Groundwater Recharge Assessment in the Upper Limpopo River Basin: A Case Study in Ramotswa Dolomitic Aquifer

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Declaration

I Simamkele Siyambonga Baqa declare that ***Groundwater Recharge Assessment in the upper Limpopo River Basin with a case study in Ramotswa Dolomitic Aquifer*** is my own investigation and covers no section copied in whole or in part from any source unless it is clearly acknowledged in quotation marks and with detailed, complete and precise referencing. Further, the report has not been submitted before for any degree or examination at any university.

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Abstract

Hydrogeological research was undertaken in the transboundary Ramotswa dolomitic aquifer to provide understanding and quantification of the processes governing recharge mechanism and rates, in order to promote efficient and sustainable groundwater resource utilization and development, as well as to improve the Ramotswa transboundary aquifer management. Hydrochemical and tracer approaches were utilized to evaluate the processes governing the recharge mechanism while the chloride mass balance approach was further applied to assess groundwater recharge rates.

Results indicated that all groundwater samples contained detectable amounts of tritium highlighting the renewability of the transboundary Ramotswa aquifer resources. Two distinct water types were characterised: sub-modern waters approximately recharge prior to the 1950s and a mixture of modern and sub-modern waters of recently recharge rainfall indicative of active recharge in the area through intensive rainfall. Correlation between $\delta^{18}O$ and $\delta^{2}H$, and soil Cl^{-} indicated that groundwater recharge in the Ramotswa dolomitic aquifer takes place mainly by two flow mechanisms: a displacement of moisture through a diffuse or piston flow through permeable soils and from concentrated runoffs due to surface depressions, and a preferential flow component through fractures that outcrop at surface and riverbed infiltration along the ephemeral Notwane River. Annual groundwater recharge estimates varied from 0.4% MAP to 12% MAP and from 5% MAP to 14% MAP within the northern parts and the southern parts of the study area, respectively. Recharge estimates correlated well with the proposed mechanism of flow both in the southern and in the northern parts of the study area as well as with the previous studies conducted within the greater Ramotswa area.

A way forward to ensure the long-term sustainability of the transboundary Ramotswa aquifer resources is recommended, such as to preserve and protect potential recharge areas through carefully controlled land use planning and development, and to equate abstraction rates to average recharge rates, which has to be subjected to the Limpopo Watercourse Commission.

Dedication

Specially, I would like to send my devotions as well as to thank my family ooBaqa amaTshatshu noMtambeka amaQwathi for their guidance, encouragement, love and support they have showed to me throughout the journey of my life, and for raising me to be a better person in life. Enkosi.

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Acronyms and Notations

˚C	Degrees Celsius
‰	Parts per million
Δ	Delta symbolizing change between two points
µS/cm	Micro Siemens per centimetre
%MAP	Percent of mean annual precipitation
ℓ/s	Litres per second
EC	Electrical Conductivity
pH	Measure of acidity or basicity of an aqueous solution
TDS	Total Dissolved Solids
Km	Kilometre
m	Metre(s)
mm	Millimetres
mm/a	Millimeters per annum
m^3/d	Cubic metres per day
m.a.s.l	Metres above sea level
δ	Stable isotope notation
∑	Sum of
φ	Porosity

λ	Radionuclide decay constant
θ	Volumetric moisture content
~	Approximately (equivalent to)
S_y	Specific Yield

Abbreviations

BWA: Botswana country code

CGIAR: Consultative Group on International Agricultural Research

CGS: Council of Geosciences

CMB: Chloride Mass Balance

CRD: Cumulative Rainfall Departure

CWB: Channel Water Budget

DGS: Department of Geological Survey

DWA: Department of Water Affairs

DWS: Department of Water and Sanitation

EARTH: Extended model for Aquifer Recharge and Moisture Transport through Unsaturated Hard-rock

EC: Electrical Conductivity

ET: Evapotranspiration

EV-SF: Equal Volume – Spring Flow

GD: Groundwater Dating

GIS: Geographical Information System

GMA: Groundwater Management Area

GMWL: Global Meteoric Water Line

GM: Groundwater Modelling

GNIP: Global Network of Isotope in Precipitation

HS: Hydrograph Separation

IAEA: International Atomic Energy Agency

IWMI: International Water Management Institute

LMWL: Lobatse Meteoric Water Line

LRB: Limpopo River Basin

MAP: Mean Annual Precipitation

NRF: National Research Foundation

PET: Polyethylene

QGIS: Quantum Geographical Information System

RQIS: Resource Quality Information System

RSA: Republic of South Africa

RTBAA: Ramotswa Transboundary Aquifer Area

SADC: Southern Africa Development Community

SAWS: South Africa Weather Service

SVF: Saturated Volume Fluctuation

TAM: Transboundary Aquifer Management

TDS: Total Dissolved Solids

UFM: Unsaturated Flow Modelling

UoB: University of Botswana

Wits: University of the Witwatersrand

WM: Water Modelling

WMA: Water Management Area

WRC: Water Research Commission

WTF: Water Table Fluctuation

WUC: Water Utilities Corporation

ZFP: Zero Flux Plane

ZA: South Africa`s country code

Chapter 1

Introduction

1.1. General Introduction

Water, particularly the liquid freshwater, is one of earth's valuable natural resources. It is fundamental to the survival of humankind and to the resilience of the ecosystem. Surprisingly, of the 35 000 000 km^3 total estimated volume of freshwater resources on earth, groundwater, constitutes about 30.8% and approximately 97% of all accessible freshwater if only the liquid freshwater is considered (Shiklomanov, 1999 as cited in UNEP, 2002). Groundwater, though not understood as much as surface water as a result of its concealed nature, it remains as the most feasible source of large quantities of water to improve the current water supply schemes at a reasonably low cost.

The significance of groundwater has long been recognised approximately since the 1440 BC. Since then, groundwater has become one of the most valuable natural resources essential for municipal, agricultural, industrial developments and for environmental sustainability. Some of the advantages for developing groundwater include its relative widespread occurrence and accessibility (Puri, 2001; Braune and Xu, 2005; Jia, 2007), the ability to provide buffer during long dry years (Braune et al., 2010; Wang et al., 2010), its reasonably good water quality due to its partial separation from surface influences (Puri, 2001; Wang et al., 2010), and its reasonably low cost of exploration and exploitation over a long-term compared to its relative competitors such as surface water.

In relatively dry regions of the world that are characterized by limited surface water resources, groundwater often supply large quantities of water throughout the year (Allison and Hughes, 1975; Wang et al., 2008). In such arid regions the use and availability of groundwater resources is fundamental for sustainable socio-economic development and poverty alleviation as most of the arid regions largely depend on groundwater for producing food and for provisioning of basic health and sanitation (Scanlon et al., 2010).

Increasing population and agricultural activities (both linked to the increase in water demand), pollution, climate variability and change have become one of major key issues influencing the availability of groundwater resources. Further, the issue of climate variability and change is more likely to intensify the role or use of groundwater resources for water supply as groundwater resources are more resilient to the effects of climate change than surface water resources (van der Gun, 2012).

Ensuring the sustainability of groundwater resources is complex and needs to be inclusive of all major stakeholders either interested or affected within the catchment. In the case of internationally shared groundwater resources, the trade-offs between the water demands becomes more and more complex due to the poor understanding of transboundary groundwater resources and lack of legal framework among the riparian countries (Puri, 2001; van der Gun, 2012). Transboundary Aquifer Management (TAM) through international cooperation and governance (or cooperative governance) among the riparian countries is highly influential in ensuring the sustainability of the groundwater resources and promoting approaches and tools for its proper management (van der Gun, 2012).

An efficient and sustainable groundwater resource utilization and management plan depends on an understanding and quantification of the rates of groundwater recharge (Sophocleous, 1991; Healy and Cook, 2002; Braune and Xu, 2005; Marechal et al., 2006). Understanding and quantification of the rates of natural recharge is instrumental for quantifying the sustainable yield (or use) of underground water resources (Xu and Beekman, 2003; Obuobie, 2008). Therefore, quantifying the amount of recharge is vital to equitably use groundwater stored in the transboundary aquifer.

The present study focuses on the understanding and quantification of groundwater recharge processes as well as groundwater recharge rates on shared aquifer between South Africa and Botswana.

1.2. Background

The transboundary Ramotswa dolomitic aquifer is one of the three internationally recognized aquifers located within the Limpopo River Basin (LRB). The resources (both surface-water and groundwater) of the transboundary Ramotswa dolomitic aquifer are shared between South Africa and Botswana. The transboundary Ramotswa dolomitic aquifer exists within the upper parts of the Limpopo River Basin beneath the Marico and the Notwane sub-catchments in South Africa and Botswana, respectively (Figure 1).

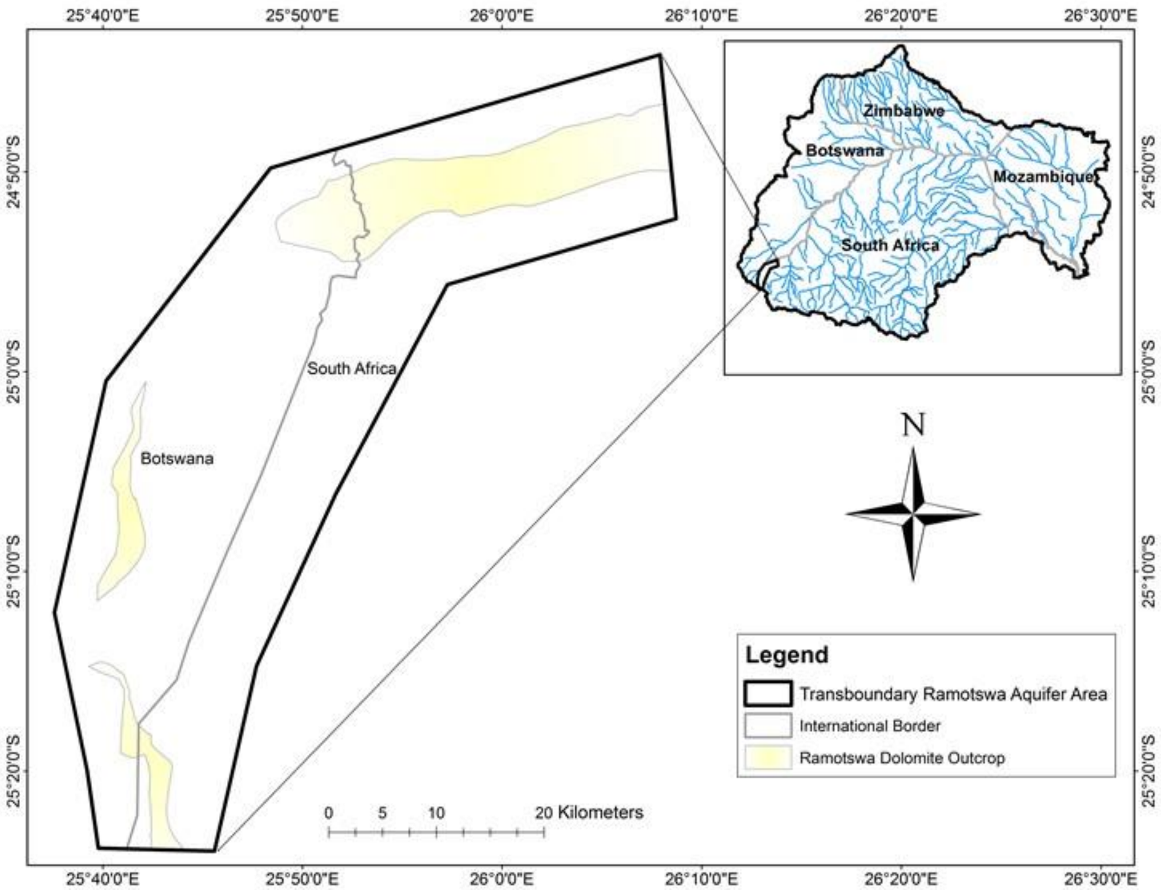


Figure 1: Ramotswa transboundary aquifer area in both South Africa and Botswana.

The greater Ramotswa area lies within a relatively dry land region. Rainfall is sporadic and erratically distributed with high seasonal variations between wet and dry seasons (Staudt, 2003; Altchenko et al., 2016). Unevenly distributed rainfall worsened by high evapotranspiration rates is among the key dominating factors contributing to the limited surface water resources. Hence, the limited surface water resources result in groundwater being the most feasible water source for water supply throughout the year, especially for rural communities including farming areas.

Groundwater occurs mainly through secondary porosity by which it further progresses through geological and morphological evolution such as dolerite dykes compartments, fracturing and Karstification (Gieske, 1992; Ranganai et al., 2001; Pietersen et al., 2011; Altchenko et al., 2016). Attributed to the karst development and high degree fracturing of dolomites, the aquifer has high capacities to store and to transmit water (storativity (S) and transmissivity (T)), which characterizes it as one of most productive and often, the best feasible source for large quantities of water to meet the water demands for socio-economic development within the greater Ramotswa area.

Conversely, the vulnerability of the transboundary Ramotswa dolomitic aquifer to meet the water demands for socio-economic development has long been stressed due to pollution from pit latrines in Botswana, which lead to the abandoning of the wellfields in late 1990`s to early 2000`s (Ranganai et al., 2001). Following the abandonment of the wellfields in late 1990`s, further, fears rose as a result of increased water demands, especially by commercial agriculture in South Africa (Owen, 2011).

Despite the raised concerns, the wellfield has been restored since 2014 to augment the Gaborone dam with large quantities of water in order to balance the water demands for the city of Gaborone and its surroundings. Recently, there has been considerable interest from the South African government to abstract water from the dolomitic aquifer in and around the transboundary Ramotswa aquifer to augment its municipal water supply (Pietersen et al., 2011). Considering the current extreme climate variability and change, including the drought experienced in 2015/16, it is highly anticipated that the future of the transboundary Ramotswa aquifer (groundwater resources) is likely to be more strained, thereby posing a greater threat to meet the future water supply demands and also making it very difficult to balance the trade-offs between several water demands.

Managing the groundwater resources of the transboundary Ramotswa aquifer is of strategic significance to minimize the potential conflicts that could arise due to inequitable sharing of groundwater resources. However, managing groundwater resource is often challenging due to the concealed nature of groundwater, large number of groundwater users and different regulatory regimes from the two countries (IWMI, 2016). Therefore, understanding and quantification of groundwater recharge rates within the transboundary Ramotswa aquifer area is decisive to equitably share, protect and promote sustainable use of groundwater resource of the transboundary Ramotswa aquifer.

However, understanding and quantification of the groundwater resource and recharge rates is challenging due to the low moisture availability, sporadic recharge producing events (Bredenkamp et al. 1995 as cited in Xu and Beekman, 2003; van Wyk, 2010). Further limiting the ability to observe and analyse aquifer dynamics related to groundwater recharge events is the lack of basic hydrological monitoring data, recharge variability compounded by the variation in soil and geology, surface topography, land use and land cover. These data need to be assessed in order to provide better understanding on the processes governing recharge and to accurately quantify groundwater recharge rates.

As part of the ongoing GRECHLIM project, the study provides a better understanding and assessment of groundwater recharge rates and processes within the greater Ramotswa area to promote sustainable groundwater resource utilization, development and transboundary cooperation between South Africa and Botswana.

1.3. Hypothesis

The study has put forward a hypothesis that understanding groundwater recharge rates and processes governing recharge mechanisms will promote efficient groundwater resource utilization and hence, improve on the transboundary aquifer management (TAM) between South Africa and Botswana.

1.4. Research Questions

The overall study is guided by two key research questions as follows:

- What are the rates of groundwater recharge within the transboundary Ramotswa aquifer?
- What are the processes governing recharge mechanisms within the transboundary Ramotswa aquifer?

1.5. Aim and Objectives

The study aims to promote efficient groundwater resource utilization, development and transboundary aquifer management through the provision of a better understanding and assessment of groundwater recharge and the processes governing recharge mechanisms within the transboundary Ramotswa dolomitic aquifer.

Specific objectives are:

- To quantify groundwater recharge rates, and
- To evaluate the processes governing recharge mechanism in the transboundary Ramotswa dolomitic aquifer.

1.6. The structure of the thesis

Chapter 2: provides an in-depth understanding of groundwater recharge assessment through reviewing and examining the relevant literature on recharge estimation techniques. The chapter begins with theoretical aspect of groundwater recharge. Further, an overview of groundwater recharge assessment is provided as well as the recharge assessment in semi-arid regions. The chapter ends with a review of recharge assessment techniques in Ramotswa dolomitic aquifer.

Chapter 3: describes the environmental setting of the greater Ramotswa area with focus on climate, vegetation and land use, soil and geology including hydrogeology. The chapter is aimed to provide a detailed understanding on factors and/ or processes governing recharge mechanism.

Chapter 4: describes the methodology used for data collection and analyses.

Chapter 5: presents and analyse, results of evaluated processes governing recharge and the estimated groundwater recharge rates. Further, comparison of the research findings with previous studies is presented.

Chapter 6: conclude and recommends on the research findings; put forward measures to improve transboundary aquifer management and recommends on future groundwater recharge assessment studies in transboundary Ramotswa dolomitic aquifer. The last two sections provide references and appendices.

Chapter 2

Literature Review

2.1. Introduction

An understanding of the processes governing recharge mechanism and the quantification of groundwater recharge rates in the Ramotswa dolomitic aquifer is essential to equitably share the groundwater resources of the aquifer within the greater Ramotswa area. Furthermore, it is essential for land-use and groundwater resource planning through an understanding of groundwater recharge and discharge zones and the characteristics of those areas. Section below provides an in-depth analysis of the appropriate techniques applicable for groundwater recharge estimation and understandings of the processes governing recharge mechanism in Ramotswa dolomitic aquifer.

The analyses begin with the notions (definition and concepts) into groundwater recharge such as the flow mechanisms defining recharge (processes governing recharge mechanism). Further, the analyses provide an overview (a global outlook) of the commonly used techniques to assess groundwater recharge rates, a summary of the commonly used techniques to assessing groundwater recharge rates and the most feasible techniques for groundwater recharge estimation across various climatic regions of Southern Africa. The last section of this chapter evaluates groundwater recharge estimation methods applied in the transboundary Ramotswa Dolomitic Aquifer.

2.2. Definition and Concept(s) of Groundwater Recharge

Groundwater Recharge can be defined as an entry of water into the groundwater reservoir (Lerner et al., 1990). It is a process by which groundwater reservoirs are replenished and one of the principles behind groundwater level fluctuation (Todd and Mays, 2005). In a broad sense, the entry of water into the saturated zone entails a three dimensional movement of water into a groundwater reservoir either in an upward, downward, and/ or lateral direction as described by Lerner et al. (1990). Hence, it is vital to precisely define the concept of recharge on the basis of the current investigation.

Conceptually groundwater recharge has been identified as a downward movement of water through the vadose zone and made available into the water table (Lerner, 1997; de Vries and Simmer, 2002; Healy and Cook, 2002; Scanlon et al., 2002; Xu and Beekman, 2003); inter-aquifer flow (Hantush and Jacob, 1955); enhanced recharge from nearby surface water bodies as a result to groundwater pumping, and artificial aquifer recharge from man-made infiltration ponds or injections wells (Healy and Cook, 2002).

Commonly, in arid and semi-arid regions like that of the greater Ramotswa area, recharge through the downward flow of water through the subsurface zone reaching the saturated zone is often the most significant means of replenishing the aquifers (Xu and Beekman, 2003). For this reason, the basis of this report rests on the first mode of recharge that occurs by downward water flow through the subsurface zone to reach the surface of water table.

Recharge through downward flow reaching the water table occurs from various sources, either natural and/ or as a human induced phenomena (Sophocleous, 2003). Natural sources of recharge include infiltrated water from precipitation, surface water bodies, and from other aquifers (leaking aquifers), whereas human induced sources of recharge comprises irrigation returns, recharge caused by pumping on the nearby surface water bodies such as dams or lakes. In either of the two sources of recharge, the water that successively reaches the surface of the water table through the subsurface zone is mainly from atmospheric (or meteoric) origin (Simmers, 1988).

In general, when the overall water input is supplied into the soil from precipitation, three prevailing set-ups that can be distinguished. Firstly, a fraction of the overall water input may be lost through evaporation. Secondly, some fraction of water may generate horizontal overland flow(s) which may later infiltrate through stream beds and contribute to recharge downstream, and thirdly, some fraction of water might flow through the subsurface zone and successively reach the surface of the water table (Brouwer et al., 1988). From these surface processes, various conceptual problems exist as a result of subsurface complexities governing the recharge.

As there are various sources of recharge, also, are the processes governing recharge rates. In agreement with Lloyd (1986), Lerner et al. (1990) conceptually defined recharge mechanism from various sources as *direct or diffuse recharge* from rainfall that in succession reaches the water table surface as a residual of soil moisture and evapotranspiration (de Vries and Simmers, 2002; Scanlon et al., 2002*),* *indirect recharge* from infiltrated rain water through the stream beds to the surface of the water table, and *localized or focused recharge* from concentrated surface-water infiltration due to topographic depressions (Scanlon et al., 2002; Sophocleous, 2003). Sophocleous (2003) in addition appended recharge mechanisms from these sources as mountain front recharge, induced recharge and preferential recharge.

Direct recharge is defined as a fraction of the total volume of rainfall added into the groundwater storage in surplus of soil-moisture deficits and evapotranspiration in a direct downward water movement through the subsurface zone (de Vries and Simmers, 2002; Scanlon et al., 2002; Sophocleous, 2003). This usually occurs in homogeneously disseminated areas from uniformly distributed percolation through the entire subsurface soil zone. This commonly occurs on irrigated land or in regions characterised by humid climate conditions (de Vries and Simmers, 2002; Sophocleous, 2003).

Indirect recharge occurs as a result of drainage to the surface of the water table through beds of the surface water courses (de Vries and Simmers, 2002; Sophocleous, 2003). Two distinct forms of indirect recharge are distinguished (Sophocleous, 2003); as those that are associated with surface water bodies or paths, and those that result from horizontal or near surface water concentration due to surface depressions.

Localized (Focused or Eventual) recharge occurs as a result of concentrated runoffs or concentrated surface-water infiltration due to topographic depressions (Scanlon et al., 2002). This mode of recharge is also known as depression-focused recharge. Sophocleous (2003) characterizes localized recharge as one of the two distinct recharge forms distinguished to prevail under indirect recharge mechanism.

It is evident from the literature that recharge takes place to some degree even in most dry parts of the world. Simmers (1998), and de Vries and Simmers (2002) proclaimed that with increasing aridity, it is more likely that the significance of direct recharge becomes less important than the localized recharge and indirect recharge in relations to total aquifer replenishment. However, in most settings various types of recharge may occur simultaneously although one mechanism may dominate compared to the other. According to Sophocleous (2003), in arid and semi-arid regions, recharge through beds of the surface water courses (localized recharge) and/ or recharge from concentrated runoffs or surface-water infiltration due to topographic depressions often provides the most prominent sources of recharge into the saturated zone.

Independently from surface complexities governing recharge mechanism, also various conceptual problems exist within the subsurface zone influencing recharge mechanism into the saturated zone. De Vries and Simmers (2002) defined three subsurface zone flow mechanisms governing recharge as;

(i) “Long-winded percolation one or the other as an unsaturated instability or saturated front (piston-like flow)”,
(ii) “Macro-pore flow through root conduits, withering cracks, and fissures”; or
(iii) “Preferential movement caused by unsteady soaking fronts and discrepancies in soil physical characteristics, notably between sandy and clayey sediments”.

As an alternative, Sophocleous (2003) categorized two mechanisms of flow driving potential recharge within subsurface zone; the *capillary flow* occurring in pore diameter of less than 3 mm due to capillary forces and gravity, and the *macropore flow* which occurs in pore diameter greater than 3 mm primarily caused by viscous forces and gravity. The movement of water through macropore(s) is well-known as a preferential or bypass flow with the resultant recharge characterized as preferential recharge which occurs through preferential pathways as opposed to uniformly distributed recharge through the vadose zone which normally occurs in a piston-like flow.

However, according to Obuobie (2008), the arid and semi-arid regions of the world are characterized by two mechanisms driving the potential recharge into saturated zones. These are a piston-like flow (uniform flow) and preferential flow as described above. Comparable to the surface complexities governing recharge, in many locations a combination of both mechanisms is likely to occur although one mechanism may dominate when compared to the other.

It is clearly evident that the quantity of water reaching the surface of the water table is a resultant of various surface and subsurface complexities. Further, DWAF (2006) stipulated that the magnitude of water that enters into the aquifer is a function of the aquifer`s ability to store or accept water depending on the aquifer`s capacity to store and to transmit water. This on its own makes it extremely challenging to quantify recharge rates accurately and precisely.

2.3. An overview of Recharge Estimation Techniques

Challenges arise when quantifying the rates of natural recharge and often associated with these challenges is the high areal variability in the distribution of groundwater recharge, the limited capability to identify and quantify the probable recharge mechanism, and the features influencing recharge for a given area (Sophocleous, 2003).

To date, a number of techniques have been developed to quantify recharge rates, however, only few can be applied successfully in the field particularly in arid and semi-arid hard-rock terrains (Sophocleous, 2003). These techniques produce recharge rates at different time scales largely encompassing a wide range of complexities which could result on incorrect values due to errors in parameter estimation and spatial coverage of the processes (Bredenkamp et al., 1995 as cited in Xu and Beekman, 2003; Healy and Cook, 2002).

In relatively dry land regions, recharge estimation through direct measurements are often challenging as recharge fluxes are generally low compared to the hydrological parameters (such as evapotranspiration) (Simmers, 1988). Successively, a number of studies have largely attained the rates of groundwater recharge through the use of indirect techniques. However, the use of indirect techniques has its own limitations which may result in inaccurate recharge values. Associated with these techniques, Lerner et al. (1990) identified four types of errors that may be encountered when estimating groundwater recharge rates: (i) inappropriate conceptual model(s), (ii) negligence of spatial and temporal variability, (iii) measurement errors, and, (iv) recharge estimation errors. This however signifies that no matter what technique is implemented in a study still various uncertainties associated with the technique remain. Therefore, choosing an appropriate technique to apply in a study is of utmost significance in any groundwater resource evaluation study.

A major decision criterion for choosing appropriate recharge estimation techniques is the expected direct and indirect recharge relationship, and possible identification of prevailing flow mechanism and factors influencing recharge (Lerner et al., 1990; Wanke et al., 2007). Further understanding the concept, assumption, limitations and applicability (both spatial and temporal) of the proposed techniques is of valuable importance including clearly defined project objectives. Healy (2011) endorsed the development of a conceptual groundwater recharge model in combination with the water budget model at initial stages of the research. He further specified that the use of a water budget model is often to guide on the decision on the technique to be utilized for the project and not to be utilized to quantify recharge rates. Various authors (Lerner et al., 1990; Healy and Cook, 2002; Kinzelbach et al., 2002; Scanlon et al., 2002; Sophocleous, 2003; Xu and Beekman, 2003) have largely endorsed the use of a variety of techniques when assessing recharge rates in order to reduce results uncertainties and to compare reliability of the results.

Scanlon et al. (2002) categorized recharge estimation techniques based on three hydrologic zones or sources from which data can be obtained namely; surface water, unsaturated zone and saturated zone. These zones provide a unique data set of which data can be obtained and further, provide varying recharge rates temporally and spatially in each zone (Obuobie, 2008). Within each zone, recharge estimation methods can be further classified into tracer, physical, or numerical modelling approaches. Beekman and Xu (2003) prepared a simplified classification model for recharge estimation techniques based on spatial scales. The model categorizes recharge into; hydrogeological provinces, hydrological zones (in agreement with Scanlon et al., 2002), and physical and tracer approaches (also referred to as the general approaches to recharge estimation).

2.4. General approaches to recharge estimation

Various methods exist to estimate groundwater recharge. General approaches for groundwater recharge assessments are (Lerner et al., 1990): direct versus indirect, water budget, Darcian physical, and chemical, isotopic and gaseous tracer methods.

Direct techniques: measure the potential recharge through the vadose zone. Therefore these provide an indication of recharge. Such techniques include the use of neutron probe, lysimeters, direct observation, infiltration coefficients etc. Generally, direct techniques provide point or local estimates of recharge thus overlooking the spatial variability in recharge flux as a result of land use/cover variability, soil and geology (Kinzelbach et al., 2002). Normally direct techniques assume a piston flow (diffuse recharge).

Indirect techniques: link recharge to other measurements such as precipitation, stream discharge through the use of models (Entekhabi and Moghaddam, 2007). Such techniques include soil moisture budget method, Zero-Flux Plane method etc. Some techniques like subsurface water balances, estimate recharge as a residual of moisture fluxes. Subsurface water balances are subjected to errors in parameters estimation such evapotranspiration rate.

Water Budget/Balance methods: make use of existing (available) data. Mostly applicable to everywhere if data is available as they account for all water entering the system (inflow), storage change and water leaving the system (outflow) (Lerner et al.,1990; Healy, 2011). Recharge is estimated as a surplus of other processes. The advantages of using the Water Budget methods is that additional water into the system can be identified by balancing input and the out water in the system (catchment). Normally it is difficult to quantify other fluxes such evapotranspiration.

Darcian Physical methods: combine Darcy`s law and the equation of mass conservation. It tries to mimic the flow of water and the actual processes of particular interest (Sun, 2005). Works well in saturated zone (flow) while it is often difficult to apply in an unsaturated zone. In an unsaturated zone, the Darcian approach (Darcy`s law) is based on the measurements of matrix potential, while in saturated zone it normally based on aquifer pumping test and head measurements.

Chemical, isotopic and gaseous tracer techniques: can be subdivided into signature and throughput methods (Lerner et al., 1990). Tracers are mostly used in arid and semi-arid areas. Applied tracers such as tritium (^{3}H) and or carbon-14 (^{14}C) are widely used to date or track a fraction of water containing a tracer. Throughput methods include mass balance methods such as chloride mass balance and these techniques rely on measuring the tracer concentration in the atmosphere and comparing to that measured on the ground. Mostly the techniques assume a piston like flow.

2.5. Commonly used techniques to assess recharge in semi-arid regions of Southern Africa

Beekman and Xu (2003) reviewed groundwater recharge estimation techniques used for three decades in Southern Africa. Techniques were reviewed in terms of applicability (flux, area and time), limitations and ratings (accuracy, easy to use and cost). From the review they came up with the commonly used techniques in arid and semi-arid regions of Southern Africa (Table 1). These techniques were then grouped into three hydrological zones; surface water, unsaturated zone, unsaturated and saturated zone, and saturated zone. From the review, promising methods for groundwater recharge estimation in arid and semi-arid regions of Southern Africa were identified as follows: Chloride Mass Balance (CMB), Cumulative Rainfall Departure (CRD), Water Table Fluctuation (WTF), Groundwater Modelling (GM), Saturated Volume Fluctuation (SVF), and Extend model for aquifer Recharge and moisture Transport through unsaturated Hard rock (EARTH). These were identified to be functional and can be applied with better confidence in arid and semi-arid regions Southern Africa.

Table 1: Review of commonly used methods for groundwater recharge estimation in arid and semi-arid regions of Southern Africa (Beekman and Xu, 2003).

