Hydrogeological Model for the Ramotswa Transboundary Aquifer Area

March 2018

Submitted by:

Implemented by:

In partnership with:

This report was made possible by the support of the American people through the United States Agency for International Development (USAID). Its contents are the sole responsibility of IWMI, and do not necessarily reflect the views of USAID or the United States Government.

Contributors
Girma Y. Ebrahim (IWMI)
Karen G. Villholth (IWMI)
Jonathan Lautze (IWMI)
Geert-Jan Nijsten (IGRAC)
Manuel Magombeyi (IWMI)
Keetile Keodumetse (DWA-Botswana)
Piet Kenabatho (University of Botswana)
Sakhile Mndaweni (DWS - South Africa)
Moses Moehadu (WUC, Botswana)
Modreck Gomo (University of Free State –South Africa)
Bochengedu Somolekae (DWA-Botswana)
Charles Nkile (DWA-Botswana)
Daniel Kealeboga (BGI)

Acknowledgments
The contribution and support from the Department of Water and Sanitation (DWS) – South Africa, The Department of Water Affairs (DWA) - Botswana, Botswana Geoscience Institute (BGI) – Botswana, and the Water Utilities Corporation (WUC) – Botswana is acknowledge.

Executive Summary

Increasing groundwater resource potential In the face of increasing population, urbanization, climate change and unreliability of surface water supplies in arid and semi-arid regions, groundwater resources are being increasingly used. The relative reliability of groundwater for long-term supply and its potential to serve as a buffer against drought make groundwater a critical water source. Managed Aquifer Recharge (MAR) is an alternative water resource management option that is gaining attention as a way to increase the quantity of water that is stored underground when there is excess during wet periods for use during dry period. Strategic implementation of MAR can enhance the benefits derived from groundwater use. However, determining the feasibility of MAR involves careful assessment of: i) aquifer response to additional recharge, ii) capacity of the aquifer to store water, iii) optimally locating MAR infiltration/ injection and recovery systems.

Hydrogeological models can be used to contextualize the role and feasibility of MAR. Hydrologeological models can be used to gain a better understanding of the aquifer system such as recharge dynamics, groundwater-surface water interaction, the effect of groundwater pumping, and storage processes and overall water budget of the aquifer system. With respect to MAR, hydrogeological models enable us to determine the storage capacity of the aquifer, aquifer response to additional recharge, the location of recharge and recovery systems and recovery efficiency. More broadly, hydrogeological modelling can provide a practical indication of the contribution and impact of MAR in an aquifer system.

Objective: The objective of the present study is to develop and calibrate three-dimensional (3D) steady sate hydrogeological model which can be used to establish initial condition for 3D transient hydrogeological model in compartment 3 of the Ramotswa Transboundary Aquifer Area (RTBAA). It is the 3D transient model that can be used for the upcoming MAR feasibility assessment.

Approach: A steady state hydrogeological model was developed using MODFLOW 2005 in MODELMUSE modelling environment. The karst aquifer was modelled using Equivalent Porous Media approach. The modelled area covers a 61 km^2 compartment that encompasses the Ramotswa village and wellfield area. The hydrogeological model focused on the Ramotswa dolomite but including also areas underlain by other formations such as Lephala Formation. The model is calibrated against average water level data observed during the period of 2000-2012. This period was selected for steady state calibration because there was no pumping in the aquifer in this period. Hence, this period was assumed to represent the steady state condition. In order to understand recharge dynamics, independent recharge estimation was accomplished using the Water Table Fluctuation method (WTF). The WTF method is applied using water level data at two observation wells, one close to the river and far away from the river. Two dimensional (2D) transient profile modelling was also carried out along the Ngotwane River. The 2D profile model was calibrated against four observation wells' water level data for the period 2000-2012. The main purpose of the 2D prolife modelling was to estimate focused recharge due to river leakage and better constrain hydraulic parameter estimation such as hydraulic conductivity and storage coefficients along the river.

Results: The steady state hydrological model produced comparable groundwater level gradient with the overall regional groundwater flow direction. Results obtained also showed that: (i) diffuse recharge from rainfall was about 28 mm/d. Focused recharge is almost twice of the diffused recharge rate (50 mm/a). (ii) Approximately 80% of the recharge entering the aquifer exits as Groundwater Evapotranspiration (GWET) and only 20% of the recharge leaves through the Northern model boundary where Ngotwane River crosses the Black Reef formation. Spatially averaged diffuse recharge estimates from the model compared well to previous estimates in the study area. Groundwater replenishment from focused recharge is associated with large uncertainty as a result of its dependence on the magnitude and frequency of flooding. WTF method produced diffused recharge which is approximately 20% of rainfall values. Both WTF method and 2D transient profile modelling produced focused recharge estimate of more than 80% of rainfall values.

Summing Up: The present model represents an important first step towards a 3D transient model development and for comprehensive effort for MAR feasibility assessment. The current model provides an initial condition or reference level for transient model calibration. While additional steps remain, the results of this report indicate that the calibrated steady state model is representative to be used as initial condition for subsequent transient model calibration and the hydraulic parameters are within the reasonable range. Recharge estimated by the model suggests good groundwater potential.

Next Steps Next steps are: 1) transient 3D model calibration and validation, 2) sensitivity analysis of model parameters, 3) MAR scenario analysis using a calibrated and validated 3D transient hydrogeological model. The forthcoming feasibility assessment, to be undertaken through the hydrogeological model scenario analysis, will help to: determine the volume of water to be added to storage in the aquifer, identify suitable sites for MAR application and to optimally site the MAR system, understand how the aquifer system react for additional recharge, and to determine the length of time the recharged water remain in storage in a particular area. In addition to hydrogeological model scenario analysis forthcoming feasibility assessment include two more important objectives: i) assessing water resources availability that can be used as a recharge source and its quality, ii) evaluating the geochemical implication of mixing recharge water used as a source and native groundwater in the aquifer.

Table of Contents

List of Figures

List of Tables

Acronyms

amsl	Above Mean Sea Level
AET	actual evapotranspiration
BGI	Botswana Geoscience Institute
DEM	Digital Elevation Model.
DWA	Department of Water Affairs (Botswana)
DWS	Department of Water and Sanitation (South Africa)
ET	Evapotranspiration
EPM	Equivalent Porous Medium
GCS	Geotechnical Consulting Services, Botswana
GIS	Geographic Information System
GPS	Global Positioning System
GWET	Groundwater Evapotranspiration
IGRAC	International Groundwater Resources Assessment Centre
IWMI	International Water Management Institute
IDW	Inverse Distance Weighted
LULC-	Land Use Land Cover
MAR	Managed Aquifer Recharge
mbgl	metres below ground level
ModelMuse	Graphic user interface for MODFLOW-2005, MODFLOW-OWHM, and others
MODFLOW	Modular Three-Dimensional Finite-Difference Groundwater Flow Model
SRTM	Shuttle Radar Topography Mission
SWL	Static Water Level
USDA	United States Department of Agriculture
USAID	United States Agency for International Development
USGS	U.S. Geological survey
WLE	Water, Land and Ecosystems
WUC	Water Utilities Corporation (Botswana)

1. Introduction

1.1 Background

In arid/semi-arid regions, surface water resources are generally scarce and unreliable, hence, groundwater is being increasingly used as a critical source of water. The reliability of groundwater for long-term supply and its buffering capacity during drought periods make groundwater a critical water source. Managed Aquifer Recharge (MAR) practices are increasingly being used as a means of increasing groundwater availability and improving the overall reliability of water supplies. Storing water in an aquifer during times of excess supply and recovering the same water for use when the demand is high is becoming an attractive water management option. Hydrogeological models can be used to assess the feasibility of MAR prior to conducting expensive field test. They provide a quantitative technique for analysing the effect of groundwater pumping, recharge dynamics, groundwater-surface water interaction, and can be used to develop a more reliable estimate of aquifer water budget. With respect to MAR assessment they provide important information such as storage capacity of the aquifer, the response of the aquifer to induced recharge, determining the location of recharge and recovery, and assessing the recovery efficiency (Mansouri and Mezouary, 2015; Woolfenden and Koczot, 2001).

1.2 Modelling Objectives

The **main objectives** of the Ramotswa Aquifer hydrogeological modeling are:

1) to investigate the use and movement of groundwater, recharge, discharge and storage process.
2) to assess the feasibility of MAR, through model scenario analysis

The specific modeling objectives are:

- to develop a 3D hydrogeological model that describes the movement of groundwater and current water budget, recharge, discharge and storage process,
- to predict the aquifer response to induced recharge through MAR such as building up of groundwater level (mound),
- to determine storage capacity of the aquifer for additional recharge through MAR,
- to identify suitable sites for MAR application and optimally site the MAR schemes.

This report summarizes the data, methods used to develop, and calibration of the steady sate 3D hydrogeological model of the Ramotswa Transboundary Aquifer in the area surrounding the Ramotswa village and potential limitations and source of uncertainties in the model. The remainder of this report is organized as follows: First, the scope of the work, activities completed and activities which are part of the second phase of the modelling work is presented. Second, study area including geology and hydrogeology briefly described, available water level data is presented. Third, previous modelling works are reviewed. Fourth, hydrogeological model development methodologies are described. Fifth, aquifer properties are described and past pumping test results are presented. Six, model calibration period is described. Seven, steady state model calibration results are presented. Eight, model calibration as well as water budget analysis results are presented. Nine, discussion and model limitation are presented. Finally, conclusions are presented. In Annex 1, independent recharge estimation using the Water Table Fluctuation Methods is described and results are presented. In Annex 2, 2D profile model construction and calibration results are presented.

1.3 Scope of Work

In order to achieve the above objectives the overall scope of the work is divided into eight major activities. The scope of the present work covers the first three activities and part of activity four.

1. Data collection, compilation and analysis
2. Conceptual model development
    - Develop an updated conceptual model of the groundwater flow for the study area
3. Model construction:
    - Construct a 3D hydrogeological model
4. Model calibration and validation
    - Calibrate the 3D hydrogeological model for the steady state condition
    - Calibrate the 3D hydrogeological model for the transient condition
    - Validate the 3D hydrogeological model for the transient condition
5. Model scenario analysis
    - Optimize MAR location through model simulation
    - Simulate the additional storage capacity of the aquifer system for MAR
6. Sensitivity analysis
7. Assess water source availability and its quality
8. Evaluate the geochemical implication of mixing of recharge water with native groundwater using Geochemical model

2. Study Area

2.1 The Ramotswa Transboundary Aquifer Area

Ramotswa Transboundary Aquifer Area (RTBAA) The Ramotswa Aquifer is located in the Upper Limpopo River Basin encompasses an aquifer shared between South Africa and Botswana. The Ramotswa Aquifer corresponds to the Ramotswa dolomitic aquifer extent mapped based on surface geology. The RTBAA is a slightly broader term than the strict boundary. RTBAA is used to capture areas in the subsurface that are hydrologically linked to the aquifer, but which lie outside the dolomitic aquifer boundaries delineated based solely on surface geology (Figure 1).

Ramotswa Aquifer Flight Area The flight area (area about 1,500 km^2) was commonly used as an encompassing boundary within which the aquifer was found. It was used to overcome ambiguities of a precise boundary for the aquifer in phase 1 of the RAMOTSWA project. Airborne geophysical surveys were indeed conducted in within this flight area in 2016 (Figure 1).

Gaborone Dam Catchment The Gaborone catchment area, located in the Upper Limpopo River Basin (Area ~4,318 km^2, Figure 1), reflects the immediate surface water boundaries within which the Ramotswa Aquifer is located. Given the linkages between surface and groundwater, the catchment is a very relevant scale. Phase 2 of the RAMOTSWA project treats the Gaborone Dam Catchment as its project study area.

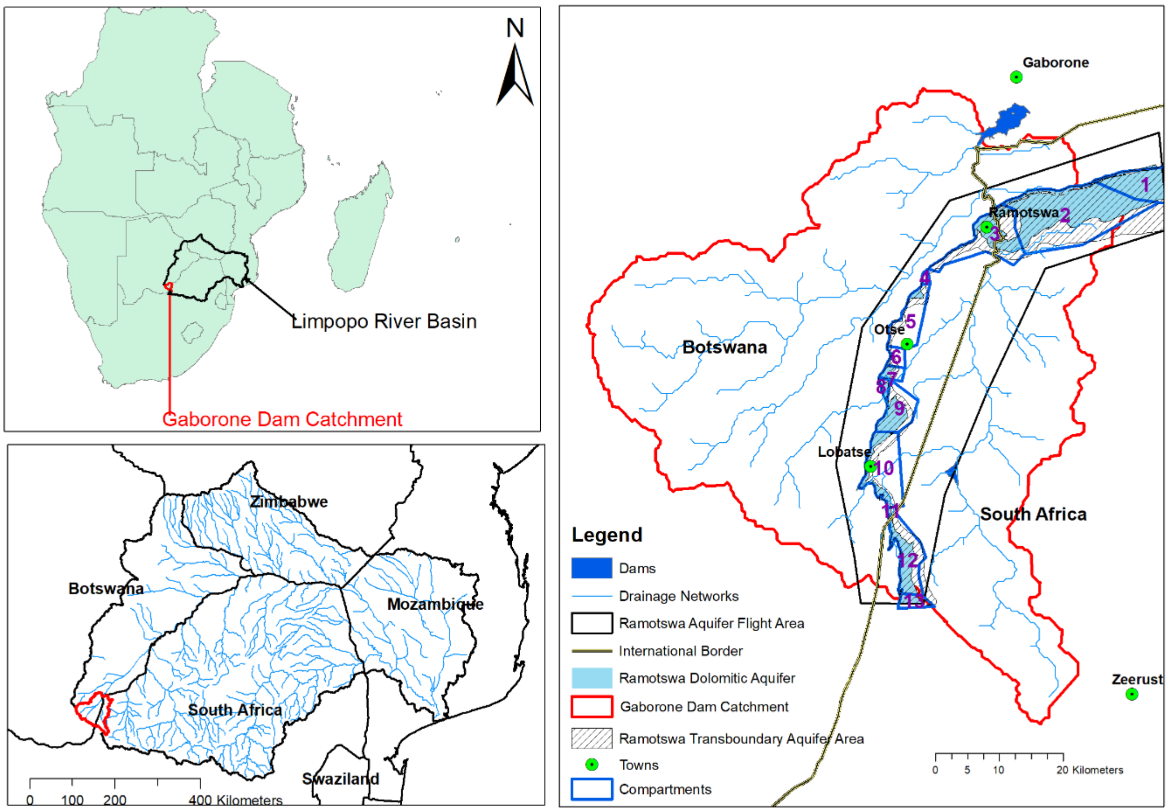

Figure 1: Study area location

2.2 Selected Modelling Area

During the airborne geophysics survey (XRI BLUE, 2016) 13 compartments were identified in the RTBAA (Figure 1). These compartments were delineated by connecting dikes but also by permeability contrast, by identifying less permeable formations. Out of the 13 compartments, only four compartments (3, 10, 11 & 12] are transboundary. Due to availability of data for model calibration, its transboundary nature and its size compared to the other three-transboundary compartments, compartment 3 is selected for modeling purpose. Compartment 3 comprises the Ramotswa wellfield area that supply water for the Ramotswa village. It covers an area of 61 km^2. The Ngotwane River, also known as the Notwane River in Botswana is the largest ephemeral river that crosses the study area. The main vegetation in area is shrub savannah that is characterized by thorn trees with thickets occurring along the river courses (WUC, 2014). As can be seen in Figure 2, the aquifer is highly urbanized on the Botswana side, while the South African side is less developed. The Hillshed image of the study area is shown in Figure 3. Catchment elevation in the modelling area ranges from 1019-1150 m above mean sea level. The annual precipitation ranges from 86-915 mm/a (Figure 4). The mean annual precipitation is 493 mm/a, standard deviation of ± 222mm [1995-2015]. There is high inter annual variability in annual rainfall (coefficient of variation of 45%). The maximum and minimum annual rainfall occur in year 2013 and 2009 respectively. Precipitation mainly takes place from October –March (Figure 5).

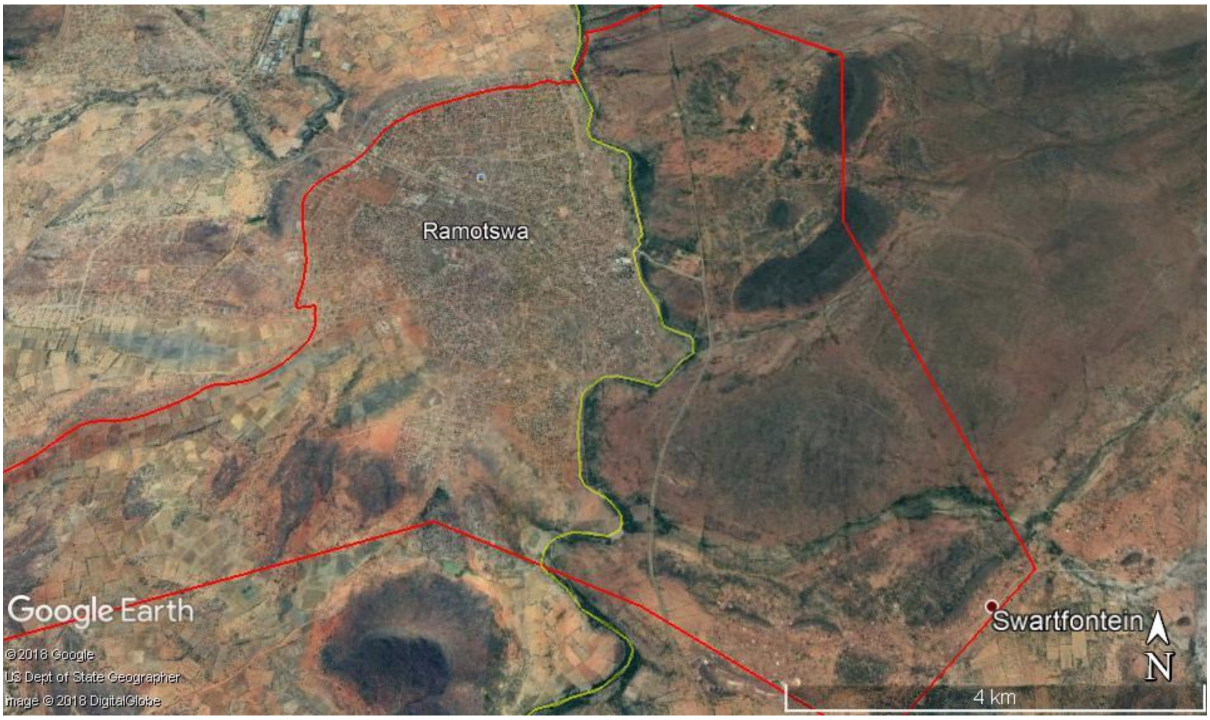

Figure 2: Ramotswa aquifer, Google Earth image. Red polygon shows the boundary of compartment 3 while the yellow line represents the international border which is also matching with the Ngotwane River. Left of the Ngotwane River is Botswana and on the right side is South Africa.

