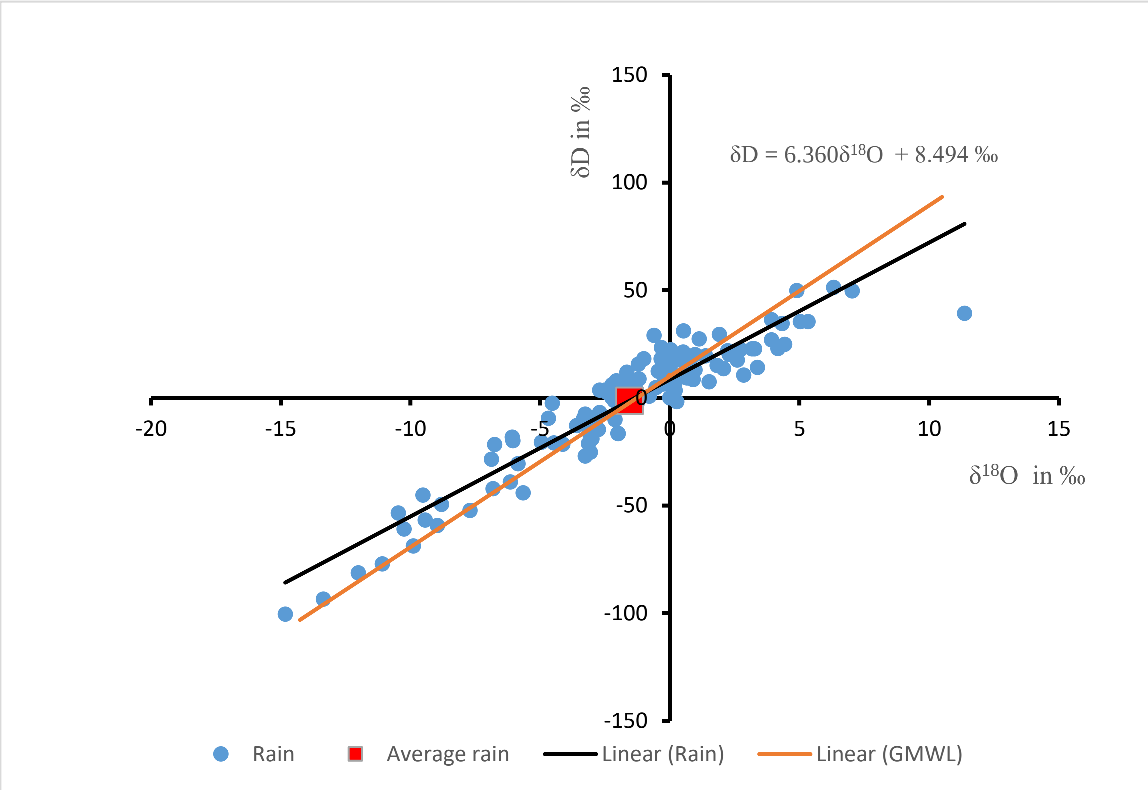

Figure 22: The δD vs $\delta^{18}O$ distribution in the rainfall of the Johannesburg area.

Figure 22 shows the O and H isotopic composition of all the rainfall events sampled since 2013 (Appendix C). The average isotopic composition of the Johannesburg rain for $\delta^{18}O$ and δD is -1.55 and -1.35‰, respectively, while the range of $\delta^{18}O$ and δD is -14.83 to 11.36 ‰ and -100.42 to 51.36‰, respectively.

The highly depleted isotopic composition of rain can be attributed to the rainout effect and consequently the altitude effect. Alternatively, the isotopically depleted rain samples could indicate recharge occurring from colder winter months, from an isotopically depleted vapour mass.

Table 10: Stable environmental isotope results.

ID	Date	Latitude	Longitude	$\delta^{18}O$	δD	d-excess
AF1 (SPRING)	06 Jun 2016	-25.1553	27.97026	-5.27	-19.29	14.23
AF2 (SPRING)	06 Jul 2016	-25.1553	27.97026	-6.01	-21.6	16.62
AF3 (SPRING)	06 Aug 2016	-25.1553	27.97026	-4.3	-19.08	8.27
AF4 (SPRING)	06 Sep 2016	-25.1553	27.97026	-2.86	-14.5	3.69
AF5 (SPRING)	06 Oct 2016	-25.1553	27.97026	-3.79	-13.71	10.39
AF6 (SPRING)	07 Nov 2016	-25.1553	27.97026	-3.6	-14.3	8.41
AVERAGE SPRING WATER						10.27
MIN RAINFALL				-14.83	-100.42	-32.93
MAX RAINFALL				11.36	51.36	32.898
AVERAGE RAINFALL				-1.55	-1.35	8.49

The enriched isotopic compositions indicate that the rain was subjected to evaporation processes during a rainfall event or before the rain sample was collected under dry and warm conditions. Isotopically enriched samples are dominant during months of sparce rainfall or during rain events of low inensity and low humidity (Dansgaard 1964).

D-excess is a useful parameter for determining the source region of air masses and it also reflects fractionation processes taking place as the vapour mass evolves (Dansgaard 1964; Froehlich et al. 2002; Hoefs 2009). The rainwater samples show a broad variation of d-excess, indicative of variable air mass sources (Figure 23). The d-excess ranges from a minimum of -32.93‰ to a maximum of 32.9‰, the widespread variation suggests that rainfall is influenced by both a local and regional moisture source.

The main parameters responsible for the variation in d-excess include relative humidity and temperature. The lower d-excess values can be attributed to sub-cloud evaporation, driven by kinetic isotope fractionation, under low relative humidity and high temperature conditions (Clark & Fritz 1997; Hoefs 2007). The low d-excess of rain samples are derived from a regional air circulation, formed from a maritime source. Rain with higher d-excess values can be interpreted as rainfall that originated from a local interior moisture source, formed under low temperature and high humidity conditions.