Zone	Method	Limitations	Applicability[2]			Rating[3]		
			Flux (mm/yr)	Area (km^2)	Time (yrs)	Acc.[4]	Ease	Cost
Surface Water	**HS (Base flow)**	Ephemeral Rivers	400-4000 (0.1-1000)	10^4-1300 (10-1000)	0.3-50 (1-100)	2-3	1-2	1-2
	CWB	Inaccurate flow measurements	100-5000	10^{-3}-10	1d-1yr	2-3	2	3
	WM	Ephemeral rivers	1-400	10^{-1}-5*10^5	1d-10yr	2	2-3	3
Unsaturated[1]	**Lysimeter**	Surface Runoffs	1-500 (0-200)	0.1-30m^2	0.1-6	2	3	3
	UFM	Poorly known relationship hydraulic conductivity – moisture content	20-500	0.1-1m^2	0.1-400	3	2	2
	ZFP	Subsurface heterogeneity, periods of high infiltration	30-500	0.1-1m^2	0.1-6	3	2	2
	CMB	Long-term atmospheric deposition unknown	0.1-300 (0.6-300)	0.1-1m^2	5-10000	2	1	1
	Historical	Poorly known porosity, present ^3H levels almost undetectable	10-50 (10-80)	0.1-1m^2	1.5-50	2-3	2-3	3

Zone	Method	Limitations	Applicability[2]			Rating[3]		
			Flux (mm/yr)	Area (km^2)	Time (yrs)	Acc.[4]	Ease	Cost
Saturated – Unsaturated	**CRD**	Deep (multi-layer) aquifer, sensitive to specific yield (S_y)	(0.1-1000)	(1-1000)	(0.1-20)	1-2	1-2	2
	EARTH	Poorly known (S_y)	(1-80)	(1-10m^2)	(1-5)	1-2	2	1
	WTF	In/outflow and (S_y) usually unknown	5-500	5×10^{-5} - $>10^{-3}$	0.1-5	2	1	1
	CMB	Long-term atmospheric deposition unknown	0.1-500	2×10^{-6} - $>10^{-2}$	5->10000	2	1	1
Saturated	**GM**	Time consuming; poorly known transmissivity, sensitive to boundary conditions	(0.1-1000)	(10^{-6} – 10^{6})	(1d-20yr)	1-2	3	3
	SVF	Flow-through region, multi-layered aquifer	(0.1-1000)	(1-1000)	(0.1-20)	1-2	1-2	2
	EV-SF	Confined aquifer	(0.1-1000)	(1-100)	(1-100)	1-2	1-2	1-2
	GD	^{14}C, $^{3}H/^{3}He$, CFC: poorly known porosity / correction for dead carbon contribution	^{14}C: 1-100 $^{3}H/^{3}He$, CFC: 30-1000	^{14}C,$^{3}H/^{3}He$, CFC: $2*10^{-6}$ ->10^{-3}	^{14}C: 200-200000 , $^{3}H/^{3}He$,CFC:2-40	3	2-3	3

[1] All methods for estimating fluxes through the unsaturated zone assume diffuse vertical flow whereas in reality flow along preferred pathways is the rule rather than the exception. These methods therefore tend to overestimate the diffuse flux.

[2] Data in brackets are estimates from southern Africa; Rainfall may be up to 2000 mm/year; other data represent global values and are taken from Scanlon et al. (2002).

[3] Ratings for methods applied to semi-arid Southern Africa (based on authors experience).

[4] Accuracy

Accuracy ratings were (Xu and Beekman, 2003) given in three classes based on regional recharge estimates adopted from Kinzelbach et al. (2002); with Class 1: within a factor of 2 (difference from the true value), Class 2: within a factor of 5 (of the same magnitude) and Class 3: within a factor of 10 or more (large errors likely). Ease of application is related to data requirements and data availability and is rated from 1: easy to use to 3: difficult to use. Cost is rated from 1: inexpensive to 3: expensive (Xu and Beekman, 2003).

Recently, Abiye (2016) assembled published and unpublished reports and papers across different weather conditions in Southern Africa countries to evaluate the feasibility of various recharge estimation methods. Succeeding substantial examination, results on each paper/report were then classified based on the timespan, recharge estimation method applied and on the quality of data, where contrasting recharge estimates were encountered within the same area with the same technique. Furthermore, the results were grouped into geological and geomorphological environments: Table Mountain and Karoo aquifers, fractured basement aquifer, and the Kalahari aquifer. From the review, recharge estimates within the array of ±10 mm were said to be within an acceptable range for the interpretation of the study (Abiye, 2016).

Further from the review, the most feasible methods for recharge estimation were identified as follows: the chloride mass balance method considering that both dry and wet chloride depositions on the precipitation sample are used, the saturated volume fluctuation and the water table fluctuation method as both techniques accounts for the specific yield (the water yielding capacity of the rock (Johnson, 1963). These techniques were understood to be realistic across various climate settings in Southern Africa countries. However, some alertness was raised to consider the role and/ or the volume of water released from the wastewater treatment works into the streams, which could affect the baseflow separation and as result on overestimated recharge rates.

2.6. Recharge Assessment in Ramotswa Dolomitic Aquifer

Recharge estimation techniques in Ramotswa have been categorized into two specific objectives: (i) To evaluate the processes governing recharge mechanism, and (ii) To quantify groundwater recharge rates.

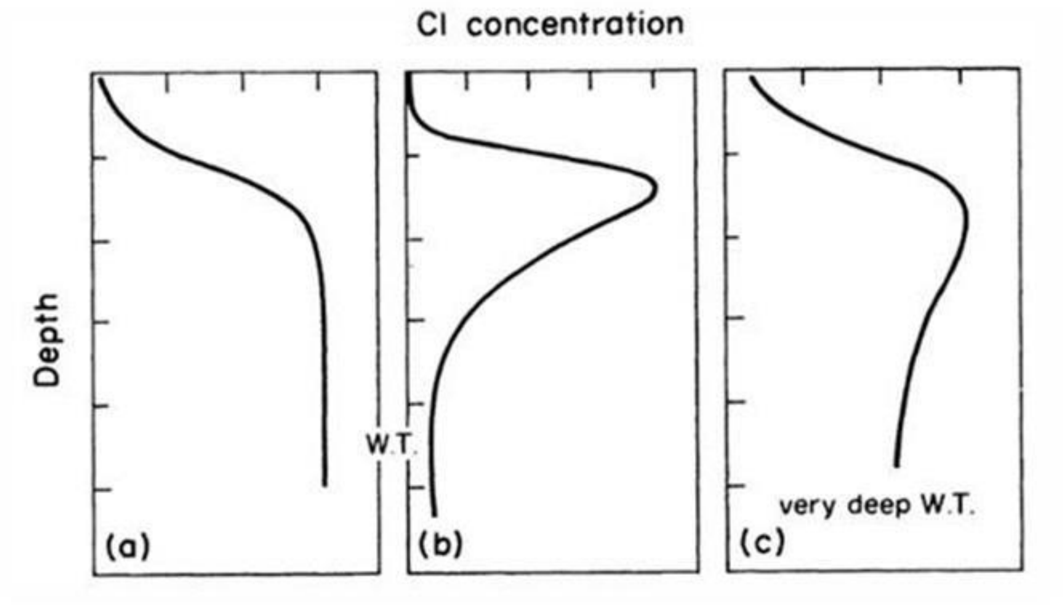
2.6.1. Evaluation of the processes governing recharge mechanism

Unsaturated zone Chloride profiling

The unsaturated zone chloride profiling technique is based on the Chloride Mass Balance principles such as that the chloride ion is highly soluble and geochemically conservative, therefore the chloride tracer movement is not decelerated nor its concentration is reduced by the adsorption processes in the permeable material or plants consumption. As a result, the chloride concentration in subsurface soil zone is mainly determined by the loss of water or moisture through evapotranspiration (Simmers, 1988; Gieske, 1992; Kinzelbach et al., 2002; Adams et al., 2005). Generally, the technique is based on the sluggish movement of natural chloride ion in the soil profile that dissolves in precipitation and infiltration (Signhal and Gupta, 2010).

The transient chloride flux is evaluated by plotting the chloride content from the pore water as a function of depth. The projected chloride configuration within the soil profile, at best is that the chloride content increases with increasing soil profile as the water is lost through evapotranspiration until the zone where a one-dimensional flux occurs (root zone/zero flux plane where evapotranspiration is minimal or does not occur at all) (Adams et al., 2005; Signhal and Gupta, 2010). Below this zone (or depth) the chloride content in pore water is expected to remain constant until it reaches the surface of the water table (Adams et al., 2005; Signhal and Gupta, 2010).

Three conceptual chloride concentrations profiles are shown in Figure 2



A. Piston-like flow with some abstraction of water by roots. B. Abstraction of water by roots, but with preferential flow paths of water beneath the root zone, or diffuse loss of chloride to the water table. C. Profile that may reflect the recharge history of a site.

Figure 2: Three conceptual chloride depth versus concentration profiles (Simmers, 1988; Beekman et al., 1997)

The basic form of a piston like flow in soil water in unsaturated zone is that the chloride concentration will increase to a constant value in the root zone. Under circumstance that the water table is deep or the chloride content in soil water is the same as in groundwater, Figure 2(a) should prevail/result (Simmers, 1988; Beekman et al., 1997). A diffuse loss of chloride results to an increase in chloride content in soil water due to evaporative enrichment and such prevail to a point where preferential flow is more dominant or prevail resulting in a decrease in chloride content beneath the root zone where evapotranspiration is ineffective. The diffuse loss of chloride and preferential flow of moisture by-passing the soil zone result in the so-called bulge type profile shown in Figure 2(b) (Beekman et al., 1997). Figure 2(c) indicates a more complex profile attributed to paleo-climate induced changes in recharge (Beekman et al., 1997).

The unsaturated zone chloride profiling approach provides an understanding on the mechanism under which recharge occurs based on chloride profiles. However, if macro-pores exist within the area under investigation, the technique would not yield reliable results (Adams et al., 2005). Generally the technique assumes a piston like flow (diffuse flow) or works well in homogeneous unconsolidated soil or permeable media with minimal macro-pores or preferential flow paths. The method assumes no additional (man-made) chloride into the system, thus assuming only atmospheric chloride deposition.

Environmental stable isotope signature

The environmental stable isotopes of ^{18}O and ^{2}H are largely used by hydrogeologist to depict and to trace water of different origins. Stable isotope application is based on the different behaviour of heavy and light isotopes during the evaporation and condensation (Adelana, 2010). Stable isotopes in rainwater provide distinctive signatures, which are manifested by atmospheric processes, altitude and latitude, and the characteristic weather configuration during a year. Differences on ^{18}O and/ ^{2}H (Deuterium or D) caused by these factors are be used as indicators for mixing ratios of waters from different origins (Figure 3).

The stable isotopes ^{18}O and ^{2}H are used to provide an insight into the processes governing recharge mechanism (the characteristics under which the water was recharged into the aquifer) and possible estimates of recharge. Craig (1961) observed that even though the isotopic content varies exceptionally in precipitation, precipitation that has not suffered from significant evaporation shows a specific relationship between oxygen and deuterium (^{2}H), so that:

$$\delta^2H = 8\ \delta^{18}O + 10\ (‰) \ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots\ (1)$$

The equation 1 is well-known as the Global Meteoric Water Line (GMWL). Globally, the isotopic composition of ^{18}O and ^{2}H is mostly determined or influenced by the evaporation from the oceanic surface and by the liberal rainout process from the ocean towards continental (or inland) regions (Mazor, 2005; Aggarwal et al., 2009). During the rainout process from the cloud to the ground evaporation may occur and most significantly during overland flows and within the subsurface soil zone during infiltration. Generally waters that had been subjected to evaporation had a slope in the order of 2-5 and these do not obey equation 1 by plotting or deviating from the GMWL (Aggarwal et al., 2009). Under circumstances where groundwater samples signify similar isotopic signature with rainfall, direct rainfall infiltration through preferential flow such as fissures, fractures, and decayed root channels is assumed based on the isotopic composition. Thus, it is possible to observe if there was direct recharge from rainfall or indirect or induced recharge from surface water bodies after evaporation has occurred through observing and analysing the ^{2}H versus ^{2}O plot in Figure 3 (Mazor, 2005).

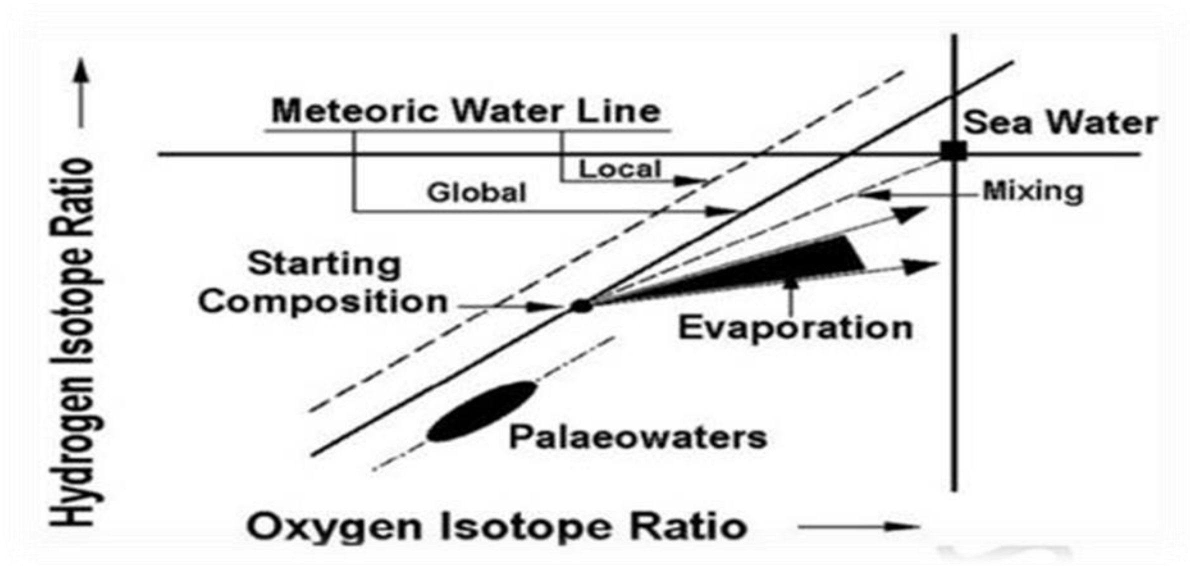


Figure 3: A schematic representation of $\delta^{2}H$ versus $\delta^{18}O$ in natural waters. Precipitation plots along the MWL with a slope close to 8, while evaporated samples deviate from the MWL with a slope from 2-5. The origin of vapour determines the value of the intercepts with the average value of d in oceanic precipitation close to 10 (source: Aggarwal et al., 2009).

Deviations of δ^2H and $\delta^{18}O$ from the standard are measured and plotted against one another. The equivalent diagram for rainwater produces the meteoric water line. The meteoric water line is normally used as an appropriate reference line for understanding and tracing local groundwater origins and movements (Mazor, 2005). The common practice involves plotting groundwater data on as δ^2H and $\delta^{18}O$ diagrams, along with the meteoric line of local precipitation as reference line. The method requires rainfall time series data of δ^2H and $\delta^{18}O$ to obtain the local meteoric water line, even though already available data can be used. This method acts as a supportive tool to other techniques such the chloride mass balance.

^3H- tritium content

Tritium is a radioactive isotope of hydrogen expressed as tritium unit (TU) where 1 TU equals 10^{18} atoms of hydrogen. Tritium dating technique is based on that all precipitation contains a small amount of radioactive tritium. Therefore, the rain water that successively recharge the aquifers is made up of tritiated $((3H)^1H^{16}O)$ water in which has a concentration decreasing with a half-life of 12.43 years (Vogel et al., 1974; Aggarwal et al., 2009; Levin and Verhagen, 2013a in Abiye, 2013). The advantage of using the technique is that tritium (^3H) its original value decreases only through radioactive decay.

The technique is useful in identifying or characterising between old water probably recharged prior 1950`s and the water that has been recently recharged into the aquifer (post 1952 waters). The disadvantages are that tritium has a low half-life (12.43 years) and its levels in the atmosphere have decreased since the bomb test stopped in 1960s. Therefore, appropriate measures at which the tritium dating technique is relevant is only about 4 to 5 half-lives, as a result limiting the applicability of the technique to only to recently recharged waters (Levin and Verhagen, 2013b in Abiye, 2013).

Based on the above discussion the time taken since the tritium rich water was recharged into the aquifer, hence being isolated from its source (precipitation) can be estimated from the tracer concentrations (TU). The estimated residence time or age is normally equated to the apparent age of groundwater. According to Levin and Verhagen, 2013b in Abiye (2013) quantifiable mount of tritium in groundwater (boreholes) are an indication of active or recent recharge events, while if certainly there are no detectable amounts of tritium in groundwater it is normally equated to no recharge or to very old water probably recharged some long time ago (implicated to residence time).

Infiltration Test

Infiltration test is normally conducted to provide better understanding of the hydrological functioning or to quantify soil intake properties. Various techniques are available and include Single ring infiltration test, Tensionometers and Double ring infiltration test (DRIT). According to Johnson (1963), no general technique of the available that is applicable at all field conditions or problems irrespective of the research that has been done to measure infiltration rates. Double Ring Infiltration Test (DRIT) is one of the widely accepted techniques to estimate infiltration rates and the technique is describe Johnson (1963); Brouwer et al. (1988); ASTM (2003).

DRIT is applied through driving the double infiltrometer rings into the soil to a depth of 6 cm without disturbing the soil surface. Normally a thirty cm and sixty cm diameter rings are used. The dual ring infiltrometers ensure a vertical ‘one-dimensional” flow of water into the soil. Normally, two operational techniques that can be utilized with the DRIT; the constant head experiment and the falling head experiment, with the latter being utilized in the study

When conducting the constant head experiment, water is added into the inner and the outer ring continuously at known flow rate to sustain a constant head in both rings during the course of the experiment. The volume of water used during the course of the experiment is then used to estimate infiltration rates. On the other hand, when conducting the falling head experiment, water is added to both rings (inner and the outer ring) to a desired level or depth and the time it takes for the water level in the inner ring to reach or fall to a certain depth is recorded. Then, the elapsed time since initial water level depth dropped to a maximum depth in conjunction with the changes in water level depth are used to estimate infiltration rates.

DRIT involves the use of idealistic water depths. The weight of the water above enforces water through soil surface depending on the applied pressure. Constant head experiments largely produce abnormal infiltration rates, while the falling head experiments result on varying infiltration rates due to the changes in water level depth over time (Gregory et al., 2005; Nichols et al., 2014).

2.6.2. Quantification of groundwater recharge rates

Chloride Mass Balance approach

Chloride is hydrogeochemical conservative among common ions. Generally, when the overall water input is supplied into the soil, some of the water will be lost through evapotranspiration, some used by plants, and only a portion percolates to successively reach the surface of the water table. During the process whereby water is lost through evaporation and transpiration, the chloride content is conserved within the soil and in the fraction of water that remains. A fraction of this water percolates through the subsurface zone to reach the water table surface. The fraction that successively reaches the water table surface contains dissolved chloride and therefore is highly saline. The chloride levels at each location reflect the relative amount of water returned to the atmosphere by evapotranspiration (Mazor, 2005).

Therefore, the Chloride Mass Balance (CMB) approach is based on the assumptions that the initial input of atmospheric chloride content is conserved during rainfall-infiltration in subsurface zone (Xu and Beekman, 2003; van Wyk, 2010). These assumptions are mainly based on the specifics that the chloride ion is highly soluble and non-absorbing and chemically conservative during recharge processes.

As a result three approaches have been developed to quantify recharge rates using the technique as follows (Adams et al., 2005):

- “To measure the chloride content in rainfall samples and in unsaturated zone to estimate the moisture flux”.
- “To measure the chloride content in groundwater and in rainfall samples to estimate the total groundwater recharge”.
- “To compare the above techniques to describe the recharge mechanism”.

For the purpose of the project only the second approach will be investigated to provide quantitative estimates of groundwater recharge. The approach is outlined as follows.

The CMB approach utilizes the chloride concentrations in precipitation to that in groundwater. Total Recharge (R_T) is estimated by dividing the input flux (the product of the chloride concentration in the precipitation multiplied by the amount of precipitation over the study area) divided by the chloride concentration in groundwater. The Equation is as follows:

$$R_T = \frac{PCl_{rfl}}{Cl_{gw}} = mm/yr \quad \text{...} \textbf{(2)}$$

Where:

P precipitation in (mm/yr)

$\boldsymbol{Cl_{rfl}}$ is the chloride concentration in rainfall (mg/ℓ)

$\boldsymbol{Cl_{gw}}$ is the chloride concentration in upper groundwater (mg/ℓ).

The advantage of using the method (equation 2) is that the method is independent of recharge mechanism and that the chloride concentration during rainfall-infiltration is conserved. Uncertainty arises due to unknown chemistry of rainfall from the previous rain events and the contribution from other sources.

2.7. Summary

In-depth analyses of groundwater recharge estimation techniques that can be used in Ramotswa transboundary aquifer were evaluated. This began with the classifications and concepts of groundwater recharge; further, seconded by the commonly used methods to assess groundwater recharge in arid and semi-arid regions of Southern Africa. The most feasible techniques for groundwater recharge estimation were acknowledged. Six techniques were identified and reviewed to evaluate groundwater recharge process and to quantify groundwater recharge rates. However, only one of the six techniques was studied with regards to groundwater recharge estimation in the study area. Basically the review of the techniques discussed the theory behind each technique such as applicability and reliability of the technique with regards to achieving the study objectives.

Chapter 3

Study Site

3.1. Introduction

In order to develop an understanding on the processes governing recharge and to quantifying reliable groundwater recharge estimates in the study area, it is essential to understand both surface and sub-surface complexities defining the rate (movement) and the magnitude of water recharged into the aquifer. Such complexities include the lithology, structure and understanding the hydrogeological settings of the study area. The section below evaluates the characteristics of the Transboundary Ramotswa Dolomitic Aquifer through reviewing the physiography, climate, land-use and vegetation, surface-water hydrology, geology, and the hydrogeology of the Ramotswa Dolomitic Aquifer. The review begins with a short overview of the Limpopo River Basin (LRB) as follows.

3.2. An overview of the Limpopo River Basin

The Limpopo River Basin (LRB) is one of fourteen major shared river basins located in the SADC region. It is one of the largest drainage areas in the region with an estimated total catchment area of 416 296 km^2 and is shared among four riparian countries (Figure 4): Botswana, Mozambique, South Africa and Zimbabwe (Zhu and Ringler, 2010). South Africa occupies the largest share of the basin followed by Mozambique, Botswana and Zimbabwe (SADC, 2003; LBPTC, 2010; Zhu and Ringler, 2010) as shown in Figure 4 and Figure 5. Geographically, the basin is located at latitudes 22° and 26° S and longitudes 26° and 35° E.

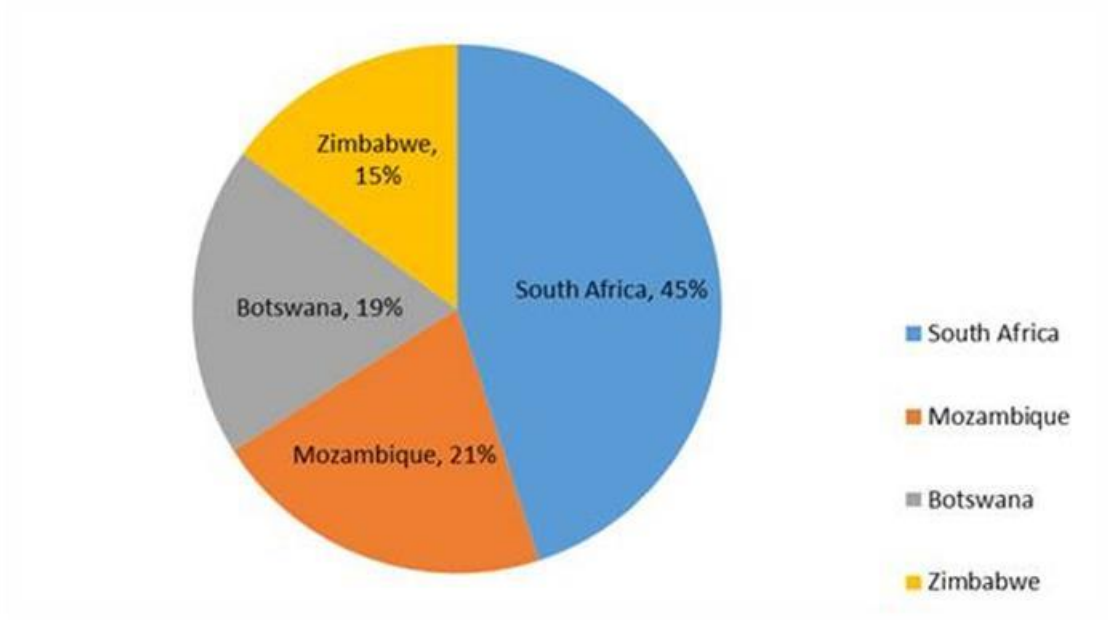


Figure 4: The percentage of total catchment area of the Limpopo River Basin within each of the four riparian countries (Source: Zhu and Ringler, 2010).

The climate of the LRB varies from arid (desiccated desert like environments further inland south of Zimbabwe) to semi-arid (tropical dry savannah within the South East Districts of Botswana) and to sub-humid (alongside the coastline of Mozambique) (Zhu and Ringler, 2010). The climate varies as a result of the seasonal movements of the Inter Tropical Convergence Zone (ITCZ) and associated high pressure systems, the distance from the Indian Ocean and the altitude above sea-level (SADC, 2003). Annual rainfall varies from 200 mm/a to 1200 mm/a, with a mean annual of about 530 mm, while evaporation is triple the amount of rainfall per annum in most areas within the catchment. Further, the rainfall is unequally distributed spatially with high seasonal variation during wet and dry seasons, and about 95% of these rainfall events occurring during summer months from October-March. Summer months (wet season) are characterized by high and short thunderstorm summer rainfalls, while winter months are characterized by cold, dry weather conditions (FAO, 2004; LBPTC, 2010; Altchenko et al., 2016). The region is prone to droughts and floods as a result to anti-cyclones or short intense thunderstorm rainfalls.

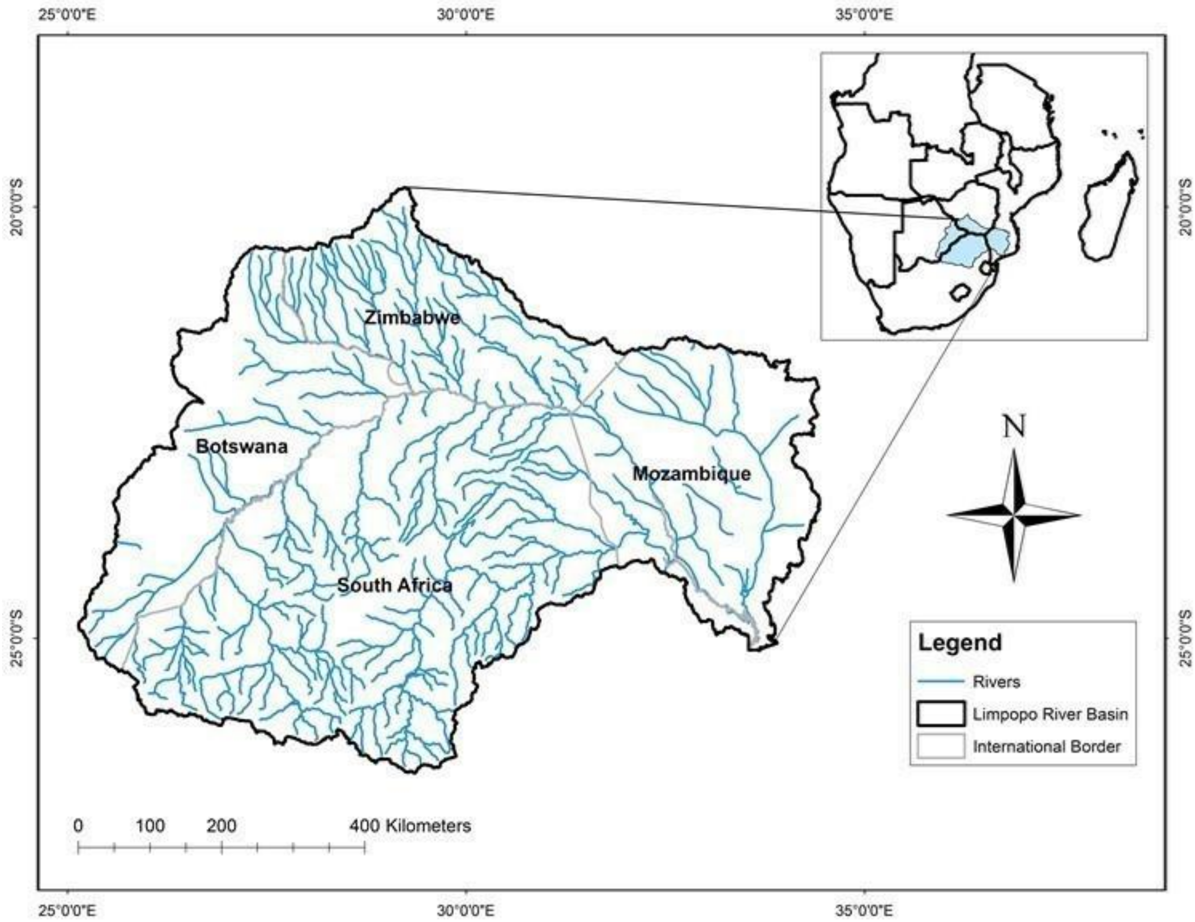


Figure 5: Limpopo River Basin shared among the four riparian countries

The basin is a home to more than 14 million people with more than 50% of the population residing in rural or undeveloped areas where more than 70% of the population lacks access to improved sanitation (LBPTC, 2010). The basin`s water resources are mainly utilized for irrigation, mining, power generation, and for both urban and rural water supply (FAO, 2004; LBPTC, 2010). Irrigation consumes the largest share of the basin`s water resources. Most of this water comes from groundwater through irrigation boreholes and dug wells (LBPTC, 2010).

The state of surface water quality is regarded as impacted although not severe. According to McMullen and Jabbour (2009) the basin is well-thought-out to be in a form that water demand or withdrawal exceed or are close to exceeding water availability. Increasing water scarcity compounded by climate variability and change, is one of the limiting factors for sustainable economic development (Zhu and Ringler, 2010).

Groundwater occurrence is mainly through alluvial sand aquifers along the Limpopo River. Independently from these aquifers, a series of aquifers where the water resources of such particular aquifers are shared between two or more countries exist and these are (Owen, 2011):

- Ramotswa Dolomitic Aquifer; (karst dolomitic: shared between South Africa and Botswana) (under the current investigation)
- Tuli Karoo Sub-Basin (sandstone units: shared among Botswana/South Africa/Zimbabwe) and
- Limpopo Pafuri Basin (alluvium: shared among South Africa/Zimbabwe/Mozambique).

The transboundary Ramotswa Dolomitic Aquifer is one of most high yielding aquifers within the basin and the most productive aquifer in Botswana. As part of the ongoing GRECHLIM project, which intends to explore the potential of the transboundary Ramotswa dolomitic aquifer, section below evaluates the characteristics of the transboundary Ramotswa Dolomitic aquifer to provide better understanding on the factors and on the processes governing recharge mechanism within the greater Ramotswa area.

3.3. Location and Coverage of the Transboundary Ramotswa Dolomitic Aquifer

The transboundary Ramotswa dolomitic aquifer, often karstic, lies within the upper catchments of the LRB at latitude 24˚ 40'0" S to 25˚ 30'0" S and longitude 25˚ 30'0" E to 26˚ 10'0" E (Figure 1).The aquifer is shared between South Africa and Botswana. In South Africa, the aquifer is located within the northern part of the country precisely in the North West Province in and around the Supingstand and (upper) Dinokana area, and in Botswana; the aquifer locates within the South-East Districts (SED), immediately from the Ramotswa border between South Africa and Botswana all the way to the Lobatse international border gate with South Africa. The ephemeral Ngotwane River is one of the major tributaries of the Limpopo River and it forms an international border between South Africa and Botswana within the upper parts of the transboundary Ramotswa dolomitic aquifer.