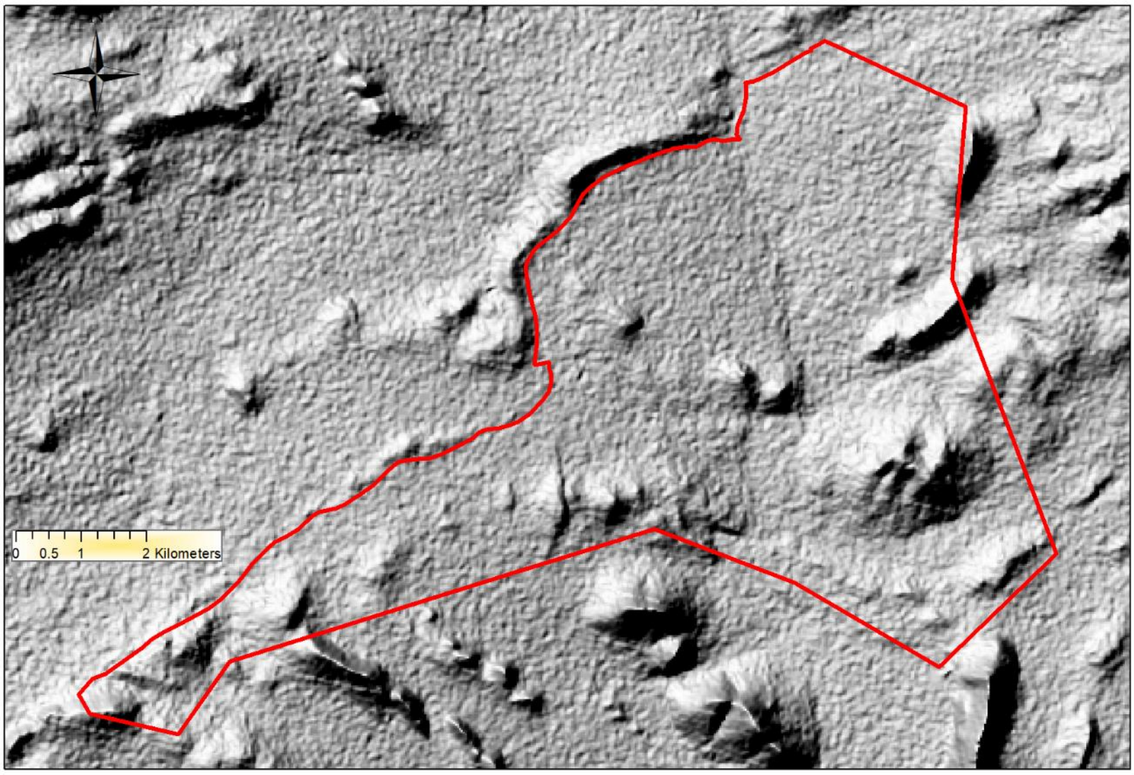

Figure 3: Hillshed image of compartment 3 (Z factor 10)

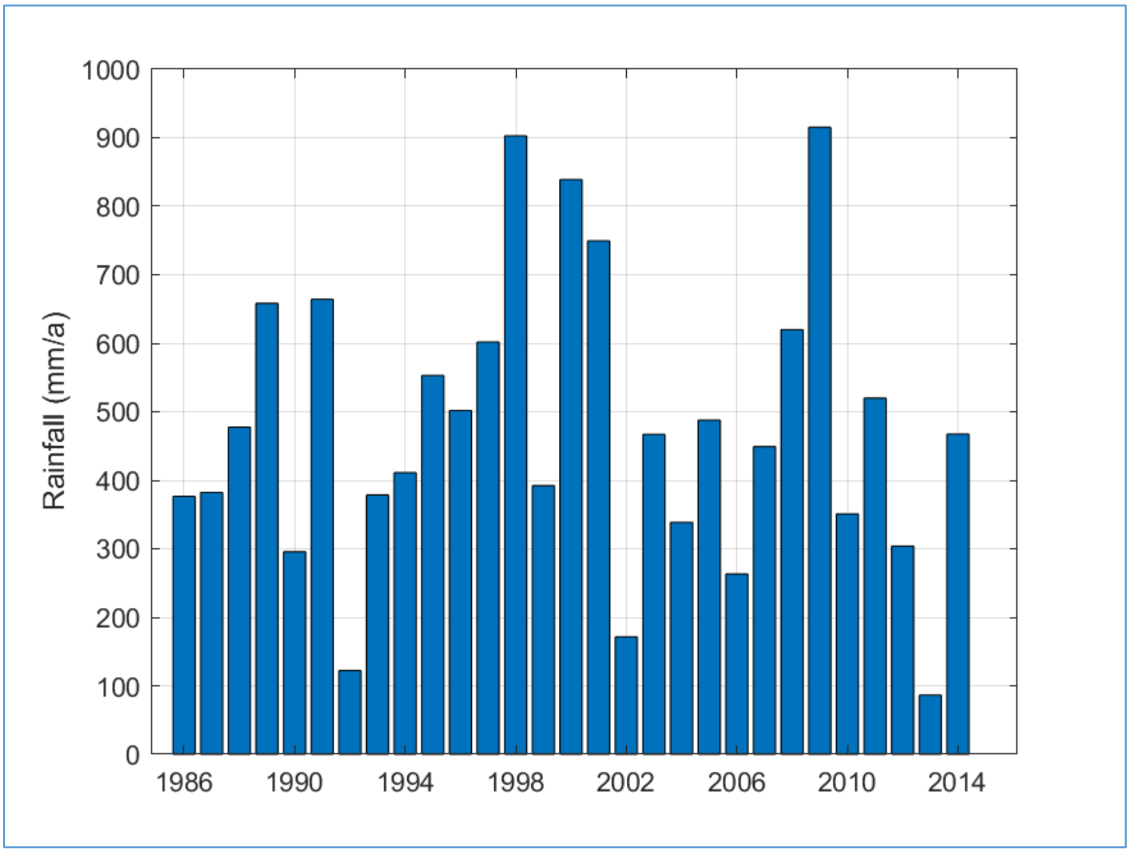

Figure 4: Annual rainfall in Ramotswa station

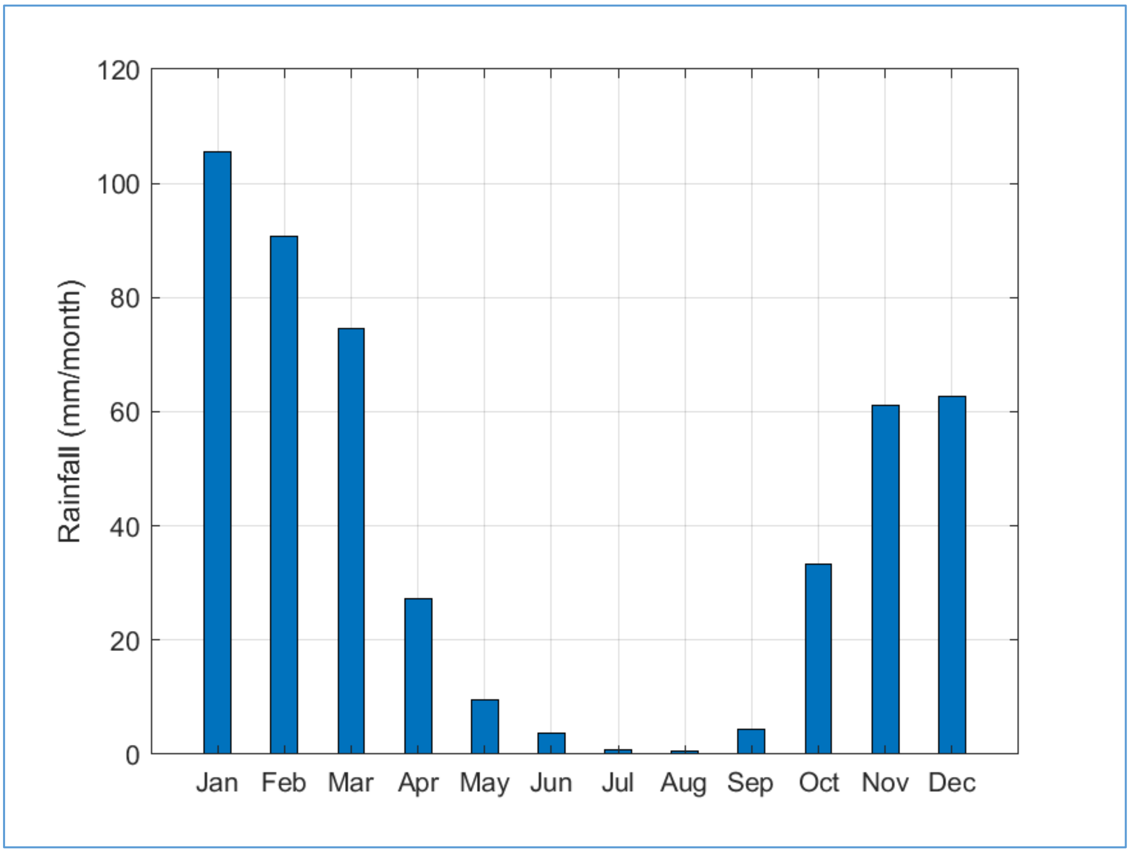

Figure 5: Mean monthly rainfall of Ramotswa station [1986-2014]

2.2.1 Geology

Geologically, the Ramotswa wellfield area is formed from the Transvaal Supergroup. The Transvaal Supergroup is divided into four lithostratigraphic units, i.e. the Protobasinal rocks, Black Reef Formation, Chuniespoort Group and Pretoria group (Catuneanu and Eriksson, 1999; Eriksson and Reczko, 1995). Figure 6 presents these lithostratigraphic units and their chronology. The simplified geology of the study area is shown in Figure 7.The lithology of the Black Reef formation is dominated by clastic rocks ranging from conglomerate to sand stones and mudstones (Catuneanu and Eriksson, 1999). The Chuniespoort Group comprises of seven formations, the five dolomite formation in the Malmani Subgroup (i.e, Oakatree, Monte Cristo, Lyttelton, Eccles and Frisco), Penge and Duitschland Formations (Catuneanu and Eriksson, 1999). The five dolomite formations are differentiated based on their chert content and type of stormatolie as well as by interbedded subordinate carbonaceous mudstones and rare quartzite (Catuneanu and Eriksson, 1999; Eriksson and Altermann, 1998). The Malmani dolomite formations starts from the lower most Okatree, succeeding Monte Christo, Lyttelton, Eccles and upper most Frisco Formation (Eriksson and Altermann, 1998). The Frisco Formation is overlain by iron-rich facies of the Penge Formation, also known as Ramotswa Formation in the Botswana (Catuneanu and Eriksson, 1999). The Penge Formation consists of micro –to macro-banded iron formations with shard structures and subordinate interbeds of carbonaceous mudstone and intraclastic iron formation breccias (Catuneanu and Eriksson, 1999). The upper most Chuniespoort Group formation is the Duitschland Formation overlying the Penge Formation. The Duitschland Formation comprise predominant dolomitic mudstones with interbedded dolomites and quartzites (Catuneanu and Eriksson, 1999). The Pretoria Group consists of 14 formations dominated by clastic and volcanic lithologies (Catuneanu and Eriksson, 1999).

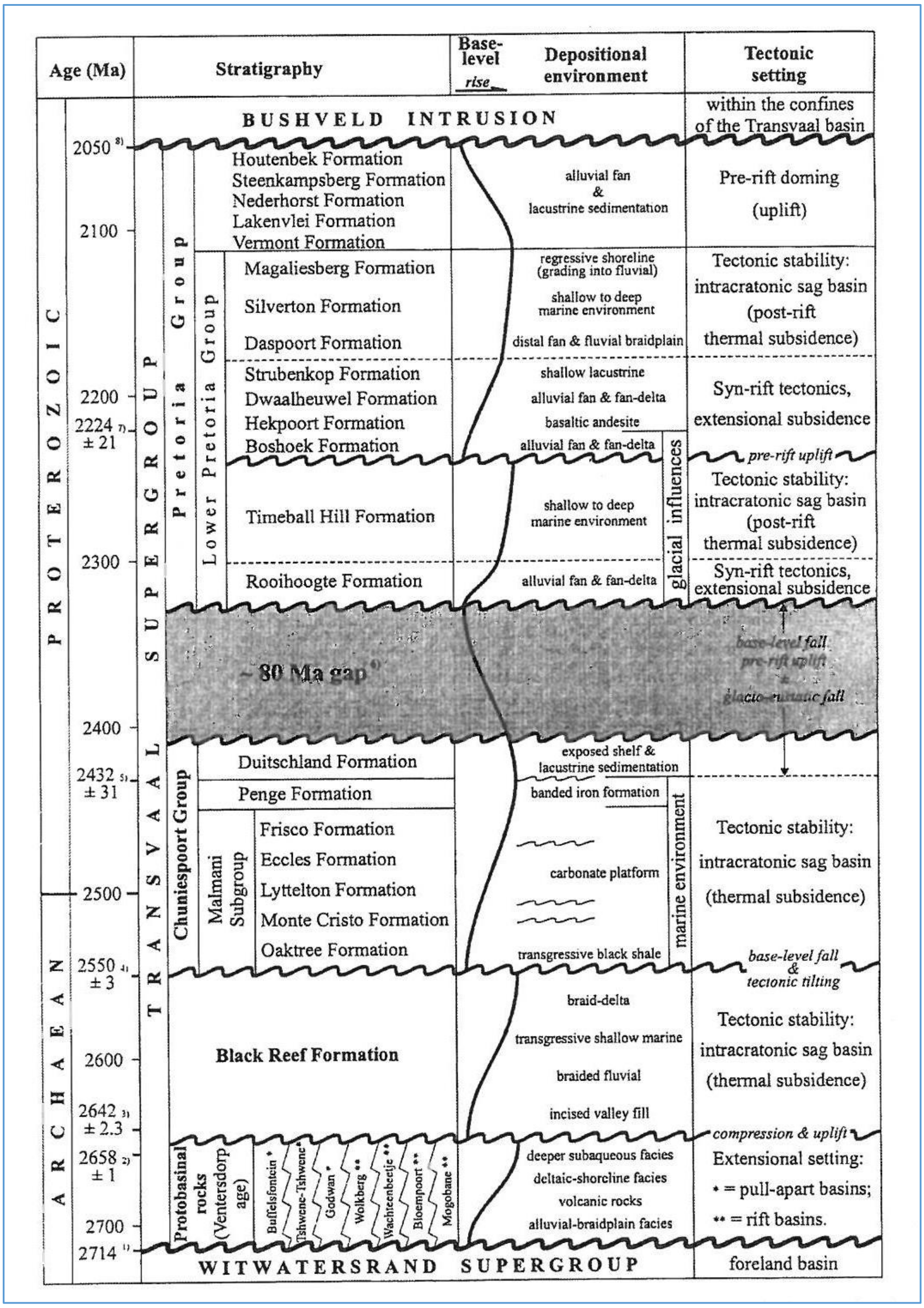

Figure 6: Lithostratigraphy of the Transvaal Supergroup obtained from Catuneanu and Eriksson (1999)

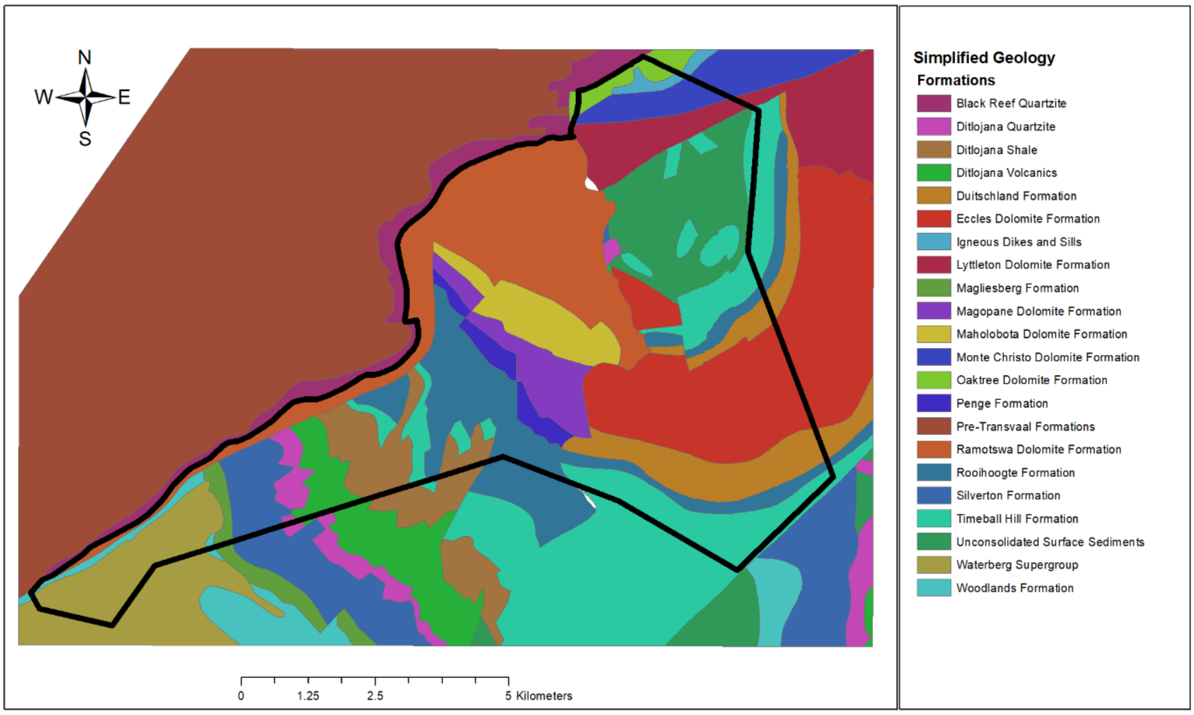

Figure 7: Simplified surficial geology of compartment 3

2.2.2 Hydrogeology

The main aquifer in Ramotswa Transboundary Aquifer is the Ramotswa Dolomite formation. However, a second aquifer systems exists in the RTBAA, the most productive dolomite aquifer and the Lephala formation with relatively low yielding although some high yielding wells are drilled in this formation (GCS, 2000; Selaolo, 1985). Groundwater in the Ramotswa Dolomite is located in parts of the formation where karstification has occurred. The Ramotswa Dolomite comprises five carbonate formations referred to as either “chert-free” or “chert-rich” dolomite. While the chert rich formation **Eccles** and **Mont Chisto** are classified as a good aquifer, the other three chert poor dolomite formations, Oaktree, Lyttelton, and Frisco are regarded as poor aquifers. According to GCS (2000) the Dolomite Aquifer presents two zone of karst development and fissuring.

1) **The upper karstic zone** This zone has a thickness that varies from 20 to 50 m and is thought to be the results of fluctuations in the present water level. Dolomite dissiolution appears preferntially along fractures but the less cherty dolomite also presents karstic features at outcrop. The solution cavities are usually filled with mud or wad.
2) **The deeper karstic zone**. This zone resutls from older fluctuations in the water elvel and has a thickness that ranges from 25 to 50m. The solution cavities are open and generally do not contian mud or wad filings.

The main groundwater flow direction is from South to North and generally follows the natural topography (Selaolo, 1985; Staudt, 2003; WUC, 1989). The aquifer is criss-crossed by impermeable dikes, which affect groundwater flow. The role of the dikes in relation to the groundwater flow is largely unknown. Based on pumping test in the cave sandstone aquifer, Morpulae, Botswana, Morel and Wikramaratna (1982) found that the transmissivity of dolerite dikes is at least hundred times smaller than the transmissivity of the cave sandstone aquifer. In another study in Botswana (Bromley et al., 1994) reported that dolerite dikes less than 10 m thickness are tend to be permeable due to cooling joints and fractures that generate hydraulic continuity across the intrusion whereas, thicker dolerite dikes serve as groundwater barrier Based on the water level differences observed across compartment or dike boundaries in Groundwater Region 10, South Africa, Meyer (2014) concluded that, the dikes at deeper depth are impermeable while some flow across the boundaries may occur within the upper weathered section of the dike.

2.2.3 Groundwater Level Data

The spatial distribution of observation wells in the study area shown in Figure 8. As shown in this Figure most of the observation wells are located in compartment 3 along the Ngotwane River channel. The depth to groundwater from the ground surface ranges form 2.3-25.0 m (Figure 9). Figure 10 shows groundwater level time series data from old monitoring boreholes. Figure 11 presents all groundwater level time series data in the study area. Figure 12 presents the groundwater level time series data for observation wells along the Ngotwane River. Figure 13 presents groundwater level time series data close and far away from the River.

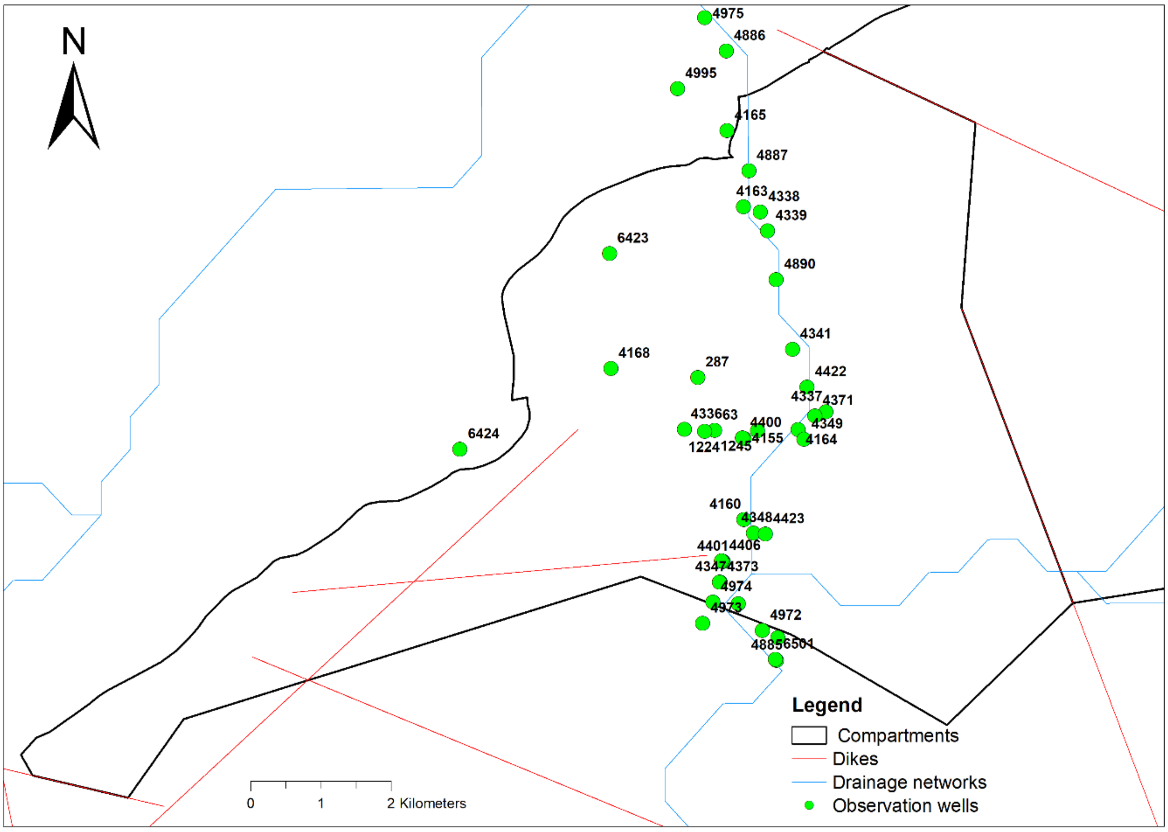

Figure 8: Location of observation boreholes in the Ramotswa (Coordinates for each observation boreholes were obtained from GCS 2000 modelling report)

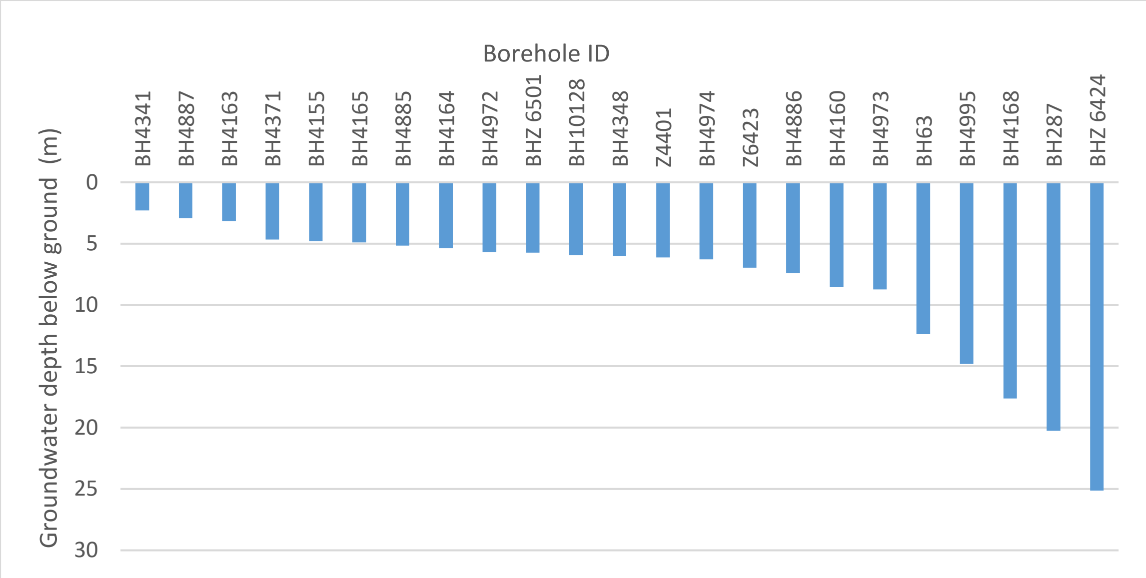

Figure 9: Groundwater depth below the ground surface for 23 observation wells in the study area