Figure 23 assesses the relationship between d-excess and $\delta^{18}O$. Enriched rainfall samples with a low d-excess indicate that rain was close to the moisture source (oceanic air mass), rainfall was subjected to sub-cloud evaporation, hence enrichment. Isotopically depleted rain samples with high d-excess are influenced by the rainout effect, driven by decreasing temperatures. It is responsible for the progressive depletion of isotopes as the air mass is transported away from oceanic sources towards the interior.

Variations in d-excess are caused by different origins of the air masses. Relative humidity, temperature, vapour pressure and fractionation processes taking place below the cloud base like evaporation and condensation also play a role in the variability of d-excess (Clark & Fritz 1997).

The Alberts Farm spring (Figure 24; Figure 26) issuing on the contact between the quartzite and the shale can be found south of the study area; its location is given in Table 10. Figure 25 shows the isotopic composition of the spring water samples which were sampled monthly between June and November.

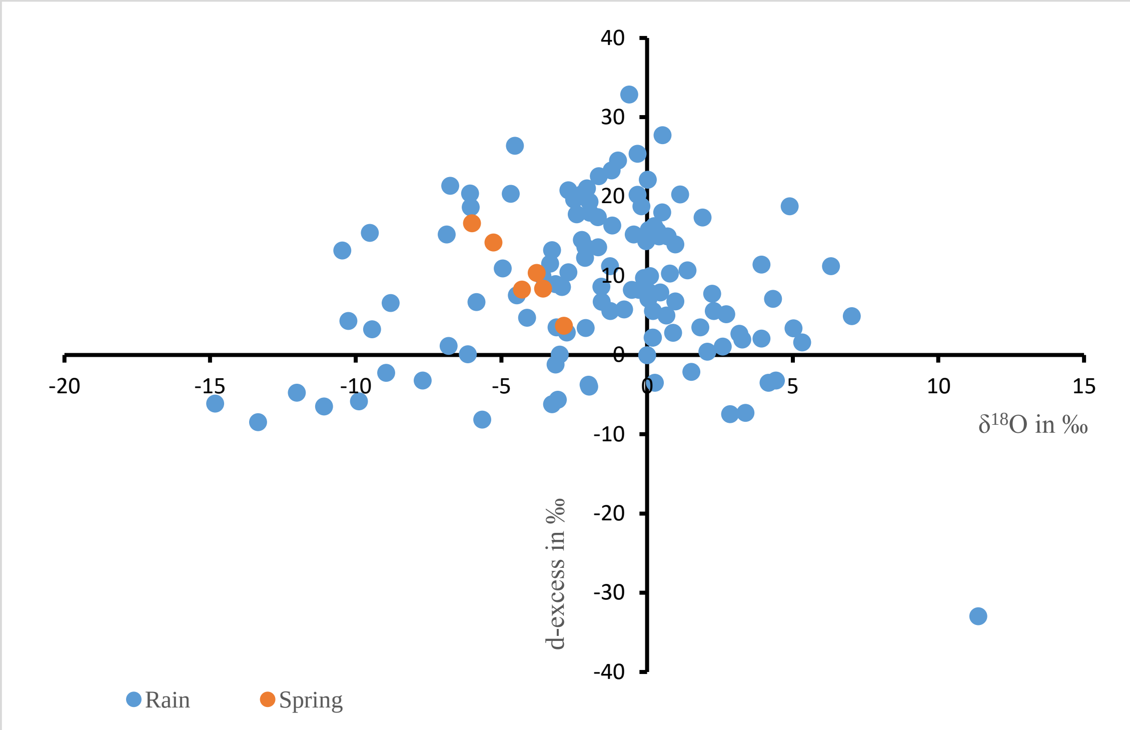

Figure 23: d-excess vs δ^{18}O for the rain and spring water samples within the study area.

The isotopic composition of the spring appears to be depleted, with isotopic compositions ranging between -6.01 and -2.86‰ for δ^{18}O and a δD minimum and maximum of -21.6 and -13.71‰, respectively (Table 10). The depleted isotopic compositions of the spring water samples could be indicative of recharge that took place at higher altitudes, owing to the rainout effect. Additionally, the depleted isotopic signatures can be interpreted as the spring receiving water from deep circulating groundwater. Alternatively, spring water samples with highly depleted isotopic signatures can be interpreted as the spring being recharged by rainwater from a colder climate, derived from an isotopically depleted air mass. The less depleted spring water isotopic ratios were recharged by rainfall from a warmer season, originating from an isotopically enriched vapour mass. The isotopic variation of the spring water samples suggests that the rainwater recharging the spring was influenced by a seasonal effect.

Figure 23 shows the d-excess distribution of the spring which ranges between 3.69‰ and 16.62‰. The variation in d-excess suggests that the recharging waters originate from variable moisture sources, the d-excess values at the lower end of the spectrum indicate that the rainfall that recharged the spring derived from a regional moisture source, that formed under conditions of high temperature and low humidity. And samples with high-d-excess values were recharged by rainwater originating from local moisture sources.

The spring has a seasonal distribution of d-excess, samples AF3 – AF4 present enriched isotopic compositions with lower d-excess values typical of summer recharging rainwater that has been subjected to evaporation. Samples AF1 and AF2 are isotopically depleted and have higher d-excess values, similar to those expected of winter rainfall. The study area is controlled by two climate systems, the Subtropical HP (SBHP) system in winter and the ITCZ in summer; these two systems are responsible for the seasonal variation of isotopic compositions and consequently d-excess. In winter, the SBHP system is responsible for the cold westerly winds carrying isotopically depleted air masses from the polar regions an in summer the ITCZ is responsible for bringing isotopically enriched maritime tropical air masses.