The area under investigation comprises of the dolomite outcrop, as well-defined by the existing surface geology maps in and around the Supingstand, Ramotswa, Lobatse, and the upper Dinokana area. A 5 km buffer zone around the rim of the dolomite outcrop was made in accordance to the airborne geophysics flying zone (Altchenko et al., 2016). Precisely, the extent of the aquifer is not well defined; however, Altchenko et al. (2016) approximated the aquifer extent to be about 1500 km^2.

3.3.1. Topography

The greater Ramotswa area is characterized by relatively flat topography with fairly hilly slopes ranging from 1000 m.a.s.l. to 1550 m.a.s.l. The hills and escarpment in and around the upper Dinokana area (South Africa) form the highest peak within the study area, while Otse hills (1491 m) and Monalanong hills (1494 m) form the highest peak within the area in Botswana. In overall, the elevation ranges from ±1000 m.a.s.l to ±1550 m.a.s.l. The elevation appears to be decreasing from high in the Lower Ramotswa study area (Upper Dinokana and Lobatse) toward the upper Ramotswa study sites in and around Ramotswa and the Supingstand area. Hills and escarpments are believed to be fragments of erosion successions which initiated in late Tertiary (Staudt, 2003).

Elevation within the dolomite outcrops is relatively flat throughout the study ranging from ±1000 m.a.s.l to ±1400 m.a.s.l. The surface topographic map for the greater Ramotswa area is shown in Figure 6. The surface contours are plotted at 100 m contour interval. Hilly slopes such as shown in Plate 1 could act as potential sources of recharge where permeable fracture exist, while also they could also enhance surface flows (runoffs), which will then recharge on the stream bed or concentrate where surface depression prevail and resulting on recharge.

Plate 1: Hill slope in the south of Lobatse.

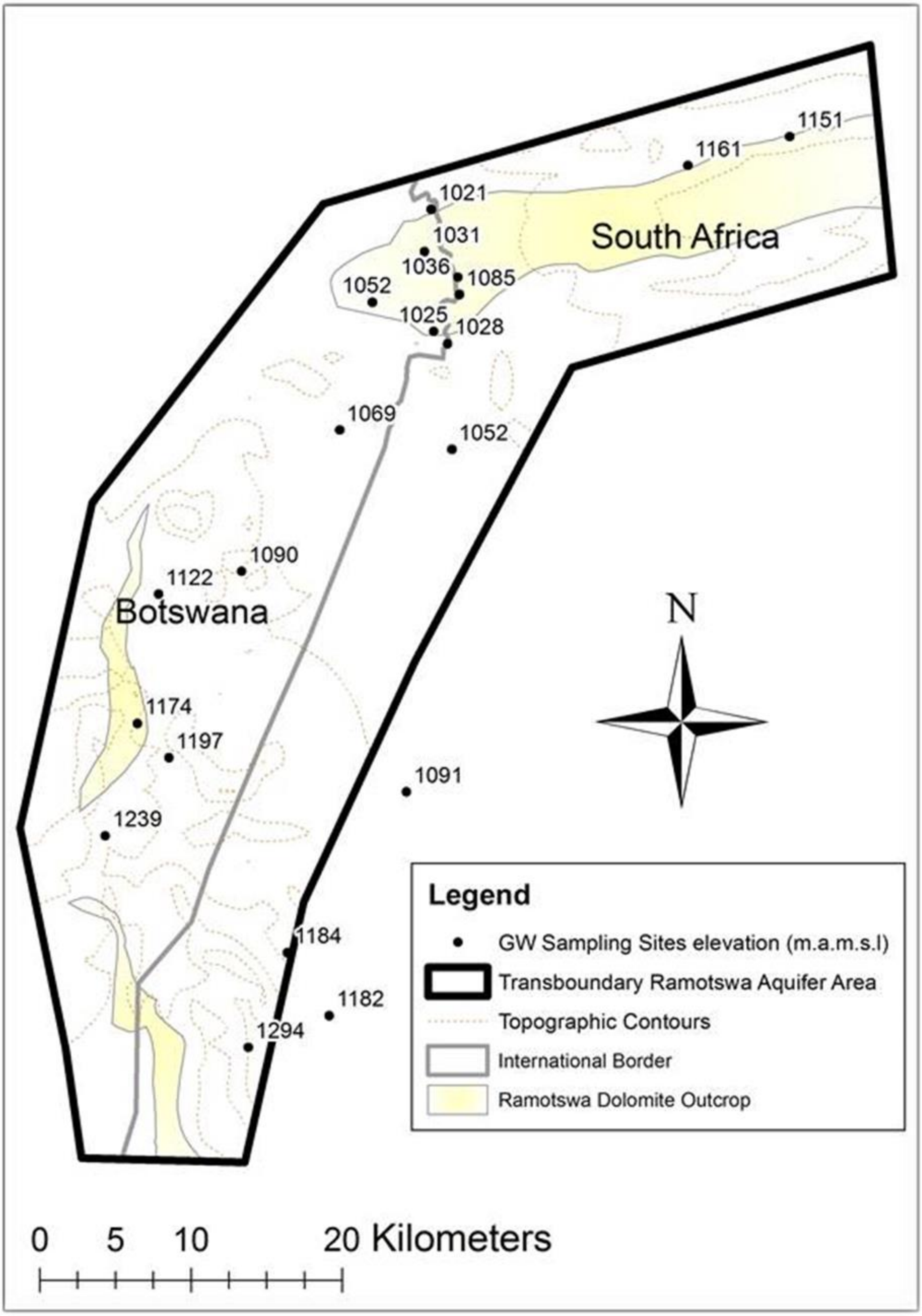


Figure 6: Elevation contours at 100 m contour interval within the greater Ramotswa area

3.3.2. Climate

The greater Ramotswa area falls within a semi-arid climate region of Southern Africa, where rainfall is sparse with high seasonal variations during wet and dry seasons. Wet (or rainy) season occurs during summer months, October to March and is characterized by short, intense convective storms. Dry seasons occur during winter time (April - September) and are characterized by dry cold weather conditions. Governing the variation in seasonal rainfall is the latitudinal movement of the ITCZ, which migrates to south of the equator during summer months and back to the north of the equator in winter (Mphale et al., 2014).

The results presented below were generated from climatological data from the Department of Meteorological Services of Botswana, the South African Weather Service and the RESILIM Project: Potential Role of the Transboundary Ramotswa Aquifer; Baseline Report - 15th June 2016 Altchenko et al. (2016).

Precipitation

A number of rainfall stations exist within the greater Ramotswa area, however, only three of these locate within the Transboundary Ramotswa Aquifer area: the Lobatse weather station, Witkleigat weather station and the Ramotswa weather station. The Madikwe and the Boschrand weather stations were also used to cover up the Supingstand area which had no rainfall data available.

Daily rainfall varies across the study area (Figure 7 – Figure 12). Significant daily rainfall was recorded around the year 2000, which exceeded 200 mm in Witkleigat and Ramotswa in a day (Figure 7 and Figure 9). Furthermore, rainfall of up to 170 mm was also recorded far North East of Ramotswa and Boschrand (Figure 8). However, the Lobatse station (Figure 10) far south of the study did not record any significant rainfalls around the period where other stations recoded significant rainfalls. Generally, daily rainfall varies from 0.5 mm to 50 mm/day with some years having extreme rainfall above 100 mm up to 240 mm. Such high rainfall contributes to significant parts of recharge into the aquifers (Braune and Xu, 2005).

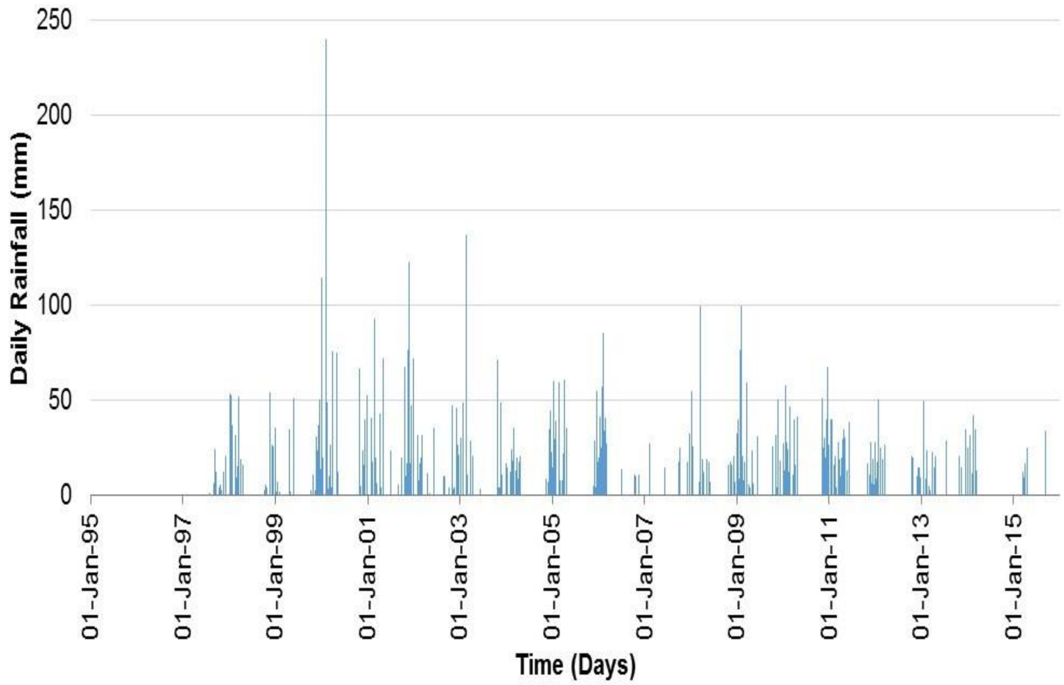


Figure 7: Witkleigat daily rainfall data from 1995 to 2015

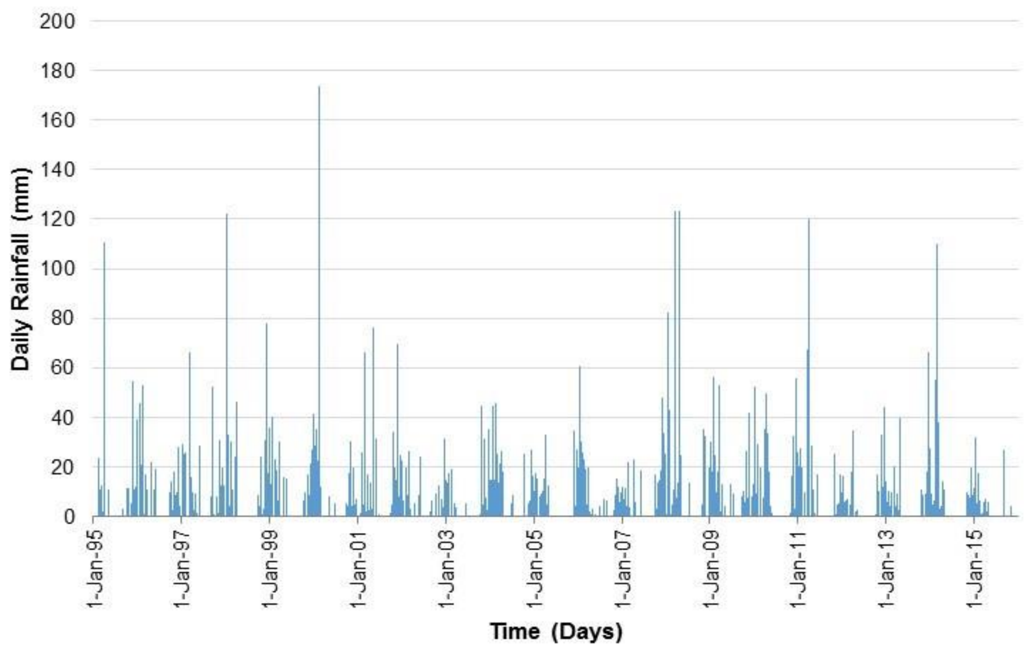


Figure 8: Boschrand daily rainfall data from 1995 to 2015

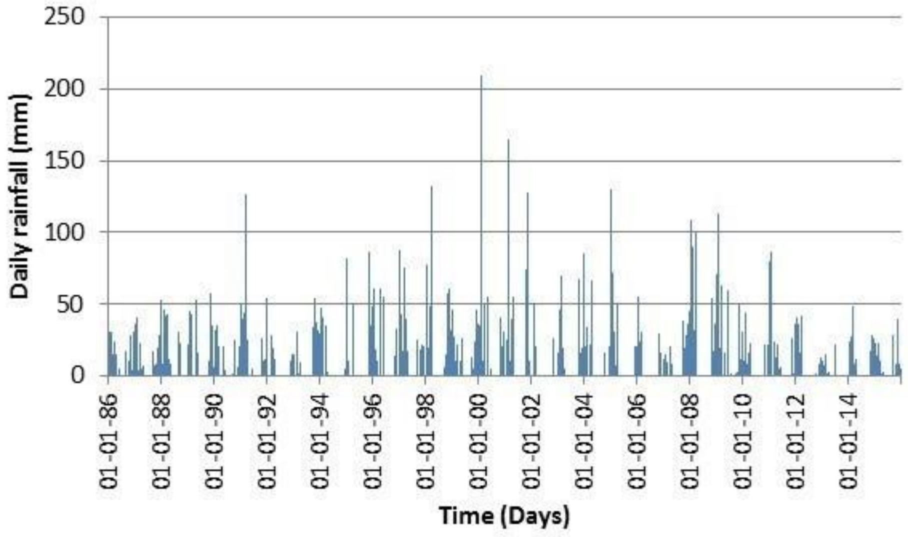


Figure 9: Ramotswa daily rainfall data from 1986 to 2014

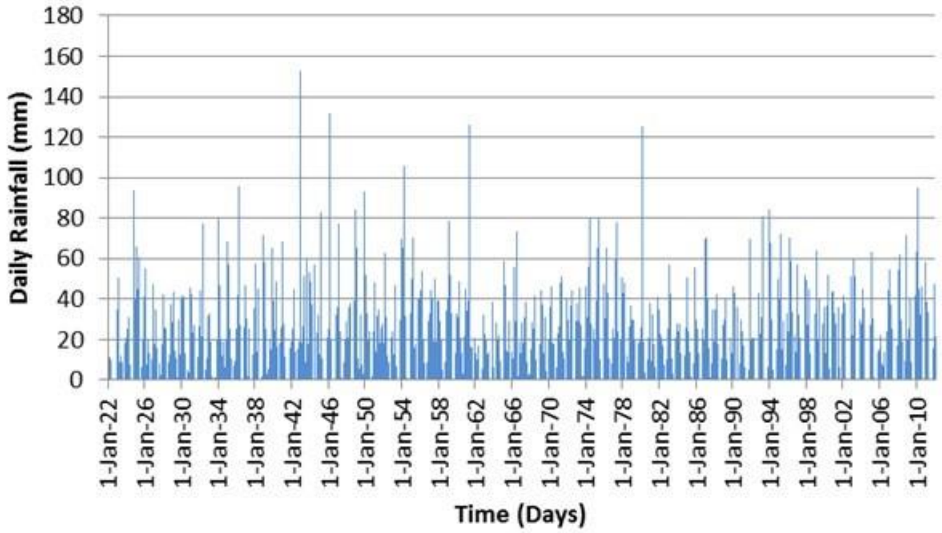


Figure 10: Lobatse daily rainfall data from 1922 to 2010

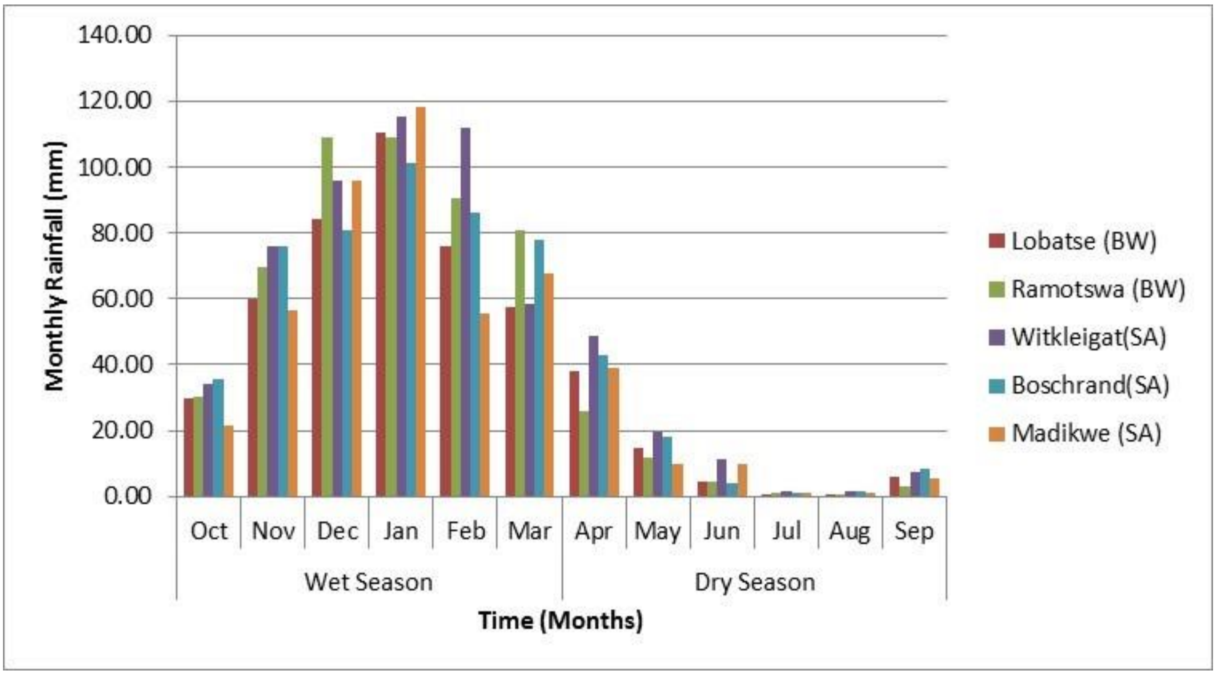


Figure 11: Monthly average rainfall data in a sequence of a hydrological calendar.

Monthly rainfall varies from about 0.1 mm (Lobatse) – 48.52 mm (Witkleigat) in winter and from 21.33 mm to 118.24 mm (both at Madikwe station) in summer months across the basin. Summer months are characterized by short, intense convective storms. Most of the rainfall occurs during the summer months, where highest temperatures are recorded; hence, significant portions of this rainfall are lost into the atmosphere through evaporation. However, these types of rainfall play a significant role in recharge the aquifer in a short space of time. Winter months are characterized by dry cold weather condition. In some stations the records reveal that rainfall does not occur at all during these months (winter), while some stations record a maximum rainfall of less than 20 mm monthly.

Plate 2: Hail storm (white cover) during the time of sampling in South Africa (upper Dinokana area) August 2016.

Annual rainfall varies on a yearly basis across the basin (Figure 12) with annual rainfall ranging from ±200 mm to ±800 mm. The Ramotswa rainfall data and the Boschrand rainfall data show significant years (1998 – 2009) of high rainfall mostly above 400 mm a year while in some years it reached up to 800 mm in a year (1998, 2000, 2001, 2008 and 2009). While at the Lobatse rainfall station significant rainfall were also recorded, however, none of the recorded rainfall data could reach up to 800 mm a year as recorded in Ramotswa and Boschrand station.

After the rainy years from 1995 to 2001, annual rainfall decreased across the basin (all three stations) for about 5-6 years and during this time annual rainfall ranged from ±200 mm to ±600 mm across the study. While in 2008 and 2009 high rainfalls were recorded again. This clearly shows how the rainfall varies across the basin and an example would be the year 2000 and 2001, where annual rainfall exceeded 800 mm at Ramotswa while at Boschrand and Lobatse annual rainfall could not reach the 800 mm mark. However a year later in Boschrand annual rainfall reached 800 mm while in Ramotswa it did not and Lobatse rainfall decreased drastically from 698.2 mm to 291.3 mm for the year 2000 and 2001 respectively.

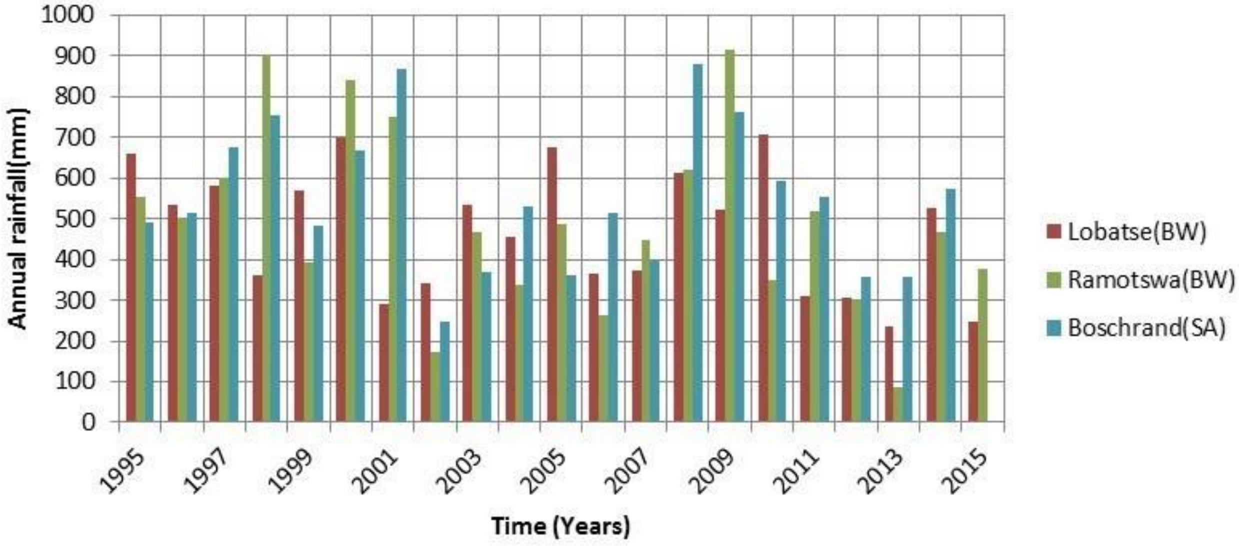


Figure 12: Annual rainfall data from three stations within the study area

Mean annual rainfall ranges (Figure 13) from 472 mm per annum in Lobatse up to 547 mm per year at Boschrand station which characterizes the area as a relatively dry land region. The mean at Witkleigat is about 607 mm per annum. However, Altchenko et al. (2016) gave a warning of possible misinterpretation of rainfall data due to missing data.

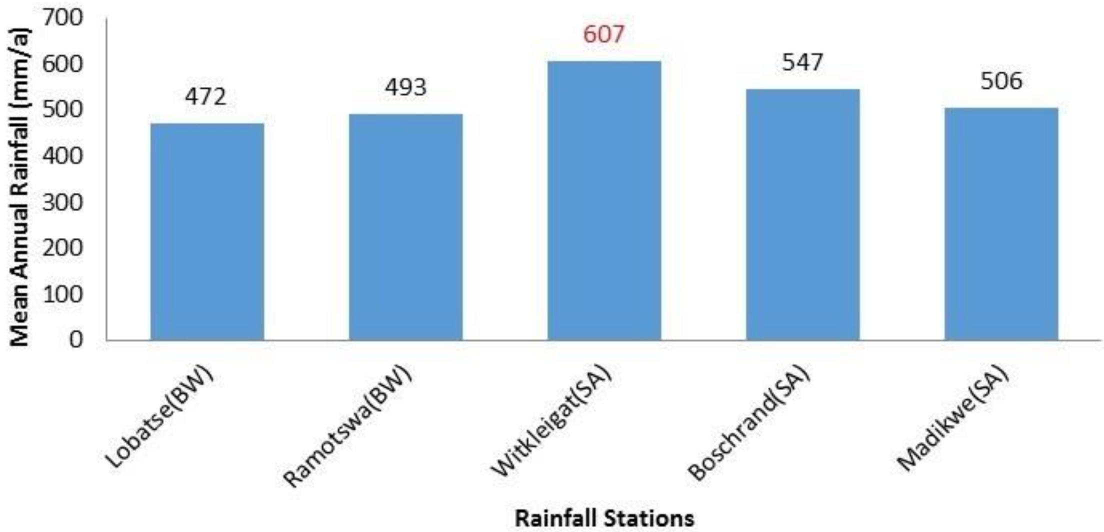


Figure 13: RTBAA Mean annual rainfall

Temperatures

Temperature variation is seasonal. Average daily temperatures vary from 30 ˚C – 32 ˚C during the day to about 16 ˚C – 20 ˚C at night (Altchenko et al., 2016). However, daily temperatures may reach up to 40 ˚C during summer months (October – February), while it may fall below 20 ˚C in winter (June – August). Monthly evaporation data ranges from 100 mm to 200 mm at Li-Maricopoort Dam, and from 100 mm to 250 mm at Molatedi Dam, where data is available. Annual evaporation ranges from 1000 mm to 2500 mm in Li-Maricopoort dam and from 1500 mm to ±3000 mm at Molatedi dam. Generally, evaporation exceeds mean annual rainfall by a factor of 2 to 3 times which could mean that rainfall recharge into the aquifer could only be possible in times where rainfall is high and evaporation rates are low. This is one of the major factors resulting on dry streams and also on low moisture fluxes recharge the aquifers.

3.3.3. Land-use and Vegetation cover

Land-use

The land use of the study area ranges from rural (undeveloped land) in and around the upper Dinokana area and the Supingstand area in South African to a semi-developed land (rural-urban transition zone) in and around Ramotswa – Otse – Lobatse located in the South-East Districts of Botswana.

A number of commercial and subsistence farming activities are found along the border between the two countries and along the Notwane River in both South Africa and Botswana. These farming activities largely depend on groundwater abstraction through dug wells (Plate 3 and Plate 4) and private boreholes, while some depend on rainfall (rain fed) (Plate 5).

Plate 3: A hand dug well in Notwane River, where the river forms an international border between South Africa and Botswana. The black pipe is used to abstract water for irrigation using water pump to a field 100 m away from the river.

In Botswana, specifically in the South-East Districts, which includes Ramotswa and Lobatse, most of the population resides around the transboundary Ramotswa area. In South Africa, the population is scattered along the area from the Dinokana, Maphephane and Gopane up to Moshana and Supingstand area. The upper Ramotswa area on the South African border is classified as protected area which includes the Madikwe nature reserve.

Plate 4: Subsistence farmers fed by the Notwane dug well, the black pipe abstract water from the Notwane dug well in plate 3.

Plate 5: Rain fed crop farming in Ramotswa, Botswana

Vegetation cover

In Botswana the study area lies within the Hardveld region characterized by acacia bush (such as tortilis, mellifera, erabescens) wooded grassland and the dry savannah vegetation such as mixed shrub and tree savannah vegetation (Staudt 2003; Altchenko et al., 2016). In South Africa, the predominant vegetation is more or less the same to that of Botswana and it is characterized by grassland and savannah biome (Altchenko et al., 2016). Based on the biogeographical classification regions in Appendix A; the transboundary Ramotswa aquifer lies within the Bankeveld and Bushveld basin, which extends into Botswana. Climate and soil type is the major influencing type controlling vegetation type in the region.

3.3.4. Surface Water Hydrology

The transboundary Ramotswa aquifer lies within the upper catchments of the LRB, precisely within the Notwane and Marico sub-catchments in Botswana and South Africa, respectively. The region lacks physical natural resource of surface water.

The Marico sub-basin originates in South Africa within the dolomitic plateau region of the North West province. The Marico River flows in a northerly direction and then north-easterly direction and joins the Crocodile River. Its major tributaries include the Klein Marico, Groot-Marico and the Ngotwane river(s). The upper Marico River is perennial and crosses the dolomite outcrop far north of the study area just outside the transboundary Ramotswa aquifer area. The dolomite springs feed into the Ngotwane River and Molopo River within the upper Dinokana area (or Lower Ramotswa study area).

The ephemeral Notwane River flows in a North – South direction forming an international border between South Africa and Botswana. The Gaborone dam catchment falls within the Notwane catchment although it falls outside the study area. The Dinokana springs are located far south of the study area in South Africa and play a major role in augmenting water supply to the upper Dinokana area where about 50% of the spring flow is used for supply for the upper Dinokana area and the remainder is left as an ecological reserve. Some dams are found within the area on the Notwane tributaries.

Plate 6: Vegetation cover at south Ramotswa study area (Lobatse)

3.3.5. Reginal Geology

The greater Ramotswa regional geology dates back to late Archaean platform progression of the Transvaal Supergroup. These were deposited on three structural basins of the Kaapvaal craton (Erikson and Alterman, 1998; Catuneanu and Erikson, 1999; Erikson et al., 2001; Moore et al., 2001), Griqualand West Basin, Kanye Basin and Transvaal Basin.

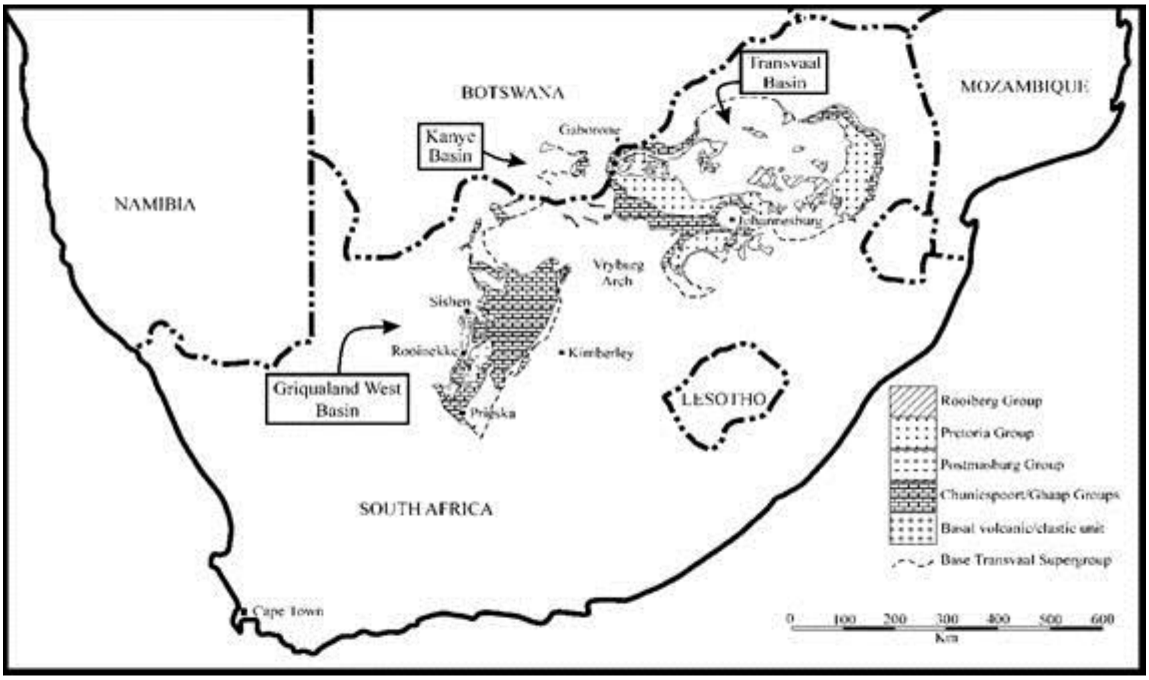


Figure 14: Distribution of the three structural basins of the Kaapvaal craton in South Africa and Botswana (Source: Moore et al., 2001).