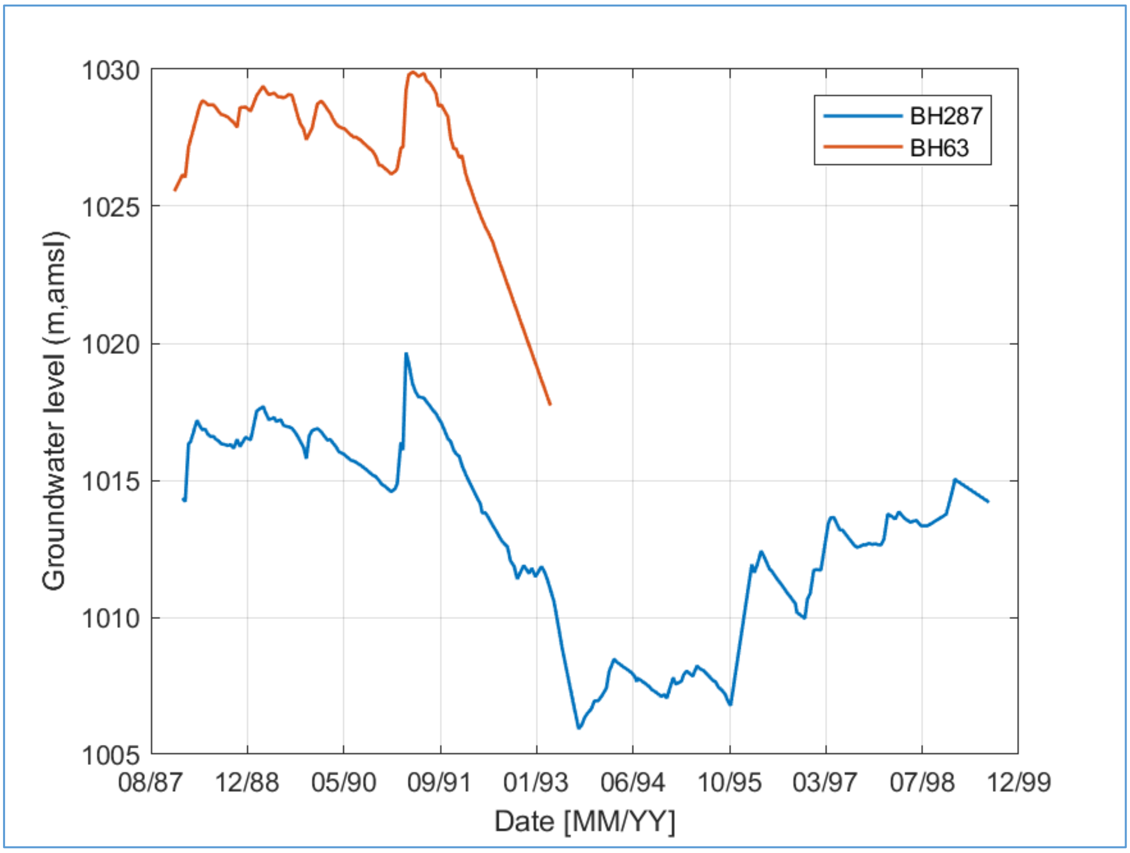

Figure 10: observed water level in two monitoring observation wells with old water level data

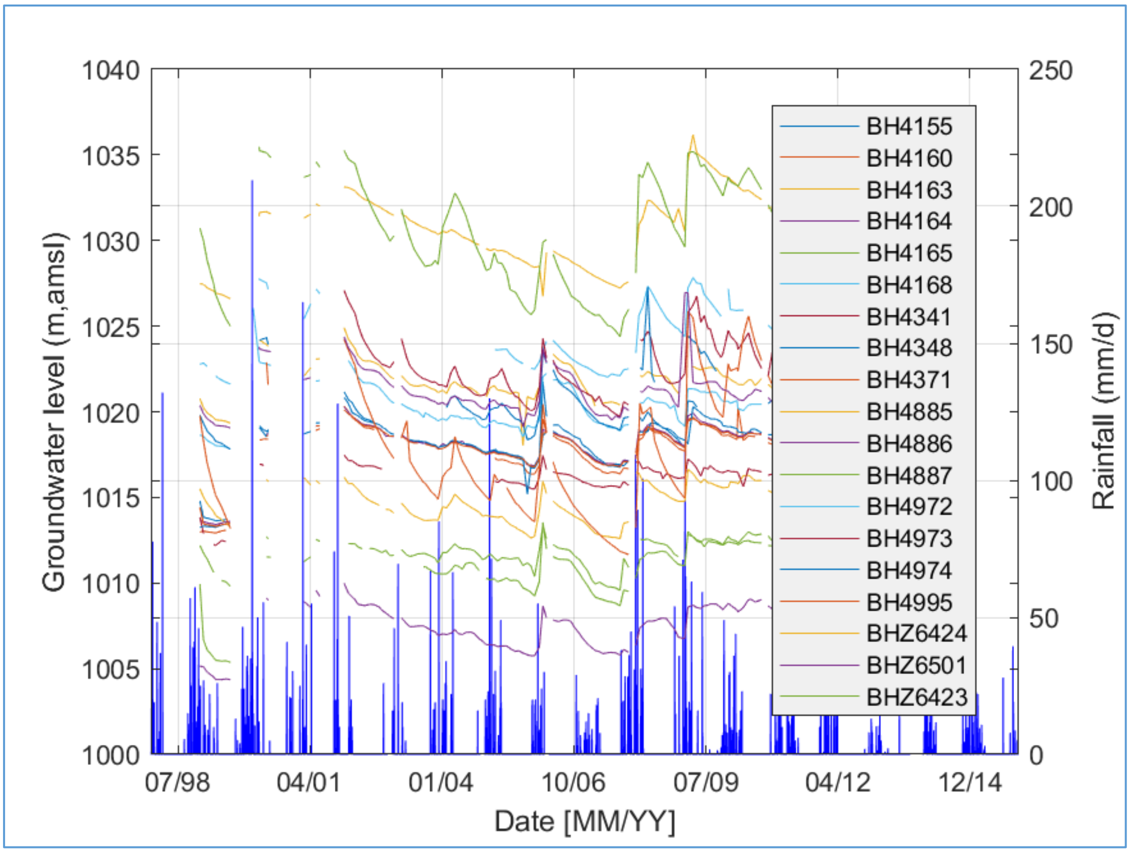

Figure 11: observed water level in 19 observation wells with data from 1999-2012

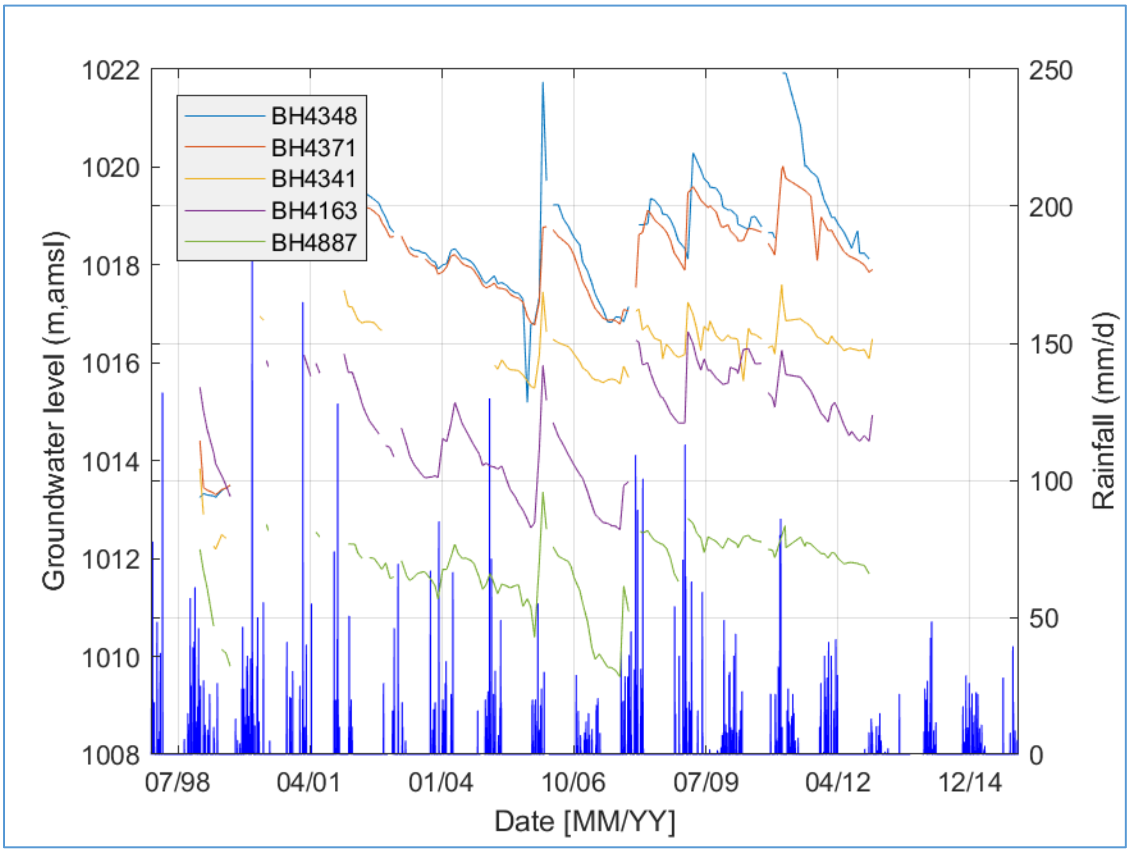

Figure 12: Observed water level data in five monitoring observation wells located along Ngotwane River

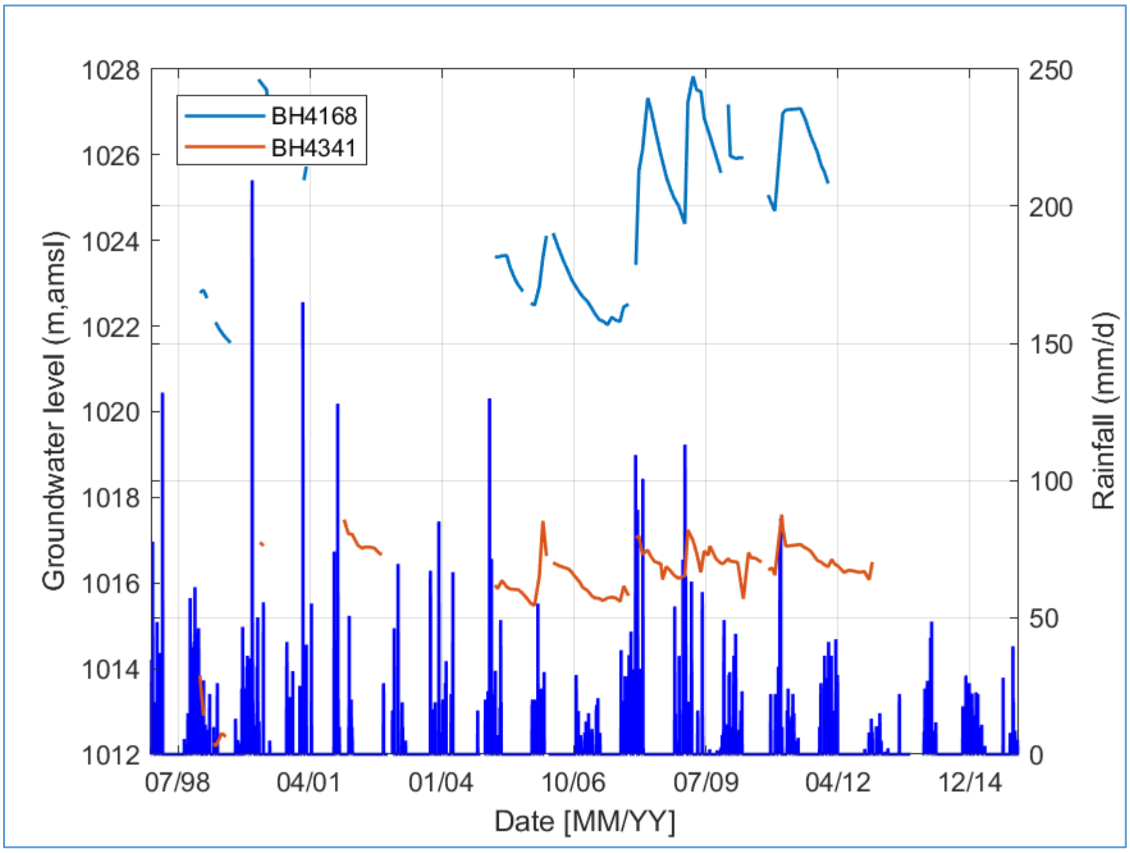

Figure 13: Observed water level near to the river (BH4341) and distant from the river (BH4168)

3. Review of Previous Modelling Studies

The Ramotswa aquifer has been the subject of more than three hydrogeological modelling (GCS, 2000; IoH, 1986; WUC, 1989) and other hydrogeological studies (DWA, 2006; Selaolo, 1985; Staudt, 2003; WUC, 2014). All the past hydrogeological modelling studies were: 1) undertaken with perspective of Botswana not in the transboundary context, 2) focused on groundwater availability assessment and none of the modelling studies were dealing with MAR potential assessment. The present modelling study is undertaken in the transboundary context, focusing on both understanding current aquifer water budget and MAR scenario analysis using more recent information, and data collected in the past two decades since the last modelling studies has been completed. The hydrogeological model boundaries are also defined based on the geophysical study carried out during the first phase of the Ramotswa project. This report reviews two of the modelling studies conducted in the past. WUC (1989) modelling study was not included in the review because of lack of modelling report that describe the first phase of the modelling work that the WUC (1989) used as a basis for modelling.

Institute of Hydrology, University of Bloemfontein, South Africa Assessment

The Institute of hydrology University of Bloemfontein, South Africa (IoH, 1986) modelled the Ramotswa aquifer using the Galrekine finite element method. The modelling area is presented in Figure 14. Rectangular elements were used to represent the different hydrogeological conditions. The rectangular elements have a length of 1 km. Finer grids were used to represent the main linear feature and along the Ngotwane River. Vertically the aquifer is represented using three layers. Spatially, the aquifer area was divided into three zones representing: 1) the massive dolomite rock, 2) the Lephala formation, and 3) the main E-W linear features together with its immediate area. All geologic formations, including the Black Reef Quartzites other than the Ramotswa dolomite formation were included as Lephala formation. The E-W linear feature represents areas of dense fractures. This area is located along the principal E-W karst feature in the Southern area where the main production boreholes are located. The study indicated that fractures together with recharge from the Ngotwane and the intersections with other minor valley from east and west have produced favourable conditions for enhanced secondary permeability by chemical dissolution. The width of the E-W linear feature was assumed to be 250m. According to the study the linear feature extends 2km on the Botswana side of the international boundary and it may extend NE several kilometre into the South Africa. However, it is unclear if the eastern karst has similar groundwater conditions to the zone in the Botswana.

The study adopted no flow boundary in the North and West, and constant head boundary in the South and Eastern sides. An unconfined aquifer is simulated. The aquifer thickness was assumed to be 85m. This was divided into three layers with thickness of 15m for the upper aquifer and 30m for the middle aquifer and 30 m for the lower Aquifer. The hydraulic properties used for the three zones in the three layers are presented in Table 1. To avoid model convergence due to change in hydraulic conductivity an intermediate zone of 300 m width was introduced into the grid between the main linear feature and unfractured rock massive dolomite. Abstraction from four production wells (BH4336m BHZ4400 BH4349 and BH4337) located along the linear feature of the dolomite aquifer were included in the model. The average abstraction rates used in the model for the four boreholes was 2400 m^3/d. Initial conditions representing the pre-abstraction period June 1984 was approximated using constant water level of 1015m. Recharge was not included in the model. The model simulation was compared with drawdown obtained with the long-term pumping test undertaken from July to September 1984. No calibration was attempted during this period except adjusting the simulated drawdown by 30% to account well loss. Although, not statistical performance is reported it was argued that without optimizing the aquifer parameters a good match between observed and model simulated drawdown was observed. Based on the area for each zone and assumed saturated thickness and storage coefficients values, total aquifer storage was computed to be 16.5 Mm^3. This was obtained by multiplying area by saturated thickness and storage coefficient and not by the model.

Table 1: hydraulic parameters used in the study

Aquifer zones	Hydraulic properties	Aquifer vertical layers		
		Upper aquifer	Middle aquifer	Lower Aquifer
Massive dolomite	Permeability (m/d)	0.1	0.1	0.1
	Storativity	0.003	0.003	0.003
E-W linear feature	Permeability (m/d)	65	7	7
	Storativity	0.02	0.02	0.02
Intermediate zone of the E-W linear feature	Permeability (m/d)	10	5	5
Massive Lephala formation	Permeability (m/d)	0.05	0.05	0.05

	Storage coefficient	$3x10^{-4}$	$3x10^{-4}$	$3x10^{-4}$

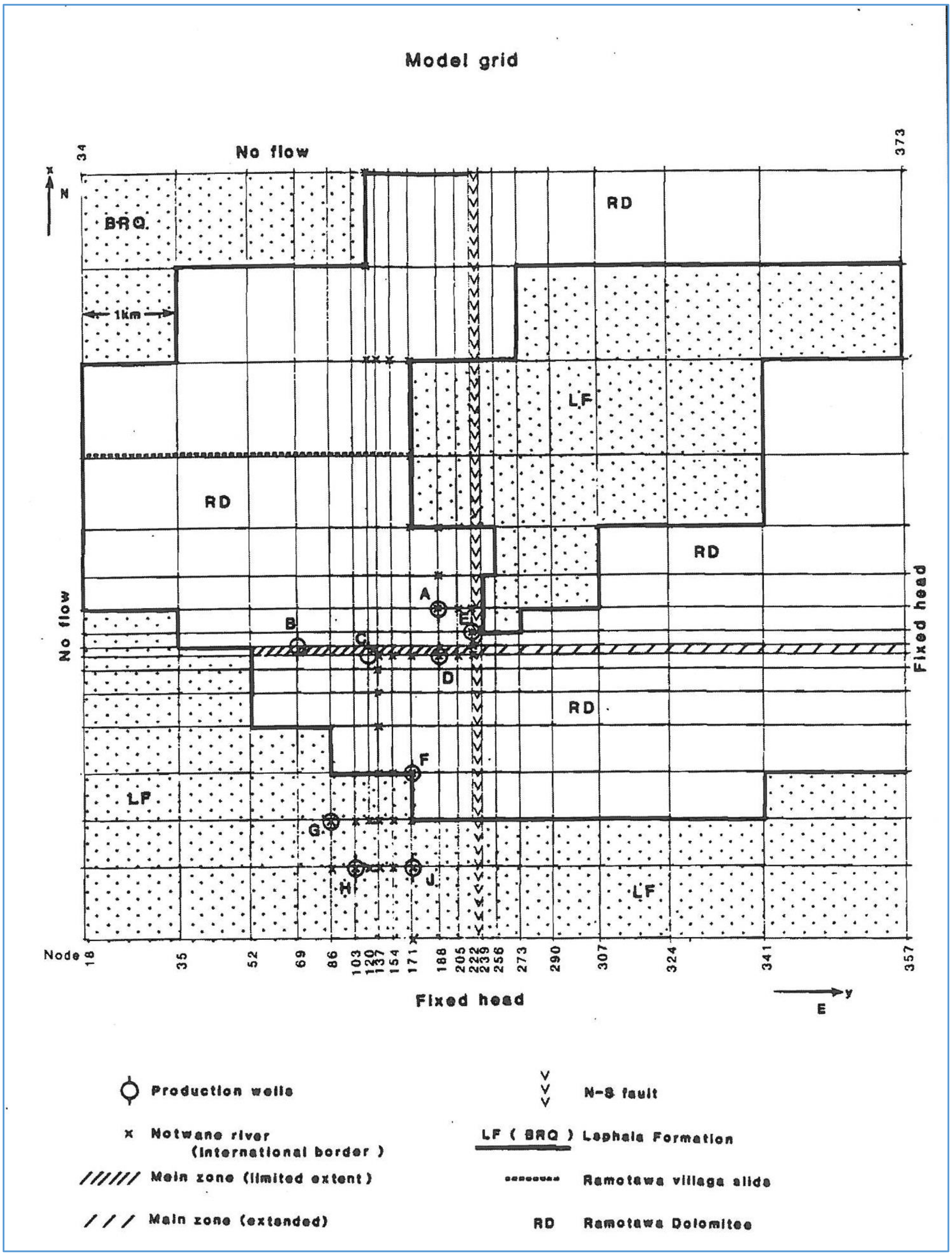

Figure 14: Model area, boundary condition and model grid used in IoH (1986)

Geotechnical Consulting Services (Botswana) Report

The Geotechnical Consulting Services team (GCS 2000) modelled the Ramtoswa Aquifer area in 2000 using the Processing Modflow, PM Version 5. The aim of the modelling work was to reproduce the previous model developed by the Water Utilities Cooperation (WUC) Botswana in 1989 using GWATER software, to improve the modelling work using the latest dataset and to derive some conclusion regarding the sustainability of groundwater abstraction in the aquifer. The model area which was modelled by WUC (1989) and later modified by GCS (2000) is presented in Figure 15. GCS (2000) modelled the area using a two –dimensional model, which is a single layer model. The model area was divided into 500 m grid cells and refined in the proximity of the Ngotwane River using 200 m grid. The model area extends 10km in the EW direction and 12 Km in the NS direction, however, it lacks geographical reference. The WUC (1989) model was bounded by no flow boundary in the West along the Black reef Quartzite, no flow boundary defined in the east approximately 4km from the Ngotwane River along the flow line, No flow boundary in the Southern boundary along the surface water divide, and Fixed head boundary of 1000 m amsl along the contact to the Black Reef Quartizite and Ngotwane river. In the GCS (2000) modelling work the aquifer was modelled as no flow along the perimeter. The fixed head boundary at the North was replaced by no flow boundary. The reason for eliminating fixed head boundary was that, as it is very close to the wellfield and may act as an unlimited source of water for the wellfield. The Eastern boundary also moved 12 km east of the Ngotwane River so that it coincides with the watershed divided and zone of enhanced faulting in the dolomite.