An isotopic comparison of the average rainfall and the spring water samples demonstrated that the isotopic composition of spring water doesn't match that of recent rainfall indicating that recent rainwater is not the primary source of the spring. It can therefore be assumed that the spring is primarily recharged by older rainfall originating from higher altitudes that has been a part of a deeper regional circulation. According to the average d-excess of the spring water, 10.27‰, and rainfall, 8.50‰, it shows that the dominant moisture source originates from a regional circulation with the local recycled continental air masses having a limited role.

Figure 24: Alberts Farm spring.

Spring water samples falling below the LMWL can be used for the Allison et al. (1984) isotopic shift method, where the vertical and horizontal isotopic shift between the water sample and the LMWL can be used to estimate mean monthly recharge. Using the spring sample AF4 recharge was computed as 10.19 and 23.90 mm/month for the $\delta^{18}O$ and δD isotopic shift, respectively. These values represent recharge that occurred in the fractured aquifer(s) of the Witwatersrand quartzitic rocks. The quartzites, south of the study area, represent the highest point of the catchment area therefore locally the spring is being replenished by mountain front recharge. The depleted isotopic signatures of the spring suggest that the spring water was a part of a deeper regional circulation, inferring that recharge is taking place via preferential flow through a series of faults and fractures in the quartzites.

Since the spring was not sampled throughout the year annual recharge cannot be inferred hence recharge was calculated as a monthly estimate.

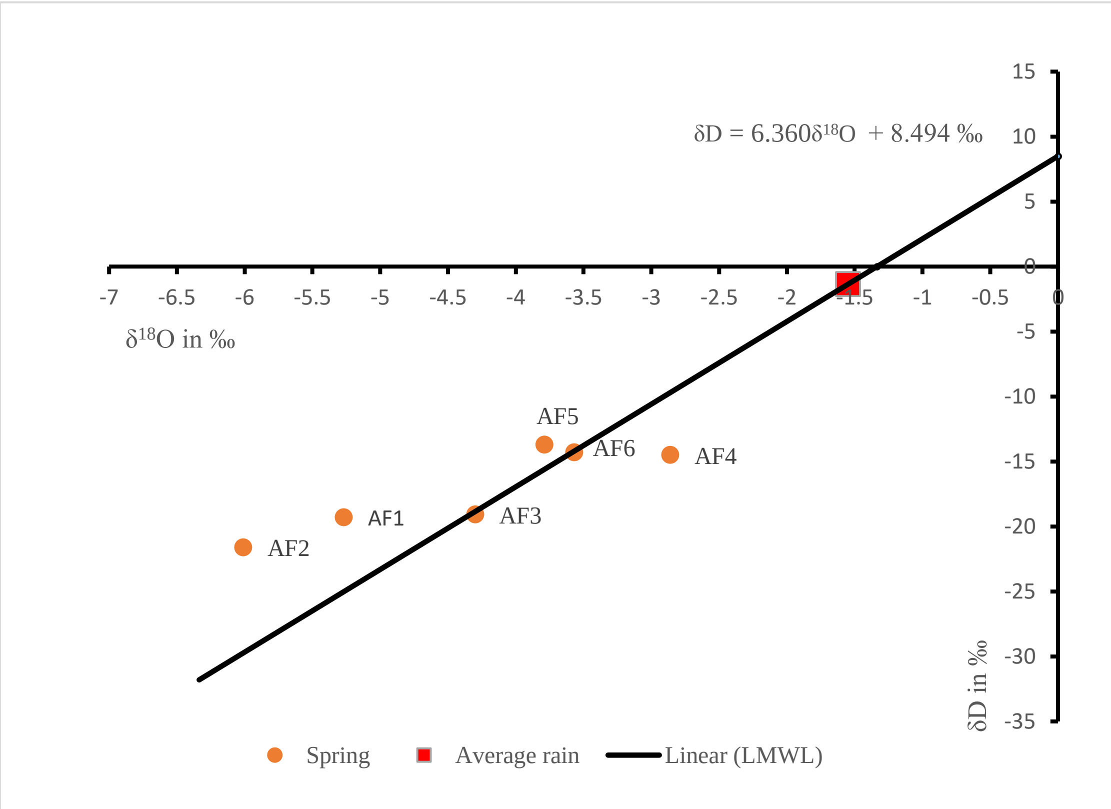

Figure 25: δD VS $\delta^{18}O$ for Alberts Farm spring.

5.4.2 TRITIUM RESULTS

Tritium is useful in groundwater studies as it can provide the residence time of groundwater. Residence time can be easily calculated provided that the input function of tritium is known.

Eight water samples were collected for the analysis of tritium on 04/06/2016 (during winter) (Figure 26), four were collected on the Braamfontein Spruit, three on the Jukskei River and one on the Crocodile River. The water samples were collected in winter because that is when the rivers are sustained by baseflow, which is representative of groundwater.

The tritium results are summarised in Table 11. Seven of the eight samples (S1 – S7) all had tritium units falling in the range of 0.8 – 4TU suggesting that the recharge waters are a mixture of sub-modern and recent recharge.

Table 11: A summary of tritium results for the streamwater samples.

SAMPLE NAME	Latitude	Longitude	TU	Residence time
S1	-26.16333	27.99969	2.3 ± 0.3	16
S2	-26.14861	27.99814	2.5 ± 0.3	14
S3	-26.1375	28.01114	2.1 ± 0.3	18
S4	-26.11361	28.01969	2.6 ± 0.3	14
S5	-25.98086	28.01589	3.2 ± 0.3	10
S6	-25.97978	28.01572	3.2 ± 04	10
S7	-25.91167	27.94719	3.6 ± 0.4	8
S8	-25.90558	27.93472	4.8 ± 0.4	3

Figure 27 shows tritium values increasing with increasing distance, from that relationship the water samples can be grouped into three types of water.