The age of the Transvaal succession span approximately a period of 2.65 to 2.05 Ga and constitutes the sedimentary floor of the Bushveld igneous complex (Catuneanu and Erikson, 1999; Moore et al., 2001). The Transvaal Supergroup comprises the lowermost basinal Protobasinal rocks of the Ventersdorp age succeeded by the earliest carbonate platforms as well as the earliest superior-type banded iron formation and significant clastic sedimentary and volcanic rocks (Erikson and Alterman, 1998; Erikson et al., 2001). These rocks are preserved within three structural units of the Transvaal Supergroup: Probasinal rocks, Chuniespoort group and the Pretoria group (Foster, 1988; Catuneanu and Erikson, 1999; Erikson et al., 2001; Moore et al., 2001; Meyer, 2014).

Lithostratigraphy

An overview of the geology of the greater Ramotswa area within the Transvaal Supergroup is presented in Appendix B: momentarily from oldest formation to the youngest.

Protobasinal Rocks

A thin layer of the Protobasinal rocks occurs within the upper parts of the study along the Supingstand area, Ramotswa village and north of Otse Mountains (Appendix B). The Protobasinal rocks of the Ventersdorp age overlie the Witwatersrand Supergroup forming the basal formation of the Transvaal Supergroup with study area. These rocks were deposited approximately in 2714 – 2658 Ma in a deeper sub-aqueous facies, and alluvia- braided plain facies.

Chuniespoort Group

A linear outcrop of the Chuniespoort Group is found within the greater parts of the Ramotswa aquifer extending from Thabazimbi-Supingstand-Ramotswa-Lobatse-upper Dinokana area (Appendix B). The group subdivided into four stratigraphic formations: Black Reef Formation, Malmani (Ramotswa) Dolomites, Penge Formation and the Duitschland Formation.

Black Reef Formation overlies the Protobasinal rocks of the Ventersdorp age and forms the basal clastic rock formation of the Transvaal Supergroup mainly dominated by conglomerates to sandstones and mudstones (Foster, 1988; Catuneanu and Erikson, 1999). It is the oldest rock formation within the study area. The Black Reef Formation is subdivided into the lower finning-upward succession comprising basal conglomerates grading into mature quartz arenites and mudstones followed by an upper upward-coarsening sand succession (Catuneanu and Erikson, 1999; Erikson et al., 2001). The lower succession displays irregular thickness and tends to fill the basement palaeotopography, while the upper succession is transitional and conformable with the Dolomite formation (Foster, 1988; Erikson et al., 2001). Interbedded shale, dolomite, quartzite commonly occur within this zone (upper succession).

Malmani (Ramotswa) Dolomite comprises dolomitic rock formations which are subdivided into five formations based on the whether they consist of chert as well as the variety, absence or presence of stromatolite structure, intercalated shales and erosion surface (Button, 1973; Obbes, 2001; Erikson et al., 2006). Succeeding the Malmani dolomites is the banded iron stone of the Penge Formation.

Penge Formation comprises of micro-macro banded ironstones or formation with shard structures and subordinate interbeds of carbonaceous mudstones and intraclastic iron formation and breccias. These bended ironstones are succeeded by the carbonaceous mudrock and limestones of the upper most Chuniespoort group; Duitschland formation.

Duitschland Formation is the upper most (or the youngest) formation of the Chuniespoort Group and is made up of carbonaceous mudrocks, limestones and dolomite or dolomite mudstones with interbedded dolomites and quartzites, a thick erosional chert breccia body, two paleosols and two lava beds (Catuneanu and Erikson, 1999).

Pretoria Group

The Pretoria group sediments overlies the Chuniespoort group with numerous alternating mudrocks and sandstones units consisting of lesser diamictite or conglomerate members and volcanic units that present an alternating of alluvial sedimentation with epeiric sedimentation (Moore et al., 2001). The group is subdivided into 12-14 formations with lowermost two formation being part of the transboundary Ramotswa aquifer. The lowermost two formations include the basal alluvial and lake deposit of the Rooihoogte formation and the shallow deep marine environments/deposits of the Timeball Hill formation also known as the Lephala formation (naming convention in Botswana).

Rooihoogte Formation comprises the basal in-situ karst-fill, alluvial fan, fan-delta and shallow-rift tectonics (Catuneaunu and Erikson, 1999; Erikson et al., 2001). Lacustrine settings and braided river systems.

Timeball Hill (Lephala) Formation is the second lowermost formation after the Rooihoogte Formation within the Pretoria Group and forms as the uppermost (youngest) formation within the Transboundary Ramotswa aquifer. The Timeball Hill or Lephala formation is subdivided into four successions of which only two are formalized within the formation (Key, 1983):

- The upper Argillites
- Polo ground Quartzite member
- Lower Argillites
- Bevert`s conglomerate member

The basal Bevert`s conglomerate consist of tabular, rounded chert pebbles in ferruginous (quartzite) grit matrix. Overlying the Bevert`s formation is the lower Argillites formation comprising the argillaceous rocks with subsidiary arenites, quartzites, siltstone, shales, mudstones and sandstones with rare Chert and grit layer (Ramotswa, Undated). The lower Argillites are overlain by the Polo ground quartzite member comprising manganiferous and manganese sediments (Key, 1983; Selaolo, 1985; Staudt, 2003). The uppermost (upper Argillites) part of the Timeball Hill formation consist of interlayered argillites with sandstones.

3.3.6. The Geology of the Transboundary Ramotswa Aquifer

Introduction

The transboundary Ramotswa aquifer lies within the western parts of the Transvaal Supergroup. The Transvaal Supergroup is characterized by three stratigraphic succession preserved within the Transvaal Supergroup (Altchenko et al., 2016):

- Rooiberg Group
- Pretoria Group
- Chuniespoort Group

The transboundary Ramotswa aquifer developed within the two lowermost groups of the Transvaal Supergroup; Chuniespoort group and Pretoria group (Figure 15 and Table 2). Within these groups two lithological formations exist within which the two aquifers are to be found: the transboundary Ramotswa Dolomitic Aquifer and the Lephala Aquifer which extend over a wide area as part of the Bushveld Basin

(Institute of Hydrology, 1986). Both aquifers are thought to be of low hydraulic connection via a predominately N-S trending fracture zone with the dominant feature of the system marked by anisotropy associated with high density of fracturing. The Ramotswa dolomite aquifer develops as the most productive and main water bearing rock formation within the region (Selaolo, 1985; Altchenko et al., 2016). Section below mainly discusses the geology and hydrogeology of the main water bearing rock formation, the Ramotswa Dolomite formation.

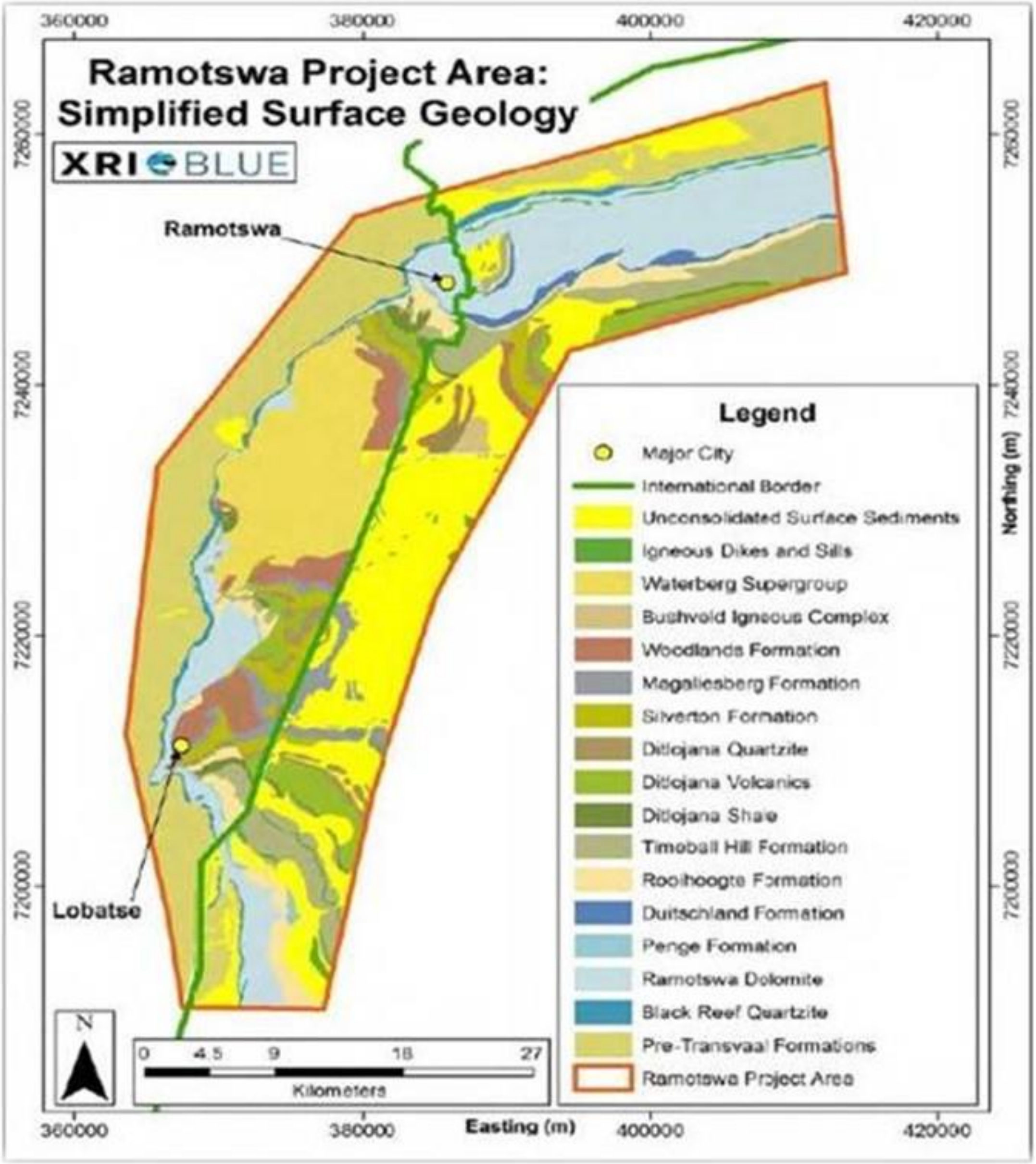


Figure 15: Simplified surface geology of the study based on the 1/25000 digitalized map of South Africa and Botswana, and AEM survey (Source: XRI Blue as cited in Altchenko et al., 2016).

Ramotswa Dolomite

The Ramotswa dolomites (Transvaal Supergroup/Chuniespoort group) correspond to the Malmani dolomites in South Africa (Table 2). A linear outcrop is seen across the basin extending from the Supingstand area (South Africa) to the South East Districts in Botswana which include areas of Ramotswa and Lobatse towards the border of South Africa, in the north of Zeerust or Mahikeng Town. The dolomite formation is subdivided into five stratigraphic formations based on whether consist of chert content (here referred to as chert rich) or does not contain chert (referred to as chert free) (Button, 1973; Obbes, 2001; Erikson et al., 2006). These sediments were deposited in the large Transvaal Basin during the Vaalian Erathum of the Proterozoic, while Chert-free units were being deposited in subtidal environments and Chert-rich units were deposited or are from the intertidal to supratidal zone (Meyer, 2014). The stratigraphic sequence of the Ramotswa dolomite is showed in Table 2.

Table 2: Lithostratigraphy of the greater Ramotswa area within the Transvaal Supergroup (Catuneanu and Erikson, 1999; Erikson et al., 2001; Altchenko et al., 2016)

Age (Ma)		Thickness (m)	Stratigraphy				Lithology
PROTEROZOIC		~450-15000	TRANSVAAL SUPERGROUP	Pretoria Group	Timeball Hill Formation (**Lephala Formation**)		Mudrock, quartzite,
	~2720-2050	~10-150			Rooihoogte Formation		Quartzite, mudrock, Bevets conglomerate/breccia member
		270-660		Chuniespoort Group	Duitschland Formation		Carbonaceous mudrocks, limestone and dolomite
	~2500	10-150			Penge Formation		Banded iron stone
	2640-2500	~400			Malmani Subgroup (**Ramotswa Dolomite**)	Frisco Formation	Chert-free dolomite
		~600				Eccles Formation	Dolomite and Chert
		100-200				Lyttelton Formation	Chert-poor dolomite
ARCHEAN		300-500				Monte Cristo Formation	Chert-rich dolomite
		10-200				Oaktree Formation	Chert-free dolomite
		25-30			**Black Reef Formation**		Quartzite, conglomerate and shale
	~2714 - 2658			**Protobasinal rocks** (Ventersdorp age)			Volcanic rocks

Stratigraphy

Overlying the basal clastic rocks of the Black Reef Formation within the Chuniespoort group and forming the basal formation of the Ramotswa dolomites are the blueish grey to grey dolomite or recrystallized dolomites of the Oaktree formation. The Oaktree formation are characterized as chert-free dolomites with some distinct wrinkled grey or brown appearance whether that is reminiscent of the elephant`s skin (Plate 7).

Plate 7: Weathered dolomite rock with an unusual pattern of cracks and channels, similar to the wrinkled elephant skin “elephant hide”; South Lobatse.

Overlying the chert-free dolomites of the Oaktree formation is the chert-rich dolomites of the Monte Christo formation, which consists approximately 300 m – 500 m chert-rich dolomites which can be micritic or recrystalised and can vary from brown, grey, cream, pink and blueish grey (Foster, 1988). Interstratification of chert is occurs at all scales. Algal sediments feature such as docimal stromatolites, columna stromatolites and spheroidal oncolites prevail including sedimentary structure such as ripple marks, interference ripples and oolites (Foster, 1988).

Succeeding the chert-rich Monte Christo formation is the 100 m – 200 m chert-poor dolomites forming the Lyttleton formation. The Lyttleton formation consists of shale and quartzite rock types. Foster (1988) described the formation as lithological similar to the Oaktree formation. Some grey/blueish dolomite or recrystalised dolomite which weathers to chocolate brown or grey brown colour is present.

Succeeding the Lyttleton formation are the chert-rich dolomites of Eccles formation. The characteristic of the chert-rich Eccles formation consist of columnar and domical stromatolites algal laminations and abundant ripple marks. The formation is about

600 m thickness. The Eccles formation is overlain by the Chert-free dolomites forming the Frisco formation. The Frisco formation ~400 m thickness forms at the upper most rock formation within Ramotswa dolomites.

Structure

Dykes

Several studies have been conducted to characterize the dolomite compartment within greater Ramotswa area. Holland (2009) (as cited in Meyer (2014)), interpreted the available information up to 2009 within the lower Ramotswa study (Groundwater Region 10 in South African GMA). The study was further subdivided into 3 compartments in which were mapped.

The Ramotswa dolomitic aquifer locates within the western parts of Groundwater region 10 in and around the Dinokana area (catchment). Further, the Region 10 is subdivided into 10 groundwater management areas (GMA) in which the study lies within the Dinokana GMA. The Dinokana GMA has an estimated total area of 273 km^2 which is further subdivided into 4 management areas. Subdivisions are established based on three different dyke compartments (Meyer, 2014) as follows:

- “The Tweefontein dyke (Tweefontein – Dinokana)”
- “The Dinokana dyke (Dinokana – Skilpad eye)”
- “An unnamed short dyke forms the boundary between the Skilpad eye and the Skilpadhek GMUs”.

These dykes are thought to be in an upright or near vertical mafic bearing. These are layered intrusion characterized by low permeability and as a result acting as impermeable barrier to groundwater flow across the dolomite aquifers (Meyer, 2014). Bredenkamp (2002) suggests fracturing at certain depth due to tectonic activity does occur thereby enhancing some trans-compartment flow (Meyer, 2014). Also dykes near surface are suspected to be weathered which could possible allow groundwater to flow across them.

Crockett (1969) and Key (1983) as cited in Staudt (2002), analysed the complex faulting in the area between Ramotswa to Lobatse (Botswana). Further, Dievorst (1988) reinterpreted those structures in a more detailed mapping approach and as a result two phase folding were suggested; the first with an approximately East to East-North-East trending axes and the second with a South West trending axes. Tilted block faulting occurred concurrently with the second phase of folding resulting in a reorientation of the first phase folding axes due to the rotations and tilting of blocks along normal fault planes. From this the tectonic history of the study area can be divided into five phases (Staudt, 2002):

- “Deposition of the Transvaal Supergroup over a more extensive basin than the present outcrop distribution”,
- “Intrusion of basal dolerite sills of the Bushveld igneous complex”
- “Folding along easterly plunging axes”
- “Refolding along South-South westerly trending axes”
- “Anti-Clockwise rotation of the Upper Transvaal strata and tilting to the North-West.”

Based on the Airborne Electromagnetic (AEM) data (Appendix C) collected by XRI Blue (2016), intrusive dykes were revealed from the magnetic data as linear magnetic anomalies. These dykes are thought to serve as either preferential flow paths for groundwater or as an aquitard that acts as a barrier or no flow boundary (intrude upwards in long sheet). Additionally, anomalies within the AEM data were interpreted to be possible dykes throughout the study area. Most of the interpreted dykes with the AEM data (Appendix C and Appendix D) were close to or in similar location as the linear magnetic anomalies, substantiating that the interpreted linear magnetic anomalies are dykes. Only a few were not in close proximity to the linear magnetic anomalies while some interpreted appeared to be offset by faulting, which may alter the expected groundwater flow regime near the dike in the faulted area. According to WRC (2003) most of these dykes acts as leaky barriers which allow only a limited amount of flow through them, and these are about 10 m to 30 m wide. Also these dykes are a major factor facilitating the spring development within the region and this occurs when a permeable layer with groundwater meets an impermeable layer (dykes) and the surrounding rock forcing groundwater to discharge at the surface.

Faults

Also from the AEM data collected by XRI Blue (2016), faults were interpreted (Appendix D) supplementing data from Botswana DGS and CGS South Africa. Some interpreted faults were present in the subsurface zone while they did not appear to extend to the surface. Only a few did extend to the surface.

Hydrogeology of the Ramotswa Dolomitic Aquifer

Two aquifer systems were identified within the greater Ramotswa area: Ramotswa Dolomitic Aquifer and the Lephala Aquifer. Both aquifers belong to the Transvaal Supergroup: Chuniespoort group and the Pretoria group, respectively. These aquifers are highly permeable and also high yielding aquifers thought to be linear features with high Storativity (Storage Coefficients) and Transmissivity values. The Ramotswa dolomitic aquifer progresses as the most productive, hence the most significant aquifer for development and while also it progresses as the most vulnerable to over abstraction and pollution due to its karst nature.

Chert-rich dolomites are often associated with high aquifer productivity (Altchenko et al., 2016), while chert-free dolomite formation are classified to be poor aquifer formations (Altchenko et al., 2016). Chert-rich dolomite layers like the Monte Christo and Eccles formation are best aquifers within the sequence with high permeability properties (WRC, 2003).

Groundwater occurrence is mainly through secondary porosity. Active groundwater circulation has favoured local Karstification along the structural lineaments producing high transmissivity and Storativity values (Institute of Hydrology, 1986; Staudt, 2002). Two karst zones have developed due to groundwater circulation: the upper karst zone and the lower karst zone (Institute of Hydrology, 1986; Staudt, 2002; Staudt, 2003; DWA, 2004; Altchenko et al., 2016):

- “The upper karst zone”
  - “Is approximately 20 m – 50 m (thickness)”
  - “Characterized by highly dense fractures and/ fissures”
  - “Consist of both shallow and deeps zones of Karstification zone due to fluctuations on the present day water level”
  - “Dolomite dissolution occurs preferable along the fracture but less chert dolomite also presents karst feature outcrops. Solution cavities have been filled with soil and separated by a layer of limited karst development and fissuring”
- “The lower karst zone”
  - “Thickness varies from 25 m – 50 m”
  - “Comprises ordinarily free infill material”
  - “Karstification occurs or is a resultant of old groundwater circulation or fluctuation in the water level. Solution cavities are open and generally do not contain mud or wad filing”.

Groundwater Flow

Groundwater level measurements were conducted during a two-fold field trips from the 25th – 29th of August 2016 and from the 1st – 5th of September 2016 in both South Africa and Botswana, respectively. However, due to inaccessibility of the sampling wells within the northern parts of the study, no groundwater level measurements were conducted and as a result the groundwater level flow direction map or potentiometric surface map presented below (Figure 16) is based on the sampling data completed in and around the upper Dinokana area in South Africa, and in Lobatse and Ramotswa area in Botswana respectively.

Figure 16 shows that groundwater levels are generally high (shallow) in the southern parts of the study and gradually decreases towards north and then towards the north-east direction within the upper Ramotswa study area. This shows that groundwater flows from the south in upper Dinokana area (South Africa) towards Lobatse in a northerly direction and towards the Ramotswa and Supingstand area in a north-easterly direction in Botswana and South Africa, respectively. Rather not often in Karst environments (Altchenko et al., 2017), the groundwater flow direction seems to follow the surface topography, which is relatively high the southern parts of the study and decrease north towards Lobatse Ramotswa and Supingstand area. However, some concealed form of connections between the upper and the lower Ramotswa study area is believed to have been established resulting on groundwater to flow in a NNE direction (Altchenko et al., 2017).

In agreement with the above, Beger (2001) constructed a potentiometric surface map for the area around Lobatse and depicted a groundwater flow direction from the south in Lobatse towards north to Ramotswa. Further correlating well with Beger (2001), Staudt (2003) produce a piezometric map in and around Ramotswa area and depicted a north – north-easterly groundwater flow direction imitating the slope and the course of the Ngotwane River (hereto referred to as Notwane River).

Aquifer Properties

Selaolo (1985) conducted both a short and long pumping test in Ramotswa wellfields and concluded that the results are of the upper karst formation within the dolomite can be characterized as unconfined aquifer. High Transmissivity (T) values we recorded in highly fissured and karstified dolomites. Exceptionally high T-values such as 4336 m^2/d were not considered to be representative of the linear feature as whole; these were mainly attributed by localized karstification especially along the Notwane River. High T-values associated with the short aquifer test were mainly due to water being pumped from the rock fissures while for the long tests, the T values are an expression of both the fissures and the rock matrix.

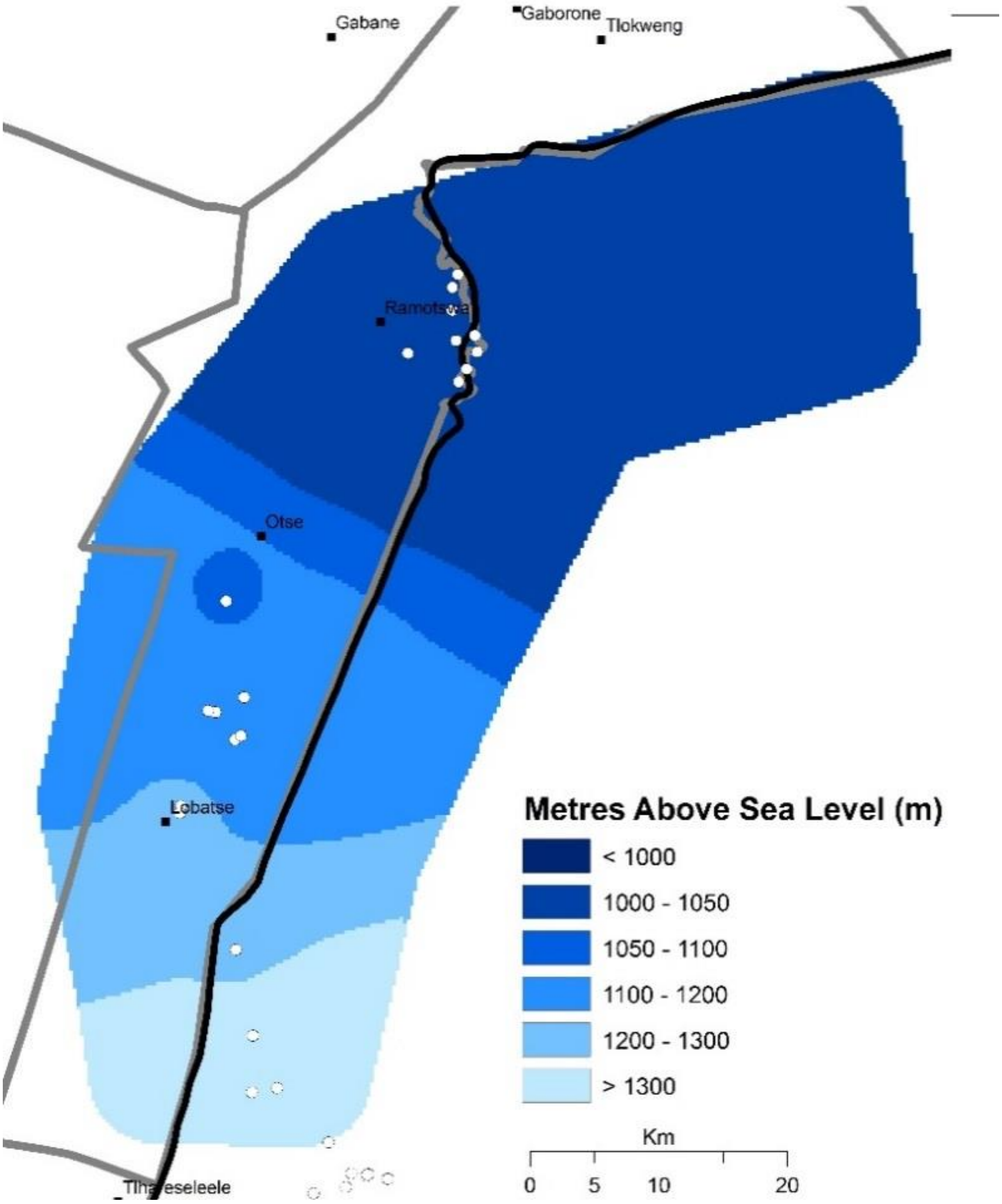


Figure 16: Transboundary Ramotswa piezometric map (Source: Altchenko et al., 2017)

Summarized Aquifer Properties

Table 2: Aquifer properties from short aquifer pumping test (Source: Selaolo, 1985).

Well	Short Test		Long Test		
	T (m^2/d)	Specific Capacity (m^2/d)	T Pumping (m^2/d)	T Recovery (m^2/d)	Specific Capacity (m^2/d)
4422	510	87		255	
4337	1600	600	300	455	300
4349	1320-1435	1030	900	470	900
4400	540		330	240	90~100
4336	4650-5730	2735	(poor data)	120	515
4423	1530-1760	1280	110	120	
4155			120		

Table 3: Storage co-efficient properties from the cone of influence technique (Source: Selaolo, 1985)

Well	Storage Co-efficient	Aquifer Type
4337BWA	0.026	Unconfined conditions
4349BWA	0.073	Unconfined conditions
440BWA	0.024	Unconfined conditions
4336BWA	0.02	Unconfined conditions
4422BWA	0.042	Unconfined conditions

Recharge and Discharge zones

To date, no detailed studies have been undertaken to assess groundwater recharge within the study area. However, recharge is thought to occur through the outcrops and river channels, mainly the Notwane River. River infiltration is thought to be the major source of recharge along the Notwane River and this has been observed where water level rose to about 11.34 m from 12.50 m to 1.16 m below the surface before the river flooded (Selaolo, 1985). Also, recharge is thought to occur through rainfall infiltration, more like diffuse recharge or as a result of surface depressions. Some studies suggest recharge from the South African side into the aquifer system in Botswana (Staudt, 2003). A series of springs are to be found in the South African side mostly in the south parts of the study area while there is no available information on discharge sources in Botswana. Discharge from the springs may also play a significant role in recharging the aquifer further downstream along the flow paths.

3.4. Summary

Physical and geological characteristics of the transboundary Ramotswa dolomitic aquifer have been reviewed and discussed, based on previous studies and on-going GRECHLIM project 2015-2018.

The Ramotswa dolomites correspond to the Malmani dolomites in South Africa. A linear outcrop was seen across the greater Ramotswa area lengthening from the Supingstand area (South Africa) to the South East Districts in Botswana, which include areas of Ramotswa and Lobatse towards the border of South Africa, in the upper Dinokana area. Five stratigraphic formations prevails based on whether consists of chert content or does not contain. Chert rich dolomites such as the Monte Christo and the Eccles formations are characterised as best aquifer formation with high permeable values.

The hydrogeological setting of the transboundary Ramotswa dolomitic aquifer seems to be very complex and based on the reviewed literature, the Ramotswa dolomitic aquifer can be characterised as an unconfined to semi-confined fractured and karstified dolomitic aquifer. Groundwater mainly occurs through secondary openings such as active groundwater circulation/karst development and fracturing. Active groundwater circulation is thought to have favoured local karstification along structural lineaments producing high storativity and transmissivity values. Dolerite dykes have compartmentalised (divide) the aquifer into various hydrogeological units while also facilitating spring development particularly in the southern parts of the study area. Groundwater recharge is mainly thought or seemingly occurs through the fractures that outcrop at surface and those are partially covered with permeable soil, dissolutions cavities and the ephemeral Notwane River. However, the complex nature of the geology and hydrogeology of the area including the spatially distributed rainfall which is worsened by high moisture deficits makes it very challenging to estimate recharge rates accurately.

Chapter 4
Methodology

4.1. Introduction

It is evident from the literature that quantification of groundwater recharge rates is practically complex, particularly in arid and semi-arid hard rock terrains where recharge rates are generally low compared to moisture fluxes such as precipitation and evapotranspiration. As a result the quantification of the rates of natural recharge has been largely achieved by means of indirect techniques.

To provide a detailed understanding on the quantification of groundwater recharge rates and on the processes governing recharge mechanisms in Ramotswa Dolomitic Aquifer, the study utilizes a combination of both qualitative and quantitative research approaches. Qualitative assessment provides insight into the governing processes controlling recharge mechanism, while quantitative assessment provides quantitative estimate of groundwater recharge within the aquifer. The physical and tracer approaches presented in Chapter 2 are undertaken in this study.

4.2. Research Approach

Three key strategic phases were implemented in the study as follows:

- Desktop survey
- Fieldwork
- Presentation of the results and analyses
    - Evaluation of the processes governing recharge mechanism
    - Quantification of groundwater recharge rates

4.2.1. Desktop survey

Desktop survey encompassed secondary data collection through the review of records as follows.

Review of Records

A review of the literature conducted within the greater Ramotswa provided an understanding on the processes governing recharge mechanism, factors influencing recharge, and on the groundwater recharge rates. Furthermore, the review assisted in identifying the appropriate techniques for estimating groundwater recharge in transboundary Ramotswa dolomitic aquifer. The reviewed literature included geological and hydrogeological maps, reports, unpublished university thesis and other database from the Southern Africa Development Community (SADC) groundwater archives, International Water Management Institute (IWMI), Limpopo Watercourse Commission (LIMCOM), Department of Water Affairs (Botswana), Department of Water and Sanitation (South Africa) and from Water Research Commission.

4.2.2. Field Work

Sampling site selection

Sampling sites based on the known dolomitic areas as well-defined by the existing surface geology maps were selected in and around Surpingstand, Ramotswa, Lobatse and the upper Dinokana area. A 5 km buffer zone around the rim of the dolomite outcrop was created as part of geophysical investigation conducted by XRI Blue. Within the buffer zone and the fly area zone (under geophysical investigation), borehole(s) drilled on to the karst or dolomite rock were selected for sampling and those that contained no dolomite or not regarded as karst were removed.

However, due to inaccessibility of some of the selected boreholes, we further decided to include some boreholes which were outside the 5 km buffer zone some of which were also drilled within the dolomite formation including the dolomite springs which are thought to be discharge sources from the dolomitic aquifers.

Data collection

A two fold field trip was conducted from 25th – 29th of August 2016 and from the 1st - 5th of September 2016 in South Africa and Botswana, respectively. A total of 31 groundwater samples were collected from the municipal water supply boreholes, monitoring boreholes, windmill powered boreholes, private (household) boreholes, dug well (Notwane dug well) and from the springs (eyes) (Figure 17). Monitoring boreholes were sampled using Grundfos submersible pump which was powered by a petrol generator (Appendix E).