The aquifer was modelled using two zones: Dolomite and Lephala formations. Two sources of recharge were included in the model. Recharge from the rainfall was assumed to be 10mm/a, and recharge from the Ngotwane river was set to be 50mm/a. Both values were assumed to be constant throughout the simulation period. The Ngotwane River is considered as discharge zone and modelled using a drain package assuming approximate depth of the drain equal to 5m.The conductance value was calibrated during the steady state calibration.

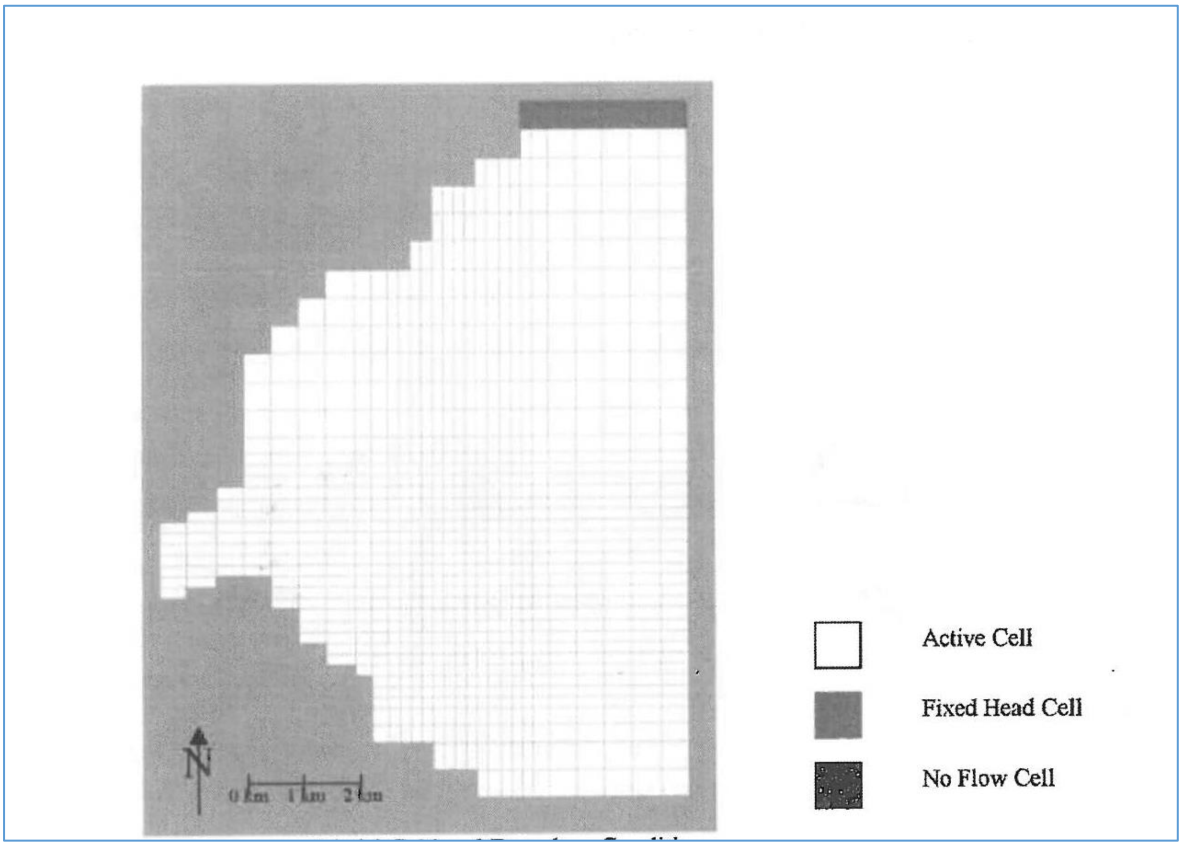

Figure 15: Model area, model grid and boundary condition used by WLPU Consultants and later modified by Geological Consulting Services (2000)

4. Methodology

4.1 Modelling Approach

Groundwater flow in karst aquifers is due to a combination of diffuse, fracture, and conduit flow (Quinn et al., 2006). Three approaches can be used to simulate flow in a karst aquifer. These approaches are: Equivalent porous medium approach (EPM) (Long et al., 1982), dual porosity model, and discrete fracture network models (Cook, 2003; Singhal and Gupta, 2010). The EPM concept assume that by averaging highly fractured and interconnected rocks over a large volume, the average flow resemble flow through a porous medium (Hassan et al., 2014).This approach is based on the assumption that the groundwater flow is not controlled by a small number of fractures, instead, the fractures are assumed to form a network of interconnected conduits, similar to pore space within a granular medium (Kaehler and Hsieh, 1994) This simplification is necessary because accounting individual fractures is a complex process. The dual porosity model approach assume that the porous medium consists of fracture network and matrix block which is less permeable, on the other hand the discrete fracture network approach assume that fracture network form path for groundwater flow and the matrix is impervious (Lee et al., 1999). EPM approach has been successfully applied in Karst aquifer (Ghasemizadeh et al., 2015; Larocque et al., 1999; Panagopoulos, 2012; Scanlon et al., 2003). Kuniansky (2016) tested the application of EPM approach based on MODFLOW 2005 and MODFLOW-CFP (MODFLOW-Conduit flow process model) in Florida Karstic aquifer and reported that, for monthly and seasonal time scales both models reproduced well the observed data. Based on this comparison, Kuniansky (2016) concluded that for monthly and seasonal time scale flow modelling in karst aquifer the increased effort required such as the collection of data on conduit location, and increased computational burden is not necessary.

In the present study we used the EPM approach as it is less data intensive and modelled the groundwater flow system using MODFLOW 2005 (Harbaugh, 2005) using the ModelMuse modelling environment/user interface (Winston, 2009). MODFLOW was selected for the present study for the following reasons:

1) It is available in the public domain and freeware that facilitate the use and further refinement of the model by the project partners,
2) It is one of the industry standard groundwater model developed by the USGS (United State Geological Survey),
3) It is well documented and adequately tested, and previously applied to model groundwater flow in karst aquifer.

MODFLOW is a three-dimensional finite-difference hydrogeological model for simulating and predicting groundwater conditions and groundwater/surface-water interactions. The transient 3D groundwater flow equation in MODFLOW is described using Equation 1. Equation 1, when combined with boundary and initial conditions, describes transient three-dimensional ground-water flow in a heterogeneous and anisotropic medium, provided that the principal axes of hydraulic conductivity are aligned with the coordinate directions (Harbaugh, 2005).

$$\frac{\partial}{\partial x}\left(Kx\frac{\partial h}{\partial x}\right)+\frac{\partial}{\partial y}\left(Ky\frac{\partial h}{\partial y}\right)+\frac{\partial}{\partial z}\left(Kz\frac{\partial h}{\partial z}\right)-W=Ss\frac{\partial h}{\partial t} \quad (1)$$

Where kx, ky, and Kz are hydraulic conductivity values in x, y, and z direction [L/T], h is hydraulic head [L], t is time [T], W is a source-sink term [1/T] representing recharge, pumping, evaporation, and Ss is specific storage [1/L), which when multiplied by the saturated thickness gives the confined aquifer storage coefficient, S[-], or the unconfined aquifer specific yield Sy [-]. In general kx, ky, kz and Ss are a function of space and W may be a function of space and time.

4.2 Conceptual Model

Conceptual model is the foundation of the model analysis (Bredehoeft, 2005). Generally, it includes information about the water budget that defines flow into and out of the aquifer system and boundary conditions. Conceptual model development involves an iterative process (Bredehoeft, 2005). The conceptual model may be redesigned as more data become available. During the conceptual model development, simplification is necessary to meet data limitations. However, such assumptions should be reasonable and consistent to the hydrogeology of the aquifer system. According to Bredehoeft (2005) the appropriateness of the conceptual model cannot be tested until a numerical model is built and comparisons made between observations and model simulation results. As such, the best approach is presumed to be concurrent development of conceptual and numerical models rather than waiting until a 'perfect 'conceptual model is formulated before starting to assemble the numerical model.

Developing a conceptual model consists of the following five steps (ASTM Standard D5447, 2010):

1. Specify the physical extent of the aquifer systems that impact or control the groundwater flow system, and analysis of groundwater flow directions
2. Determine appropriate physical and hydrological model boundaries
3. Define distribution and configuration of aquifer and confining unites. Of primary interest are the thickness, continuity, lithology and geological structures
4. Determine hydraulic properties for each aquifer unit. This include specifying hydraulic conductivity, storage characterises of the aquifer system such as specific yield and storage coefficients
5. Determine the source and sinks of water to the aquifer system, including their rate and temporal variability. Source and sinks include pumping wells, infiltration, evapotranspiration, drains, leakage across confining layers, and flow to or from surface water bodies.

Conceptual models of carbonate aquifers can be found (White, 1969, 2012). The conceptual model for the present study is developed based on a previous study referred to above (GCS, 2000). Figure 16 shows a piezometric map interpolated using Inverse Distance Weight (IDW) method using Feb 2006 water level data monitored in 12 observation wells. Although the general trend of the groundwater level contours is slope from South to North, local variation is considerable. Recharge in the area occur as diffuse recharge from rainfall and concentrated or focused recharge form the Ngotwane ephemeral stream. Leakage from sewer pipes and wastewater discharge also contribute to aquifer recharge but they are assumed less important. The outflow from the model domain is mainly through domestic pumping and groundwater evapotranspiration by deep-rooted plants. The GCS (2000) modeling study adopted two –dimensional (2D) Areal flow model. The 2D areal flow model assume that the aquifer is a 2D planar feature where groundwater flow is predominantly in the horizontal plane. However, as pointed out by Merz (2012) this assumption is usually valid for aquifers that have a horizontal extent that is much larger than the aquifer thickness and for aquifers that have high vertical hydraulic conductivity so that vertical head gradients within the aquifer are negligible. These two assumptions are difficult to achieve in the Ramotswa aquifer. Hence, in the present study we used 3D flow model. Unlike the previous study (e.g. GCS (2000), IoH (1989), we assumed the Ngotwane river as a recharge points instead of discharge points. This is based on the recent pumping test carried out by DWA (2006) and assessment by WUC (2014).

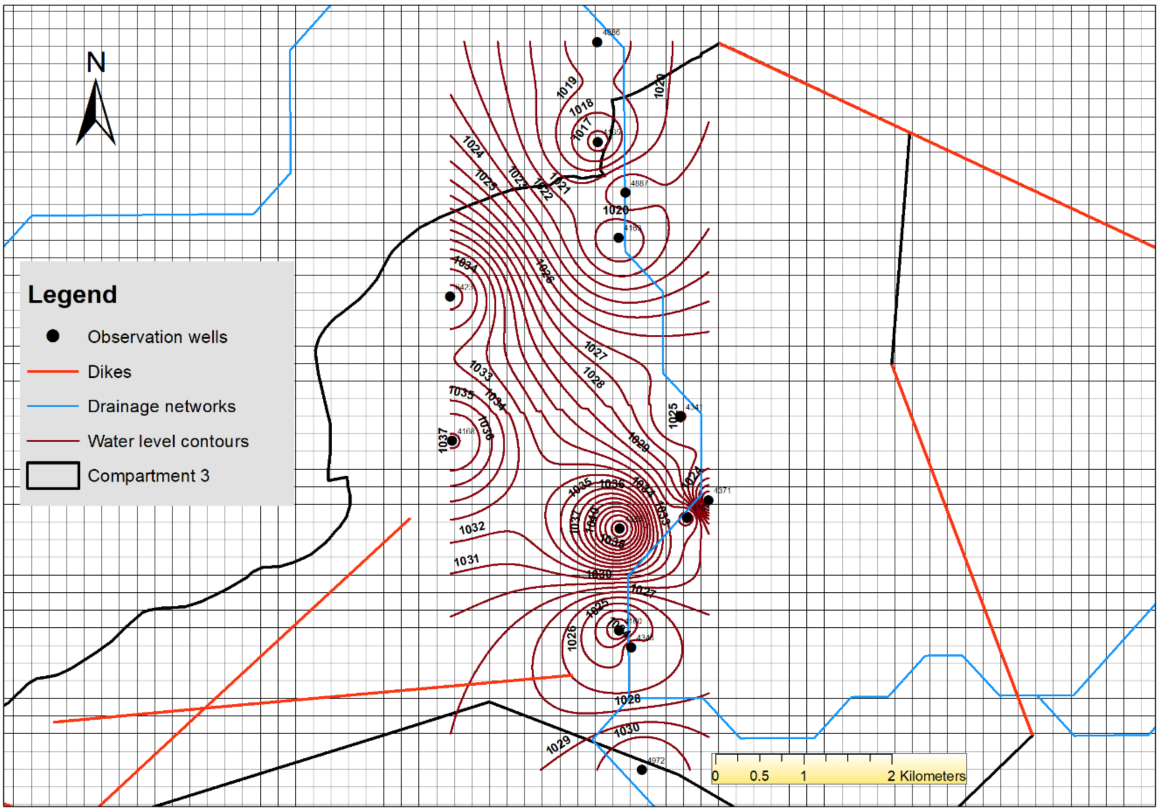

Figure 16: Groundwater level contours interpolated using IDW based on 12 observation wells water level data monitored on 17 FEB 20006

4.3 Model Construction

4.3.1 Spatial Model Discretization

Spatially, the model area was divided into a gird cell size of 90 x 90 m (Figure 17). Vertically the model is discretized into two layers representing the upper karstic zone and the deeper layer. The thickness of the aquifer is specified based on the deepest borehole depth drilled in the area and assumed 150m. This represents the level at which groundwater flow becomes negligible underlying the lower water-bearing unit. According to Williams (2008) in the arid and semi-arid region the highly weathered and karstified layer below the soil layer may reach 30 m while in other climatic region it is typically 3-10m. Hence, the thickness of the top layer of the aquifer is assumed 30 m while the deeper aquifer has a thickness of 120 m. The water-bearing units are unconfined to partly confined and are in hydraulic connection with each other. The upper layer was assumed to be unconfined while the second layer was treated as confined. The elevation of the top layer was defined based on Shuttle Radar Topography Mission (SRTM) 90 x 90 Digital Elevation Model (https://earthexplorer.usgs.gov/).

4.3.2 Temporal Model Discretization

Temporally, the 18 year's simulation period (2000-2017) was divided into 216 monthly stress periods. A stress period is an interval of time over which specified inputs are assumed to be constant. Monthly stress period was selected in order to allow simulation of seasonal change in groundwater use, recharge, evapotranspiration and seasonal aquifer storage and recovery options through MAR. Each monthly stress periods were further divided into bimonthly time steps, which are units of time for which water levels and flows are calculated throughout the model cells. The transient model was set up with a monthly stress period (216 stress periods) and weekly time step.

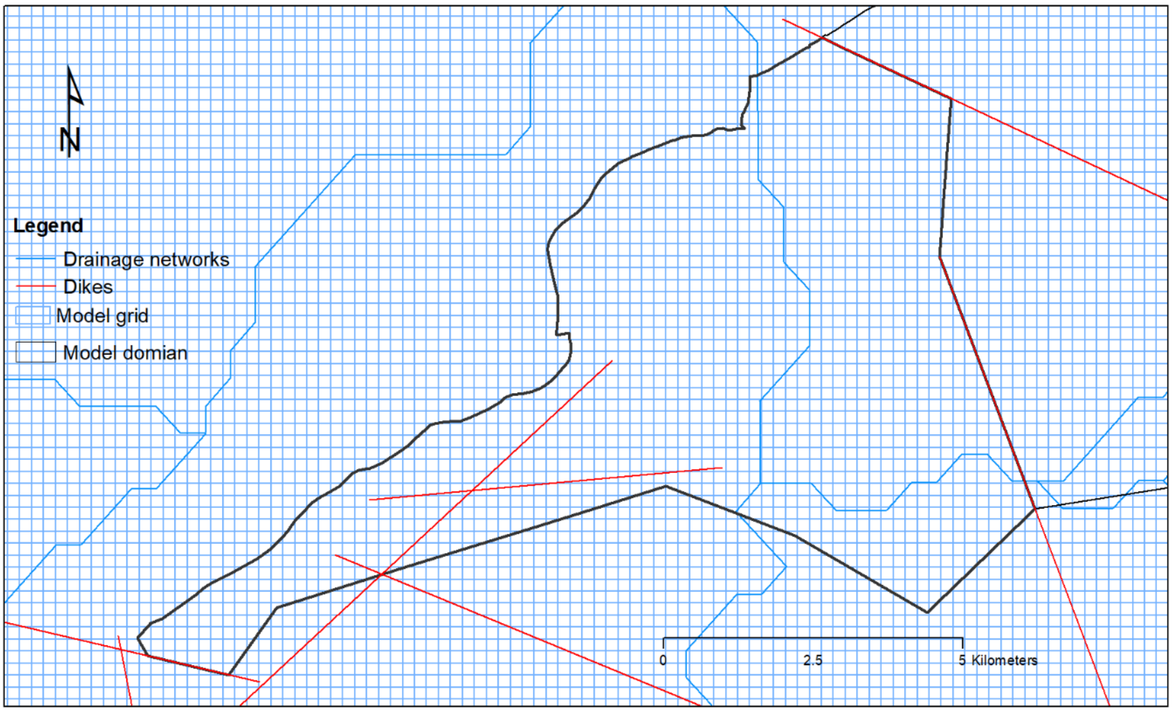

Figure 17: Model domain and grid (grid size of 200m by 200 m was used here for illustration purpose)

4.3.3 Boundary Conditions

The identification and assignment of appropriate boundary conditions is instrumental in selecting the proper conceptual model. The model boundaries should be specified in a consistent manner with natural hydrologic features. The model domain is bounded by no flow boundary in all direction except at the North where the Ngotwane River crosses the Black Reef Quartzite defined as General Head Boundary (GHB). GHB Package in MODFLOW is used to simulate flow into or out of a cell, from an external source in proportion to the difference between the head in the cell and the head assigned to the external source. The constant of proportionality is called the boundary conductance. Head at the external source, which is at the outside model boundary was assigned based on observed water level at observation well BH4165. The following boundary conditions were assigned along the perimeter of the model domain (Figure 18).

- **North:** This boundary was represented by a no flow boundary along the contact between the dolomite aquifer and the Black Reef Quartzite and GHB where the Ngotwane River crosses the Black Reef Quartzite.
- **North East:** no flow boundary along the dike

- **East**: Represented by a no flow boundary along the contact between the unconsolidated surface sediment formation with the Tsokwnae Quartzite of the Timball Hill formation
- **South** The southern boundaries were represented by no flow boundaries along the contact between the Lephala Formation and the Tsokwnae Quartzite
- **Southeast**: The Southeast boundary was specified as no flow along the dikes
- **Northwest**: The Northwest boundary was set as a no flow boundary along the Black Reef Quartzite.
- **Southwest**: A no flow boundary along the dikes

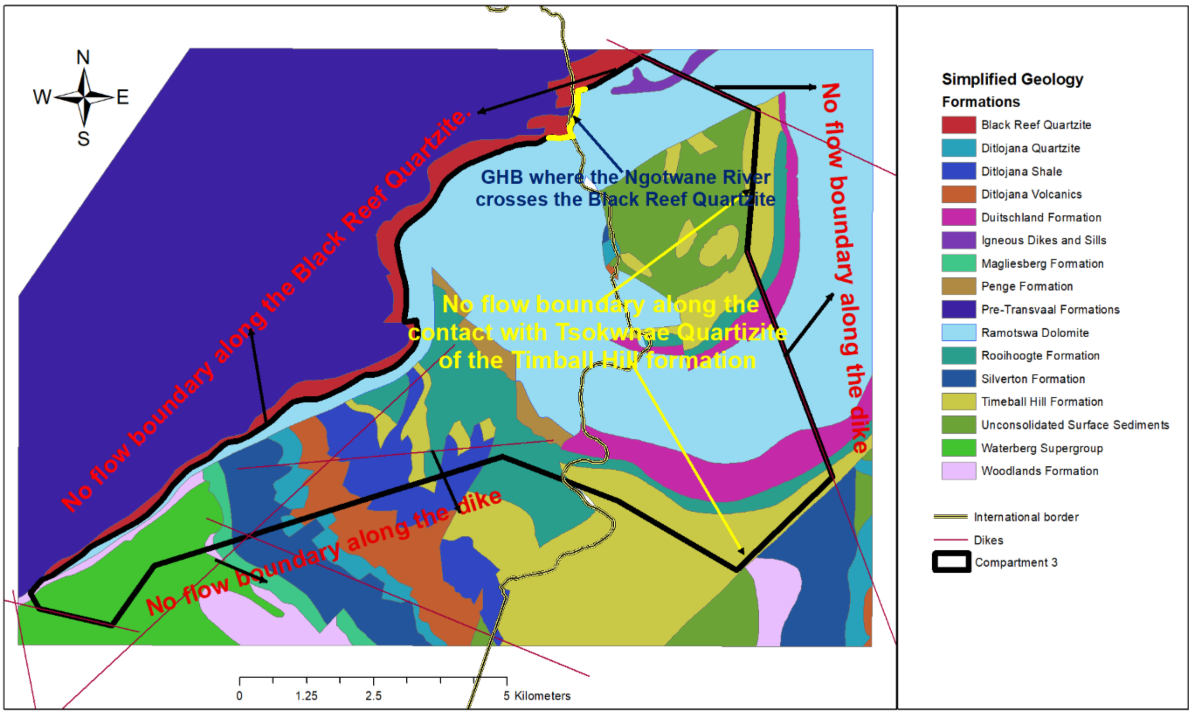

Figure 18: Boundary conditions

Dikes inside the model domain (Figure 17) are simulated using Horizontal Flow Barrier Package (HFB) of MODFLOW (Hsieh and Freckleton, 1993) assuming less permeable as demonstrated by Gebreyohannes et al. (2017) and Faunt et al. (2004). In the HFB Package, dikes are conceptualized as being located in the boundary between two adjacent finite difference cells in the same layer. HFB then adjust the horizontal hydraulic conductance computed by Layer property Flow package or Block centred Flow package to account for the barriers. The key assumption underlying HFB is that the width of the barrier is negligible as compared to the horizontal dimensions of the cells in grid so that the original conductance calculation is not affected (Harbaugh, 2005). The HFB package requires two parameters to be specified; the hydraulic conductivity and the thickness of the barrier. Barrier thickness is not explicitly considered in the package, but included implicitly in a hydraulic characteristics which is the ratio of the hydraulic conductivity to the thickness of the barrier. Hence, HFB allows the capability to use hydraulic characteristics as calibration parameters. The geometry of the dikes are assumed to be vertical, and extend from the surface of the ground to the bottom of the model layer.