Type water 1 is characteristic of relatively deep circulating old water, type water 2 is a mixture of old deeper circulating water with more recent rainwater and type water 3 is contaminated water by rain or other sources. Type 1 water includes samples S1 to S4, which were collected from the Braamfontein Spruit these samples have the lowest tritium values and subsequently the highest residence times, in aquifers before discharging, with sample S3 having the highest residence time of eighteen years (Table 11).

Type 2 water includes samples S5 – S7 which were sampled from the Jukskei River these samples have higher tritium values and a lower residence time of eight years. Sample S8 falls under type 3 water; it was sampled from the Crocodile River. Sample S8 appears to be an outlier; it is the only sample with a tritium value closer to that of rainfall (input function 5.6 TU).

The samples with a longer residence time could be from waters that were part of a deeper circulation suggesting that the recharge is not from current rainfall and that the aquifer in which the groundwater was flowing through has a low permeability, thus recharge was direct. Samples S5,6 and 7 have a shorter residence time, a possible explanation could be that older deeper circulating water is mixing with recent rainwater and or a local source along its flow path hence the tritium values are slightly higher. The higher tritium values of type 2 water could be indicative of preferential (indirect) recharge.

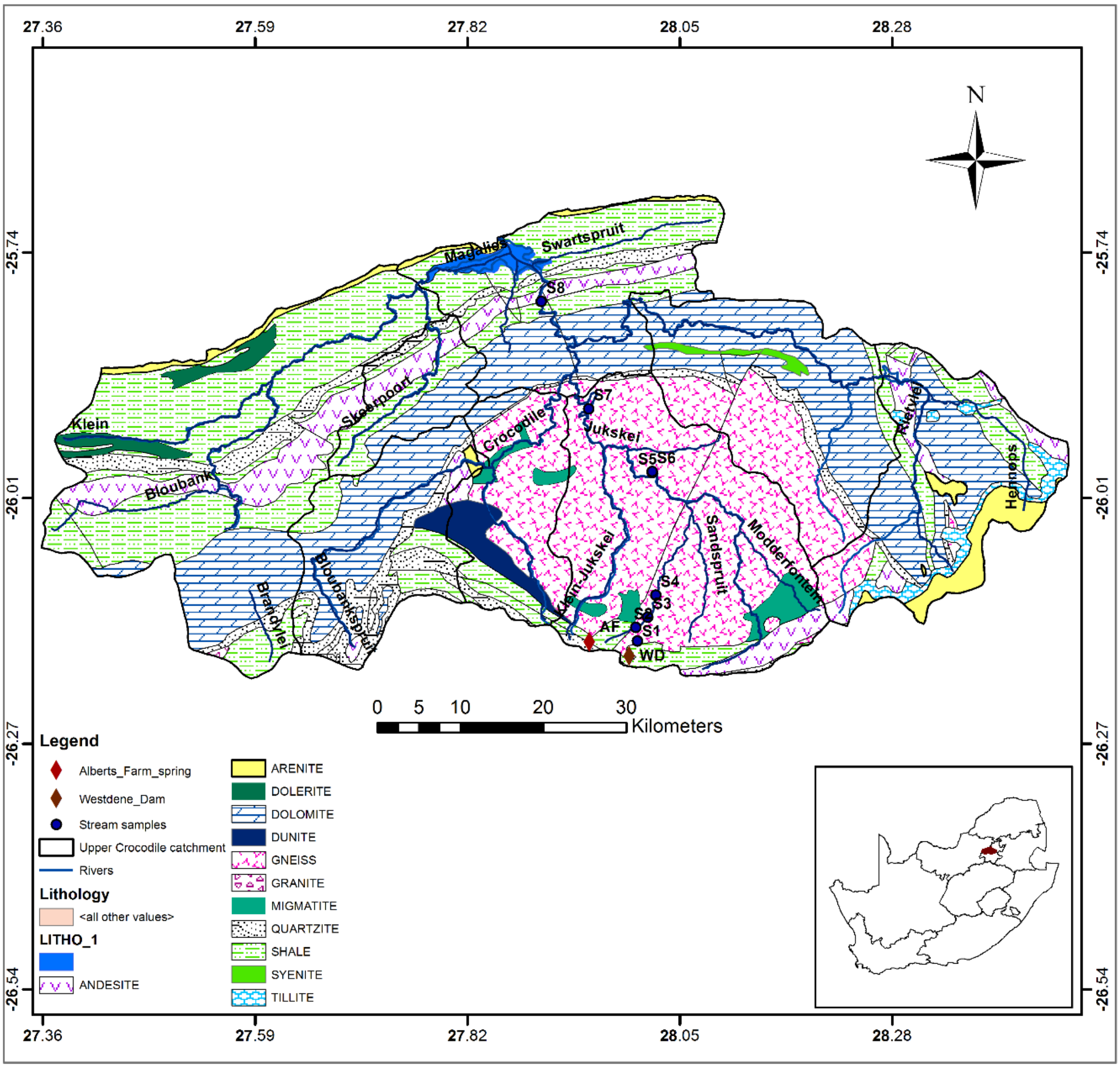

Figure 26: Location of the stream water samples, the Alberts Farm spring and Westdene Dam.

It is important to note that the residence time of the water samples reflects the amount of time the groundwater has been circulating for before joining the rivers as baseflow.

Possible sources for the collected samples are the Alberts Farm spring and the Westdene Dam (Figure 26). The spring and the water from the dam flow into the Braamfontein Spruit, which eventually merges with the Jukskei River where samples S5-S7 are located. Samples S1 – S4 would have lower tritium units because they are closer to the source while samples S5 – S7 have higher tritium units because of mixing between older waters with more recent rainwater along the way. Considering that the entire flow path of the Jukskei River is over igneous rocks, it can be assumed that recharge or baseflow would be of relatively recent rainwater from the fractured aquifers of the basement rocks.