Furthermore, a total of 65 soil samples at five sampling sites and at different depths at each site were excavated with a hand Dutch auger for chloride analysis (profiling) and soil moisture analyses. Sampling at each of the sites varied from 1 m to 2.5 m depth due to the underlying rock formation. Potential recharge sites such as the Notwane River were selected and excavated.

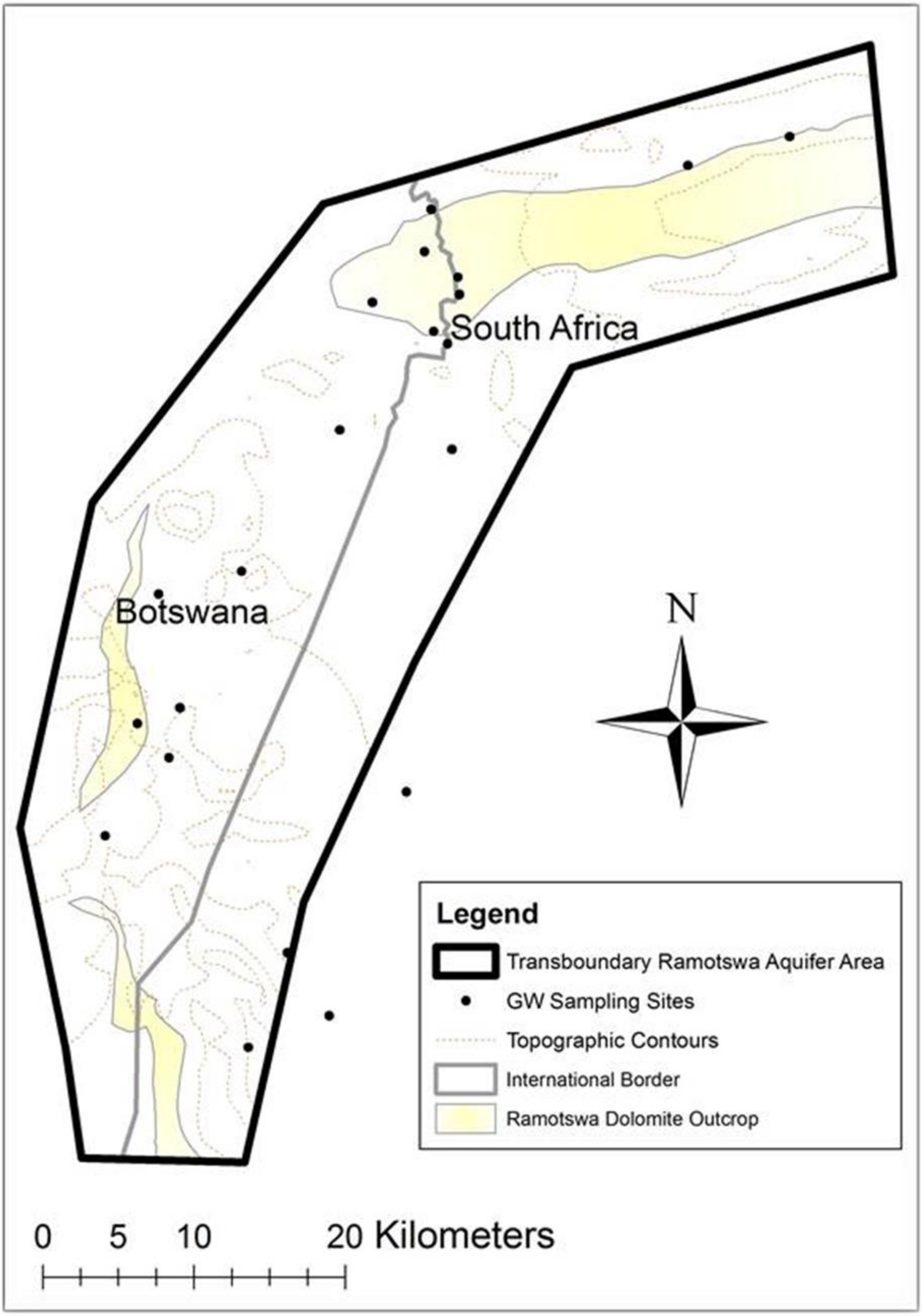


Figure 17: Groundwater sampling sites within the greater Ramotswa area (RTBAA)

Preservation and Analytical description

Hydrochemical Analyses

Samples were collected in tightly sealed 500 mℓ PET bottles from the municipal supply boreholes, monitoring boreholes and private boreholes. To preserve chemical signature the samples were stored in a cooler box at about 4^{o}C temperature immediately after sampling in the field. Samples were sent for analyses within 24 hours of sampling. Analyses were conducted at accredited labs: Water Utilities Corporations in Botswana and Resource Quality Information System at DWS in South Africa. Samples were analysed for major ions: Ca^{+2}, Mg^{+2}, Na^{+}, K^{+}, CO_3^{2-}, HCO_3^{-}, Cl^{-}, SO_4^{-2}, $CaCO_3$, F^{-}, N^{+}, P^{-3}, Si, NH_4^{+} and Total Dissolved Solids (TDS). In-situ water quality parameters such temperature, pH, TDS, EC were measured (with a Multi-parameter probe) immediately at each of the sampling sites as these parameters are subjected to drastic changes with time after water sample collection. Two sets of identical samples were sent to both labs for comparison of the results from the two labs, as a quality control check.

Environmental Isotope analyses

A total of 37 stable isotope samples were collected in 10 mℓ glass bottles and immediately covered with a black plastic to prevent sunlight which can cause evaporation of the samples thus causing isotope fractionation. The sampled sites included the dolomite springs, rainfall, dams and groundwater samples (monitoring boreholes, municipal supply boreholes, windmill powered boreholes and private boreholes). Furthermore, the samples were kept in a cooler box at approximately 4^{o}C and transported to the laboratory. Samples were analysed at Wits Hydrogeology laboratory using the LGR liquid Water Isotope analyser for $\delta^{2}H$ and $\delta^{18}O$ stable isotopes.

Tritium analyses

A total of 31 tritium samples were collected in 1000 mℓ PET bottles from an undeveloped area to semi-developed area in South Africa and Botswana respectively. Samples were then stored in a cooler box and sent for analyses on the second week of sampling. The analyses were performed at iThemba labs (Johannesburg) in South Africa.

Samples were distilled and subsequently enriched by electrolysis. After distillation, a volume of some 500 mℓ of water sample was mixed with 4g Sodium peroxide and introduced into the electrolyte cell. A direct current of approximately 10 – 15 ampere at 12V was then passed through the cell, which is cooled because of process heat generation. For liquid scintillation counting the samples are prepared by directly distilling the enriched water sample from the now highly concentrated electrolyte. A 10 mℓ volume of this distilled water sample is mixed with 11mℓ Ultima Gold LLT LSC cocktail in a counting vial. The sample is then placed in the Packard Tri-carb2770 TR/SC Low-level liquid scintillation analyser and counted 2 to 3 cycles of 4 hours. Detection limits for enriched samples were 0.2 TU (Butler et al., 2016).

Soil chloride analyses

Soil samples were stored in industrial plastic bags and kept in a box in a cool dry place to prevent sunlight (evaporation). Samples were analysed at Water-Lab (Pretoria) in South Africa. Samples were analysed with the Ion-Selective Electrode (ISE) for total chloride as chloride(s). Total chloride content in soil samples was determined by combusting a weighed sample in an oxygen bomb or vessel with a diluted base solution to absorb the chloride vapours. The bomb or vessel was then rinsed into a beaker with water and following the addition of ionic strength adjuster, the chloride was determined (E. Botha 2017, pers. Comm., 9 Jan).

Soil Moisture analyses

Soil samples were weighed and dried in an oven for 24 hours at 110 ^{0}C to determine the amount of moisture lost through evaporation (heating) on each of the profiles.

Section below describes how the collected data was analysed to achieve the study objectives. This includes both the reviewed records (secondary data) and field work data (primary data).

4.2.3. Presentation of the results and analyses

4.2.3.1. Evaluation of the processes governing recharge rates

Stable isotope approach

A δ^2H versus $\delta^{18}O$ diagram alongside the meteoric water line (MWL) of local precipitation (Lobatse MWL) as well for global precipitation were utilized to understand or trace local groundwater origin and movements. Further details on the interpretation are provided in Chapter 2. Groundwater isotopic signature signifying attributes or isotopic signature with the rainfall and plotting on the meteoric water line were attributed/interpreted to be indicative of direct rainfall recharge with minimal or no surface or subsurface residence time due to preferential recharge or flow of rainfall into the aquifer through the outcrops (fracture or fissures). While groundwater isotopic composition showing signs of deviation from the meteoric water line and plotting below it, were interpreted to be subjected to evaporation prior evaporation and/ or during overland flow. These were interpreted to be indicative of indirect recharge mechanism such as from undefined surface channel concentration.

Tritium dating and analysis

Tritium concentration in groundwater in conjunction with the use of historical records of tritium in precipitation was used to distinguish between old groundwater and the recently recharge water, and as well as for dating such groundwater samples. The following three assumptions were made:

- Tritium concentration in the Pretoria precipitation has been considered as representative of the tritium concentration in the Transboundary Ramotswa aquifer area.
- Precipitation openly enters the subsurface zone with minimal residence time within the subsurface zone. This was assumed to be a reasonable assumption in Transboundary Ramotswa considering the highly fracture and Karstified dolomites which outcrops at surface.
- Current tritium content in precipitation is natural occurring, thus tritium content in recently (or actively) recharged aquifers is the same as the concentration in precipitation. The effect other processes is insignificant in groundwater.

A semi-quantitative approach based on the approximate groundwater recharge age for continental areas based on tritium values Clark and Fritz (1997) was adopted as follows:

- Water with tritium values smaller than 0.8 TU were interpreted as indicative of sub-modern waters approximately recharged prior to 1950s
- Water with tritium values in the order of 0.8 TU to 4 TU were interpreted to be a mixture of sub-modern and modern water assumed to be composed of recently recharged waters.

Double Ring Infiltration Test (DRIT)

DRIT were conducted at each of the five sampling sites in Botswana side to provide an insight into the hydrological functioning of the sites. The falling head tests or approach was conducted with the double ring infiltrometer from the Department of Geology, University of Botswana. The minimum duration of test lasted for 40 minutes while the maximum duration of test lasted for 100 minutes. Changes in water level over time were converted into depth of water per unit of time (mm per hour). A graph presenting infiltration rates vs. time was produced.

Unsaturated zone Chloride Profiling

Total Chloride contents as Chloride (s) per soil profile were plotted as a function of depth designed on an excel spreadsheet to characterize the processes governing recharge. Three conceptual chloride profiles from Figure 2 were adopted to distinguish the processes governing the movement of moisture within subsurface zone. An increase in chloride concentration within the subsurface zone were assumed or interpreted to be caused by evapotranspiration, thus suggesting a relatively slow movement of moisture through a diffused or a piston like flow of moisture or water. A gradual decrease in chloride content to a point where chloride content seemed to relatively constant with changing profile depth was attributed to preferential movement or flow of moisture by-passing the sub-surface zone. The trend (chloride concentration) seems to oscillating and this can be attributed to alternating wetter and drier periods (seasonal changes) (Gaye and Edmunds, 1996; Kinzelbach et al., 2002; Ma et al., 2008).

4.2.3.2. Quantification of groundwater recharge rates

Chloride Mass Balance approach

Total Recharge (R_T) was estimated by dividing the input flux (the product of the chloride (Cl) concentration in the precipitation multiplied by the amount of precipitation over the study area) divided by the chloride concentration in groundwater (Equation 2) (Adams et al., 2005).

4.3. Summary

An overview of the research approach applied in the study was presented. Three key strategic phases were executed; desktop survey which more like a review of existing (secondary) data, fieldwork data collection, data analyses and interpretation which included; evaluation of the processes governing recharge and quantification of groundwater recharge rates in Ramotswa dolomitic aquifer. Data collection and analyses techniques were describe under each of the phases.

Chapter 5

Results and Discussion

5.1. Evaluation of the processes governing recharge mechanism

Hydrochemical Analyses

Measured pH values within the study area varied from 6.6 to 9.2 with an average pH value of 7.9 (Table 4). A pH between 6.6 and 9.2 gave us an impression that the groundwater within the transboundary Ramotswa dolomitic aquifer falls within an acceptable range for domestic use as such pH values fall within the international standards for drinking water quality 6.5 - 9.5 (WHO, 2006). Normally, the pH of water has no direct (health consequences) impacts on consumers, however, it still remains as one the key operational water quality parameters as it could affect the taste of water, solubility and speciation of metal ions which could results on adverse effects on the consumers (DWAF, 1996; WHO, 2006).

Table 4: Summary of the primary water quality parameters

	pH	Temp (°C)	EC (µS/cm)	TDS (mg/ℓ)
Minimum	**6.6**	**9.7**	**15.6**	**33.6**
Average	**7.9**	**21.3**	**643.6**	**420.5**
Maximum	**9.22**	**26.8**	**2051**	**1325.5**

Temperature in groundwater samples varied from 9.7 °C to 26.8 °C with the average temperature of 21.3 °C. Temperatures were relatively high in Botswana, mostly above 22 °C. Such high temperatures in Botswana corresponded closely to the mean annual air temperatures which are normally high ranging from 30 °C – 32 °C during the day to about 16 °C – 20 °C at night (Altchenko et al., 2016).

The amount of the total dissolved solids (TDS) varied from 33.6 mg/ℓ to 1325.5 mg/ℓ with an average of 420.5 mg/ℓ (Measured pH values within the study area varied from 6.6 to 9.2 with an average pH value of 7.9 (Table 4). A pH between 6.6 and 9.2 gave us an impression that the groundwater within the transboundary Ramotswa dolomitic aquifer falls within an acceptable range for domestic use as such pH values fall within the international standards for drinking water quality 6.5 - 9.5 (WHO, 2006). Normally, the pH of water has no direct (health consequences) impacts on consumers, however, it still remains as one the key operational water quality parameters as it could affect the taste of water, solubility and speciation of metal ions which could results on adverse effects on the consumers (DWAF, 1996; WHO, 2006).

Table 4). An average of 420.5 mg/ℓ of the total dissolved solids across the study area characterises the transboundary Ramotswa dolomitic aquifer waters as freshwater even though BH4975BWA surpassed the 1000 mg/ℓ mark, which is universally considered as an arbitrary limit for human consumption in natural waters (DWAF, 1996; WHO, 2003; Bouchaou et al., 2008). According to DWAF (2006), waters with TDS values above 1000 mg/ℓ to 2000 mg/ℓ have a marked salty taste, though, consumption of such waters appear not to have any adverse effects in the short term.

In general, the total dissolved solids designates the amount of dissolved constituents in water and normally depends on the surface and subsurface process influencing recharge such as precipitation, regional geology and the water balance (evaporation-precipitation) (Phyllis et al., 2007). Bredenkamp (2000) is in a view that the concentration of the TDS in natural waters provides a measure of recharge and its spatial variability because evaporative processes resulting on increased TDS content. Further, he delineates between (active) recharge and discharge areas associated to low and high TDS values respectively. Therefore, the total dissolved solids concentration has greater implication on recharge and discharge areas and on the characteristics of recharge processes.

Electrical Conductivity (EC) values varied from15.5 µS/cm to 2051 µS/cm with an average EC value of 643.6 µS/cm. Generally, EC values were relatively low, mostly below 1000 µS/cm further an indicative of freshwater (Talabi, 2013) except for ODIO8ZA (1294.8 µS/cm), ODIO13ZA (1225.3 µS/cm), BH4348BWA (1268 mg/ℓ) and BH4975BWA (2051 mg/ℓ), which had high EC values exceeded 1000 µS/cm. High EC values at ODIO8ZA and ODIO13ZA were assumed to be the residue of evaporation from the house holds tanks while high EC values at BH4348BWA and BH4975BWA along the Notwane River were suggested to have resulted from evaporation enriched recharge in waters moving downstream which could have increase salinity and hence EC.

EC and TDS seemed to be low in the southern parts of the study such as the upper Dinokana area increasing towards the Ramotswa area. Similarly the chloride distribution map shows relatively low chloride levels in southern parts of the study area, which increases towards the northern parts of the study (Appendix G). An increase in EC, TDS and chloride levels from the upper Dinokana area towards Ramotswa wellfields area seems to correlate well to the groundwater flow direction as depicted by the potentiometric map on Figure 16, which reasonably characterises the southern parts of the study area as potential recharge areas.

High EC and TDS areas such as the Ramotswa wellfield area were suggested to be signifying characteristics of diffused flow of moisture within the subsurface which permits for an evaporative enrichment and/ or indicative of localized or indirect recharge from concentrated run-offs. On the other hand low EC and TDS areas such as the upper Dinokana area in the south suggested rapid (direct) recharge into the aquifer with minimal residence time on the surface and this was supposed to be true in areas where interconnected fractures covered with thin permeable soils and on areas where fractures outcrops on surface. Therefore, areas indicative of high EC and TDS values can be linked to low recharge rates while the areas with low EC and TDS values could be associated to high recharge rates.

The anion trilinear plot (Figure 18) shows enrichment in CO_3^{2-} and HCO_3 waters while the quadrilateral facies plot show that the transboundary Ramotswa aquifer waters can be classified as Ca-Mg-HCO_3 water type. Generally, a Ca-Mg-HCO_3

water type reflects the dissolution of the host rock (dolomite). This suggests that the water-rock interaction process is more prominent and the correlation between Mg and Ca (Figure 19) shows enrichment in Mg and Ca with an average ratio of 0.96 suggesting that the dolomite is near to saturation. Mg^{2+} and Ca^{2+} are normally constituents of the dolomites and Figure 19 shows relatively strong correlation between Mg^{2+} and Ca^{2+}. Generally, in dolomitic terrains Mg^{2+} and Ca^{2+} enrichment is mainly derived from the water-rock interaction through dissolution of the dolomite rocks with recently recharged acidic waters.

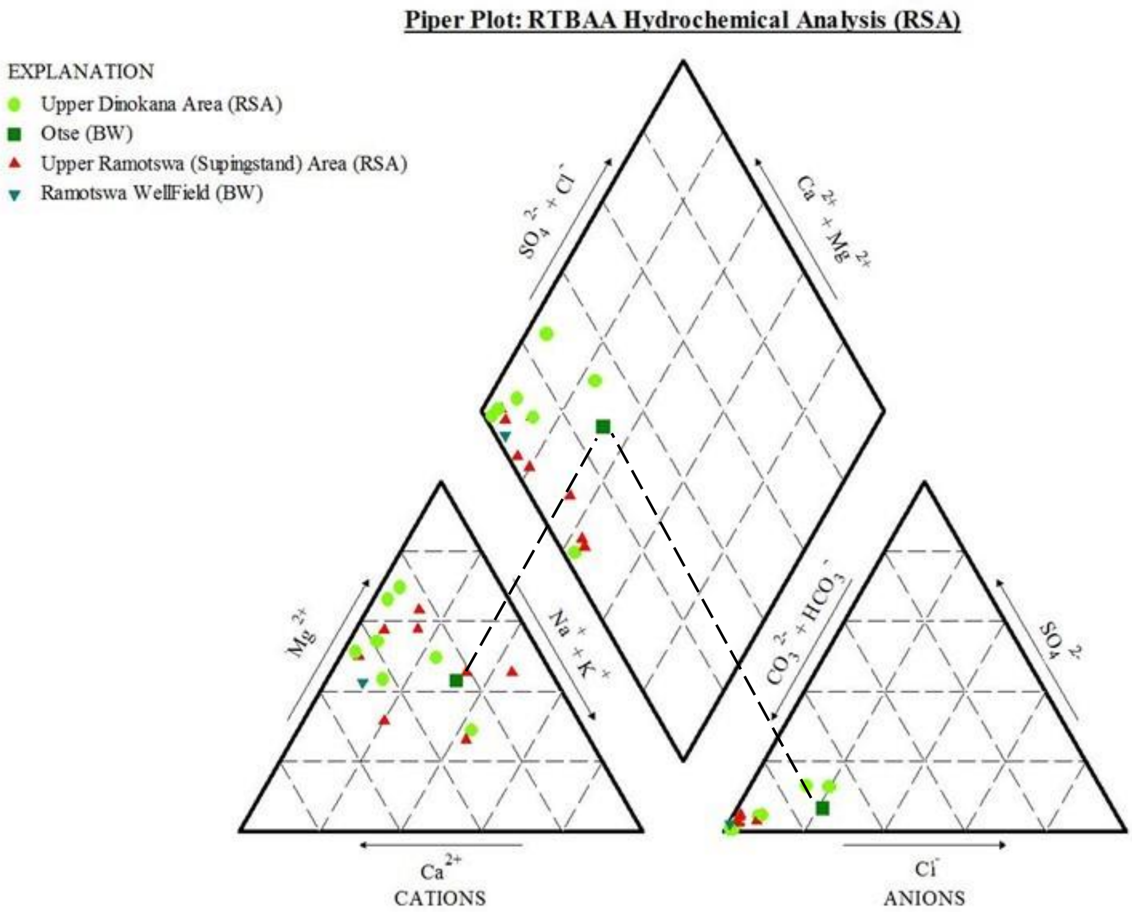


Figure 18: Piper diagram visualizing hydrochemical characteristics and water types

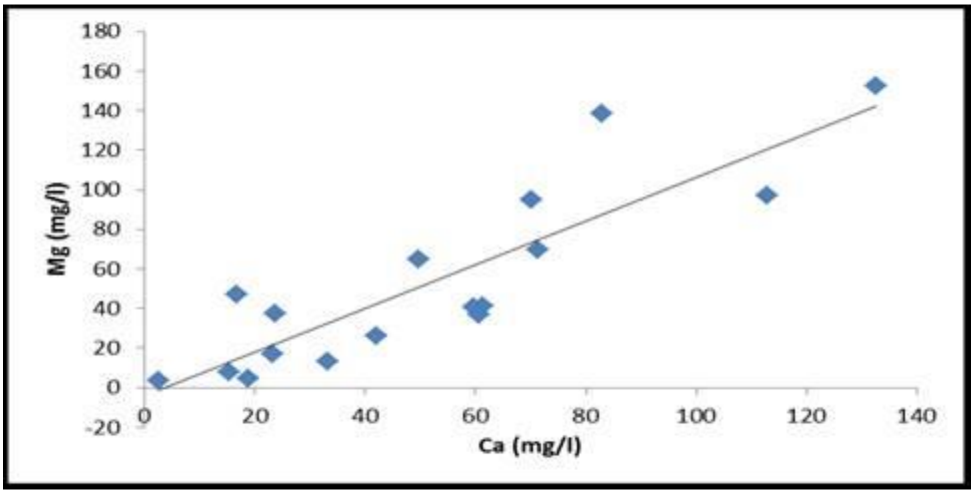


Figure 19: Correlation between Mg and Ca from groundwater samples in transboundary Ramotswa aquifer

Stable isotope signatures

Stable isotope analyses included groundwater samples (32), surface water samples (3), and precipitation samples (2). Stable isotope of $\delta^{18}O$ and δD were correlated and are presented in Figure 20 and in Appendix H. Local meteoric water line based on $\delta^{18}O$ and δD was determined based on the data by Talma and van Wyk (2013) assuming that the rainfall isotopic composition in Lobatse is representative for the greater Ramotswa aquifer area. The stable isotopes signatures were used to characterize the processes governing recharge and to trace the origin and history of groundwater.

Lobatse Meteoric Water Line

The Lobatse meteoric water line (also referred to as the Local Meteoric Water Line) based on $\delta^{18}O$ and δD data is $\delta D = 6.4\ \delta^{18}O + 6.7$ ‰. Both the slope (s) and deuterium excess (d-excess) are slightly lower than s and d-excess of the Global Meteoric Water Line (GMWL), suggestive of secondary evaporation signature or humidity variation in precipitation. According to Wu (2005), a slope less than that of the GMWL are known to have resulted from secondary evaporation during precipitation in South Africa. This phenomenon tends to cause enrichment in $\delta^{18}O$ and δD (the heavy isotope) in the residual liquid water (rainfall) and as a result causes a decrease in deuterium excess (d-excess value). Froehlich et al. (2002) defines deuterium-excess as the measure of the relative proportions of $\delta^{18}O$ and δD contained in water which signifies an evaporation effect or the deviation from the GMWL in $\delta\ ^{18}O$ and δD.

Stable isotope signature in precipitation

Precipitation data were taken from the hail storm that occurred during the sampling (July 2016) at -25.41972 S and 25.81735 E, altitude 1309 m.a.s.l. Stable isotopes of δD and $\delta^{18}O$ from precipitation 01 (P 01) and precipitation 02 (P 02) samples were -58.5 ±0.42 ‰) and -9.42 ±0.06 ‰, and -44.9 ±0.13 ‰ and -7.79 ±0.04 ‰, respectively (Figure 20).

Stable isotope signature in surface water

Surface water from the study area (Table 5 and Figure 20) showed slightly distinct isotopic composition which made it possible to distinguish if there is a link between surface water and groundwater.

Three of the samples from the dams were highly depleted 2.1 ±0 ‰, 3.38 ±0.05 ‰ and 2.65 ±0 ‰. These were the Notwane dam (N-Dam 01) and the Supingstand dam (SUD), which both locates in South Africa and the Lesetleng dam in Botswana. The δD and $\delta\ ^{18}O$ from these dams ranged from 9.61 ±0 ‰ and 2.1 ±0 ‰ (elevation 1132 m.a.s.l) at Notwane dam (N-dam 01) to -15.43 ±0 ‰ and 2.65 ±0 ‰ (elevation 1057 m.a.s.l) at Lesetleng dam, respectively (Figure 20). The Ngotwane dam (N-dam 02) showed similar characteristics to groundwater samples from the greater Ramotswa aquifer area as well as the current precipitation. As a result the samples plotted well below both the LMWL and the GMWL signifying a sign of evaporation.

Table 5: Stable isotope result from surface water sources (dams) within the study area

Sample ID	Latitude	Longitude	Elevation (m.a.s.l)	δ 2H (‰)	± ^{2}H StDev	δ ^{18}O (‰)	± ^{18}O StDev
SUD 01	-24.789347	26.05894	1132	12.90	0.31	3.38	0.05
N-DAM 01	-25.199005	25.813472	1117	9.61	0.00	2.10	0.00
N-DAM 02	-25.079697	25.936892	1053	-26.78	0.00	-4.75	0.00
LESETLENG DAM	-24.96568	25.81764	1057	-15.43	0.00	2.65	0.00

Stable isotope signature in groundwater

Largely groundwater samples were highly enriched with the heavy isotope, which could be ascribed to either heavy rains or high monthly precipitation amounts that result in more negative $\delta^{18}O$ and δD. According to Vogel and Ehhalt (1963) groundwater isotopic composition in semi-arid regions in South Africa is characterised by relatively enriched isotopic composition compared to the highly depleted rainfall (precipitation). Further, the depleted isotopic groundwater composition seem to correspond to sporadic heavy rainfalls, therefore in a view that recharge through these aquifer is mainly attributed to such episodic and to isolated periods of heavy rainfalls.

Stable isotope results from groundwater samples within the greater Ramotswa area are presented in Figure 20 and in Appendix E. The stable isotope results of $\delta^{18}O$ and δD varied from -51.74 ±0.21 ‰ and -10.33 ±0.07 ‰ to -8.67 ±0.1 ‰ and -1.5 ±0.04 ‰ in Ramotswa wellfields and in the upper Dinokana area, respectively. The stable isotope results for the Supingstand area within the upper Ramotswa area showed no sign of evaporation where most of the data plotting on and above the LMWL. Generally, samples plotting above the meteoric water line can be ascribed to low rainfall during dry air conditions (low humidity of air) where the first rain is generally depleted and the precipitation plot well above the meteoric water line (Clark and Frits, 1997), while also these could be as a result of direct (rapid) rainfall recharge through fractures that outcrop on the surface resulting in groundwater isotopic signature similar to rainfall isotopic signature (Appendix C and D).

Ramotswa wellfields samples mostly plotted below both the GMWL and LMWL. These samples showed a slight deviation from the LMWL suggesting marks of secondary evaporation. The Upper Dinokana area, Lobatse area and the Ramotswa wellfield area showed varying isotopic composition with some of the samples plotting above the GMWL and LMWL while some samples plotted below the GMWL and the LMWL respectively. This suggests that recharge within these areas occurs as a result of one or more various processes controlling recharge mechanism.

Such variation on the processes controlling recharge is shown on the variation in isotopic composition on the springs located with the southern parts of the study area. The springs such as the Dinokana spring (DNK spring) plots slightly just above the LMWL and the GMWL, while the Groot-Marico springs (GME 01 and 02) plots just below the LMWL and the GMWL. Such isotopic characteristics shown by the Dinokana spring signifies an isotopic signature similar to that of local precipitation thus suggestive of direct rainfall infiltration probable through the fractures that outcrop at the surface. This signifies recharge event prior evaporation. On the other hand the Groot-Marico springs shows a slightly deviation from the LMWL and further these suggest for secondary evaporation before and/ or during infiltration. Deviation from the LMWL was interpreted as indicative of indirect rainfall infiltration or localized recharge as a result to ponding due surface depressions from localized runoff. Possible deviation from the LWML could be as result to a diffused or sluggish water movement within the subsurface zone thus permitting for isotope fractionation.

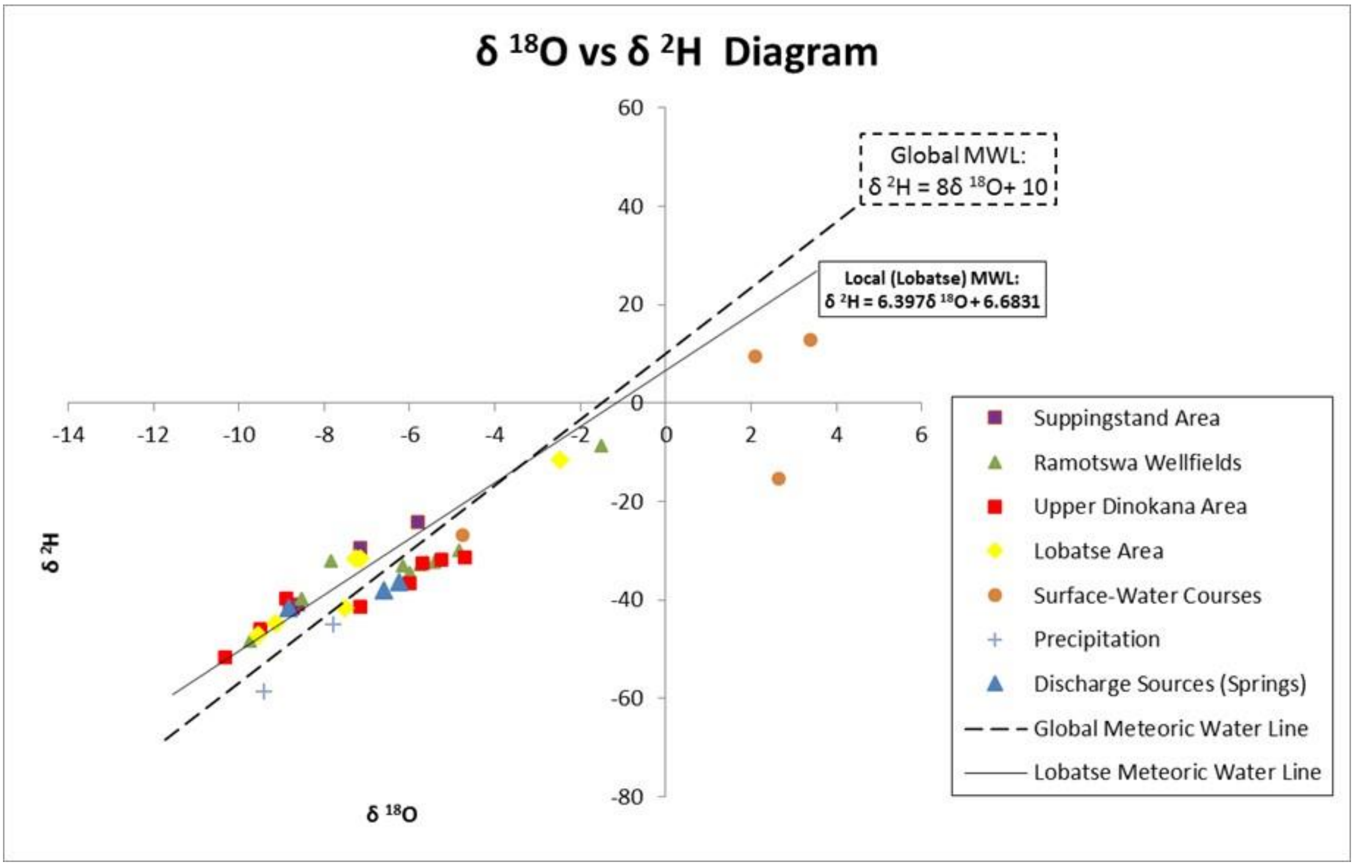


Figure 20: Correlation of $\delta^{18}O$ vs δ D (^{2}H) diagram of groundwater samples collected within the greater Ramotswa area. Rainfall isotope data from the upper Dinokana area collected during field work. Global Meteoric Water Line and Local (Lobatse) Meteoric Water Line are plotted for comparison.