4.3.4 Initial Conditions

Initial conditions are the distribution of water levels at every active cell. In a transient solution, the initial heads provide the reference elevations for the heads solution. The classical approach for defining initial head for the transient model simulation follows a two-step procedure, calibrating steady state model for pre-development time period and using the output of the steady state model as initial head for transient simulation. Using model generated initial head also ensures consistency between the initial head data and the model hydrologic inputs and parameters (Anderson and Woessner). The initial estimate should normally have no effect on the solution to the steady-state flow equation, but it may affect the number of iterations required to obtain an acceptable approximation of the solution. Initial condition for the transient model will be specified based on calibrated steady state model for the average water level conditions 2000-2012.

4.3.5 Groundwater Pumping

Groundwater pumping is a major component of the water budget of the study area. According to WUC (2014) the Ramotswa wellfield was commissioned in the 1980s for emergency water supply to the Gaborone city. The wellfield started its operation with four production boreholes. After the completion of the Gaborone dam in 1984, water supply from Ramotswa wellfield to the Gaborone city was disconnected and the Ramotswa wellfield was used to supply Ramotswa village at a rate of 1000 m^3/d (WUC, 2014). Later in 1989 the number of production boreholes were increased to ten and in the year the average daily abstraction rate also increased to 1696 m^3/d. In 1996, abstraction from the Ramotswa wellfield was totally stopped due to high nitrate level in the groundwater and re-operated in 2014. However, the re-opening date of the wellfield is not known. Figure 19 presents the abstraction data for the period 1988- 1995 obtained from GCS (2000). The total abstraction for the period 1988-1995 ranges from 0-8319 m^3/d. Abstraction data for the recent period from March 2015-Augst 2017 was obtained from Water Utility Corporation, Botswana (Figure 20). A total of 9 production wells were used to represent domestic water supply during the simulation period. The depth of these wells ranges from 102-120 m below the ground surface.

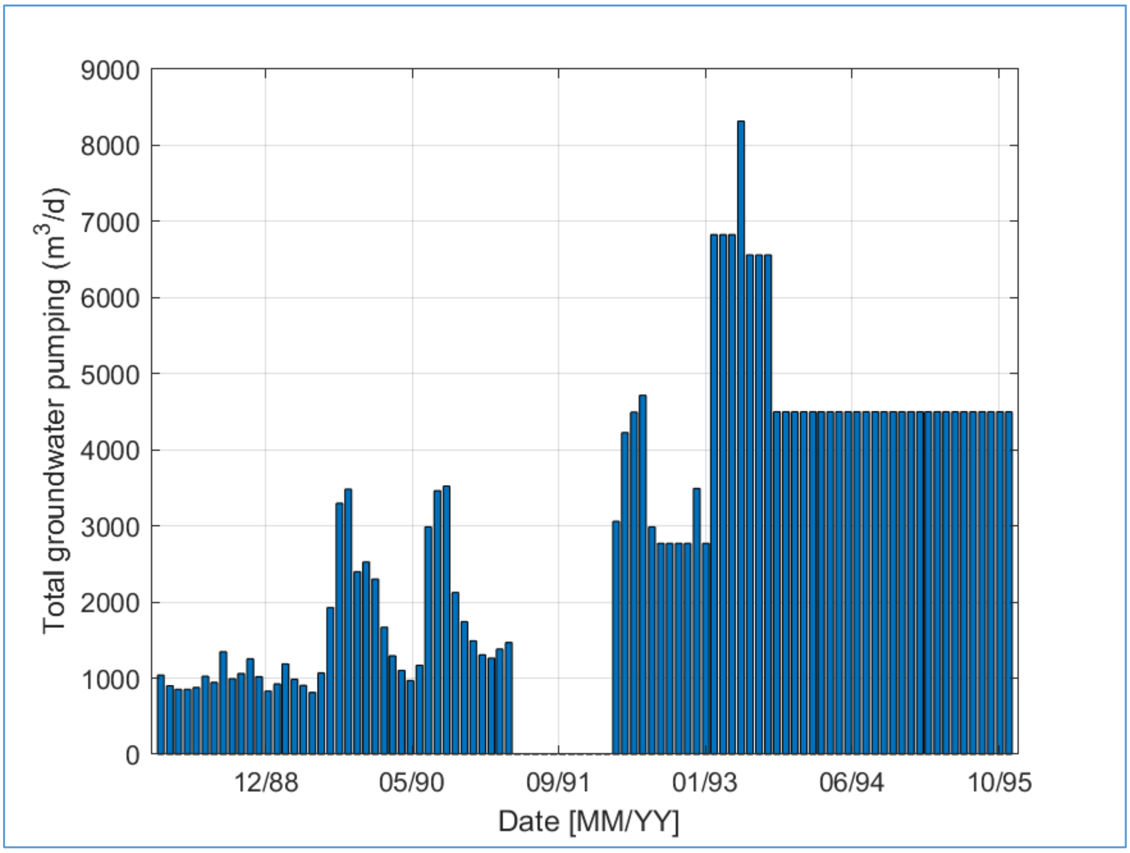

Figure 19: Monthly abstraction for the period of 1988-1995 from GCS (2000) modelling report

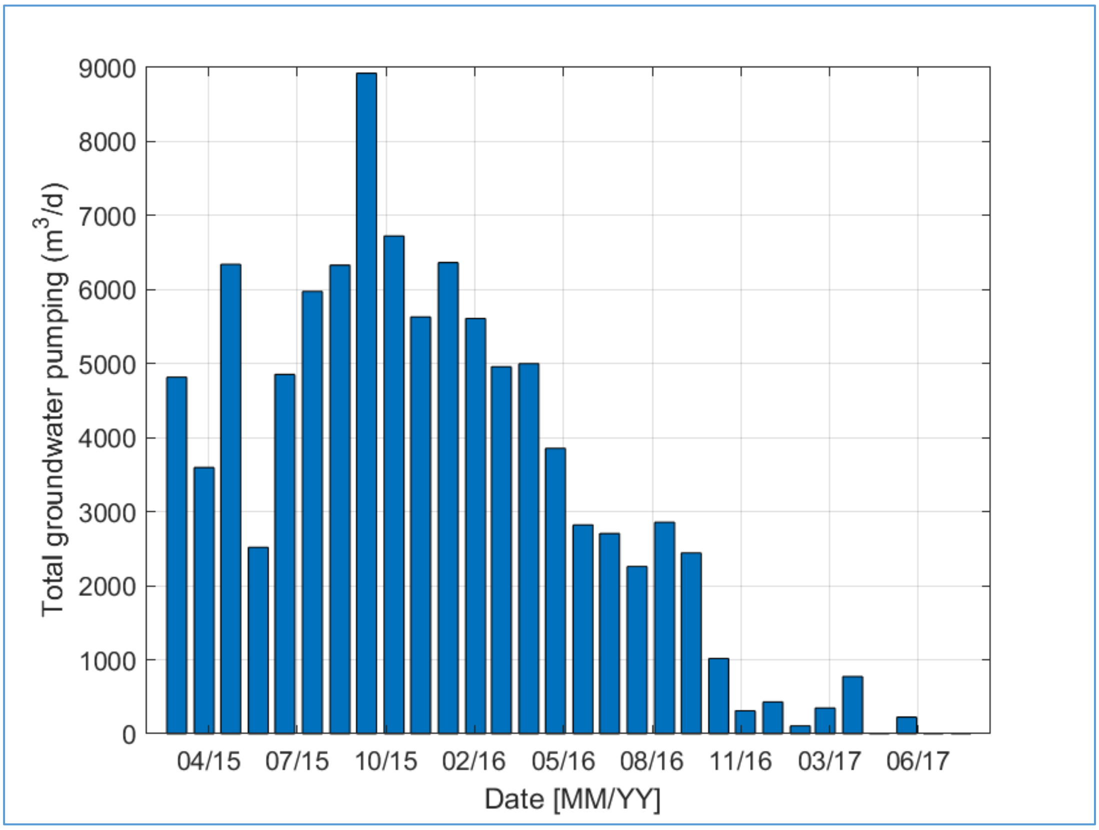

Figure 20: Total monthly abstraction for the period of March2015-Augst 2017 from WUC, master reading (daily abstraction values were calculated by dividing the monthly reading differences by the number of days in a moth)

4.3.6 Groundwater Evapotranspiration

Groundwater Evapotranspiration (GWET) occurs in areas where the depth to groundwater is shallow and plant roots penetrate to the zone of saturation. This mainly occurs along streams. The Evapotranspiration (ET) Package in MODLFOW simulates the effects of plant transpiration and direct evaporation removing water from the saturated ground-water regime. When the groundwater table is at or above the evapotranspiration surface, water is extracted at maximum evapotranspiration rate. When the groundwater table is below the extinction depth - no extraction takes place. When the groundwater table is between the evapotranspiration surface and the extinction depth the extraction varies linearly with the groundwater table elevation. To simulate GWET using the ET package, maximum evaporation rate, evapotranspiration surface and extinction depth need to be specified in the model. A maximum evapotranspiration rate is assumed to occur when the water table is at land surface. The maximum evapotranspiration rate was computed using the monthly S class pan evaporation data obtained from the Molatedi Dam station (A3E004 S class pan) and assumed pan coefficient (kp) of 0.85 using the relation ETo= $K_p E_{pan}$. Figure 21 shows the monthly S class evaporation and rainfall data. The ET extinction depth was determined based on Shah et al. (2007) study for a given soil type and land cover. The soil type in the study area is Sandy Clay loam. This was determined using soil global dataset (Soilgrid 250 m ISRC Worl Soil information). Three soil maps; sand, clay and silt representing the 2m soil depth were extracted from the Soilgrid250m. From these three maps soil texture classes were derived using the USDA soil texture class triangle. For Sandy Clay loam soil type and land cover type grass the ET extinction depth was found to be 3m.

4.3.7 Groundwater Recharge

The primary sources of recharge to the Ramotswa aquifer is via leakage from Ngotwane ephemeral stream, and more diffusely from rainfall. In drylands recharge from ephemeral streams is predominant (Goodrich et al., 2004). Recharge from precipitation occurs when rainfall infiltrates the land surface and percolates past the root zone to the water table; it generally is a small fraction of the total precipitation. The diffuse recharge processes are arguably, more widely understood but major uncertainty exist in areas where recharge from ephemeral streams dominates (Doll and Fiedler 2007). Recent review of recharge estimations methods in ephemeral streams is presented by Shanafield and Cook (2014). The seven methods reviewed in this paper include: controlled infiltration experiments, monitoring changes in water content, heat as a tracer of infiltration, reach length water balances, flood wave front tracking, groundwater mounding and groundwater dating. Since, there is no gauging station in the river it was not possible to use one of these method. In the present study recharge from precipitation and recharge from the River was simulated using the recharge package. Recharge rates were estimated as a fraction of precipitation. The monthly rainfall values were specified in the model as recharge flux and this was multiplied by two multiplying factor representing recharge zone 1 (recharge form rainfall) and recharge zone 2 (from the river). The multiplying factors were estimated during model calibration. Ebrahim et al. (2015) have used a similar approach for estimating recharge from a Wadi in Oman. Recharge is also estimated using the Water Table Fluctuation (WTF) method using water level observe in observation well close to the river. The WTF method is presented in the Annex 2.

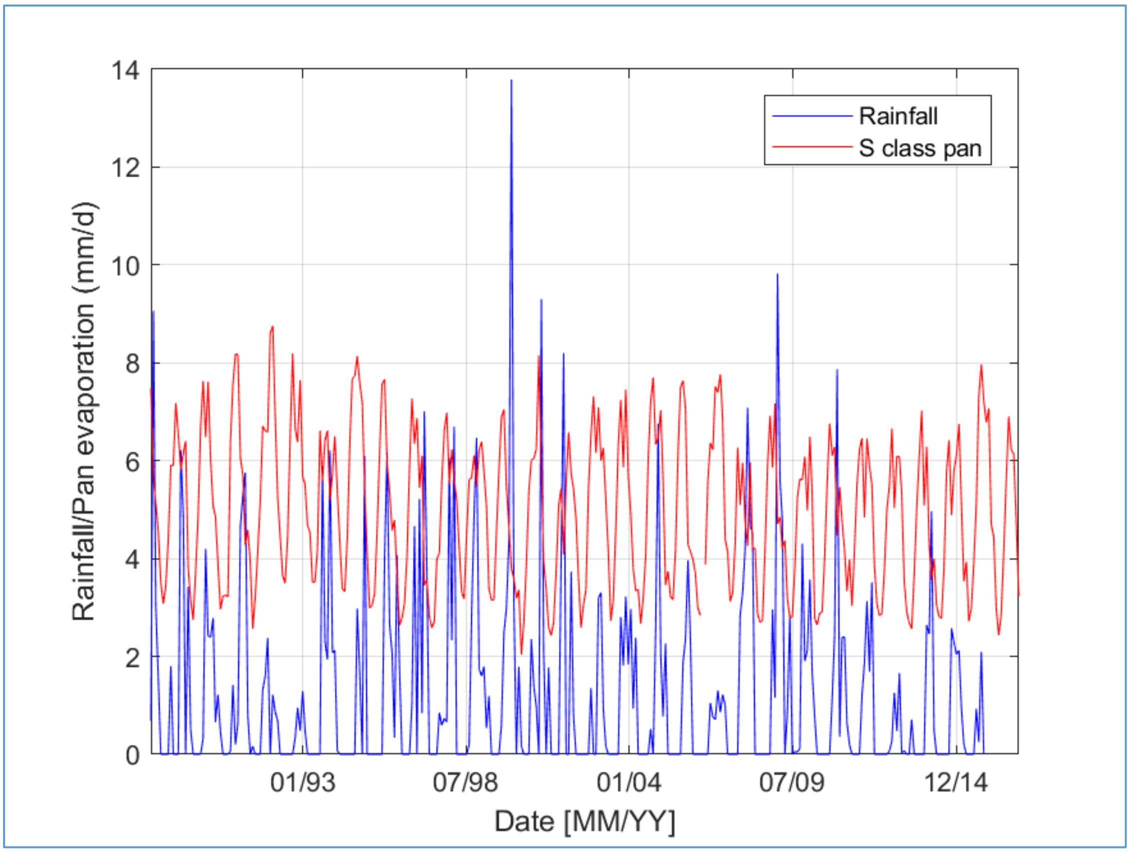

Figure 21: Mean monthly rainfall from Ramotswa station and mean monthly S-class pan evaporation from the Molatedi Dam

5. Aquifer Hydraulic Properties

Table 2 presents aquifer hydraulic properties (T=transmissivity, S=Storativity) determined using pumping test (DWA, 2006). According to Staudt (2003), the average transmissivity for the Ramotswa dolomite and Lephala formation respectively are 1170m and 492m^2/d. Staudt (2003) also reported storage coefficient of 5.7 x10^{-2} for the Ramoswa dolomite 8.7x10^{-4} for the Lephalal formation. Shevenell (1996) reported specific yield of 1 to 8X10^{-4} for conduit dominated karstic aquifer and 1 x10^{-3} for the fractured dominated karst aquifer.

Studies have shown that the hydraulic conductivity of fractured rocks can be anisotropic (Cook, 2003; Singhal and Gupta, 2010). Anisotropy is defined as having different properties in different direction. The anisotropy commonly attributed to the presence of preferred flow directions along sub-parallel set of fractures in the rock (Kaehler and Hsieh, 1994). As horizontal anisotropy increase, flow pattern becomes increasingly skewed in the direction of maximum hydraulic conductivity; however, model calculated flow patterns may be insensitive to changes in anisotropy ratios beyond some thresholds value that may be site dependent (Tucci, 1997). Vertical anisotropy ratios commonly are on the order of 10:1 (horizontal to vertical direction) or greater (Tucci, 1997). As the vertical anisotropy ratio increase, the groundwater flow path increasingly deviate from the vertical direction.

Table 2 Hydraulic conductivity calculated form DWA 2006 pumping test results (Transmissivity values are from DWA 20006 (Evaluated hydraulic parameters from Test pumping and step-drawdown Testing (C-factor))

BH/No	**Depth (m)**	**Water Strike (m)**	**Water Level (at time of Drilling) (m)**	**Estimated Yield (at time of Drilling) (m^3/hr**	**Transmissivity (m2/d)**	**Saturated depth (m)**	**K (T/b) (m/d)**
4336	102	24 – 32, 45	14.32	24, 30	800	87.68	9.124
4337	118	36, 81 - 82	5.51	12, 90	500	112.5	4.444
4340	120	14, 30, 42	7.12	15	5	112.9	0.0442
4349	120	11, 37, 112	4	24	550	116	4.741
4358	102	45	8.05	15	8	94	0.085
4373	120	40, 56, 75, 96	7.09	90	14	113	0.0124
4422	120	70 - 72	4	50 - 60	12	116	0.013
4423	120	24 - 44	6	2	210	114	1.842
4400	102	47	6.1	150	500	96	5.208

Transmissivity values are from DWA 20006 (Evaluated hydraulic parameters from Test pumping and step-drawdown Testing (C-factor)

6. Model Calibration period

The initial idea was to calibrate the model for the old period (1988-1999) using observation water level data from BH287 and BH63 and recent water level data (2000-2012). However, steady state model calibration based on two point average water level using the old data was not possible as the number of estimated parameters are greater than the number of observation. Hence, we decided to use the period 2000-2012 as a calibration period and 2015-2017 as a validation period. Table 3 presents the abstraction and water level data availability for the period 1988-2017.

Table 3: Abstraction and water level data availability

Year	**Abstraction**	**Water level**
1988-1995	Monthly abstraction data for 10 production boreholes is obtained from GCS (2000) modelling report	Available from two observation boreholes (BH63 and BH287)
1996-1998	Ramotswa wellfield was closed in 1996 due to high nitrate levels hence no pumping during this period	No groundwater level data
1999-2012	No abstraction (wellfield closed)	Water level data from 21 monitoring wells is available. Most of the observation wells are located along the Nogotwane River
2012-2014	No abstraction	Water level monitoring cessed for a while due to reform in the water sector (email communication, Alfred Petors).

2014-2015	Abstraction re-started in 2014 and data available from March 2015	No
2015-2017	Yes	Yes

7. Steady State Model Calibration

Steady state flow model was calibrated using time averaged observed heads from 12 monitoring wells measured between 2000-2012. This period was selected because there was no pumping during this period and hence groundwater levels are assumed to represent equilibrium conditions. A total of 4 parameters were subject to calibration. The adjustable parameters were horizontal and vertical hydraulic conductivity, conductance of general head boundary, recharge from rainfall and stream bed. For each parameter a reasonable physical prior range of values was specified based on past studies (DWA, 2006; GCS, 2000; IoH, 1986; Selaolo, 1985; Staudt, 2003; WUC, 2014). Four zones were defined to account the spatial variation in hydraulic conductivity representing: 1) area around the Ngotwane river, which is highly karstified due to shallow water level. This zone was specified to be approximately 200 m width, which is 100 m wide on both side of the river. 2) Zone represented by Ramotswa dolomite. No distinction in the dolomite lithological classes are made. 3) Zone representing areas covered by unconsolidated sediments. 4) Zone representing Lephala and all other geological formations. Initial horizontal hydraulic conductivity values were determined by dividing the transmissivity values obtained by pumping test (DWA, 2006) by approximate values of saturated thickness. The initial value of the hydraulic conductivity of the first layer was assumed to half of the mean horizontal hydraulic conductivity determined from pumping test, while for second layer initial value was set equal to the mean value. This is because even if the upper layer is highly karstified compared to the lower layer the solution cavities are reported to be filled by mud or wad while the second layer cavities are open and generally do not contain mud or wad fillings (GCS, 2000). The vertical hydraulic conductivity values at each layer were assumed to be one-tenth of their respective hydraulic conductivity values. Horizontal anisotropy is assumed to be one. Initial parameter values and parameter ranges used for steady state model calibration are presented in Table 4. Average water level at BH4165, which is 1014.516 was used to define external head at GHB and conductance value was estimated during model calibration.

During model calibration, parameter values were adjusted until the simulated water level matched the observed values. Both manual and automatic calibrations were used. The automatic calibration was performed using the automated parameter estimation code PEST (Doherty 2000). PEST is a model independent non-linear parameter estimator. PEST automatically minimizes the sum of square errors using the Gauss-Marquardt-Levenberg optimization algorithm in a weighted least-square sense using the measured and simulated values (Equation 2). Since its inception in the mid-1900s, PEST become the industry standard in the calibration of all kinds of environmental modelling problems (Doherty and Johnston, 2003). In the course of parameter estimation PEST required to run the model many times as part of the process of calculating the Jacobian matrix (i.e. the matrix of derivatives of observation with respect to parameter increment). During the optimization, observations were assumed to be of equal weight (equal importance in determining the optimization outcome).

$$\phi_k = \sum_{k=1}^{N_{pbs}} (w_k h_m{}^k - w_k h_s{}^k)^2 \ldots\ldots\ldots\ldots\ldots\ldots (2)$$

Where ϕ is the objective function value, w_k is the weight applied to the difference between the measured (h^k_m) and simulated (h^k_s) parameter of the same type k, and N_{pbs} is the total number of measured parameter values of the same type.