The above would validate the tritium results for samples S5 - S7 and the theory of mixing of older waters from the springs with the more recent rainwaters or a local source encountered along the way.

Sample S8 has the highest tritium unit and the lowest residence time of 3 years. One of two explanations can be given, either sample S8 is from water that has been recharged by very recent rain, or it is from water that has been contaminated by a local industrial source. The latter is more plausible as sample S8 was sampled not too far away from Pelindaba, a nuclear research centre. Abiye (2015) also had samples from the Crocodile River and Hartbeespoort Dam exhibiting high tritium units which he attributed to contamination from a local industrial source.

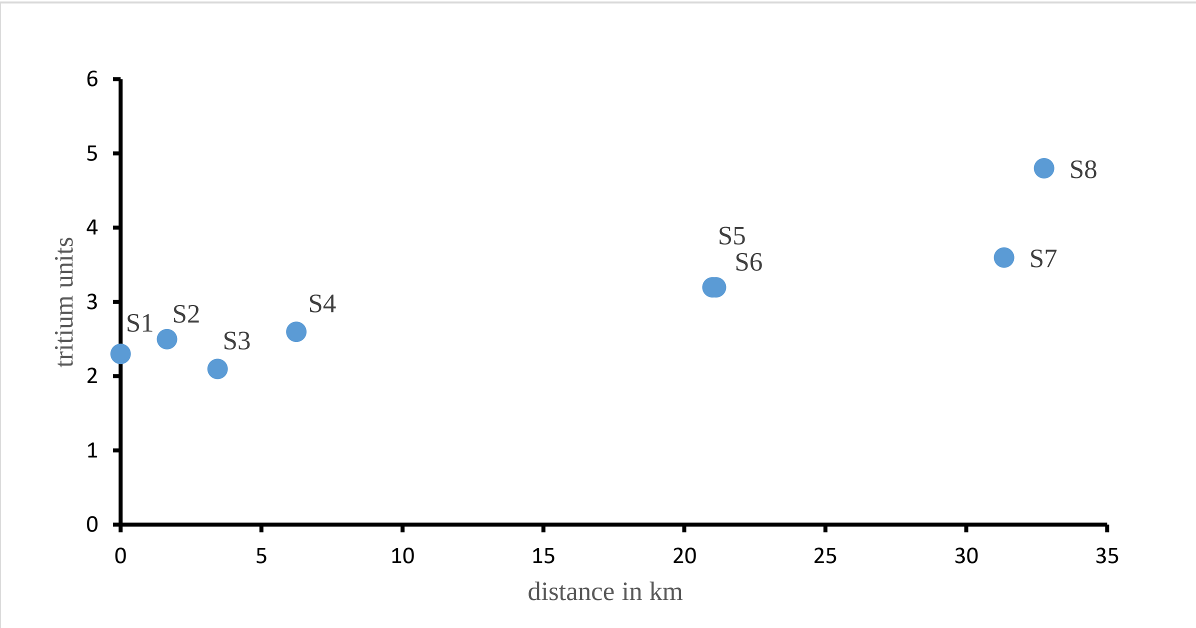

Figure 27: Tritium units vs distance for stream water samples.

6 SYNTHESIS

6.1 COMPARISION OF RECHARGE ESTIMATES

The mean annual estimates of recharge and baseflow have been summarised in Table 12. The quantitative recharge methods by large were found to give reasonable recharge values with the exception of the BFS method which was found to overestimate and underestimate baseflow (groundwater recharge) for some quaternary catchments. Mean annual recharge for the Upper Crocodile catchment was given as 4% and 5.8% of MAR by the WB and BFS methods, respectively. The slightly elevated recharge value for the BFS method can be attributed to the exclusion of quaternary catchment A21H in the recharge calculations. Overall the recharge amounts calculated for the WB and BFS method are in agreement with each other. The recharge values can be confirmed by a study conducted by de Vries and von Hoyer (1988) who obtained a recharge estimate of 4% for a catchment with similar geological characteristics.

The BFS method was also used to calculate recharge for the fractured aquifers in the catchment. Groundwater recharge was estimated as 4.9, 11 and 4.6% for the quaternary catchments of A21C, A21E and A21F, respectively. Recharge for A21C and A21F are in close agreement with each other. The higher recharge of 11% for A21E can be explained by the water transfers into the Crocodile catchment, approximately 370 x 10^6 m^3/yr of water is transferred to the Crocodile River, flowing through A21E hence the overestimation of recharge. Apart from recharge calculated for A21E, the recharge estimates are comparable with a study conducted by Sibanda et al. (2007) who had recharge values of 2.7 – 3.6% for a fractured aquifer. The above recharge values differ to those obtained by Abiye (2016), the study calculated recharge as 14 and 27% using the WTF and BFS method, respectively, for a fractured aquifer. The extremely high recharge of 27% calculated from the BFS method was attributed to the high volume of wastewater entering the streams.

Using the WTF method mean annual groundwater recharge was calculated as 14% of MAR and the BFS method calculated recharge as 1.1, 6.4, 5.9 and 20% for quaternary catchments A21A, A21B, A21D and A21G. The recharge values calculated using the WTF method differ greatly from those calculated using the BFS method except for A21G. An explanation for the lower recharge values of the BFS method is that because the exact volume of wastewater discharge was unknown the design capacity volume had to be used.

The problem with using the design capacity volume is that plants often flow below this value thus when calculating recharge, by subtracting wastewater discharge from baseflow, the final baseflow amount is underestimated hence recharge is underestimated. Studies conducted by Abiye (2016), Bredenkamp (1988), Leskiewicz (1984), Bredenkamp et al. (1986) and Kuhn (1989) in the Malmani dolomite revealed recharge values of 17, 15 and 17, 12.5, 13.9 and 10.3%, respectively. The WTF method (14%) arrived at comparable recharge values as the abovementioned studies.