Table 6: Correlation between δD and elevation within the upper Ramotswa study area

Sample Name	Latitude	Longitude	Elevation (m.a.sl)	δ 2H (‰)	± 2H StDev	Region
SWBP01	-24.8735	25.88668	1036	-24.12	0.00	Supingstand Area
MOSHANA	-24.976	25.88292	1052	-29.51	0.00	
SUB	-24.7899	26.08373	1151	-40.96	0.02	
SUB04	-24.8071	26.02321	1156	-41.73	0.26	
SUB03	-24.8071	26.02321	1161	-41.99	0.39	
BH 4975	-24.8331	25.87078	1021	-8.67	0.10	Ramotswa wellfields
BH 4348	-24.8995	25.87702	1024	-34.44	0.37	
BH 4373	-24.9057	25.87224	1025	-32.42	0.15	
BH 4340	-24.913	25.88041	1028	-32.73	0.23	
DUG WELL	-24.5202	25.52561	1030	-39.84	0.31	

BH10129	-24.8583	25.86679	1031	-32.2	0.28	
BH287	-24.8795	25.86945	1032	-48.22	0.00	
BH 6424	-24.8884	25.83583	1052	-35.28	0.14	
BH 4450	-25.0621	25.70864	1122	-30	0.00	
BH 4458	-25.1593	25.71485	1197	-33.08	0.44	

In overall, correlation between δD and $\delta^{18}O$ in groundwater within the transboundary Ramotswa aquifer mainly shows two distinct groups, which are mainly clustered in the fourth quadrant. These groups are distinguished on the basis of showing evaporation signal or not. However both evaporated and un-evaporated samples cluster around both GMWL and LMWL suggest that these waters are of atmospheric origin and this is further supported by the isotopic composition of the precipitation upper Ramotswa study area. The first group of groundwater samples plotted above the GMWL also on top of the LMWL resembling local rain events thus suggesting recharge occurs through direct rainfall recharge from high and short intense rainfall events during summer months. This is observed across the aquifer area while the second group of groundwater samples plots below the GMWL showing a slight deviation from the GMWL as a result to evaporation before or during infiltration. This is observed in Ramotswa wellfields and upper Dinokana area groundwater isotopic composition. Further, deviation from the LMWL suggest that recharge occurs through slow accumulation of rainfall (diffused rainfall infiltration) and/ or focused recharge in depressions filled from surface runoff either from high intense rainfalls (heavy rains where rainfall intensity is greater than the soil infiltration capacity resulting on surface run-offs) or as a result to topographical effects. This is could be possible in Ramotswa area where the Notwane River can acts as recharge source during floods.

^{3}H-Tritium Dating

Tritium values sampled during July-August 2016 varied from 0.2 to 2.3 TU (Appendix H). A total of 17 samples contained tritium values lower than 0.8 TU. The results was interpreted as sub-modern recharged waters most likely before the 1950`s, according to approximate groundwater recharge age for continental areas based on tritium values by Clark and Fritz (1997). The remaining 14 groundwater samples contained tritium values between 0.8 and 2.3 TU, which fall within the 0.8 and 4 TU range, which is classified as a mixture of modern and sub-modern recharged waters (Clark and Fritz, 1997; Mazor, 2005).

With the knowledge of the current tritium input in precipitation in South Africa (3 TU) (Abiye (Eds), 2013), it was possible to estimate the apparent groundwater residence time and to evaluate whether the recent rainfall is actually recharging the aquifers (*see Chapter 4*). None of the sampled boreholes or springs matched the current concentration of tritium in precipitation although some were relatively close. However, sampling depth was thought to be among the major factors influencing the quantified tritium content in groundwater and the corresponding estimated ages.

Based on these tritium values and on the assumptions made in Chapter 4, the transboundary Ramotswa groundwater can be characterized into two groups; as a mixture modern to sub modern recharged waters, and sub-modern recharged waters. Further, based on estimated apparent groundwater ages or residence time in RTBAA, recently recharged groundwater is characterized as a mixture of modern and sub-modern recharged waters dated from less than 24 years to less than 12.34 years, while the sub modern recharged water type is dated from above 24.86 years to less than or equal to 49.72 years. Assuming that the above estimated ages are considered true, then the estimated groundwater residence time should increase with depth as per borehole profile and along groundwater flow paths from recharge to discharge areas. Therefore, if sampling were to be conducted on the upper surface of the water table and assuming that no abstractions occurred, the estimated age would even be shorter than the current estimates ages.

Tritium levels varied across the transboundary aquifer area (Figure 21) with no distinct characteristics while all groundwater samples indicated detectable amounts of tritium content suggesting the occurrence of recharge. Correlation between surface elevation and tritium distribution showed that the topographical high lower Ramotswa study area (upper Dinokana and Lobatse area) contains relatively low tritium value compared to the upper topographically low part of the study area

(Ramotswa wellfields area and Supingstand area) (Figure 21). The difference in tritium levels from one area to another may be attributed to the heterogeneity of the system as a result of karstification, fracture networks as well as the processes controlling recharge mechanism from one location to another.

The relationship between the piezometric map (Figure 16) and the tritium distribution site map (Figure 21) indicates that the tritium concentration in groundwater is independent of the groundwater flow direction. This again can be attributed to heterogeneity of the Ramotswa dolomitic aquifer. High tritium values are shown within the upper parts of the study area where groundwater flows at greater depth below 1000 m.a.s.l (up to 42 m depth from the surface) compared to the southern parts of the study area where groundwater flows at shallow depths above 1300 m.a.s.l (±10 m depth from the surface), while at the same time the tritium values vary across the study area with relatively low tritium content in the upper parts compare to the northern parts of the study area.

High tritium values greater or equal to 1.5 TU indicated the presence of recharge within the past 12.43 years. These high tritium values were scattered across the basin (transboundary Ramotswa aquifer area) also indicating the inhomogeneity of the system. While low tritium values (less than 0.8) indicate a relatively long and deep groundwater flow paths suggesting hydraulic connection between the shallow and deep aquifer. Groundwater with high tritium content dated younger than 24.86 years with about 8 of those boreholes dating younger or equivalent to 12.43 years thus suggesting recent (active) recharge within the past 12.43 years. The remaining boreholes with low tritium content dated above 24.86 years but less than 49.72 years. These boreholes were interpreted to contain a mix of modern and sub-modern waters.

From the analyses above, two types of waters are distinguished as follows:

(i) A mix of sub-modern to modern recharged waters with tritium content between 0.8 – 2.3 TU. The mix water was compared to the current tritium content in precipitation (3 TU) and the result indicated active recharge within the past 12.43 years.

(ii) Relatively deep circulation through karst formation and fractures with water containing tritium levels below 0.8 TU, which were interpreted as indicative of sub-modern waters or recharged water probably prior to 1950s.

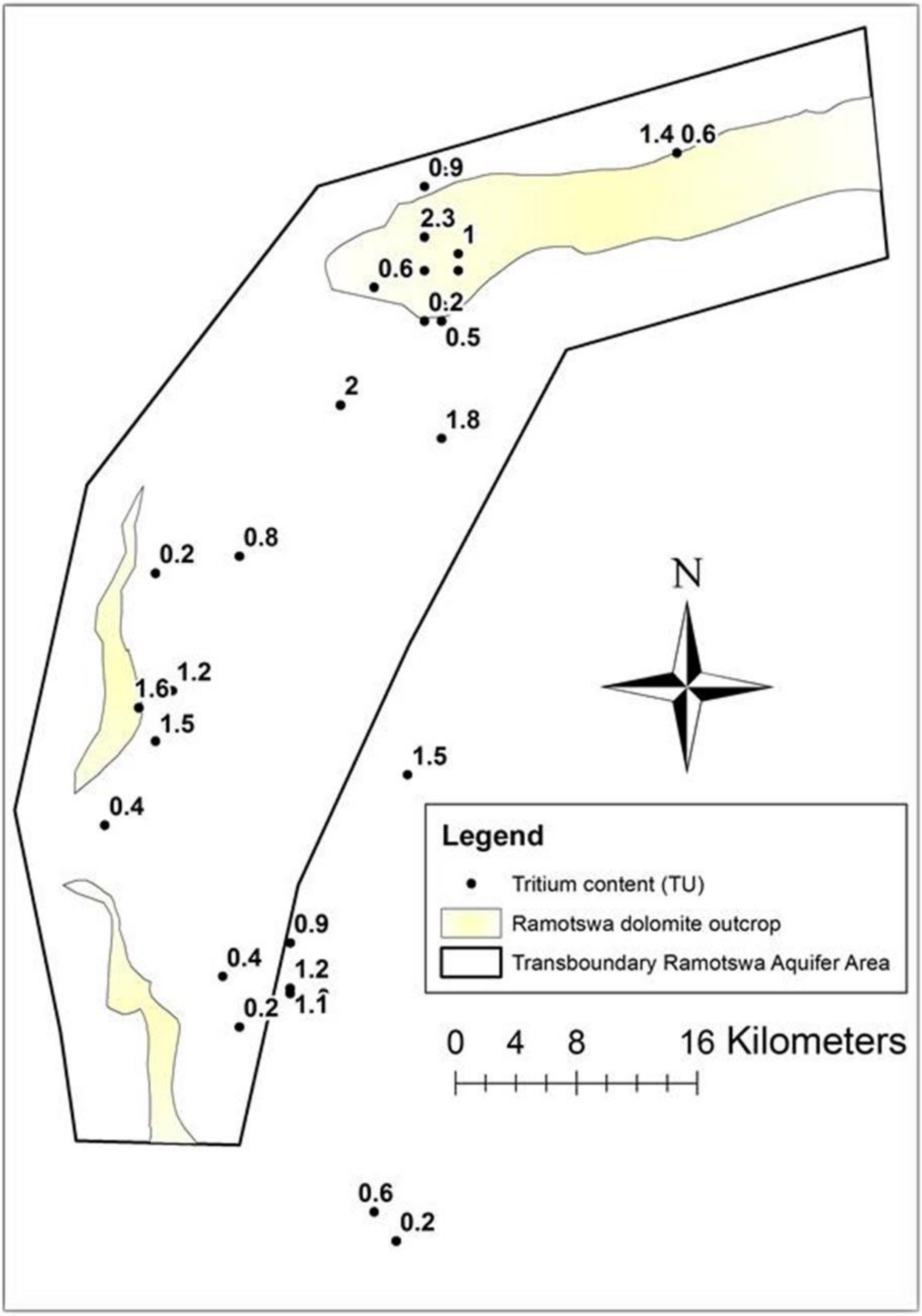


Figure 21: Tritium sites in transboundary Ramotswa aquifer area

Double Ring Infiltration Test

Infiltration rates versus time plot are presented from Figure 22 - Figure 25. Tests were conducted for a minimum of 40 minutes to a maximum of 100 minutes, the latter being at site 2.

Infiltration rates vary from 24 mm/hr to 78 mm/hr (Table 7). High infiltration rates were recorded at the Notwane River, which is located at an international border between South Africa and Botswana. While test results were relatively high compared to other sites, infiltration test conducted at the sites was the shortest test in duration (45 minutes) of all four sites in which the tests were conducted. The test was stopped immediately after the rings/wells dried at 45 minutes to proceed with the test at other sites.

Infiltration rates at North Ramotswa (T1) and North Lobatse (T4) were similar with 0.04 cm/min at both sites with test duration of 100 minutes and 71 minutes, respectively. This suggests the variability in soil permeability at each of the sites. North Ramotswa site was located within the flood prone area of the Notwane River with fine textured soils while the North Lobatse site contained relatively coarse textured soils with medium size pebbles (visualized during auguring). Another factor which could contribute to differences in the results obtained is the time dependant falling head test that was conducted which normally produce rates with varying time depending on the water level in the ring (Nichols et al., 2014). This is normally caused by generated pressure heads on the above rings. Infiltration rates above 15 mm/hr but below 50 mm/hr were classified as medium, while infiltration rates above 50 mm/hr were classified as high (Brouwer et al., 1988).

Table 7: Infiltration test results conducted within the Ramotswa study area

Test Site	Site name	Infiltration rate (mm/hr)	Infiltration classification (Brouwer et al., 1988)
IT1	North Ramotswa	24	Medium
IT2	Notwane River	78	High
IT3	Lobatse	30	Medium
IT4	North Lobatse	24	Medium

An initial drastic or sharp rise in the rate of infiltration is seen across at all sites in Figure 22 - Figure 25. This occurs at initial stages of the test where the initial water input is less than the soil`s initial infiltration capacity (Brouwer et al., 1988). This normally occurs in dry soils where the initial water input rapidly infiltrates into soil open pores until it’s saturated. This is normally referred to as the initial infiltration rate. Normally as the test proceeds, the soil`s infiltration capacity and rate decreases until a constant final rate is (Brouwer et al., 1988; Gregory et al., 2005).

Generally infiltration rates decreases gradually after the first 10 to 20 minutes suggesting that soil capacity to transmit water is low compared to the surficial water input, thus suggesting vertical flux through a diffuse like flow where recent infiltrated water displaces the water that infiltrated earlier. This was observed across all infiltration test sites. The difference in pressure head on the outer ring of the infiltrometer normally results in a decline in the infiltration rates on the inner rings (Eijkelkamp, 2015) thus affecting the estimated infiltration rates. Beside the fact that inner ring was monitored manually and that the water level in the outer ring infiltrometr was physically kept constant could have implications on the results. Further, according to the ASTM standard an inner and outer ring diameters of 30 and 60 cm compared to the use of 15 cm and 30 cm diameter rings applied in the study are normally prefered.

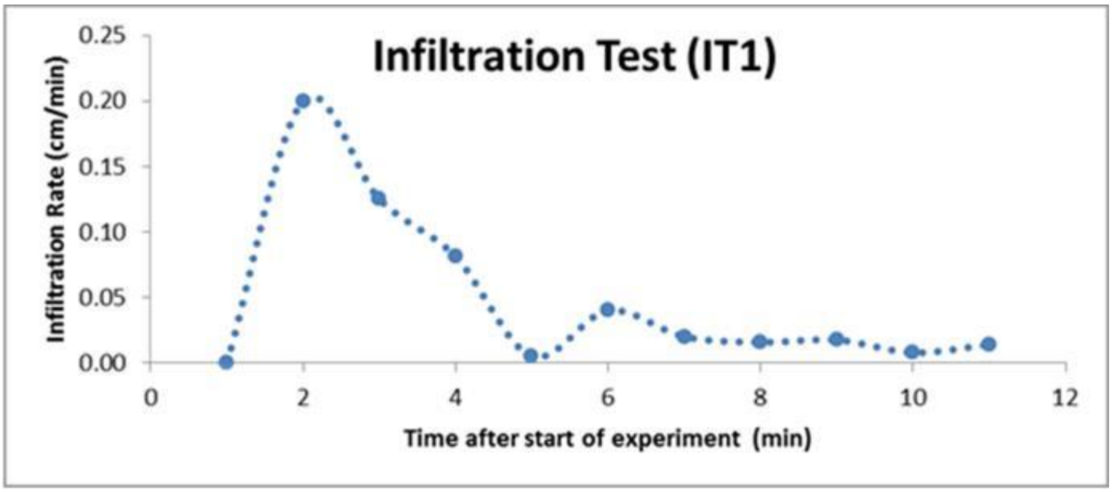


Figure 22: Time dependent infiltration rate North Ramotswa

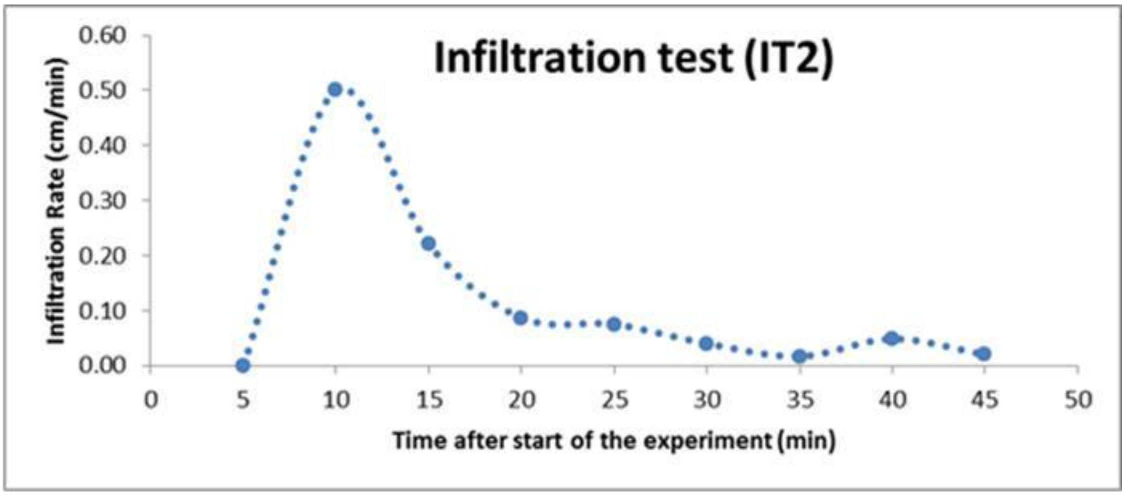


Figure 23: Time dependent infiltration at Notwane River

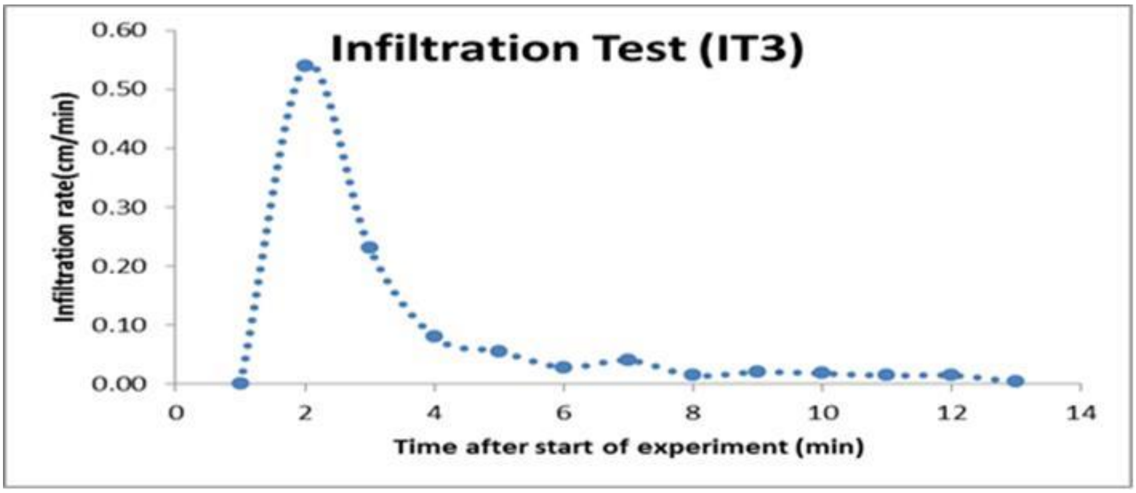


Figure 24: Time dependent infiltration at Lobatse

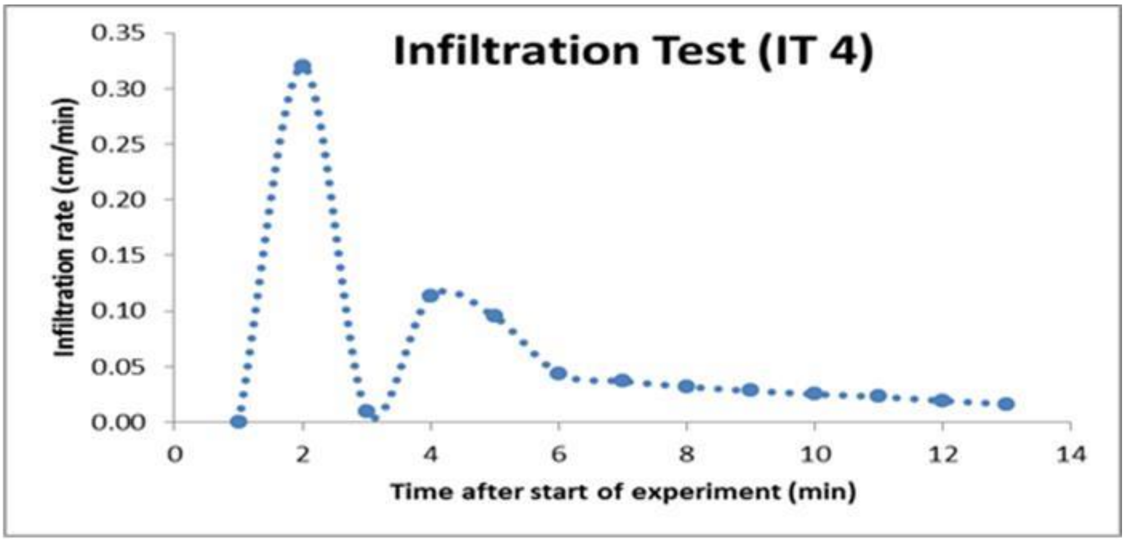


Figure 25: Time dependent infiltration at North Lobatse

Chloride Profiling

Chloride profiles and moisture analyses are presented in this section.

The chloride profiles seem to show unique attributes at each of the sites across the basin. The upper Ramotswa chloride site (Figure 26) shows an increasing trend in chloride content with increasing depth from the surface. The chloride content increases gradually with increasing moisture content. The chloride concentration seems to be oscillating and this can be attributed to alternating wetter and drier periods (seasonal changes) (Gaye and Edmunds, 1996; Kinzelbach et al., 2002; Ma et al., 2008). A similar trend was observed in Moshana (Figure 28) where chloride content increases gradually with increasing depth. However, the Moshana chloride profile seems not to be oscillating as significant as the upper Ramotswa chloride profile. At both profiles; upper Ramotswa and Moshana the moisture content also indicates an increase with increasing depth while the increase is relatively low at upper Ramotswa compared to Moshana. The increase in chloride and moisture content with depth is observed for up to 2 m depth at Moshana and upper Ramotswa area while at upper Dinokana the trend is visualized up to 1.4 m depth due limited data.

An increase in chloride can be caused by the loss of water through evapotranspiration resulting in concentrated chloride content within the profiles during wetter periods. While low chloride concentration is mainly attributed to recent soil moisture flux as a result rainfall and this may depend on the rainfall intensity and magnitude. According to Gaye and Edmunds (1996) the presence of several peaks and troughs such as in Figure 26 is proportional to the significance of evapotranspiration and past climate or climate change (Gaye and Edmunds, 1996; Ma et al., 2008). Further this suggests oscillating recharge events (Gaye and Edmunds 1996). Such events are indicative of a piston like or diffuse rainfall infiltration into the soil.

The lower Ramotswa (Figure 28), Notwane River (Figure 27), upper Dinokana (Figure 31) and the North Lobatse (Figure 30) profiles show similar trends with relatively constant chloride content or profile with depth. This may be attributed to recent rainfall-infiltration dissolving the chloride content as it progresses further down in the subsurface zone. High chloride content within the first 15-30 cm depth at Lower Ramotswa chloride profile can be as a result of evaporation resulting on the concentrated levels of chloride content.

North Lobatse chloride profile indicates relatively low and constant chloride content with an increase in depth. This is indicative of no evaporation effects on recent recharge events possible through fracture outcrops or preferential flow paths such decayed root channels or earth worm channels.

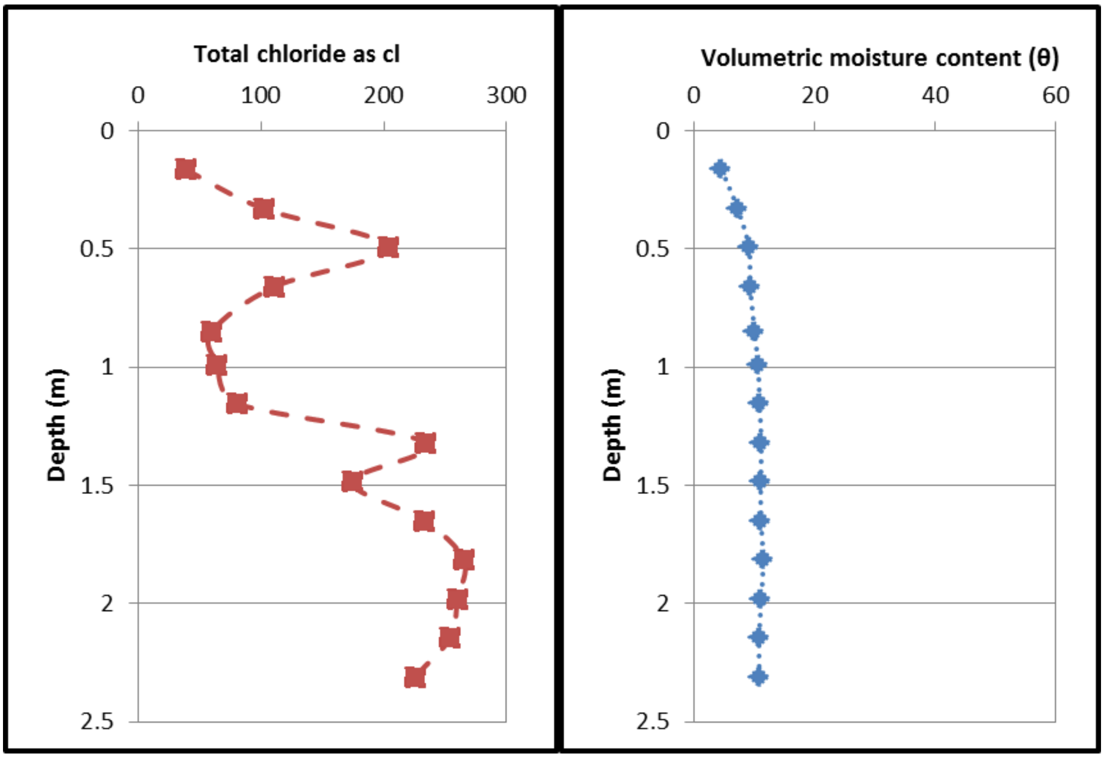


Figure 26: Moisture and chloride as a function of depth of soil profile (upper Ramotswa area)

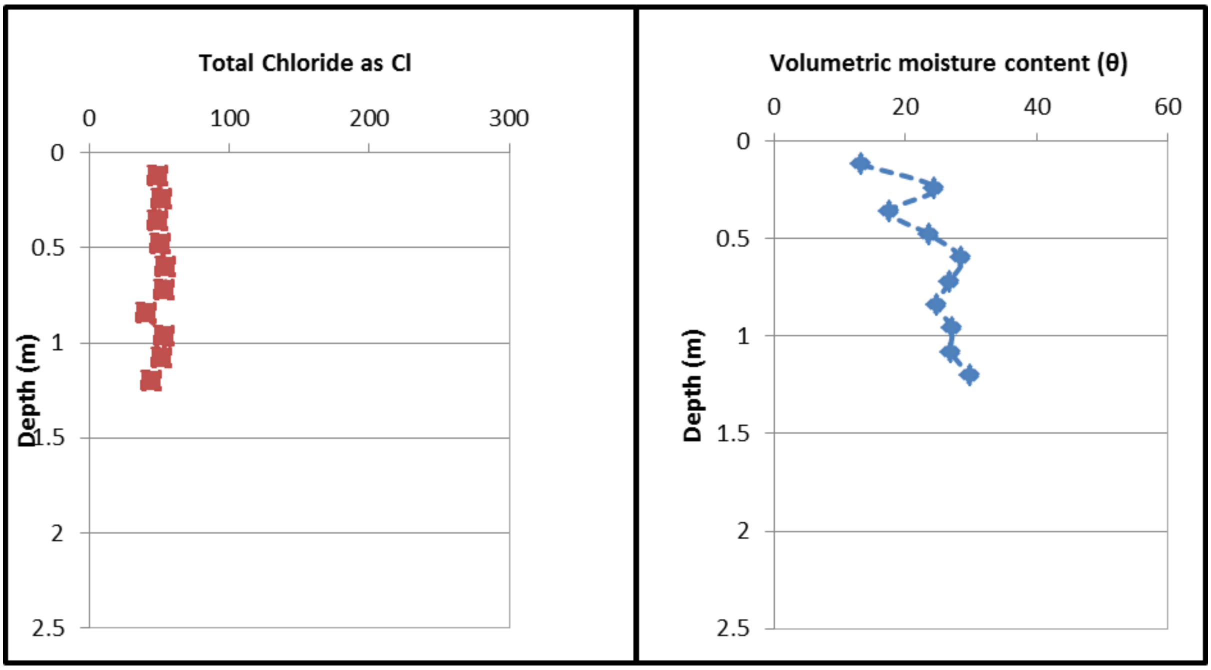


Figure 27: Moisture and chloride as a function of depth of soil profile (Notwane River: Ramotswa area)

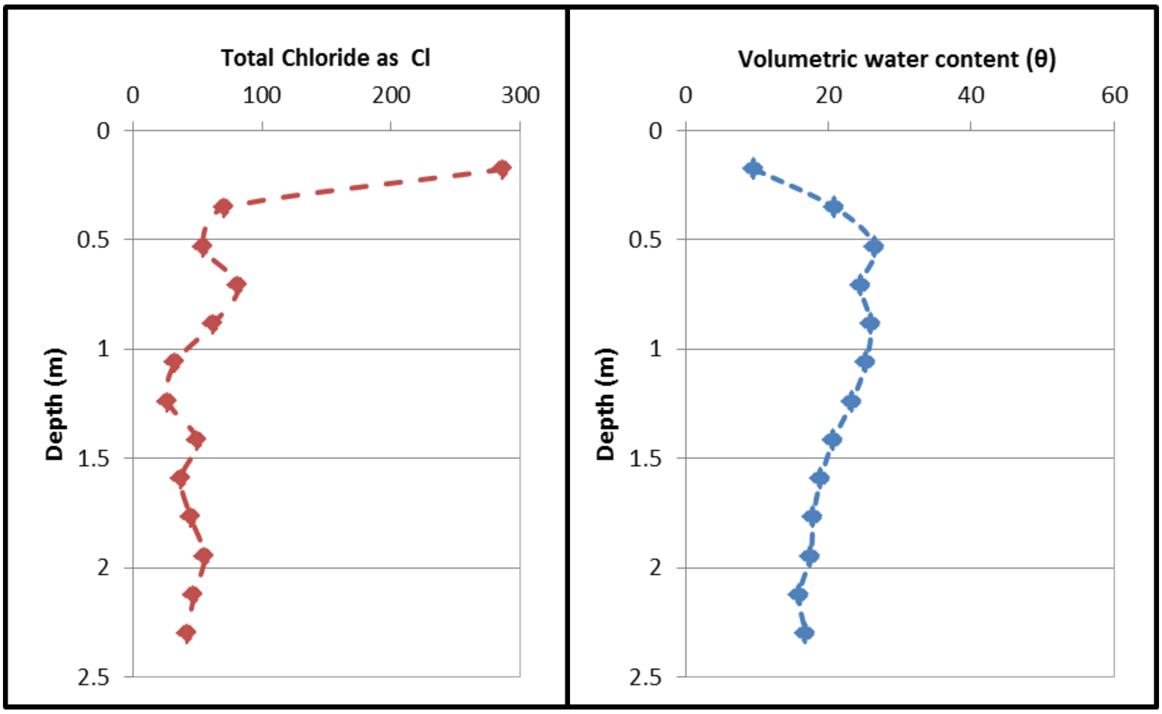


Figure 28: Moisture and chloride as a function of depth of soil profile (Lower Ramotswa)