Table 4: Parameters initial values and ranges for steady state model calibration (though parameter initial value and ranges are provided here manual calibration was preferred and used)

Parameter	Initial value	Lower bound	Upper bound	Calibrated value
Hydraulic conductivity (m/d)				
Ngotwane river zone layer 1,	1.4	0.01	9.0	1
Ngotwane river zone, layer 2,	2.8	0.01	9.0	2
Ramotswa dolomite zone layer 1	0.1	0.01	9.0	0.1
Ramotswa dolomite zone layer 2	0.1	0.01	9.0	0.5
Unconsolidated sediments zone layer 1	0.1	0.01	9.0	0.5
Unconsolidated sediments zone layer 2	0.1	0.01	9.0	0.5
Lephala and all other formations zone layer 1	0.05	0.001	2	0.01
Lephala and all other formations zone layer 2	0.05	0.001	2	0.05
Recharge (m/d)				
Ngotwane river zone approximately 200 m wide				1.374E-4
Recharge from rainfall				7.74E-5
GHB conductance				100

8. Results

8.1 Steady State Model Calibration Results

Table 5 present the observed and simulated water levels and residuals. Figure 22 shows the scatter plot of observed and simulated water levels. The obtained pattern of simulated water level contours is presented in Figure 23. In general, good correlations between the observed and simulated groundwater levels were obtained (Cor=0.79). The mean absolute error between the observed and simulated groundwater level values was 5.37 m. The simulated water level contours are also consistence with the overall regional groundwater flow direction. As indicated in Figure 22 the model simulation very much underestimate water level at observation well 4155. This is shown in big blue circle in Figure 23. Since, the model is based on the EPM approach local variation in transmissivity, fracture and conduits flows are not simulated. This could be the reason for the big difference in observed and simulated water level at this observation well. The correlation coefficient between simulated and observed water level significantly improve when we are excluding observation well 4155 (Cor=0.86). The mean absolute error also reduced to 4.27 m.

Table 5: Observation wells locations used for steady state model calibration, observed average water level, simulated water level and residual

BH	X	Y	Observed water level (amsl)	Simulated water level (amsl)	Residual (m)
4155	386488.9	7247355	1042.65	1028.48	14.17
4160	386482.7	7246197	1021.85	1028.90	-7.04
4163	386481.2	7250657	1017.89	1021.52	-3.63
4168	384588.6	7248352	1039.62	1032.92	6.70
4341	387176.2	7248627	1023.91	1023.48	0.43
4887	386556.4	7251168	1020.14	1018.82	1.32
Z6423	384569.7	7249989	1036.06	1030.72	5.34
4371	387496.2	7247678	1020.65	1024.89	-4.24
4401	386179.6	7245604	1036.86	1031.40	5.46

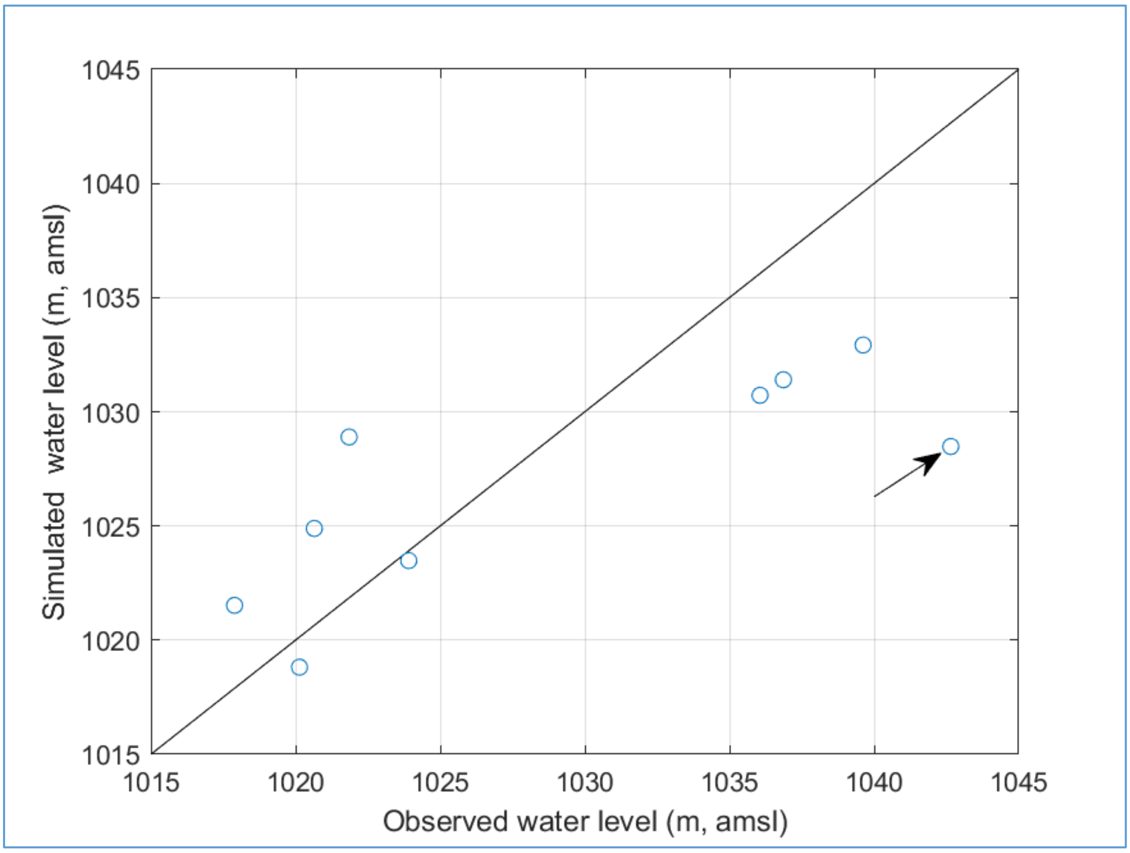

Figure 22: Observed vs simulated water level scatter plot, steady state model calibration. The arrow in the figure shows where the model simulated water level very much underestimate the observed water level in BH4155

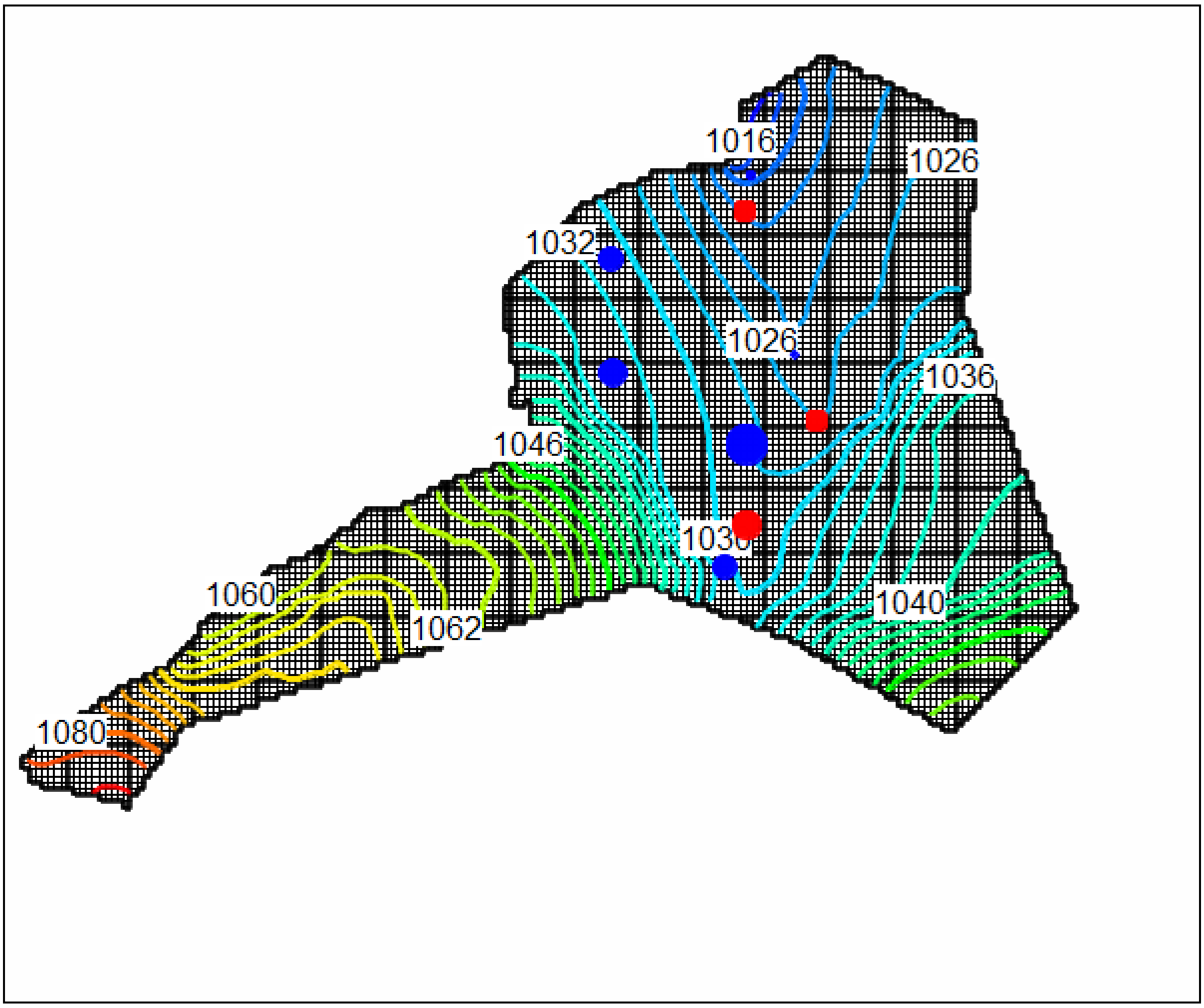

Figure 23: Steady state head distribution form the second model layer (simulated head lower than observed is shown in blue dot, while the red dot show areas where simulated head is higher than the observed value. The size of the dot represent the residual difference)

8.2 Water Budget Analysis

Groundwater budget analysis for the entire model domain was carried out using ZONEBUDGET (Harbaugh, 2009). The water budget equation in its simplest form described in Anderson et al. (2015) as: Inflow = Outflow +/- Δ in Storage

When outflow is not balanced by inflow change in storage occurs, resulting either increase or decrease in groundwater storage and accompanying change in groundwater level. For steady state model the change in storage is assumed to be zero. The difference in inflow and outflow during the model simulation is recommended not to be larger than 1%. Difference of more than 1% in the mass balance indicate possible numerical problem and may invalidate simulation results (ASTM Standard D5447, 2010). The water budget of the entire model domain from the calibrated steady state model is presented in Figure 24. Recharge is the main inflow to the model domain. Recharge occurs due to infiltration of surface water through stream channels and infiltration of precipitation through the unsaturated zone. No distinction is made between these recharge in the water budget, rather the sum is presented. The main outflow includes GWET, domestic pumping and groundwater that is leaving through the model domain through the GHB. During the calibration period domestic pumping do not occur. Hence, at dynamic equilibrium the total recharge should be equal to the sum of GWET and outflow through the GHB. As shown in Figure 24, GWET accounts about 80% of the total outflow while groundwater outflow across the model boundary accounts the remaining 20%. The diffused recharge calculated by the steady state model was about 28 mm/a. This is in the range of recent study by Baqa et al. (2018), who reported recharge rate in the area ranging from 2.1 to 73mm/a using the chloride mass balance method. The focused recharge from the river is the most uncertain parameter. In the present study, we used initial recharge estimate from GCS (2000) and set this value about 50 mm/a. Due to its high correlation with hydraulic conductivity we did not opt to vary this parameter during model calibration. Therefore, the focused recharge from the river used in this study is similar to previous study by GCS (2000).

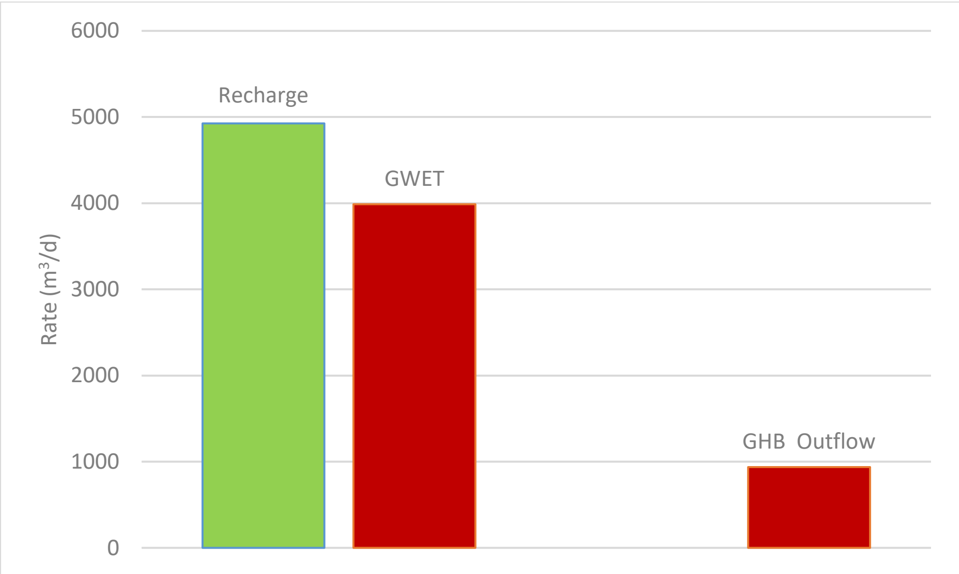

Figure 24: Water budget of the entire model domain obtained from steady state model

9. Discussion and Limitations

In view of the need for groundwater development in compartment 3, on Botswana side, in the past more than three hydrogeological models were developed to assess groundwater potential of the aquifer. The present study builds on these studies. The main purposes of the modelling work are: 1) to investigate the use and movement of groundwater, recharge, discharge and storage process 2) to assess the feasibility of MAR through model scenario analysis. Toward this end, a 3D steady state hydrogeological model was developed for compartment 3, area encompassing the Ramotswa village with catchment area of approximately 61 km^2. The aquifer was modelled using MODFLOW2005 assuming Equivalent Porous Media for flow through the karstic aquifer. In Equivalent Porous Media approach individual fractures or conduits cannot be adequately represented, rather the spatially averaged system properties are simulated (Teutsch and Sauter, 1991).

The steady state model was calibrated against average water level conditions for the period 2000-2012. In general, there is good correlation between observe and simulated water level data. The simulated water level contours are also consistent with the overall regional groundwater flow direction. Given several significant simplifying assumptions, the obtained gradient of groundwater levels (from South to North), as confirmed by the calibration results give some confidence that the developed model is overall representative of the groundwater flow in the aquifer. Further, improvements of the model is certainly necessary. Results of the calibrated steady sate model water budget shows that approximately 80% of the recharge entering the aquifer exits as GWET and only 20% of the recharge is leaving through the Northern boundary.

Independent recharge estimation was also carried out using WTF method. WTF method was applied at two observation wells, close to the river and far from the river (see annex 1). The notion is that water level dynamics for observation well located far from the river is due to diffused recharge from the river while for well close to the river is both due to diffused recharge and focused recharge from river leakage. In general, there is good correlation between the observed and simulated water level estimated using the WTF for observation well far from the river. The obtained specific yield is within the range of previous study estimate. Recharge was estimated assuming linear function of rainfall and this was found to be 20% while assuming zero for the y- intercept for the linear equation. This value is too high to consider it reasonable however, given the karstic nature of the aquifer recharge through sinkholes, this is not too high to assume realistic value. WTF method applied for observation well close to the river did not produced a good result. The correlation between observed and simulated water level is low. However, the recharge value obtained from this method (recharge 90% of rainfall) is comparable with the recharge obtained with 2D transient profile model, discussed next.

In order to support hydraulic parameters and focused recharge estimation along the Ngotwane River, a transient 2D profile model was developed and calibrated against four observation wells located along the river (see annex 2). The calibration was performed using observed water level data for the period 2000-2012. This period was selected as there are no pumping well that induce later flow and violate the assumption of no lateral flow. The 2D profile model calibration reproduced very well the observed water level dynamics in the four observation well. Correlation between observed and simulated water level as high as 0.90 was obtained in observation borehole BH4168. Combined recharge form rainfall and the river was estimated using monthly rainfall values and multiplier constant. The multiplier constant was estimated during model calibration and this was estimated to be 0.833. Given the climatic conditions, recharge of this amount may not be reasonable even if we assume the dominant portion is derived from river leakage. We rather attribute this to parameter estimation problems, due to the high correlation between recharge and hydraulic conductivity. One of these parameters needs to be fixed to get reliable parameter values. According to Hill and Tiedeman (2006) the two common model errors are due to: 1) the model does not match the observation and/or the weighted residuals are not randomly distributed in time, in space, and or relative to simulated values and 2) the optimized parameter values are unrealistic and confidence intervals on the optimized values do not include reasonable values.

The following assumptions were made in the present 3D hydrogeological model:

1) The aquifer is represented by two layer model due to lack of data on lithology and degree of karstification with depth,
2) Groundwater flow through the karstic aquifer is assumed to behave similar to flow through Equivalent Porous Media and flow through individual fractures and conduits cannot be adequately represented.
3) Dikes compartmentalizing the groundwater flow are assumed to be impermeable, hence no flow occurs across this dikes even in the weathered top part.

4) Lateral boundaries at the North and some part of the Eastern side are not physical boundaries but, where chosen as no flow boundaries based on their low permeability. Hence, flow across these boundaries are assumed to be negligible.
5) The Ngotwane River is assumed to be recharge zone
6) The dolomite is dipping to the South but this was not accounted in the model. Model layers were assumed to be vertical.
7) Evapotranspiration from the saturated zone is assumed to be a linear function of groundwater depth below the land surface. It is maximum at land surface and decreases linearly to zero at 3 m below the land surface.
8) No flow conditions exists at the bottom boundary of the model layer.
9) In hydraulic conductivity zoning no distinction in the dolomite lithological classes are made, however, dolomite with chert-rich (example, Monte Christo and Eccles formation) are productive than chert-poor dolomite (example, Okatree, Littleton, Frisco). This assumption was made to reduce the number of estimated parameters.

10. Conclusions and Next steps

The main objective of hydrogeological modelling under steady state condition is for estimating time independent aquifer parameters and for providing good initial condition for transient model calibration. Hence, the good correlation between observed and simulated water level data, and the agreement between simulated groundwater level contours and the overall regional groundwater flow direction support the correctness of the conceptual hydrogeological model, the representativeness of aquifer parameters and reasonableness of the steady state model for the transient model calibration and for subsequent use in MAR assessment.

Spatially average diffused recharge from steady state model estimated via model calibration is about 28mm/a, suggesting a high groundwater potential. Based on the calibrated steady state model water budget analysis approximately 80% of the recharge entering the aquifer leave as GWET. However, even if GWET account the larger water balance component it was not accounted in any of the previous modelling studies. Hence, recharge values used in the previous modelling work has to be viewed as net recharge values. GWET occur along the river in low land area where the groundwater is close to the ground surface. About 20% of the recharge is leaving through the Northern boundary, where the Ngotwane River crosses the Black Reef formation.

The present model represents only the first step towards a 3D transient model development and for comprehensive effort for MAR feasibility assessment. The current model provides an initial condition or reference level for transient model calibration. It is the transient model that would enable us to get a better understanding of the storage process in the aquifer and allow us to do MAR scenario analysis. The 3D transient model enable us to determine the storage capacity of the aquifer and allow us in making choice as to where the MAR system to be located for maximizing recharge and recovery.

Next steps are: 1) transient 3D model calibration and validation, 2) sensitivity analysis of model parameters, 3) MAR scenario analysis using a calibrated and validated 3D transient hydrogeological model. The forthcoming feasibility assessment, to be undertaken through the transient hydrogeological model scenario analysis, will help to: determine the volume of water to be added to storage in the aquifer, identify suitable sites for MAR application and to optimally site the MAR system, understand how the aquifer system react for additional recharge, and to determine the length of time the recharged water remain in storage in a particular area. In addition to MAR feasibility assessment using hydrogeological model scenario analysis forthcoming feasibility assessment include two more objectives: i) assessing water resources availability that can be used as a recharge source and its quality, ii) evaluating the geochemical implication of mixing recharge water used as a source and native groundwater in the aquifer.

References

Anderson, M., and Woessner, W., Applied Groundwater Modeling: Simulation of Flow and Advective Transport., 1992, Academic Press, San Diego.

Anderson, M. P., Woessner, W. W., and Hunt, R. J., 2015, Applied groundwater modeling: simulation of flow and advective transport, Academic press.

ASTM Standard D5447, 2010, Standard Guide for Application of a Groundwater Flow Model to a Site-Specific Problem.

Baqa, S. S., Magombeyi, M. S., Villholth, K. G., Abiye, T., and Butler, M., 2018, Groundwater in the Ramotswa Dolomitic Aquifer: From Quanttfying Recahrge Rates to Understading Flow Pathways, manusctip in preparation.

Beven, K., 2006, A manifesto for the equifinality thesis: Journal of hydrology, v. 320, no. 1-2, p. 18-36.

Bredehoeft, J., 2005, The conceptualization model problem—surprise: Hydrogeology journal, v. 13, no. 1, p. 37-46.

Bromley, J., Mannström, B., Nisca, D., and Jamtlid, A., 1994, Airborne Geophysics: Application to a Ground-Water Study in Botswana: Ground Water, v. 32, no. 1, p. 79-90.

Catuneanu, O., and Eriksson, P. G., 1999, The sequence stratigraphic concept and the Precambrian rock record: an example from the 2.7–2.1 Ga Transvaal Supergroup, Kaapvaal craton: Precambrian Research, v. 97, no. 3-4, p. 215-251.

Cook, P. G., 2003, A guide to regional groundwater flow in fractured rock aquifers, Citeseer.

Cuthbert, M., 2010, An improved time series approach for estimating groundwater recharge from groundwater level fluctuations: Water Resources Research, v. 46, no. 9.

Delleur, J. W., 2006, The handbook of groundwater engineering, CRC press.

Doherty, J., and Johnston, J. M., 2003, Methodologies for calibration and predictive analysis of a watershed model1, Wiley Online Library.

DWA, 2006, Ramotswa Test Pumping of Production Boreholes, Groundwater Division – Department of Water Affairs.