Regionally recharge calculated by the BFS method was reasonable but it should be noted that the BFS method can produce questionable recharge estimates if the additional input sources are unknown or are not properly accounted for.

The WTF method had the greatest mean annual recharge estimate and the WB method had the least. The reason being the WTF method was applied to the dolomitic terrain which is known for its high permeability as a result of dissolution cavities, sinkholes and conduit systems found in the dolomite. These structures promote rapid infiltration; hence, recharge was expected to be high in the dolomitic terrain. The BFS and the WB method have recharge estimates that are similar to one another, the above could be due to recharge estimates being calculated for a regional area rather than a local area.

Table 12: A summary of the methods used and their respective recharge estimates.

METHOD	Time period	Recharge(mm/yr)	MAR(mm/yr)	R% of MAR	MRT (years)
WATER BALANCE	1995-2004	28.2	709	4	
BASEFLOW	1998-2003	27.5	603	5.8	
WATERTABLE FLUCTUATION	1991-1996	97.74	676.8	14	
TRITIUM	2016				18 – 3
ISOTOPE SHIFT	2016	10.19–23.90 mm/month			
MRT = MEAN RESIDENCE TIME					

The mean annual recharge variations of the WB, BFS and WTF methods range from 0 to 9%, 1.1 to 20% and 2.6 to 39.4%, respectively. Although the time periods for each method are different, there is a commonality between them in the way they each respond to rainfall.

During periods of below average rainfall (dry periods) recharge estimates were low or no recharge occurred at all, this is seen for all the methods.

The years 2002 and 2003 are examples where this phenomenon occurred, for the WB method it shows that no recharge was recorded for those two years (Figure 14). For the BFS method the years 2002 and 2003 had the lowest recorded baseflow volumes. During high rainfall years (above average rainfall) the calculated recharge was the greatest for those years. The WB and BFS methods show that the highest recorded recharge estimates were for 1996 and 1997 and just 1996 for the WTF method.

All the quantitative methods show a general seasonal variation of recharge, where recharge or baseflow is higher in the wet summer months and lower in the dry winter months. It should be noted that the seasonal pattern is sometimes subdued for baseflow because of the wastewater contribution to streamflow which results in stream flow remaining relatively high even during winter.

6.2 LIMITATIONS OF THE RECHARGE METHODS

The primary objective of the research was to quantify recharge for the fractured crystalline and metasedimentary aquifers, and the karst aquifers. The water balance, the baseflow separation and the water table fluctuation methods were used to quantify recharge quantitatively while environmental isotopes were used to assess groundwater recharge qualitatively. Secondly, it was to identify and understand how the different recharge processes affect groundwater recharge.

6.2.1 WATER BALANCE METHOD

The water balance method considers different hydrological components, which are then used to calculate recharge. The mean annual recharge estimate calculated using the water balance is 4% of mean annual rainfall of 709 mm which is 28.36 mm/yr.

A conventional water balance model was used to estimate the areal recharge of the catchment, because the water balance method used only gives one value for recharge the spatial variation of recharge could not be assessed. The method is for the unsaturated zone and assumes recharge is direct therefore the water balance method is estimating potential recharge.

The main processes controlling recharge in the water balance method are rainfall, PET and soil moisture. For potential recharge to occur certain conditions must first be met; the rainfall must exceed PET, and the soil moisture must be at field capacity. Although in semi-arid regions this is more the exception than the rule as recharge commonly occurs even if rainfall is less than PET, in this case, recharge will occur from individual high-intensity rainfall events. Urban development is another factor that indirectly affects recharge, the built-up areas and the impervious surfaces will promote runoff thus reducing infiltration and subsequently recharge.

The limitation of using the TMWB model by McCabe & Markstrom (2007) is that it tends to underestimate recharge because of the use of monthly averaged hydrometeorological data. The use of daily values for the method could give better recharge estimates as they take into account individual rainfall events associated with recharge (Bredenkamp et al. 1995; Bakundukize et al. 2011). Another possible reason daily steps are preferred is that it is possible that a single rain event can exceed ET on a single day which would lead to recharge (Bakundikize et al. 2011). The overestimation of output values such as ET and runoff could also underestimate the recharge amount.

The accuracy of the method depends upon the accuracy of the computed components; large errors can arise from PET estimates depending on the method used to calculate PET. The Hamon method (Hamon 1963) is employed by the TMWB model to estimate PET; the Hamon method only uses the average number of daylight hours per day during the month, saturated vapour pressure and temperature to estimate PET. The level of uncertainty of the PET estimate could potentially be high because the method only uses a few parameters to estimate PET, the inaccurate PET will be carried throughout the calculation thus recharge estimates will be incorrect.

Another problem that indirectly affects the accuracy of recharge estimates is the lack of meteorological stations within the catchment area. Only four rainfall stations were used which are located in catchments A21A, A21C, A21F and A21H. The temperature data was obtained from three stations, two located within the catchment area and the third located outside the catchment area but in proximity to catchments A21F and A21H. The lack of accurate temperature data will further compound the uncertainty associated with PET.

Lack of rainfall stations within the catchment can lead to the underestimation or overestimation of mean monthly/annual rainfall, which can result in inaccurate estimates of recharge.

6.2.2 BASEFLOW SEPARATION METHOD

The baseflow separation method estimates baseflow (recharge equivalent) by separating the baseflow and the runoff components of streamflow. The method assumes steady conditions where groundwater discharge is equal to recharge over long periods, assuming any groundwater losses that occur are negligible. The BFS method gives an areal estimate representing potential recharge. The mean baseflow estimate for the entire catchment was 5.8% of 603 mm mean annual rainfall which is 35 mm/yr. The fractured meta-sedimentary aquifers of the Pretoria Group have a mean annual baseflow of 4.6% for catchment A21F. The mean baseflow for the karst aquifers of catchments A21A, A21B, A21D and A21G are 1.1, 5.9, 6.4 and 20% respectively. The fractured aquifers have a mean baseflow of 4.9 and 11% for catchments A21C and A21E, respectively.