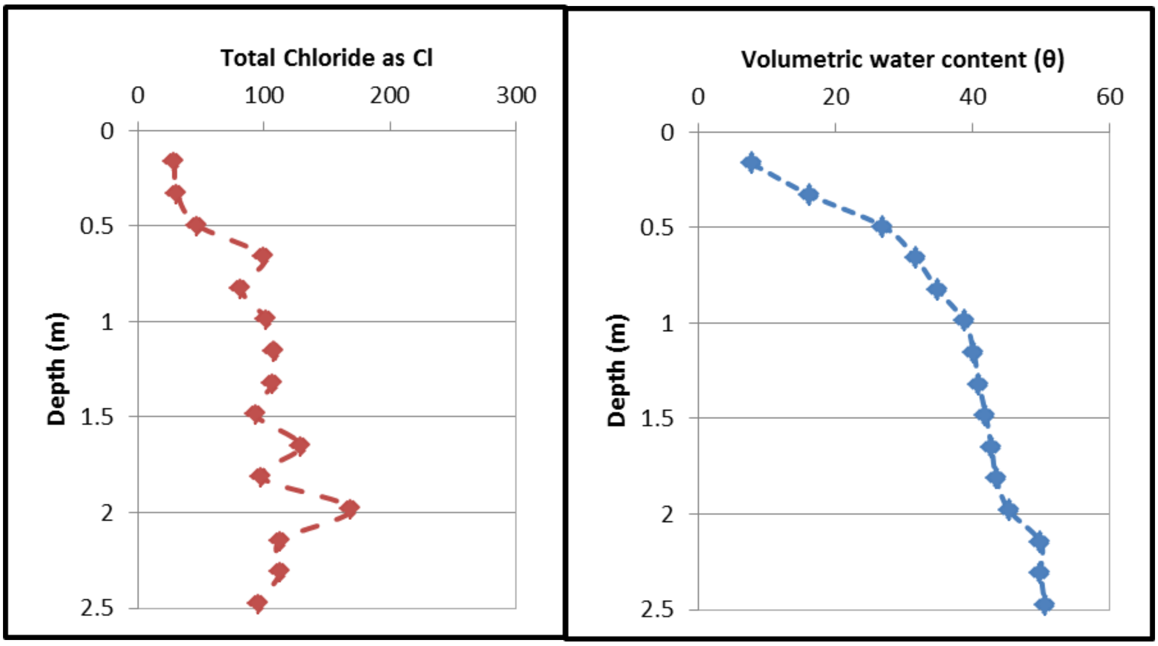


Figure 29: Moisture and chloride as a function of depth of soil profile (Moshana area)

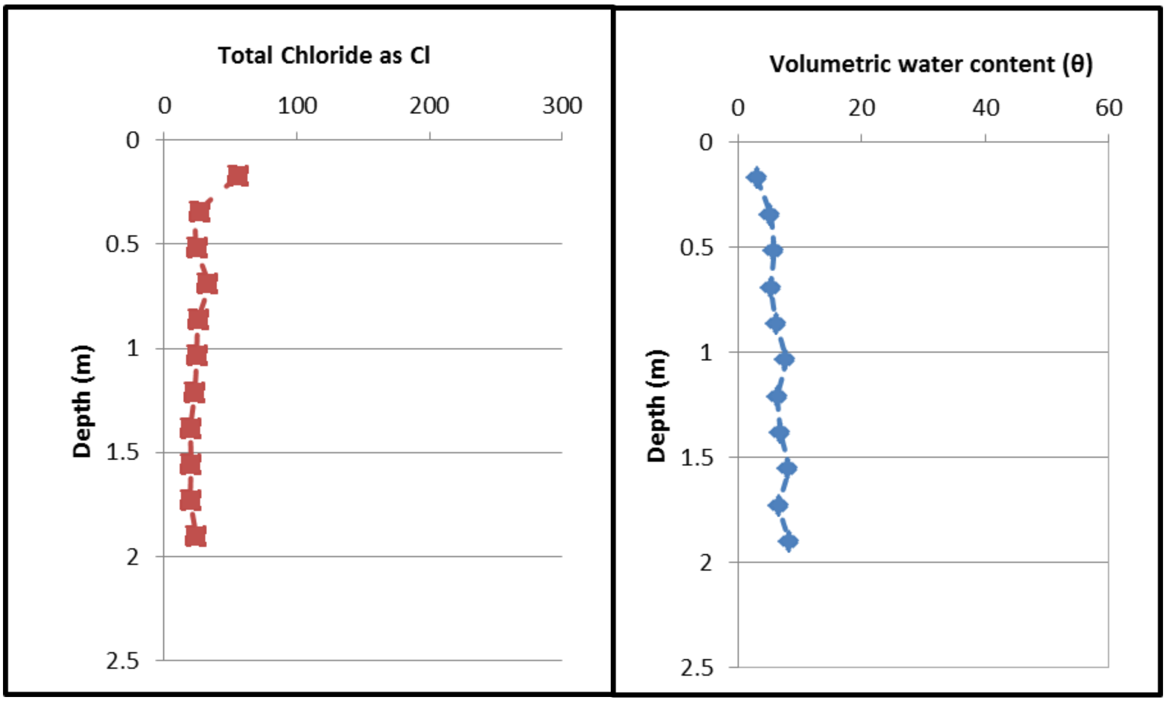


Figure 30: Moisture and chloride as function of depth of soil profile (North Lobatse)

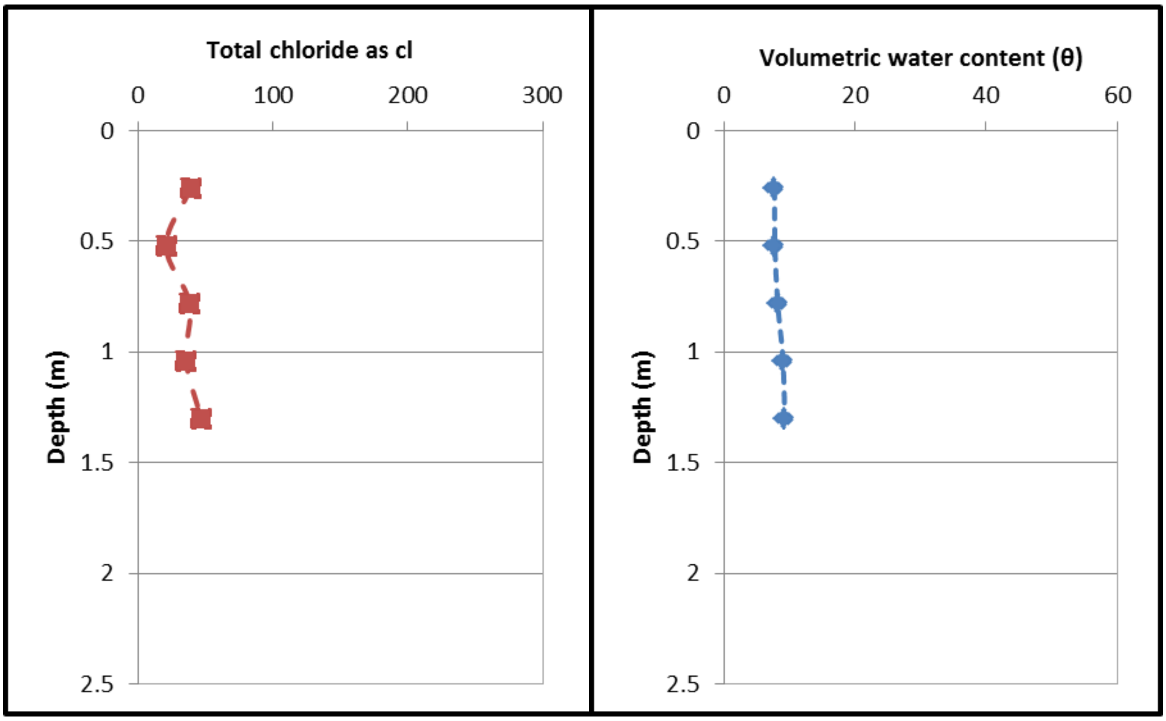


Figure 31: Moisture and chloride as a function of depth of soil profile (Upper Dinokana area)

5.2. Groundwater recharge estimation

Chloride Mass Balance (CMB) approach

Groundwater recharge estimation based on the CMB approach is presented in Appendix H and are summarized in Table 8. Approximately 77% of the boreholes analysed for chloride produced recharge rates varying from 2.4 mm/yr to 69 mm/yr, representing 0.4% and 14% of the mean annual precipitation, respectively. High recharge rates estimates ranged from 100 mm/yr to approximately 360 mm/yr which varied from 20% to 70% of the mean annual precipitation (MAP) (Table 8).

The mountainous (or topographical high) area between the upper Dinokana (South Africa) and the Otse mountains north of Lobatse in (Botswana) experienced high recharge rates of up to 360 mm/yr. However, it is uncertain if estimated recharge rates were not absolutely influenced by the nearby commercial and subsistence farming both in South Africa and Botswana, respectively. Further improvement of the results is possible if the long term data on chloride deposition are used instead of single point value sampled during the field work in upper Dinokana area. According to Abiye (2016), to obtain reliable recharge estimates with the chloride mass balance, the approach used for the study should account for dry and bulk chloride in rainfall. However, no long term chloride deposition data is available in the study area at present.

However, most of the estimated recharge rates with the area varied from 30 mm/yr to 51 mm/yr in the upper Dinokana area, in South Africa to 27 mm/yr to 69 mm/yr in the Lobatse area, in Botswana. Figure 32 indicates point recharge estimates and the results shows relatively low recharge rates within northern parts of the study area and increasing towards the southern parts of the study area. The results seems to correlate well with groundwater flow direction and topography, which both flow and decrease in a northward direction from the upper Dinokana area (high recharge estimates) to Supingstand area (low recharge estimates). Such high recharge rates are in agreement with the previous suggestion made by the use of hydrochemical analyses of relatively low TDS and EC values in the southern part of the study area, therefore suggesting recharge through direct rainfall infiltration with minimal surface residence time.

The current approach to recharge estimation with chloride mass balance was weak in the study design; hence lack of data on chloride deposition and seasonal chloride deposition or input. Adams et al. (2005) reported that if run-off contributes to recharge in a study, the use of chloride mass balance method will underestimate recharge and this is highly anticipated in transboundary Ramotswa particularly in the upper Ramotswa area where the Notwane river recharges the aquifer during high rainfalls and floods (Selaolo, 1998). This might one of the attributes resulting in low recharge rate in the upper pats of the study area. The northern part of the study area are characterized by high TDS and EC values suggesting recharge through indirect processes into the aquifer. The chloride mass balance method requires careful understanding and/ or knowledge on the potential source of chloride that can have influence on the chloride concentration within the region under investigation, wet and dry deposition, the use of both primary (present) and secondary (long-term) data to strengthen the reliability of the results.

Table 8: Summarized recharge rates estimated from the CMB approach. Overall recharge results including the rainfall chloride and mean annual precipitation for each location is presented on appendix I.

Area	**Low Recharge Rates**		**High Recharge Rates**	
	Recharge (mm/yr)	**Recharge (%MAP)**	**Recharge (mm/yr)**	**Recharge (%MAP)**
Surpingstand	2.4 – 38	0.4 - 7.5	60	11
Ramotswa	2.7 - 63	0.5 - 12	277	44
Lobatse	27 - 69	5 - 14	360	75
Upper Dinokana	30 - 51	6 - 10	101 - 288	21 - 60

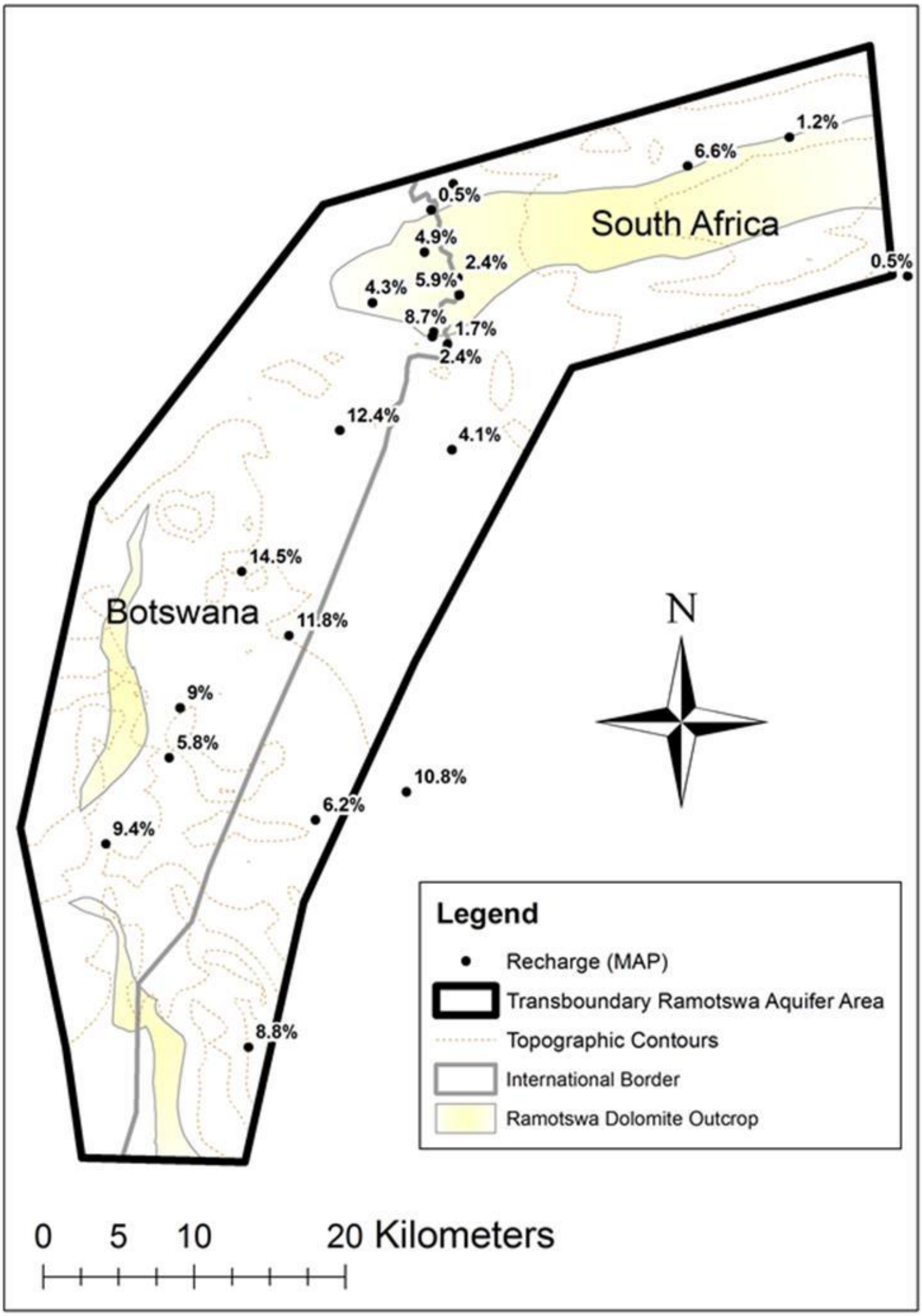


Figure 32: Point recharge estimates within the greater Ramotswa area

5.3. Comparison of current and historic groundwater recharge processes and estimates

Hydrochemical and tracer analyses indicated that groundwater recharge in the Ramotswa dolomitic aquifer takes place mainly by two mechanism of flow; a displacement of moisture in the subsurface zone by the recent infiltrated moisture (or rainfall) through a diffuse or piston like flow through permeable soils and from concentrated runoffs due to ponding in surface enhanced by surface depressions, and a preferential flow component through fractures that outcrop at surface and riverbed infiltration along the ephemeral Notwane River. Supporting the evidence for runoff percolation due to surface depressions were high EC, TDS and Cl values within the northern part of the study area, while low EC, TDS and Cl values were measured within the high topographical areas in the southern part of the study area, which also supported the supposed rapid infiltration or flow of water through preferential flow paths. According to Xu and Beekman (2003), low TDS, chloride content and Ca-HCO_3 groundwater type indicates active or renewable groundwater resources with short residence time of water near surface and high infiltration capacity.

Compatible to the above mentioned results, Gieske (1992) utilized chloride and isotope tracer profiling techniques in the shallow soil overlying the Ramotswa dolomite formation in Pitsanyane and suggested that recharge through these areas occurs either through direct infiltration through the dolomite outcrops, moisture displacement through sandy soils and river bed infiltration. Similarly, Beekman et al. (1997) applied the combine subsurface and saturated zone CMB method, ^{3}H, ^{2}H and ^{18}O in Letlhakeng-Bothlapatlou (LB) area (located within few kilometres from the study) and indicated that the subsurface zone profiles in the area can be attributed to a bimodal flow model comprising of preferential flow component and a diffused flow component.

Van Wyk (2010) estimated recharge rates in the dolomitic terrains located in the Vaalian Gaap Plateau Group dolomites in upper Kuruman in South Arica and proposed a by-pass flow of rainwater through soil zone due to the karstic nature, to be the source of recharge in those areas. Van Wyk (2010) further suggested that direct recharge through the dolomite formation contributes approximately 100% of the total recharge within the area.

Stable isotope signature in groundwater in Ramotswa dolomitic aquifer plotted along with the local precipitation, further suggesting that the high and short intense rainfall normally experienced in summer months and depleted in heavy isotopes seems to be the main source of recharge into the aquifer. Approving the above made suggestion Vogel and Ehhalt (1963) found that in semi-arid regions of South Africa, which include the greater Ramotswa area, groundwater is characterized by relatively less heavy isotopes compared to precipitation but it corresponds well with the isotopic composition of heavy rainfalls. Van Wyk (2010) was of the same view as Beekman et al. (1997) and indicated that in arid and semi-arid areas in South Africa, isotopically depleted rainfall mark the recharge producing rainfall surplus.

The overall groundwater recharge estimates from the current study were relatively high (Table 10; Figure 34; Figure 33) compared to recharge estimates from previous studies conducted within the greater Ramotswa area. However, current recharge estimates in the northern parts of the study varied from 0.4% to 14% of mean annual precipitation and these results seemed to correlate well with the recharge estimates from the previous studies (Figure 34). For example Jennings (1970) estimated recharge rates to be 12.4% to 83.63% of mean annual rainfall, while van Wyk (2010) estimated total recharge of about 24.2 mm/yr within the dolomitic areas in upper Kuruma in South Africa. Beekman and Xu (2003) suggested that annual recharge within arid regions can be averaged from less than 1 mm/yr to more than 60 mm/yr (Figure 34), which is the case in the study area although exceptional high recharge rates above 100 mm/yr were questionable considering that annual rainfall falls below 500 mm/yr in most of these regions while the loss of moisture through evapotranspiration triples the annual rainfall. While Abiye (2016) is in a view that groundwater recharge through the use of the CMB method should probable fall below 4% of rainfall below 500 mm while it may reach up to 20% if rainfall go beyond 600 mm per annum.

Gieske (1992) applied environmental tracers such as Cl, tritium and also conducted soil moisture analyses to estimate recharge and estimated recharge rates as varying from 19 mm/yr to 23 mm/yr through the outcrops and shallow soil cover in Kanye region in Lobatse. Within the same study area and for sites in Ramotswa dolomite formation, a recharge value of about 20 mm/yr was estimated through shallow soil cover overlying the dolomite formation in the Pitsanyane east area.

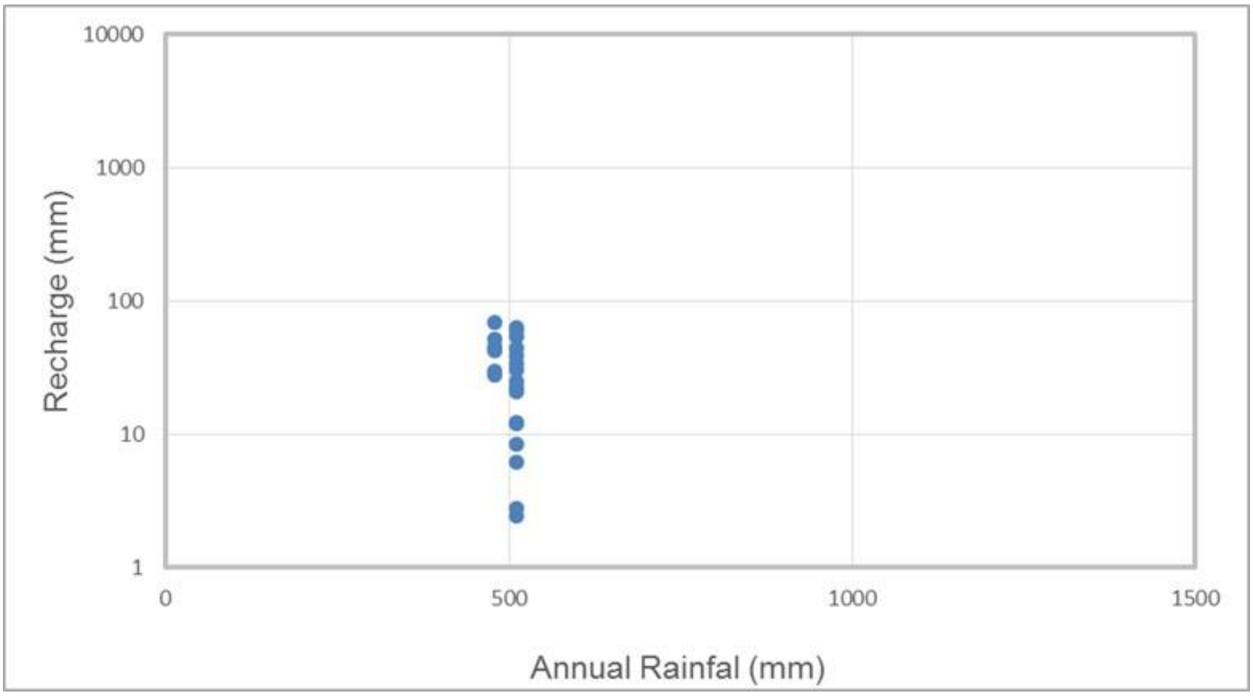


Figure 33: Annual groundwater recharge rates estimated with CMB method

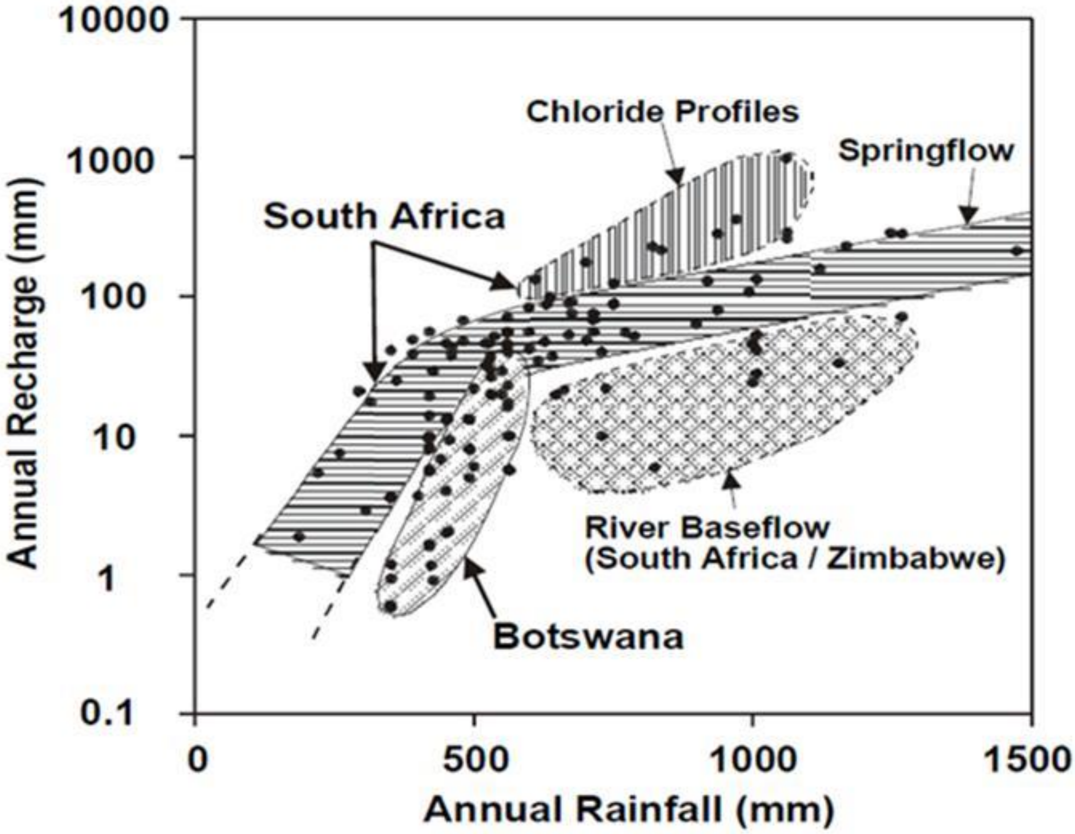


Figure 34: Groundwater recharge rates estimates from the studies conducted in Southern Africa (Xu and Beekman 2003)

5.4. Summary

Hydrochemical and tracer analyses indicated that groundwater recharge in Ramotswa dolomitic aquifer takes place mainly by two mechanism of flow; a displacement of moisture in the subsurface zone by the recent infiltrated moisture (or rainfall) through a diffuse or piston like flow through permeable soils and from concentrated runoffs due to ponding in surface enhanced by surface depressions, and a preferential flow component through fractures that outcrop at surface and riverbed infiltration along the ephemeral Notwane River. Groundwater recharge rates varied from 2.4 mm/yr to 69 mm/yr. Recharge estimates correlated well with the suggested mechanism and processes governing recharge into the aquifer.

Chapter 6

6. Conclusion and Recommendations

Evaluation of the characteristics of the transboundary Ramotswa dolomitic aquifer was made through reviewing the physiography, climate, land-use and vegetation, surface-water hydrology, geology, and the hydrogeology. Groundwater within the transboundary Ramotswa dolomitic aquifer showed similar chemical nature reflecting water-rock interaction which further suggested a relatively short groundwater travel/residence times. Groundwater was mainly characterized by alkaline condition; Ca-Mg-HCO_3 type indicative of the recently recharged water resulting from the dissolution of the host (dolomite) rocks. Correlation between EC, TDS, $\delta^{18}O$ and elevation gave us some useful ideas on the recharge and discharge zones, and also on the mechanism governing recharge. . The Low EC, TDS and Cl^- values within the mountainous regions located in the southern parts of the study area which include the upper Dinokana and Lobatse region, suggested a rapid or direct recharge into the aquifer through preferential path-ways such as fractures that outcrop on the surface or plant root channels or earth worm channel within the subsurface soil layer. The northern part of the study; Ramotswa wellfield area and the Supingstand area, which is characterized by a relatively flat topography, resulted relatively high EC, TDS and Cl^- values suggesting indirect or localized recharge from surface runoff or recharge through the stream beds such as the Notwane River.

The stable isotope signature in groundwater was characterized into two groups. The first group of samples plotted above and on the local meteoric water line, signifying un-evaporated isotopic composition, where rainfall events directly recharged the aquifers. The second group plotted just slightly below the local meteoric water line, indicating indirect or localized recharge thus allowing isotope fractionation to occur on the surface.

Groundwater samples showing an un-evaporated isotopic composition were scattered across the study area but mostly located within the southern parts of the study area, while groundwater samples showing signs of evaporation were found across the study area but mostly located in Ramotswa wellfield and upper Dinokana. All groundwater samples contained detectable amounts of tritium which indicated recharge. Tritium signature varied from 0.2 TU to 2.3 TU with low tritium values of less than 0.8 TU indicative of sub-modern waters approximately recharged before 1950s, while waters with high tritium values were considered to be a mixture of modern and sub-modern waters indicative of recently recharged waters or active recharge.

Chloride profiles within the study area varied across the basin due to subsurface heterogeneity. The Lower Ramotswa, Notwane River, North Lobatse and the upper Dinokana profile showed similar trend as per overall depth profile. These profiles lacked significant chloride peaks showing relatively constant chloride content with increasing depth indicative of direct recharge mechanism through preferential paths ways. The upper Ramotswa and Moshana profiles reflect that either diffuse or preferential flow occurs within the same profile due to the chloride peaks varying with depth on each of the profiles. The upper Ramotswa indicates some sort of sequence in the chloride oscillation such as diffuse flow followed by preferential flow, alternating up to the maximum depth, while the Moshana profile has a more complex profile but seemingly with alternating preferential and diffuse flow mechanisms up to 2 m depth. Generally, below the 2 m depth the significance of evapotranspiration seems to diminish.

Quantified groundwater recharge rates varied from 2.4 mm/yr to 69 mm/yr and these were distributed across the study area. Recharge estimates were relatively high in the upper Dinokana and Lobatse region located decreased towards Ramotswa and Supingstand area in the north of the study area. Findings on the evaluation of the processes governing recharge in the transboundary Ramotswa dolomitic aquifer indicates that recharge mainly occurs either through direct or indirect recharge processes. Direct recharge seems to be prevailing in the southern part of the study area, while indirect recharge is largely within the upper part of the study area. Such parts of the study area should be preserved in order to avoid pollution and also useful for managed aquifer recharge. Preserving these parts of the study area may include declaring them as protected areas on both countries in order to restrict any development activities that would diminish the potential value of the site in terms of recharging the aquifer.

As current recharge rates were comparable to the previous studies conducted in the greater Ramotswa area. A way forward from the quantified recharge rates would be to equate abstraction rates to average recharge rates or upper limits for groundwater exploitation and this should be monitored by the Limpopo Watercourse Commission. Equated abstraction rates based on recharge estimates should first address the basic human needs and environmental water requirements before allocating water needs to other users.

Recommendations

A quantitative and qualitative research approach was taken to investigate the processes governing recharge and to estimate groundwater recharge rates within the transboundary Ramotswa dolomitic aquifer. However, further investigations and refinement is needed to improve or to strengthen the reliability of study findings as follows:

Evaluation of the processes governing recharge mechanism

- Integrate subsurface and saturated zone CMB approach to estimate recharge
- Monitor and establish isotope tracer profiles (^{3}H, ^{2}H and ^{18}O) for longer duration
- Conduct various infiltration tests across the basin and mostly in the Notwane River streambed to estimate infiltration capacities.

Quantification of groundwater recharge rates

- Improve the chloride mass balance approach accommodate long term data to average recharge within the study area as the groundwater respond to average changes or accumulation over time.