DWAF, 2006, Vaal River System: Large Bulk Water Supply Reconciliation Strategy: Groundwater Assessment Dolomite Aquifers, Department of Water Affairs and Forestry, South Africa, December 2006. .

Ebrahim, G. Y., Jonoski, A., Al-Maktoumi, A., Ahmed, M., and Mynett, A., 2015, Simulation-optimization approach for evaluating the feasibility of managed aquifer recharge in the Samail lower catchment, Oman: Journal of Water Resources Planning and Management, v. 142, no. 2, p. 05015007.

Eriksson, P., and Altermann, W., 1998, An overview of the geology of the Transvaal Supergroup dolomites (South Africa): Environmental geology, v. 36, no. 1-2, p. 179-188.

Eriksson, P., and Reczko, B., 1995, The sedimentary and tectonic setting of the Transvaal Supergroup floor rocks to the Bushveld Complex: Journal of African Earth Sciences, v. 21, no. 4, p. 487-504.

Faunt, C. C., Blainey, J. B., Hill, M. C., D'Agnese, F. A., and O'Brien, G. M., 2004, Transient numerical model: Death Valley regional ground-water flow system, Nevada and California—hydrologic framework and transient ground-water flow model. US Geological Survey Scientific Investigations Report, v. 5205, p. 257-352.

GCS, 2000, Groundwater Monitoring, Review of Monitoring Performed by DWA and DGS, Assessment of Water Resources and Suggestion for Improvements, Phase 1 of the National Groundwater Information System TB10/3/4/96-97, Volume 19, Ramotswa Wellfield Final Report by Geotechnical Consulting Services

Gebreyohannes, T., De Smedt, F., Walraevens, K., Gebresilassie, S., Hussien, A., Hagos, M., Amare, K., Deckers, J., and Gebrehiwot, K., 2017, Regional groundwater flow modeling of the Geba Basin, Northern Ethiopia: Hydrogeology Journal, p. 1-17.

Ghasemizadeh, R., Yu, X., Butscher, C., Hellweger, F., Padilla, I., and Alshawabkeh, A., 2015, Equivalent porous media (EPM) simulation of groundwater hydraulics and contaminant transport in karst aquifers: PloS one, v. 10, no. 9, p. e0138954.

Goodrich, D. C., Williams, D. G., Unkrich, C. L., Hogan, J. F., Scott, R. L., Hultine, K. R., Pool, D., Goes, A. L., and Miller, S., 2004, Comparison of methods to estimate ephemeral channel recharge, Walnut Gulch, San Pedro River basin, Arizona: Groundwater recharge in a desert environment: the southwestern United States, p. 77-99.

Harbaugh, A., 2009, Zonebudget Version 3.01, A computer program for computing subregional water budgets for MODFLOW ground-water flow models: US Geological Survey Groundwater Software.

Harbaugh, A. W., 2005, MODFLOW-2005, the US Geological Survey modular ground-water model: the ground-water flow process, US Department of the Interior, US Geological Survey Reston.

Hassan, S. T., Lubczynski, M. W., Niswonger, R. G., and Su, Z., 2014, Surface–groundwater interactions in hard rocks in Sardon Catchment of western Spain: an integrated modeling approach: Journal of Hydrology, v. 517, p. 390-410.

Healy, R. W., and Cook, P. G., 2002, Using groundwater levels to estimate recharge: Hydrogeology Journal, v. 10, no. 1, p. 91-109.

Hill, M. C., and Tiedeman, C. R., 2006, Effective groundwater model calibration: with analysis of data, sensitivities, predictions, and uncertainty, John Wiley & Sons.

Hsieh, P. A., and Freckleton, J. R., 1993, Documentation of a computer program to simulate horizontal-flow barriers using the US Geological Survey's modular three-dimensional finite-difference ground-water flow model, US Department of the Interior, US Geological Survey.

IoH, 1986, Ramotswa Wellfield S.E. Botswana, Digital Model Study and Storage Estimates, The Institute of Hydrology, University of Blomfontein, South Africa

Kaehler, C. A., and Hsieh, P. A., 1994, Hydraulic properties of a fractured-rock aquifer, Lee Valley, San Diego County, California: US Geological Survey.

Kuniansky, E. L., 2016, Simulating groundwater flow in karst aquifers with distributed parameter models—Comparison of porous-equivalent media and hybrid flow approaches: US Geological Survey, 2328-0328.

Larocque, M., Banton, O., Ackerer, P., and Razack, M., 1999, Determining karst transmissivities with inverse modeling and an equivalent porous media: Ground water, v. 37, no. 6, p. 897-903.

Lee, J., Choi, S.-U., and Cho, W., 1999, A comparative study of dual-porosity model and discrete fracture network model: KSCE Journal of Civil Engineering, v. 3, no. 2, p. 171-180.

Long, J., Remer, J., Wilson, C., and Witherspoon, P., 1982, Porous media equivalents for networks of discontinuous fractures: Water Resources Research, v. 18, no. 3, p. 645-658.

Mansouri, B. E., and Mezouary, L. E., 2015, Enhancement of groundwater potential by aquifer artificial recharge techniques: an adaptation to climate change: Proceedings of the International Association of Hydrological Sciences, v. 366, p. 155-156.

Merz, S. K., 2012, Australian groundwater modelling guidelines: Waterlines Report Series, no. 82.

Meyer, R., 2014, Hydrogeology of Groundwater Region 10: The Karst Belt, Report to the Water Research Commission, WRC Report No. TT 553/14

Morel, E., and Wikramaratna, R., 1982, Numerical modelling of groundwater flow in regional aquifers dissected by dykes: Hydrological Sciences Journal, v. 27, no. 1, p. 63-77.

Panagopoulos, G., 2012, Application of MODFLOW for simulating groundwater flow in the Trifilia karst aquifer, Greece: Environmental Earth Sciences, v. 67, no. 7, p. 1877-1889.

Quinn, J. J., Tomasko, D., and Kuiper, J. A., 2006, Modeling complex flow in a karst aquifer: Sedimentary Geology, v. 184, no. 3-4, p. 343-351.

Scanlon, B. R., Mace, R. E., Barrett, M. E., and Smith, B., 2003, Can we simulate regional groundwater flow in a karst system using equivalent porous media models? Case study, Barton Springs Edwards aquifer, USA: Journal of hydrology, v. 276, no. 1, p. 137-158.

Selaolo, E. T., 1985, Evaluation of Aquifer Hydraulic Parameters in Ramotswa: S.E. Botswana , MSc Thesis in Hydrogeology, University College London

Shah, N., Nachabe, M., and Ross, M., 2007, Extinction depth and evapotranspiration from ground water under selected land covers: Groundwater, v. 45, no. 3, p. 329-338.

Shanafield, M., and Cook, P. G., 2014, Transmission losses, infiltration and groundwater recharge through ephemeral and intermittent streambeds: A review of applied methods: Journal of Hydrology, v. 511, p. 518-529.

Shevenell, L., 1996, Analysis of well hydrographs in a karst aquifer: estimates of specific yields and continuum transmissivities: Journal of hydrology, v. 174, no. 3-4, p. 331-355.

Singhal, B. B. S., and Gupta, R. P., 2010, Applied hydrogeology of fractured rocks, Springer Science & Business Media.

Staudt, M., 2003, Environmental Hydrogeology of Ramotswa South East District, Republic of Botswana

Teutsch, G., and Sauter, M., Groundwater modeling in karst terranes: Scale effects, data acquisition and field validation, *in* Proceedings Proc. Third Conf. Hydrogeology, Ecology, Monitoring, and Management of Ground Water in Karst Terranes, Nashville, TN1991, p. 17-35.

Tucci, P., 1997, Simulated effects of horizontal anisotropy on ground-water-flow paths and discharge to streams: Ground water protection alternatives and strategies in the USA.

White, W. B., 1969, Conceptual models for carbonate aquifers: Groundwater, v. 7, no. 3, p. 15-21.

-, 2012, Conceptual models for carbonate aquifers: Groundwater, v. 50, no. 2, p. 180-186.

Williams, P. W., 2008, The role of the epikarst in karst and cave hydrogeology: a review: International Journal of Speleology, v. 37, no. 1, p. 1.

Winston, R. B., 2009, ModelMuse: a graphical user interface for MODFLOW-2005 and PHAST, US Geological Survey.

Woolfenden, L. R., and Koczot, K. M., 2001, Numerical simulation of ground-water flow and assessment of the effects of artificial recharge in the Rialto-Colton Basin, San Bernardino County, California, US Department of the Interior, US Geological Survey.

WUC, 1989, Ramotswa Wellfiled Operation Procedures and Resource Managment Project Phase 1 Report, Republic of Botswana Water Utility Corporation.

-, 2014, Ramotswa Wellfield Monitoring Report, Water Utilities Corporation, Botswana.

Xanke, J., Jourde, H., Liesch, T., and Goldscheider, N., 2016, Numerical long-term assessment of managed aquifer recharge from a reservoir into a karst aquifer in Jordan: Journal of Hydrology, v. 540, p. 603-614.

XRI BLUE, 2016, Conceptual Model, Hydraulic Properties, and Model Dataset Support Progress Report for the Development of Subsurface Mapping for the Transboundary Ramotswa Aquifer, that is part of the Limpopo River Basin, using Airborne Geophysics,OCTOBER, 2016.

Annex1: Recharge Estimation using Water Table fluctuation Method

1.1 Theoretical Background

The Water Table Fluctuation Method (WTF) is based on the hypothesis that a rise in groundwater levels in the unconfined aquifers is due to recharge water arriving at the water table. Recharge is calculated using Equation 3 (Healy and Cook, 2002).

$$R = Sy * \frac{dh}{dt} = Sy\frac{\Delta h}{\Delta t} \qquad (3)$$

Where Sy is specific yield, Δh is the difference between the peak of the rise and low point of the extrapolated antecedent recession curve at the time of the peak and Δt is the time between those points.

Cuthbert (2010) more recently, modified the original equation of the WTF method to account for net groundwater drainage away from a given observation point. According to the author, the change in groundwater level in an aquifer through time is not only controlled by recharge but also by the net groundwater drainage away from a given observation point and expressed using Equation 4.

$$R = Sy\frac{dh}{dt} + \mathrm{D} \ldots\ldots\ldots\ldots\ldots\ldots\ldots (4)$$

Where R is recharge [LT^{-1}], Sy is the specific yield [-], and D is the net groundwater drainage away from observation point [LT^{-1}]

The advantage of this approach is its simplicity and its main limitations are difficulties of estimating specific yield and accounting for drainage term (Cuthbert, 2010). The author argue that for shallow water table conditions R may be much larger than D during a recharge event, hence the error in recharge estimation due to error in D may be negligible. Using discrete water level data for each time step Equation 4 can be written as Equation 5. After re-arranging the water level at time (t) can be calculated using Equation 6. Equation 6 is modified to account groundwater decline due to pumping and written as Equation 7.

$$R_t = Sy\frac{(h_t - h_{(t-\Delta t)})}{\Delta t} + D_t \ldots\ldots\ldots\ldots\ldots\ldots\ldots (5)$$

$$h_t = h_{(t-\Delta t)} + \left(\frac{R_t - D_t}{Sy}\right)\Delta t \qquad (6)$$

$$h_t = h_{(t-\Delta t)} + \left(\frac{R_t - D_t}{Sy} - S_t\right)\Delta t \qquad (7)$$

Where S_t is drawdown rate (LT^{-1}), h_t *and* $h_{(t-\Delta t)}$ are groundwater level (relative to the river level), and Δt is time step [T]

1.2 Estimated Recharge using the Water Table Fluctuation Method

The WTF method was used to estimate recharge at monitoring well BH4168, which is approximately 2.7 km from the river channel and for BH4371 approximately 0.12 km from the river. Recharge is assumed to be linear function of rainfall. As there is no pumping during this period the drawdown rate due to pumping was assumed to be zero. Three parameters namely, recharge fraction, Sy and Dt were used as a calibration parameter. However, it is very difficult, if not impossible to simultaneously estimate a unique parameter values for each. The result of the model calibration for BH4168 which is located far from the river is shown in Figure 25. During the calibration Sy of 0.03, recharge fraction of 0.2 and D of 0.1 provided a Nash-Sutcliff (NSE) of 0.51. The recharge values obtained ranges from 0-41.9 mm/d. The highest recharge corresponds highest rainfall of year 2000.

The simulated and observed water level values for observation well located close to the river, BH4371 is presented in Figure 26. As can be seen in the Figure 26, the simulated and observed water levels did not produced a very good fit (NSE = 0.09). The calibrated parameter values were 0.9 for fraction of recharge, 0.2 for Sy and 0.04 for D. The calibration results provided a very high value of recharge and specific yield. Even if due to river leakage following flooding events high recharge value is expected and due to high karstification along the river section high Sy is common, the optimized parameters seems very high to be acceptable. The model performance is also poor. Hence, recharge simulated using observation well close to the river was not considered reliable.

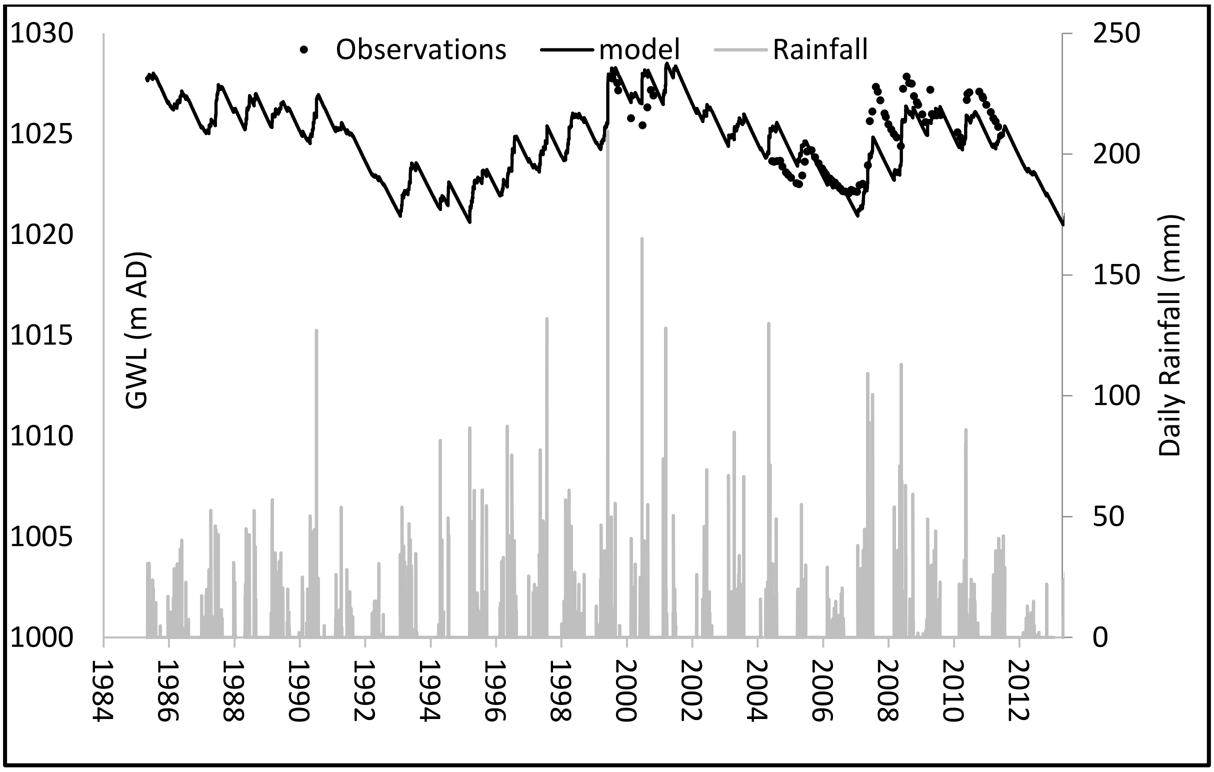

Figure 25: Observed and simulated groundwater levels using WTF for BH4168 far away from the river

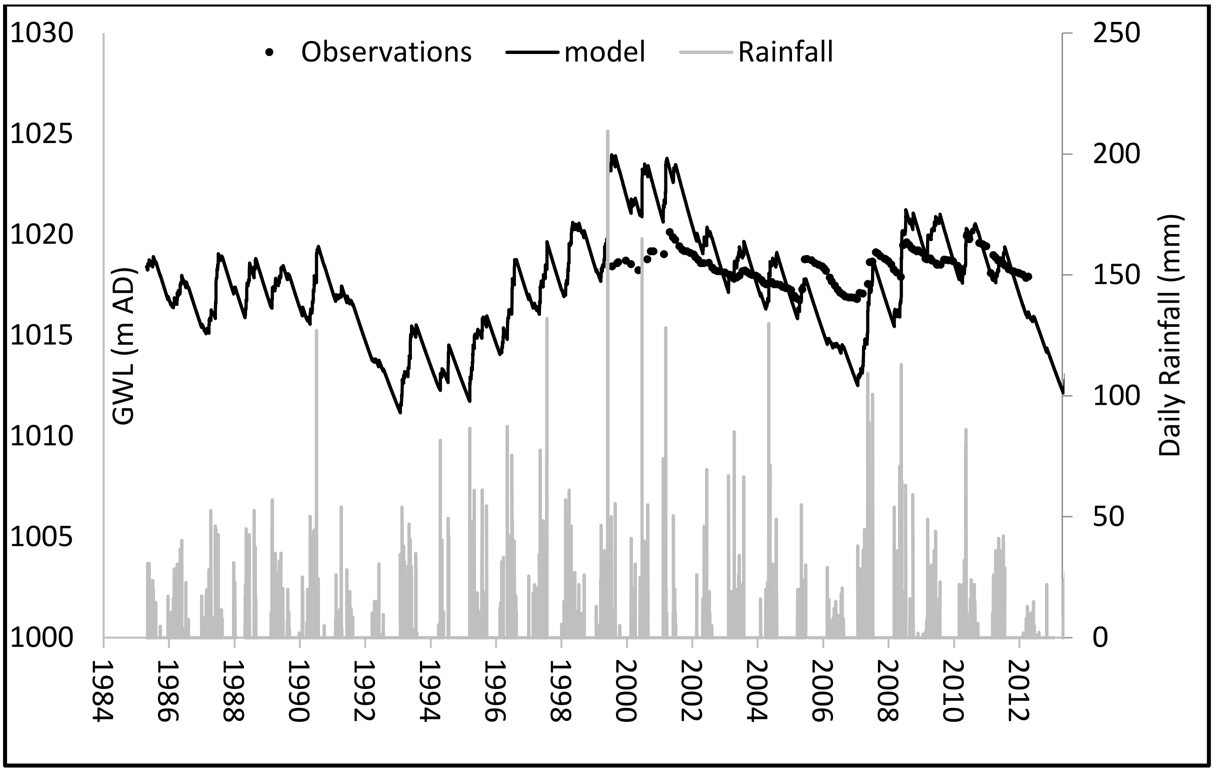

Figure 26: Observed and simulated groundwater levels using WTF for BH4371 close to the river

Annex 2: 2D profile model for Ramotswa Transboundary Aquifer

2.1 Background

Realistic representation of groundwater flow commonly require 3D models. However, due to data requirements for full 3D model such simulation may be infeasible in data scarce regions. In such cases, two-dimensional (2D) models provide an alternative solution. A 2D profile model represents 2D flow in a vertical slice of a groundwater flow system. It assumes that there is not flow through the sides of the profile and therefore cannot simulate radial flow to well (Anderson et al., 2015). In a 2D profile modelling the line of the cross-section should be aligned with streamlines and there should not be any significant later flow into or out of the plane of the cross-section (Delleur, 2006) The applicability of the 2D profile model is limited in cases where: 1) there is river where the assumption of no lateral flow is violated due to groundwater flow convergence to the river, 2) in areas where there are pumping wells, where there is radial flow in all direction, 3) fractured rock and karst aquifers are commonly modelled as equivalent porous media and this assumption is usually valid for large-scale groundwater flow models (Merz, 2012), however 2D profile model has been applied in Karstic aquifer in Jordan (Xanke et al., 2016).

2.2 Model Domain and Discretization

A 2D profile model was constructed along the Ngotwane River. Figure 27 shows the schematic profile. The 2D profile cross-section is fully inside the dolomite formation. The length of the cross-section is about 5.48 km, its orientation is South-North, such that its length is approximately parallel to the principal groundwater flow direction.

2.3 Model Discretization

Horizontally the model was discretized into 10 m x 1 m. The width of the profile was set to 1 m. The model domain consists of 584 rows and 1 column. Vertically the model is discretised into two layers representing: upper karstic zone and deeper karstic layer. The top elevation for the first layer was determined using average elevation calculated based on SRTM 30 m digital elevation model. This was set to 1023.7 m above mean sea level (amsl). Similar, the initial condition was defined based on the observed water level data on Jan 2000 and set to be 1021.96 m amsl. The upper layer was modelled as unconfined and the second layer was treated as a confined layer. Temporally the 2000-2012 model simulation period was discretised into 156 monthly stress period. Each monthly stress periods were further divided into weekly time steps. The total model thickness is assumed to be 150 m. The upper layer assumed to have thickness of 30 m and the lower layer 120m.