The BFS method results show that recharge varies both spatially and temporally, processes responsible for recharge variability are rainfall, geology, land cover and sewage. The variability of rainfall is reflected in the baseflow estimates. Karst aquifers which are characterised by their high permeability and high storage capacity have higher recharge rates overall than those of fractured aquifers, which have low permeability. The land cover for the Upper Crocodile catchment shows that the study area is highly urbanised especially south of the catchment, the repercussions of highly urbanised areas are impervious surfaces that will facilitate runoff, which translates to decreased recharge.

In the study area, where eight WWTW can be found baseflow calculations are not so straight-forward. Because wastewater is continuously being discharged into the streams, the Timeplot program cannot differentiate between baseflow and wastewater, so it treats them as one. Thus, to estimate baseflow alone sewage must be subtracted from the baseflow estimate given by the program. For recharge to be estimated accurately the precise amount of the wastewater volume needs to be known. The BFS method may underestimate or overestimate baseflow in a case where the exact volume of wastewater is not known. Such is the case for Hartbeesfontein and Olifanstfontein WWTW where the design capacity flow was used because the average estimates were not known.

The problem with using the design capacity values is that the plant might be flowing well under the given design capacity. Therefore, when wastewater discharge is subtracted from the initial baseflow estimate, the result will be the underestimation of final baseflow.

The underestimation of baseflow is likely to affect catchments A21A and A21B, as the Hartbeesfontein and Olifantsfontein treatment works are located in catchments A21A and A21B, respectively and the streams receiving discharge from these plants pass through the abovementioned catchments.

The BFS method should be applied cautiously in catchments that have additional inputs that contribute to streamflow because it can yield unreliable estimates of baseflow if the additional flow is not properly accounted for.

6.2.3 WATER TABLE FLUCTUATION METHOD

The WTF method calculates recharge by assuming the increase in water level responses is due to a rainfall event. The WTF method calculates point estimates, that represent actual recharge, but with enough boreholes, spatial estimates can be obtained. A recharge estimate of 14% of the annual rainfall of 676.8 mm was calculated for the karst aquifers. Recharge varied spatially across the dolomites with catchment A21A having the greatest recharge of 17%, A21D had a recharge of 16%, and catchment A21F had the lowest recharge of 10%. Bredenkamp, (1988) also observed a similar pattern for recharge in the dolomites, with dolomites in the east having higher recharge than the dolomitic aquifers in the west.

The scatter of recharge estimates is a result of the processes that govern groundwater level fluctuations and consequently recharge. The hydrogeological properties of the aquifers have the greatest influence on recharge.

Other aspects include; induced recharge to groundwater level from nearby surface water bodies and or springs, inter-basin transfers, regional groundwater flow, and rainfall variability.

Problems encountered with the water table fluctuation method include obtaining the specific yield for the aquifer; no pumping test data was made available for the dolomitic aquifers in the study area. Therefore, literature values were used for specific yield. The use of literature values for specific yield is likely to introduce some level of error to the calculations which consequently translate to recharge estimates being inaccurate.

Induced recharge to groundwater levels occurs predominantly from springs (catchment A21A) and substantial surface water bodies that are in proximity to boreholes. The influence of surface water bodies on groundwater levels will result in the overestimation of recharge as not all recharge to groundwater is coming from rainfall.

Groundwater abstractions have the opposite effect; it results in lowered groundwater levels therefore when recharge is calculated it will be underestimated, that seems to be the case in catchment A21F. The lack of groundwater level time series data for certain periods might be responsible for the variability of groundwater levels and subsequently recharge, amongst the boreholes. Missing data also makes it difficult to do mean monthly comparisons and inter-site comparisons as the groundwater levels are not reflective of all the changes throughout the study period (Lutz et al. 2015).

6.2.4 ENVIRONMENTAL ISOTOPES

The use of environmental isotopes provided some useful information on recharge processes, possible origins, monthly recharge estimates and the time since recharge occurred. The distribution of δD and $\delta^{18}O$ of rainfall ranges between -14.83 to 11.36‰ and -100.42 to 51.36‰, respectively. The stable isotopes of rainfall vary because of climatic processes, altitude, temperature and the rainout effect. D-excess of rainfall is widespread indicating variable moisture sources therefore rainfall originates from both local and regional air masses. An average d-excess of 8.50‰ suggests dominance from a regional moisture source.

The spring water samples have oxygen isotopic compositions ranging between -6.01 and -2.86‰, the depleted isotopic ratios can be explained by the altitude effect, as a result of the rainout effect, whereby the springs are recharged by rainwater originating from higher altitudes.

The variation of the stable isotopic compositions of the spring water can be attributed to a seasonal effect whereby the spring is recharged by rain that was isotopically influenced by different seasons. The spring water has an average d-excess of 10.27‰ suggesting that recharge waters were derived from a regional oceanic moisture source.

A comparison of spring water and average rainfall isotopic composition ruled out the possibility of recent rainwater being the primary recharge source of the spring.

It was then assumed that the spring was recharged by much older water originating from high altitudes and had been a part of a deeper circulation, based on the depleted isotopic composition of the springs.

The isotopic signatures of the water samples are influenced by meteorological parameters (temperature and relative humidity) and isotope effects namely rainout, seasonal, temperature and altitude effects. The scatter in d-excess is caused by sub-cloud processes, temperature, relative humidity and the mixing of different air masses with different source regions.