- Use both primary and secondary data such as groundwater level time series data.

General issues or future studies to consider:

- Establish rainfall isotope monitoring station to better understand isotopic characteristics of the study area.
- Conduct basic hydrological monitoring; precipitation, groundwater level, stream flow, evaporation and water use.
- Investigate the processes affecting recharge such precipitation, evapotranspiration, vegetation, soil and geology (lithology and structure, and development of the karst formations).
- Develop an integrated remote sensing and GIS approach to delineate recharge and discharge and to distinguish the characteristics of those areas.

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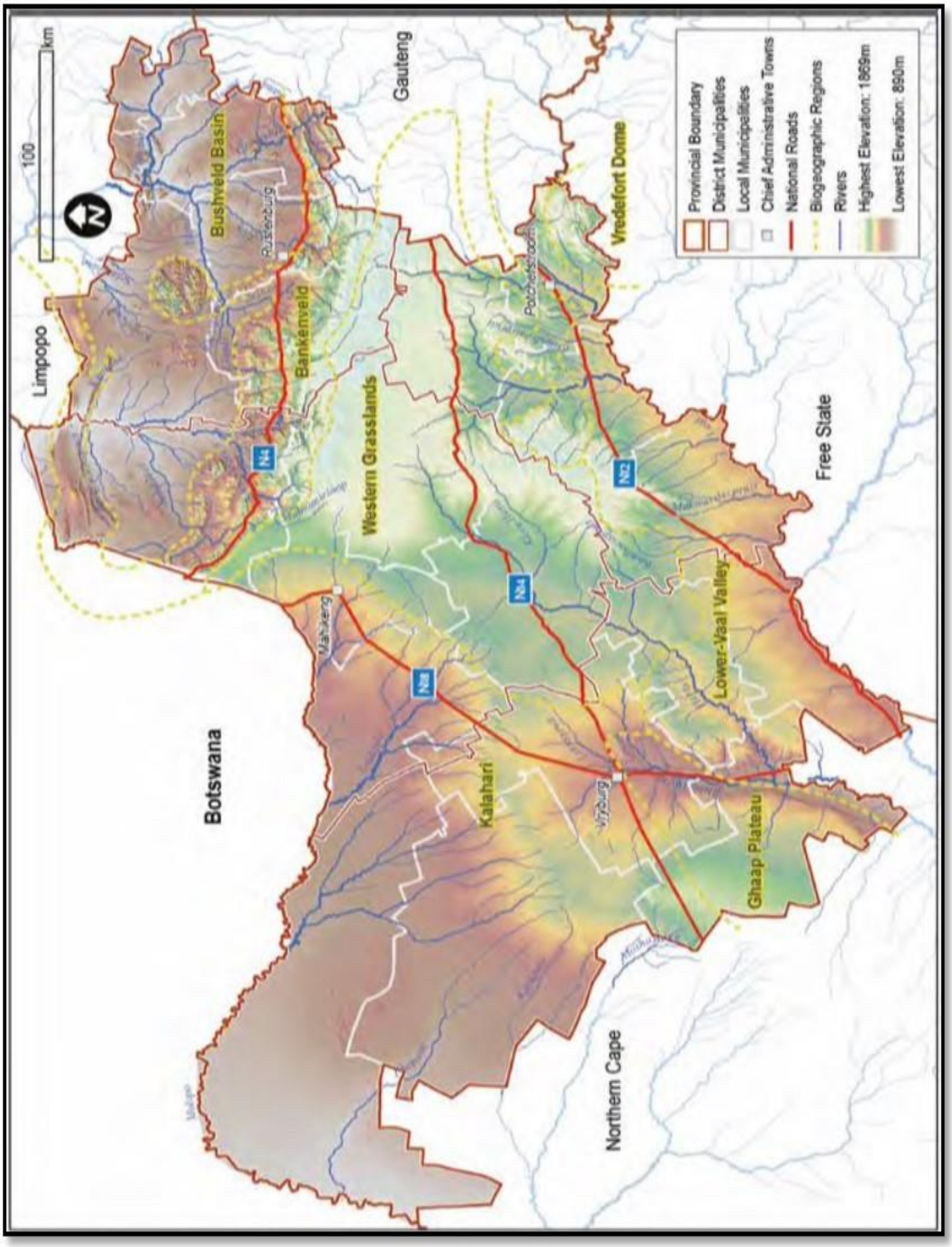
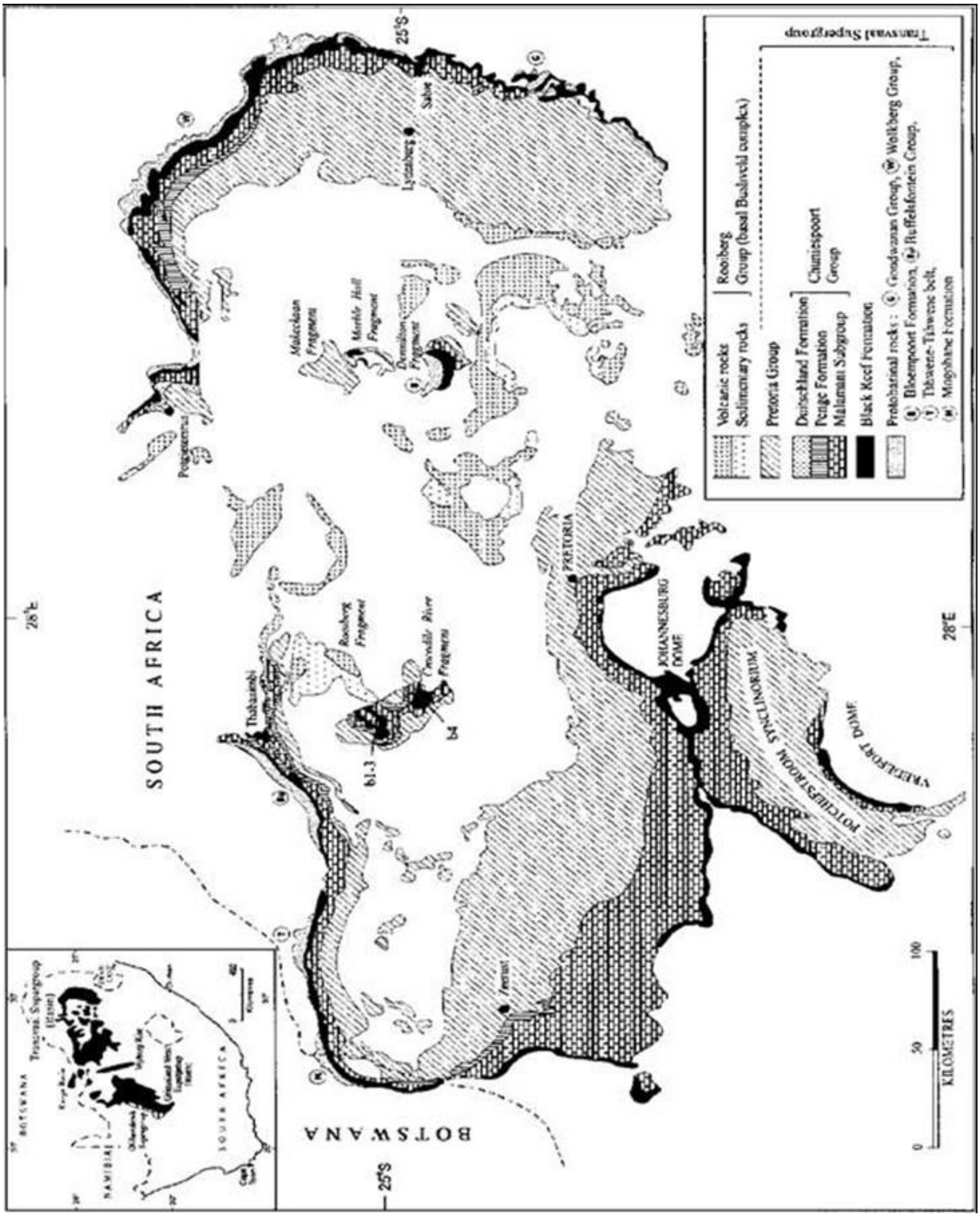
Appendices

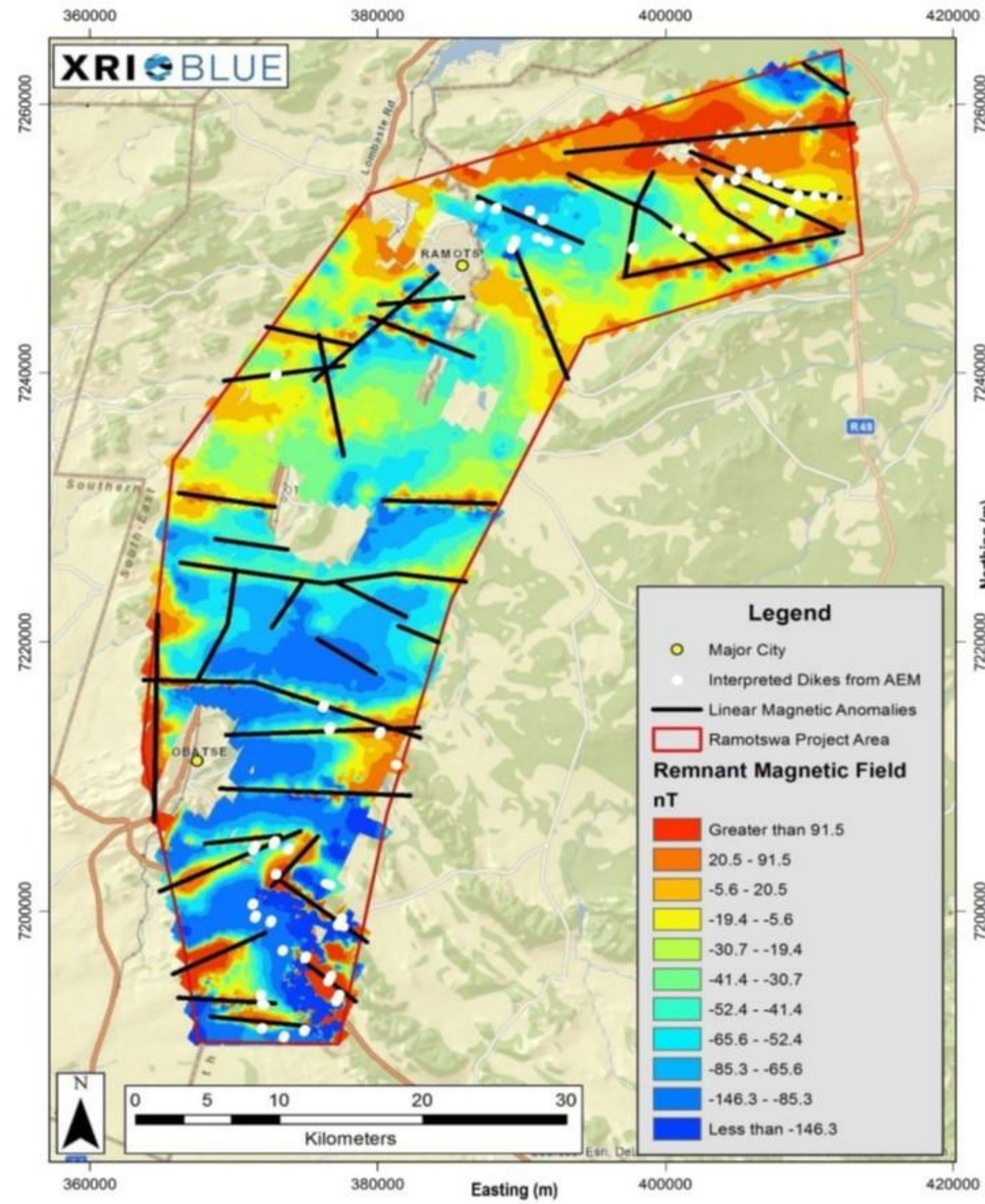
Appendix A: North West Topographical context map including the RTBAA (READ, 2015)



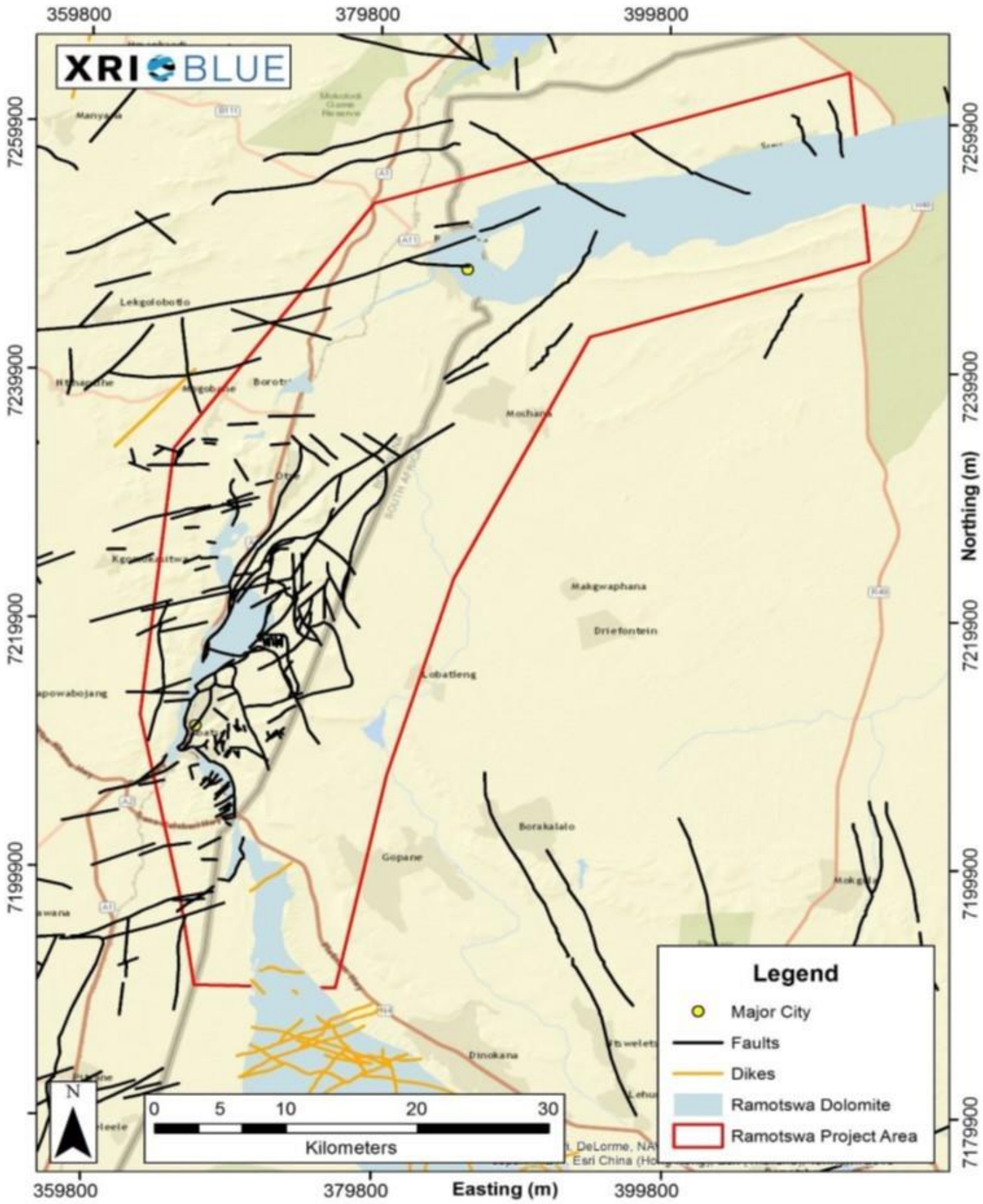
Appendix B: Geological map of the Transvaal basin showing the Transvaal Supergroup, basal Protobasinal rocks, overlain by the Black Reef Formation, Chuniespoort group, the uppermost Pretoria group (Erikson et al., 2001).



Appendix C: Remnant Magnetic Field, linear magnetic anomalies, and locations of interpretation dikes from the AEM data in the Ramotswa Project Area (XRI Blue, 2016).



Appendix D: Mapped surface extent of the Ramotswa Dolomite, faults, and dikes in and around the greater Ramotswa area (XRI Blue, 2016)



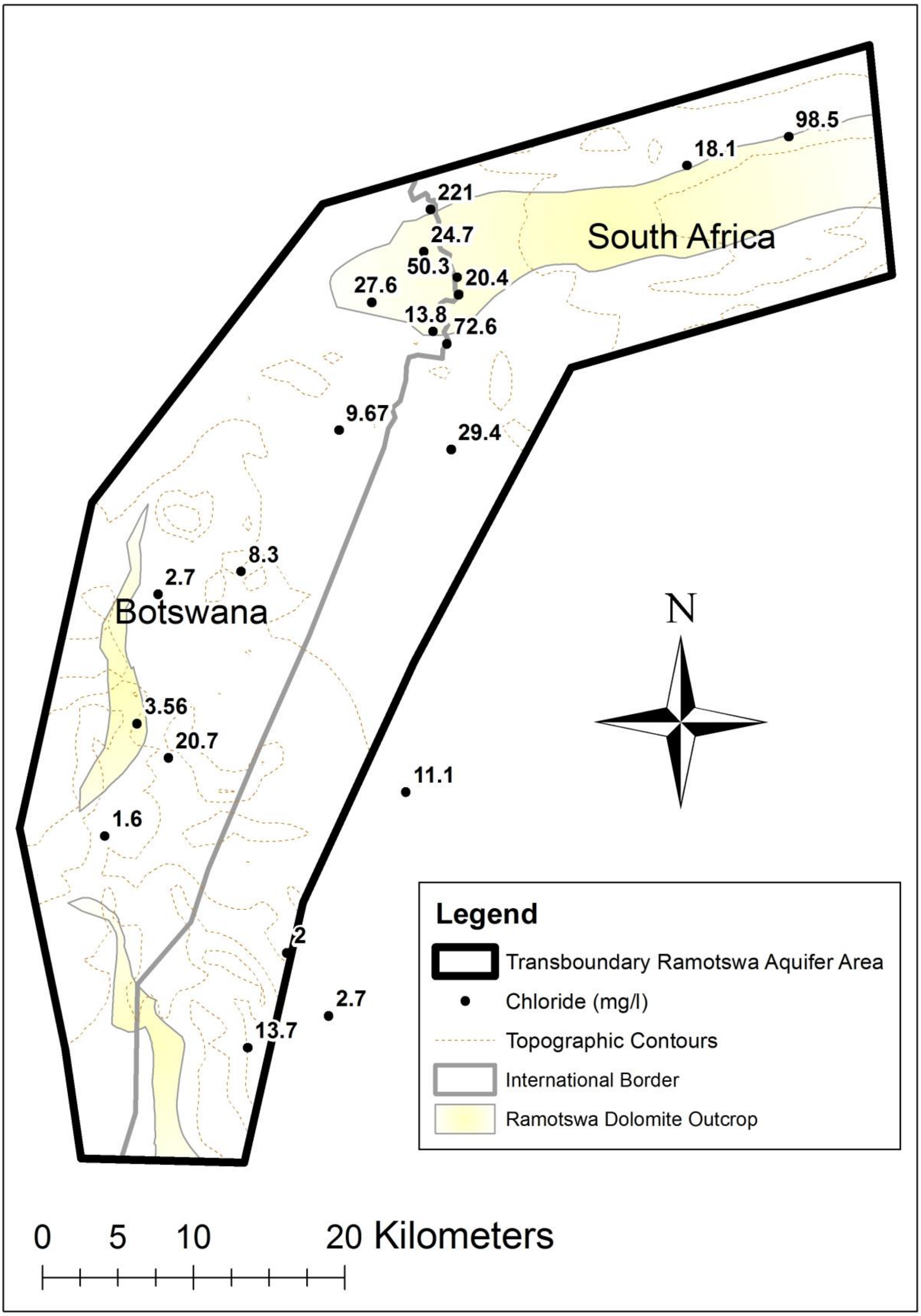
Appendix E: submersible pump

Appendix F: Primary water quality parameters

Sample_Name	EC (µS/cm)	TDS (mg/l)	pH	Temp (ºC)
ODI01ZA	15.6		-	-
ODI02ZA	34.4		-	-
ODI03ZA	299.85	217.15	7.02	17.3
ODI04ZA	32.56	110.3	6.83	26.5
ODI12ZA	586.5	354	7.65	16.3
ODI07ZA	520	310	7.62	16.8
ODI08ZA	1294.8	766	7.46	23.3
ODI09ZA	905.85	512	7.72	19
ODI10ZA	364.85	212	7.88	15
ODI11ZA	737.25	432	8	21.5
ODI13ZA	1225.2	706	7.84	9.7
ODI14ZA	303.05	177.6	8.22	13.6
ODI15ZA	93.75	33.6	9.22	22.8
ODI16ZA	147.95	88	8.68	18.3
ODI17ZA	282.2	164.5	8.1	12
ODI19ZA	540.8	319	7.39	18
ODI20ZA	358.8	208	6.6	-
BH4458BWA	970	626	8.05	22.4
BH6333BWA	880	567.5	7.77	24.4
BH919BWA	386.6	249.3	8.6	23
BH6329BWA	356.5	228.5	8.6	23
BH4975BWA	2051	1325.5	7.56	22.6
BH10129BWA	805.5	520	8.14	26.8

BH6424BWA	831.5	536.5	7.12	25.9
BH4373BWA	801.5	512	8.39	24.7
BH4340BWA	845	544	7.09	24.4
BH4337BWA	852	277		
BH310BWA	550.5	354	7.93	24.3
BHKhodaAgenciesBWA	133	85.9	8.55	23.1
BH4450BWA	623.5	404	8.14	23.5
BH4974BWA	1286	515.56	8.3	22.4
BH287BWA	1507	975	7.87	26.5
BH4348BWA	1268.5	821.5	8.03	26.5
BH2285BWA	235	470.25	8.62	22.3
BH4400BWA	398	253	7	23.4

Appendix G: Chloride distribution map



Appendix H: Stable isotope results from the greater Ramotswa area

Sample Name	Latitude	Longitude	Elevation	GW level	δ 2H (‰)	± 2H StDev	δ 18O (‰)	± 18O StDev	
ODIO20ZA	-24.976	25.88292	1052	-	-29.51	0.00	-7.14	0.00	Supingstand Area
ODIO11ZA	-24.8735	25.88668	1036	-	-24.12	0.00	-5.80	0.00	
ODIO9ZA	-24.8071	26.02321	1161	-	-41.99	0.39	-8.82	0.09	
SUB04ZA	-24.8071	26.02321	1156	-	-41.73	0.26	-8.72	0.08	
ODIO8ZA	-24.7899	26.08373	1151	-	-40.96	0.02	-8.63	0.03	
BH4337BWA	-24.8839	25.88731	1085	-	-32.28	0.25	-5.62	0.06	Ramotswa Wellfields
BH4348BWA	-24.8995	25.87702	1024	13.9	-34.44	0.37	-6.01	0.05	
BH4373BWA	-24.9057	25.87224	1025	-	-32.42	0.15	-5.41	0.05	
BH4450BWA	-25.0621	25.70864	1122	38.6	-30.00	0.00	-4.83	0.00	
BH4975BWA	-24.8331	25.87078	1021	7.03	-8.67	0.10	-1.50	0.04	
BH4340BWA	-24.913	25.88041	1028	-	-32.73	0.23	-5.76	0.07	
BH6424BWA	-24.8884	25.83583	1052	27.7	-35.28	0.14	-6.01	0.07	
BH287BWA	-24.8795	25.86945	1032	21.87	-48.22	0.00	-9.74	0.00	
BH10129BWA	-24.8583	25.86679	1031	7.9	-32.20	0.28	-7.85	0.09	
DUGWELLBWA	-24.5202	25.52561	1030	+2	-39.84	0.31	-8.53	0.06	
ODIO4ZA	-25.3317	25.76192	1292	-	-36.53	0.00	-6.01	0.00	Upper Dinokana Area
ODIO3ZA	-25.3317	25.76192	1294	-	-31.47	0.00	-4.71	0.00	
DNK01ZA	-25.4354	25.83693	1297	-	-41.40	0.21	-7.16	0.00	
ODIO12ZA	-24.8273	26.18361	1125	-	-32.60	0.20	-5.71	0.06	
ODIO19ZA	-25.1797	25.85607	1091	-	-39.77	0.16	-8.90	0.05	
ODIO2ZA	-25.3128	25.81025	1182	-	-31.74	0.00	-5.25	0.00	
ODIO15ZA	-25.2753	25.78522	1184	-	-45.93	0.00	-9.49	0.01	
GPN01ZA	-25.3065	25.79019	1229	-	-51.74	0.21	-10.33	0.07	
BHKhodaAgenciesBWA	-25.0486	25.75813	1090	-	-41.56	0.01	-7.51	0.02	Lobatse Area
BH310BWA	-24.9645	25.81624	1069	-	-11.53	0.00	-2.47	0.00	
BH6333BWA	-25.1297	25.72124	1173	32.83	-44.66	0.31	-9.13	0.10	
BH4458BWA	-24.8583	25.86679	1197	16.15	-31.66	0.12	-7.26	0.08	
BH6329BWA	-24.8583	25.86679	1174	42.32	-47.24	0.19	-9.55	0.04	
BH 919BWA	-25.206	25.67701	1239	19.7	-31.54	0.36	-7.16	0.06	

Appendix I: Tritium Data and Dating results

^{3}H Site	Lat.	Long.	Piezometer	GW level (m)	^{3}H (TU)	Detection Limits	Interpreted age (years)	Recharge classification*
T1	-24.82	25.88	ALVK 01ZA	-	0.70	±0.2	24.86	Sub-modern
T2	-24.88	25.87	BH287BWA	21.87	1.50	±0.3	12.43	Modern - Sub-modern
T3	-24.96	25.82	BH310BWA	-	2.00	±0.3	<12.43	Modern - Sub-modern
T4	-24.88	25.89	BH4337BWA	-	0.70	±0.2	24.86	Sub-modern
T5	-24.91	25.88	BH4340BWA	-	0.50	±0.2	<37.29	Sub-modern
T6	-24.90	25.88	BH4348BWA	13.89	0.40	±0.2	<37.30	Sub-modern
T7	-24.91	25.87	BH4373BWA	-	0.20	±0.2	≤49.72	Sub-modern
T8	-25.06	25.71	BH4450BWA	38.59	0.20	±0.2	≤49.73	Sub-modern
T9	-25.16	25.71	BH4458BWA	16.15	1.50	±0.3	12.43	Modern - Sub-modern
T10	-24.83	25.87	BH4975BWA	7.03	0.90	±0.3	<24.86	Modern - Sub-modern
T11	-25.14	25.70	BH6329BWA	42.32	1.60	±0.3	≤12.43	Modern - Sub-modern
T12	-25.13	25.72	BH6333BWA	32.83	1.20	±0.3	<24.86	Modern - Sub-modern
T13	-24.89	25.84	BH6424BWA	27.7	0.60	±0.2	<37.29	Sub-modern
T14	-24.86	25.87	BH10129BWA	7.9	2.30	±0.3	<12.43	Modern - Sub-modern
T15	-25.44	25.84	DNK01ZA	-	0.60	±0.2	<37.29	Sub-modern
T16	-25.85	25.46	DNKEYEZA	-	0.20	±0.2	≤49.73	Sub-modern
T17	26.43	-25.81	GME01ZA	-	0.60	±0.2	<37.29	Sub-modern
T18	-25.31	25.81	GP01ZA	-	0.20	±0.2	≤49.73	Sub-modern
T19	-25.31	25.79	GPN01ZA	-	1.10	±0.3	<24.87	Modern - Sub-modern

^3H Site	Lat.	Long.	Piezometer	GW level (m)	^3H (TU)	Detection Limits	Interpreted age (years)	Recharge classification*
T20	-25.21	25.68	LOB01BWA	-	0.40	±0.2	<37.30	Sub-modern
T21	-25.18	25.86	LobatlaZA	-	1.50	±0.3	12.43	Modern - Sub-modern
T22	-25.28	25.79	MaphephaneZA	-	0.40	±0.2	<37.30	Sub-modern
T23	-25.28	25.79	Maphephane02ZA	-	0.90	±0.3	<24.86	Sub-modern
T24	-24.83	25.18	MDKW01ZA	-	1.20	±0.3	<24.86	Modern - Sub-modern
T25	-25.30	25.75	MKQ02ZA	-	0.40	±0.2	<37.30	Sub-modern
T26	-25.33	25.76	MM01ZA	-	0.20	±0.2	≤49.73	Sub-modern
T27	-24.98	25.88	MoshaneZA	-	1.80	±0.3	<12.43	Modern - Sub-modern
T28	-25.05	25.76	OtseBWA	-	0.80	±0.2	≥24.86	Modern - Sub-modern
T29	-24.81	26.02	SUB03ZA	-	1.40	±0.3	>12.43	Modern - Sub-modern
T30	-24.81	26.02	SUBZA04	-	0.60	±0.2	<37.29	Sub-modern
T31	-24.87	25.89	SWBP01ZA	-	1.00	±0.3	<24.86	Modern - Sub-modern

Appendix J: Groundwater recharges estimates

Name	Lat	Long	Cl in GW (mg/L)	RE (mm/yr)	RE % (MAP)
ODI02ZA	-25.3128	25.81025	2.7	213.3	44.4
ODI03ZA	-25.33167	25.76192	13.7	42.0	8.8
ODI04ZA	-25.29813	25.75307	5.7	101.1	21.1
ODI12	-24.82734	26.18361	15.9	38.6	7.5
ODI07ZA	-25.08656	25.7861	10.2	60.1	11.8
ODI08ZA	-24.7899	26.08373	98.5	6.2	1.2
ODI09ZA	-24.80713	26.02321	18.1	33.9	6.6
ODI10ZA	-24.81761	25.88392	11.3	54.3	10.6
ODI11ZA	-24.87351	25.88668	50.3	12.2	2.4
ODI13ZA	-24.87273	26.15427	254.9	2.4	0.5
ODI14ZA	-25.46634	25.89235	6	102.2	20.0
ODI15ZA	-25.27526	25.78522	2	288.0	60.0
ODI16ZA	-25.22059	25.79753	3.5	164.6	34.3
ODI17ZA	-25.19637	25.80203	19.4	29.7	6.2
ODI19ZA	-25.17969	25.85607	11.1	51.9	10.8
ODI20ZA	-24.97598	25.88292	29.4	20.9	4.1
BH4458BWA	-25.15933	25.71485	20.7	27.8	5.8
BH6333BW	-25.12969	25.72124	13.3	43.3	9.0
BH919BWA	-25.20586	25.67701	1.6	360.0	75.0
BH2285BWA	-25.21076	25.67714	12.7	45.4	9.4
BH6329BWA	-25.13888	25.6962	3.56	161.8	33.7
BH4975BWA	-24.83311	25.87078	221	2.8	0.5
BH10129BWA	-24.85832	25.86679	24.7	24.8	4.9
BH4974BWA	-24.90841	25.87133	51	12.0	2.4
BH6424BWA	-24.88836	25.83583	27.6	22.2	4.3
BH4373BWA	-24.90574	25.87224	13.8	44.4	8.7
BH4340BWA	-24.91299	25.88041	72.6	8.4	1.7
BH4337BWA	-24.88393	25.88731	20.4	30.1	5.9
BH310BWA	-24.96447	25.81624	9.67	63.4	12.4
BHKhodaAgenciesBWA	-25.04858	25.75813	8.3	69.4	14.5
BH4450BWA	-25.06211	25.70864	2.7	227.1	44.4
Cl in rainfall (mg/l)	**Lobatse mean annual rainfall**	**Ramotswa mean annual rainfall**			
1.2	480	511			