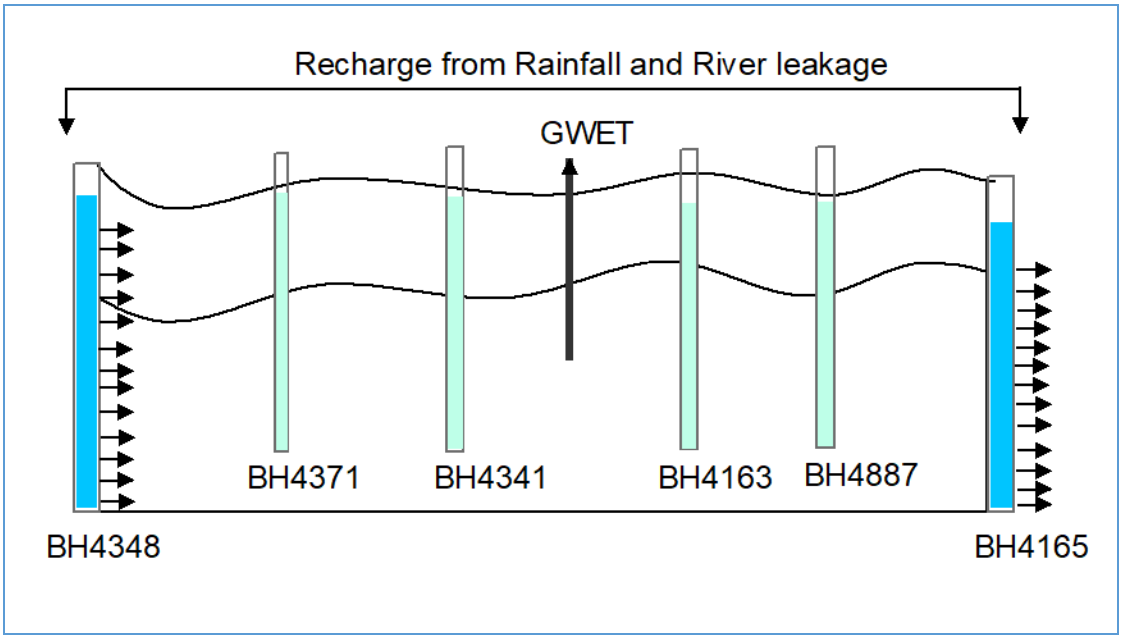

Figure 27: schematic 2D cross-section, observation wells used as boundary and intermediate observation wells used as a calibration point. GWET represent groundwater evapotranspiration. Since there was no abstraction intended simulation period is not included in the conceptual model.

2.4 Boundary conditions

Three kinds of boundary conditions were assigned. It was assumed that the left and right sides of the model boundaries represent streamline boundaries hence assigned no flow boundaries, it was also assumed that the model have inflow in the Southern and out flow at the Northern boundaries hence these boundaries were represented using GHB package, no flow boundary assigned at the model bottom because the geologic layer assumed to have very low hydraulic conductivity. The water level for GHB model boundary at the Southern and Northern were defined based on water level data measured in observation well BH4348 and BH4165 respectively. Conductance values on both Southern and Northern ends were optimized during model calibration.

2.5 Recharge and Groundwater Evapotranspiration

Recharge package was used to calculate recharge flux using monthly precipitation and multiplier constant fraction. The recharge flux applied here represents combined recharge from rainfall and river bed infiltration. The recharge multiplier constant was determined during model calibration. Groundwater evapotranspiration was simulated using Evapotranspiration package. Monthly

Evapotranspiration values were calculated based on S class pan evaporation data from Molatedi Dam. Based on the study by Shah et al (2007) Evapotranspiration extinction depth was assumed to be 3 m.

2.6 Model Calibration

The model was calibrated using observed data from four monitoring boreholes (BH4371, BH4341, BH4163, and BH4887) for the period 2000-2012. There is no pumping during this period, which reduce the effect of lateral flow. BH4371 and BH 4341 were moved to the West from their original locations to align them along the profile cross-section. Likewise, the Northern GHB boundary defined by BH4165 was moved to the East to align it along the profile cross-section. The model calibration was completed using PEST. In total six parameters were adjusted during the calibration process. These parameter include: 1) hydraulic conductivity of the upper and lower layers, 2) storage coefficients of the upper and lower layers, 3) recharge multiplier constant, 4) conductance value for GHB.

Initial parameter values, parameter ranges and calibrated values are presented in Table 6. The estimated values of storativity values for the dolomite from pumping test ranges by order of magnitude. Selaolo (1985) estimated sotrativity value of 0.03 for dolomite formation while the Institute of hydrology (IoH, 1986) reported sorativity value of one order of magnitude less for the same formation which is 0.003. Based on these two studies GCS (2000) adopted an average value, which is 0.01 for the dolomite formation. The Storativity values of dolomite formations reported for Grootfontein area, South Africa were 0.025 for Oakatree dolomite formation, 0.008 for Monte Christo and Lyttleton dolomite formation and 0.12 for Eccles dolomite formation (DWAF, 2006). The initial and parameter range for conductance value was specified based on the BH4165 saturated thickness. BH4165 has a depth of 150 m, water strike at 50 and 129 m below ground and static water level 6.1 m below ground (GCS, 2000). Based on these data the saturated thickness is calculated to be (129-6.1~ 123m). Assuming hydraulic conductivity of 2.8 m/d and distance between cell centres at the boundary equal 10 m, the initial conductance value is calculated to be 34 m^2/d. The upper and lower bound for the conductance values were calculated using the upper and lower hydraulic conductivity values.

Table 6: Parameters initial values, range and calibrated values for the 2D profile model

Parameter	Initial value	Lower bound	Upper bound	Calibrated value
Hydraulic conductivity layer 1(Kh1, m/d)	1.4	0.01	9.0	1.563
Hydraulic conductivity layer 1(Kh2, m/d)	2.8	0.01	9.0	3.335
Specific yield layer1 (Sy1, [])	0.03	0.003	0.12	3.21E-2
Specific storage layer 2 (SS2, [])	0.01	1E-5	0.12	1E-5
Recharge multiplier (Rch_mult, [])	0.2	0.05	1.0	0.833
GHB conductance North (Cond1, m^2/d)	34	0.12	110	100
GHB conductance South (Cond2, m^2/d)	34	0.12	110	10

2.7 Calibration Results

Correlation coefficient (Cor), Mean Absolute Error (MAE) and Root Mean Square Error (RMSE) between the observed and simulated water level values for each observation wells are provided in Table 7. Figure 28-31 presents the observed and simulated water levels at the four observation borehole calibration points. Figure 32 presents the scatter diagram for the observed and simulated values for all observation wells.

Table 7: 2D profile model calibration performance criteria

Observation well	Cor	MAE	RMSE
BH4887	0.86	1.77	1.82
BH4163	0.90	2.10	2.18
BH4341	0.67	2.23	2.26
BH4371	0.73	1.26	1.37
All	0.53	1.82	1.92

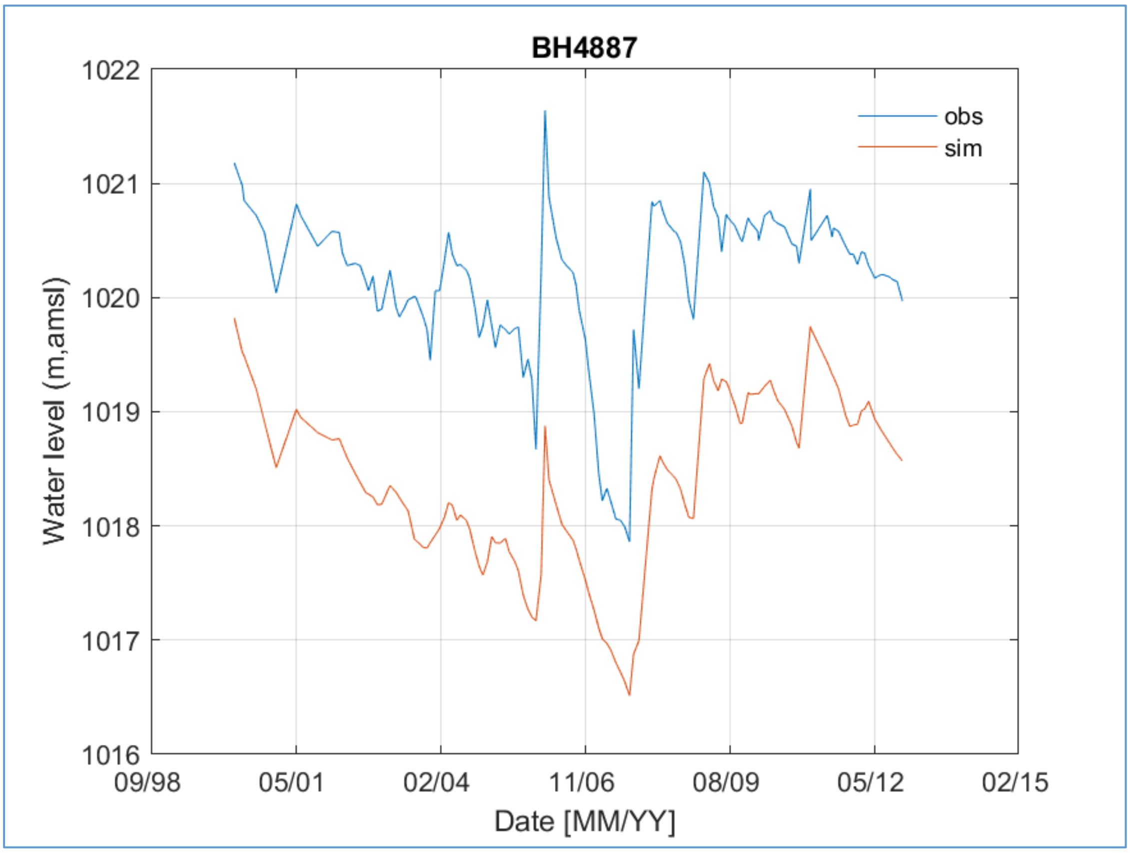

Figure 28: Observed and simulated groundwater levels during model calibration (BH4887, depth 144m)

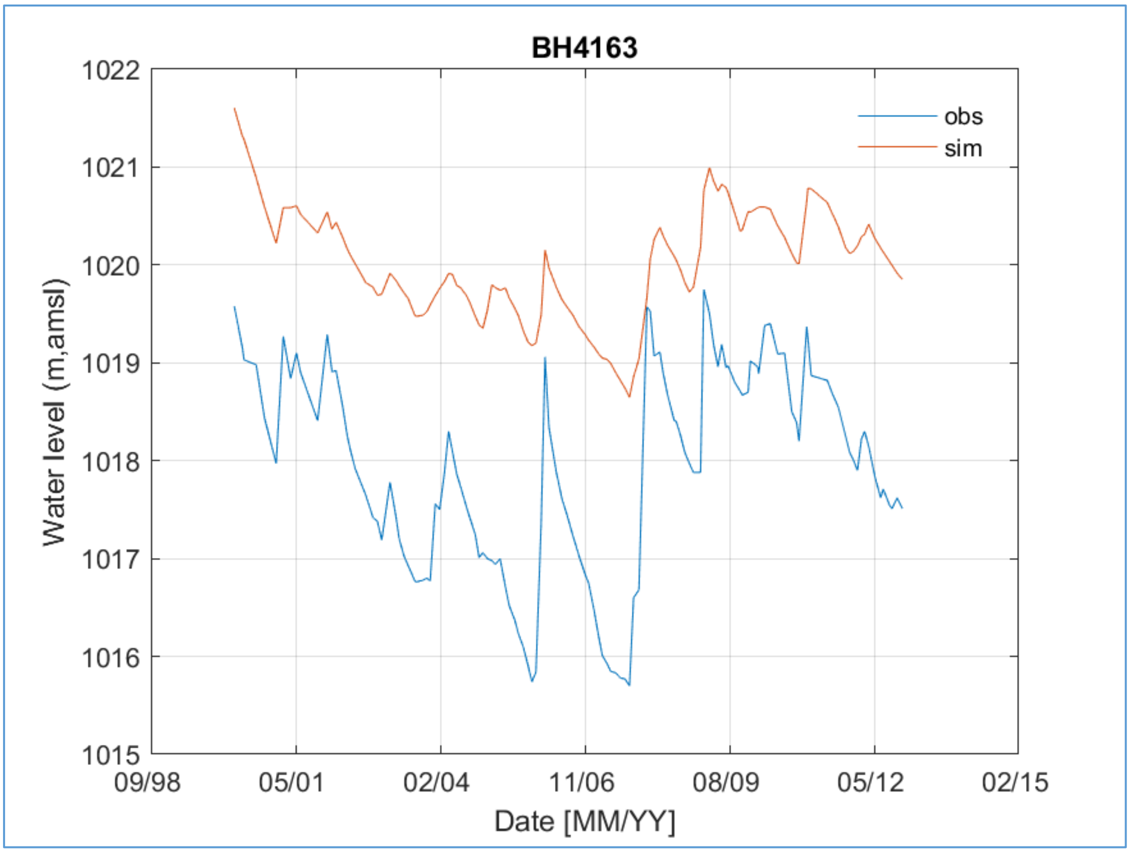

Figure 29: Observed and simulated groundwater levels during model calibration (BH4163, depth 100m)

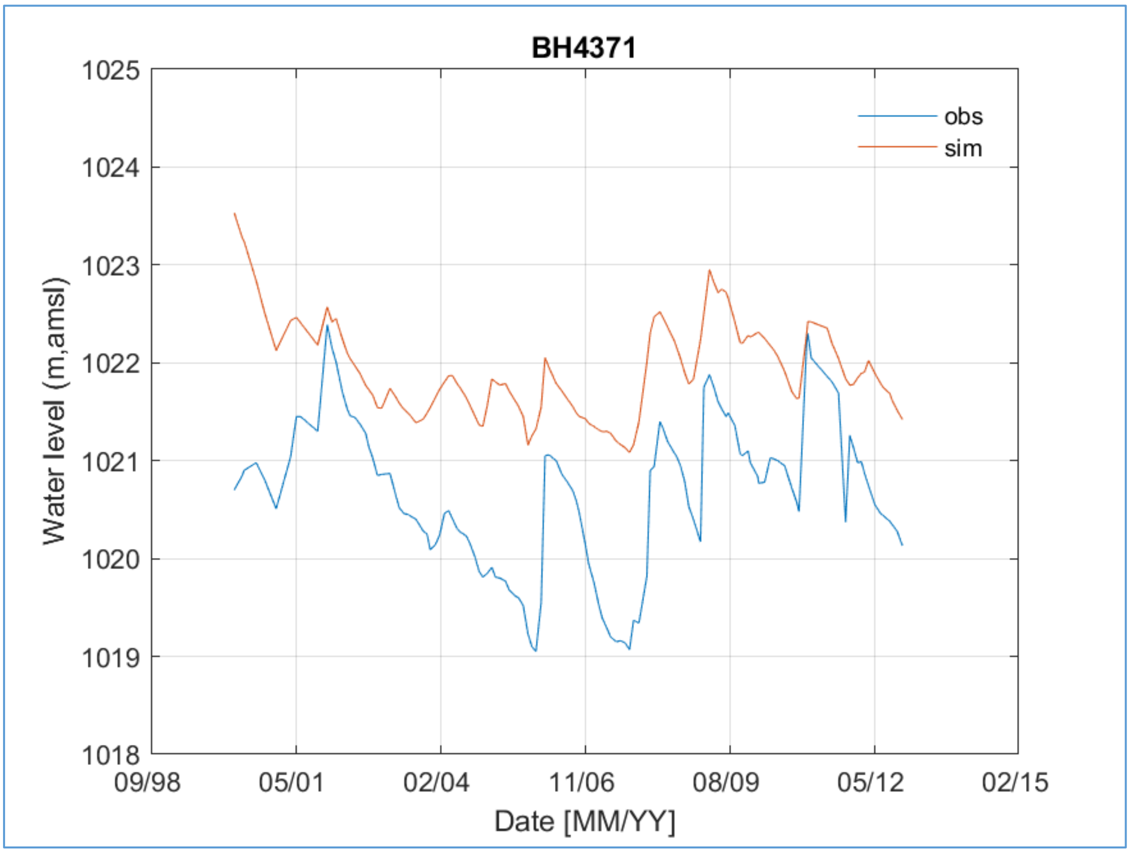

Figure 30: Observed and simulated groundwater levels during model calibration (BH4371, depth 96m)

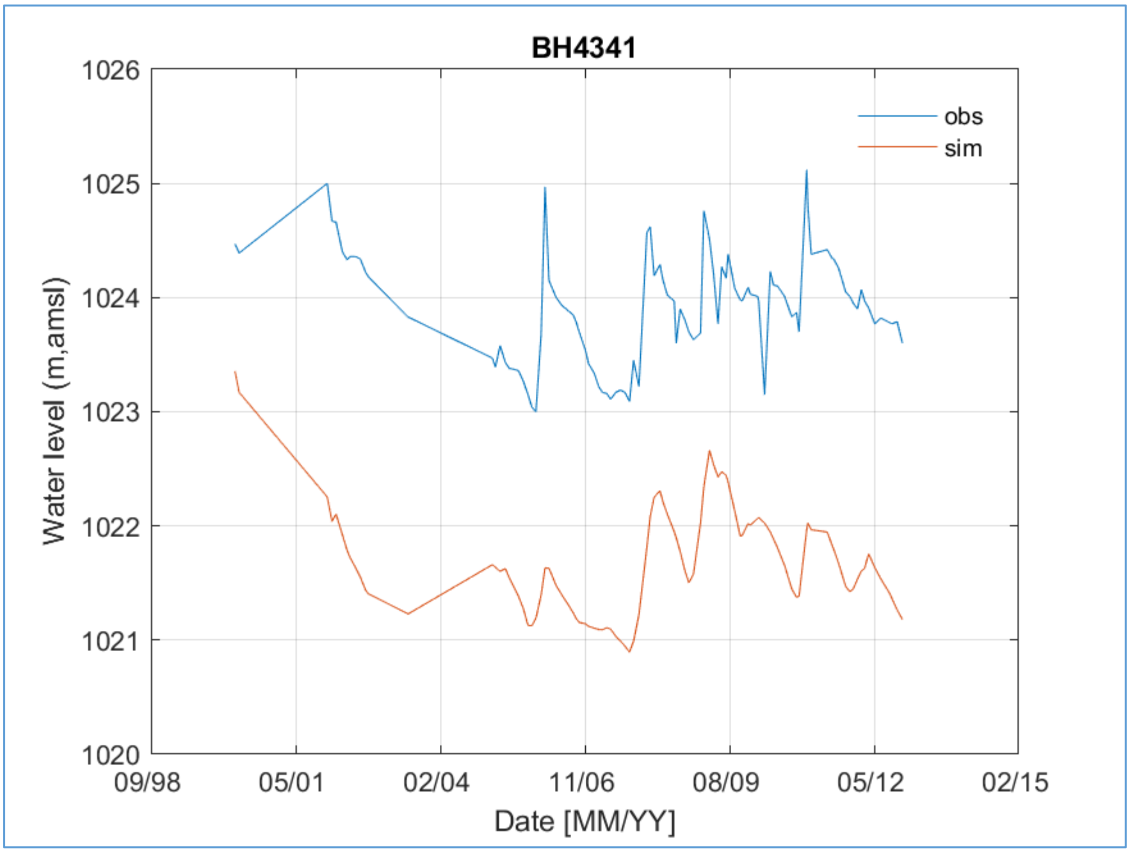

Figure 31: Observed and simulated groundwater levels during model calibration (BH4341, depth 28.5 m)

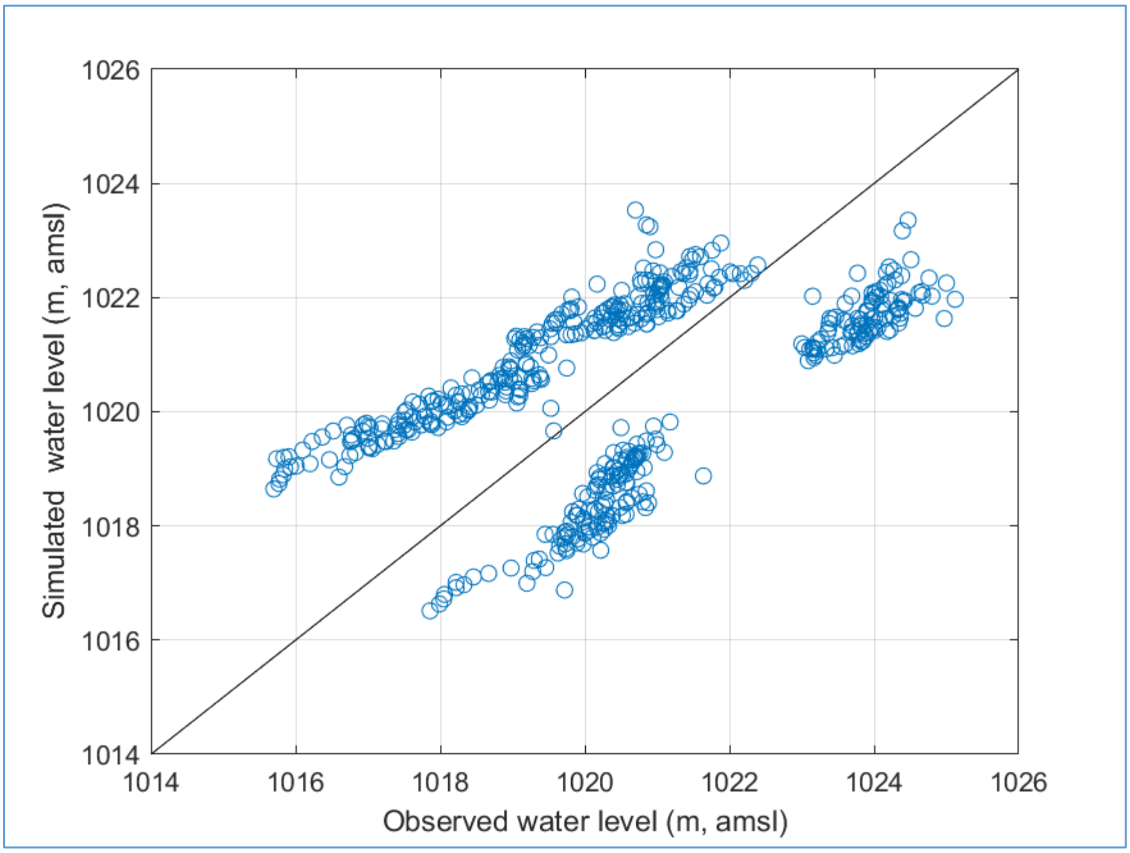

Figure 32: Observed vs simulated water level scatter plot

2.8 Discussions

In general, there is good agreement between simulated and observed groundwater levels. The calibrated values are presented in Table 5. The water level dynamics of the simulated values follow what was observed in the field. However, it should be noted that for the model to be representative the optimized values should be realistic. It obvious that unique parameter estimation may not be possible due to the known issue in model calibration which is an “ Equifinality” problem (Beven, 2006). Without independent knowledge of hydraulic conductivity and storage coefficient it is difficult to obtain representative value of recharge. Recharge is also found to be highly correlated to flow into and out of the model boundary.