The limitations of using stable isotopes is related to the complexities of interpreting stable isotopic compositions in water samples. It is sometimes difficult to distinguish the influence of the individual isotope effects on rainfall, especially over shorter time scales when the disparity in isotopic compositions is considerable.

The enrichment of the spring water allowed for the isotopic shift method to be applied to estimate recharge. Recharge amounts of 10.19 and 23.90 mm/month for $\delta^{18}O$ and δD, respectively, were obtained for the fractured aquifers of the Witwatersrand Supergroup quartzites. Recharge at the Alberts Farm spring occurs through a mixture of local mountain front recharge and recharge occurring via preferential flow through a network of connected fractures and faults, representing a regional circulation, in the Witwatersrand quartzites.

For a more accurate representation of groundwater recharge it is advisable that more than one water sample is used. Therefore, results of the isotopic shift method can be improved by further sampling of the spring, especially during the summer months where isotopic signatures of the spring water are likely to be enriched.

The radiogenic isotopes of hydrogen, tritium revealed that the stream samples could be grouped into three types of water; type 1 representing older water that has been a part of a deep circulation suggesting that the aquifer had a low permeability and thus recharge was direct. Type 2 is characteristic of older water that has been mixed with younger more recent rainwater and a local water source and type 3 representing water that has been contaminated by a local industrial source/rain.

The major drawback of using tritium to determine residence time is that ages don’t represent the real age of groundwater but rather an apparent age, representing mixing of groundwater of different ages.

7 CONCLUSION

An integrated recharge estimation method has been applied in the Upper Crocodile catchment.

Groundwater recharge was estimated quantitatively using the conventional water balance method, baseflow separation method and the water table fluctuation method. Recharge as a percentage of mean annual rainfall was estimated as 4% for the water balance method, 5.8% for the baseflow separation method and 14% for the water table fluctuation method.

The environmental isotopes were used to qualitatively estimate groundwater recharge through the use of stable isotopic signatures and groundwater ages. The stable O and H isotopes revealed that the springs were recharged by rainwater derived from different seasons and experienced progressive isotope depletion. The tritium readings were used to determine the mean residence time of groundwater that discharged into streams in the form of baseflow, which displayed three types of water, i) relatively old water, ii) old water that was mixed with more recent rainwater and iii) groundwater that was contaminated by a recent rainwater.

Using the WTF method, BFS method and the isotopic shift method groundwater recharge was estimated for the different aquifer types. Using the water table fluctuation method, recharge in karst aquifers was estimated to be 14% this closely correlated with the recharge estimate of catchment A21G which is underlain predominantly by dolomites. The recharge estimate for catchment A21G is 20% obtained from the baseflow separation method. The isotope shift method resulted in a recharge amount of 10.19 – 23.90 mm/month for the fractured meta-sedimentary aquifers of the Witwatersrand Supergroup, south of the study area. The fractured basement crystalline aquifers of quaternary catchments A21C and A21E have recharge estimates of 4.9 and 11%, respectively, which was obtained using the baseflow separation method. The fractured meta-sedimentary rocks of catchments A21F have a recharge estimate of 4.6%.

Processes responsible for the spatio-temporal variability of groundwater recharge included rainfall, geology, land cover, topography and sewage. It is clear from all the methods the role rainfall plays in recharging groundwater, the variability of rainfall in time and space is reflected in groundwater. For instance, recharge exhibits a seasonal change as a result of the wet and dry seasons; recharge also varies annually where recharge is generally high during higher than average rainfall and is less during lower than average rainfall.

The different geology found in the catchment area resulted in the spatial variability of recharge, with karst aquifers having higher recharge than the fractured crystalline and metasedimentary aquifers, because of high permeability owing to the presence of karst structures. Fractured aquifers display low recharge because recharge is limited to fractures and the weathered horizon.

The different topography will either encourage runoff or promote recharge. Areas with higher elevations are more likely to experience higher run off rates and diminished recharge but can also promote mountain front recharge. Whereas low-lying areas tend to promote recharge reducing surface runoff, an example of this is the dolomitic rocks.

Land cover in the study area is dominated by urbanisation that has resulted in an increase of impermeable surfaces such as buildings, tar roads and paving. These impervious surfaces limit infiltration subsequently limiting the amount of recharge that can occur and as a result runoff is high and direct recharge is low.

The presence of sewage has complicated the hydrological system particularly on baseflow that is related to long-term recharge. If sewage is not considered the estimated recharge will be greatly inflated because of the high sewage volume in streamflow. Inflated recharge estimates will have great implications especially if groundwater quantification is for groundwater management.

Approximately 153 x 10^6 m^3/yr of wastewater was discharged into streamflow through wastewater treatment plants as of 2008. The wastewater flow contributes to baseflow, as a component of streamflow and potentially contributes to groundwater recharge through seepage and geological structures. The consequence of the presence of wastewater is the overestimation of groundwater recharge.

7.1 RECOMMENDATIONS

The recommendation of issues to be considered in the future include:

- Renewed monitoring of groundwater levels on a daily basis to obtain a complete data set which can then be used to improve recharge estimates. Closely monitored groundwater levels can be used to assess induced recharge in aquifers and would improve the accuracy of recharge estimates.
- Making sure that each quaternary catchment has its own meteorological station, recording hydrometeorological data on a daily basis. If each quaternary catchment has its own meteorological stations, the temporal variability along with spatial variability of recharge in the catchment can be assessed.
- A quantitative study of how recharge processes affect groundwater recharge.
- Using a daily soil water balance method to quantify episodic recharge.
- The use of stable O and H isotopes could be extended to streamflow, to differentiate between the amount of baseflow and the amount of wastewater discharge.
- An analysis of spring water discharge, using the springs located in both fractured and karst aquifers, to estimate recharge rates. This would be an alternative method used to validate recharge estimates as there is an abundance of perennial springs in the area.

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9 APPENDICES

Appendices A – C (REFER TO DISK)