Spatio-Temporal Groundwater Recharge Assessment Serowe case study, Botswana

Kenneth Setimela

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by

Kenneth Setimela

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Thesis Assessment Board

Chairman:	Prof. Dr. Z. Su)
External Examiner :	Dr. O. T. Obakeng
First supervisor :	Dr. M. W. Lubczynski
Second supervisor :	Drs. R. Becht
Advisor :	A. P. Frances

INTERNATIONAL INSTITUTE FOR GEO-INFORMATION SCIENCE AND EARTH OBSERVATION
ENSCHEDE, THE NETHERLANDS

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Abstract

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1. INTRODUCTION

1.1. Background

Botswana is one of the fastest growing countries in Africa. However, the lack of surface water in most parts of this semi-arid area is a major challenge to the country. This makes groundwater a very vital resource in Botswana. Many areas in the country are reliant upon it for their water supply needs, be they domestic, agricultural or industrial. About 80% of the rural population in Botswana is supplied from groundwater, Serowe included (DWA, 2000). Also the Botswana National Water Master Plan Review (Department of Water Affairs, 2006) estimates that at least 65% of Botswana's water demand is met from groundwater resources. The mining industry, which is the backbone of the country's economy, also relies on groundwater. The Serowe well field is currently unable to satisfy the water demands of Serowe village.

According to Gieske (1992), knowledge concerning the lifespan of groundwater reservoirs and the processes governing their replenishment is indispensable for the design of regional water master plans. Semi-arid areas like Botswana are characterised by years with good rains and groundwater recharge, followed by drought periods, making management of this resource very difficult but very essential.

1.2. Problem statement

The available numerical groundwater MODFLOW model of the Serowe aquifer was developed in 1998 and calibrated in steady state and in partially transient mode (Wellfield Consulting Services, 2000). A lot of data has been collected over the past ten years which allows fully-transient calibration with spatio-temporally variable fluxes. The recharge assessment proposed in this study represents a huge step forward in a preparing for a fully transient model of the area.

1.3. Objectives

The objective of this study is to quantify Serowe aquifer recharge spatially and temporally in order to evaluate renewability of the groundwater system by applying various methods.

1.4. Assumptions

This study made assumptions as listed below.

1. Lateral flow has negligible effect on the 1-D recharge estimated by using EARTH model.
2. Groundwater recharge determined from the chloride mass balance method represents net groundwater recharge.

1.5. Hypothesis

The Serowe aquifer net recharge varies in space and time.

1.6. Research questions

- How is the net groundwater recharge of the Serowe aquifer distributed spatially and temporally?

- What is the renewability of Serowe groundwater resources?

2. THE STUDY AREA

2.1. Location

The study area is situated near Serowe village, which is the administrative capital of the Central District of Botswana; about 320 km (by road) north of the capital city Gaborone, Figure 1. It covers an area of 2444 km^2 and is bounded by UTM coordinates 400000E to 497000E and 7501000N to 7545000N, in UTM zone 35 S. The most prominent geomorphological feature is the NNE-SSW trending escarpment that divides the area into two; the Kalahari sand cover in the west (sandveld), and the eastern hardveld which is essentially devoid of sand cover. According to Swedish Geological (1988) the escarpment is between 90 and 150 metres above the surrounding countryside and reaches a maximum elevation of 1260 metres above sea level. The escarpment marks the eastern limit of the Kalahari Basin.

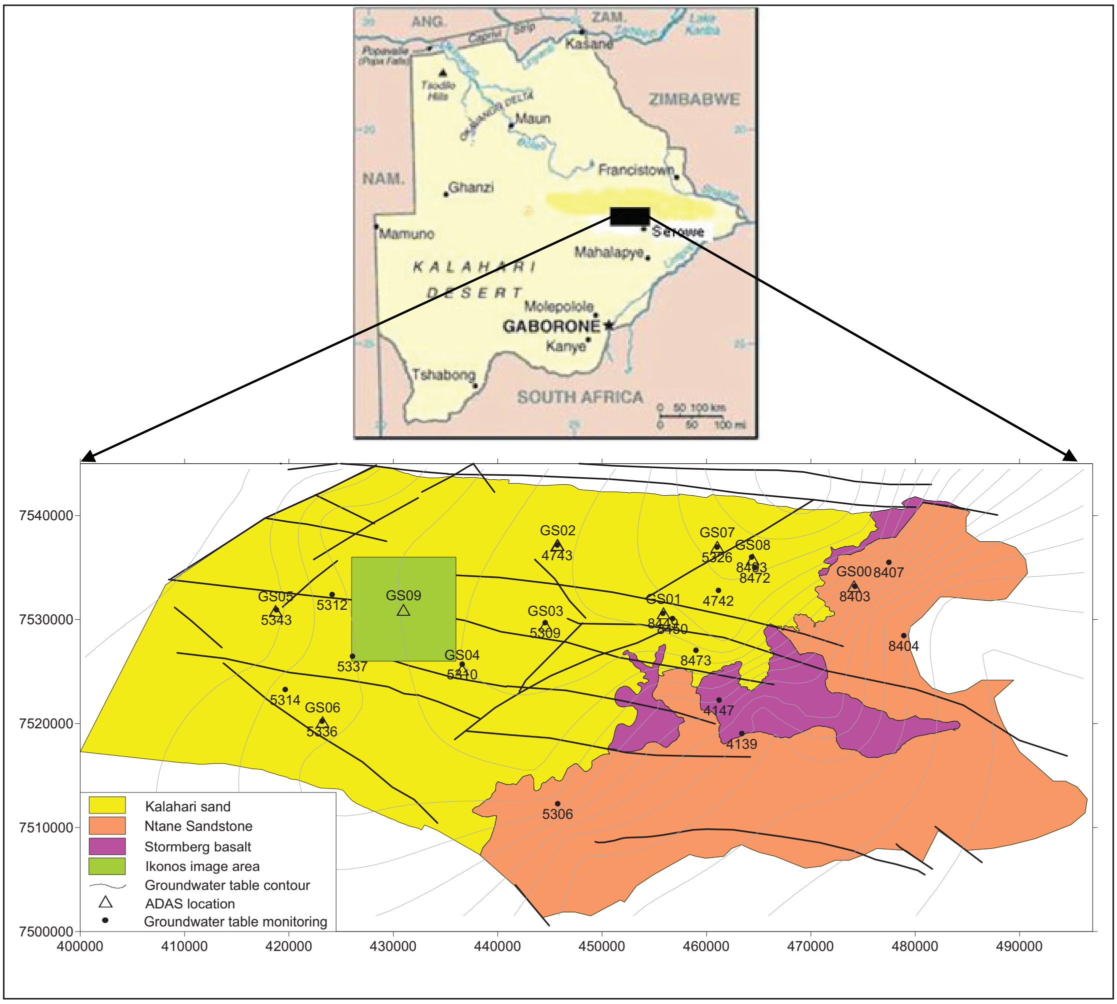


Figure 1: Project area location map

2.2. Climate and hydrology

The summer season is usually hot and wet; while winter is cold and dry.

2.2.1. Rainfall

Rainfall records of the Serowe area indicate seasonal pattern: high rains fall in the summer months of October to April winter months have very little to no rainfall. The long term mean annual rainfall for Serowe has been reported as 437mm for the period 1925 to 2004 (Obakeng, 2007).

Using the 1986-2007 rainfall records for this study, a mean annual rainfall of 452mm with a minimum of 259mm (in 2002) and a maximum of 970mm (in 2000) are obtained (Figure 2). The wettest month is December with monthly average of 100mm and the driest months are July and August with no rainfall recorded (Figure 3).

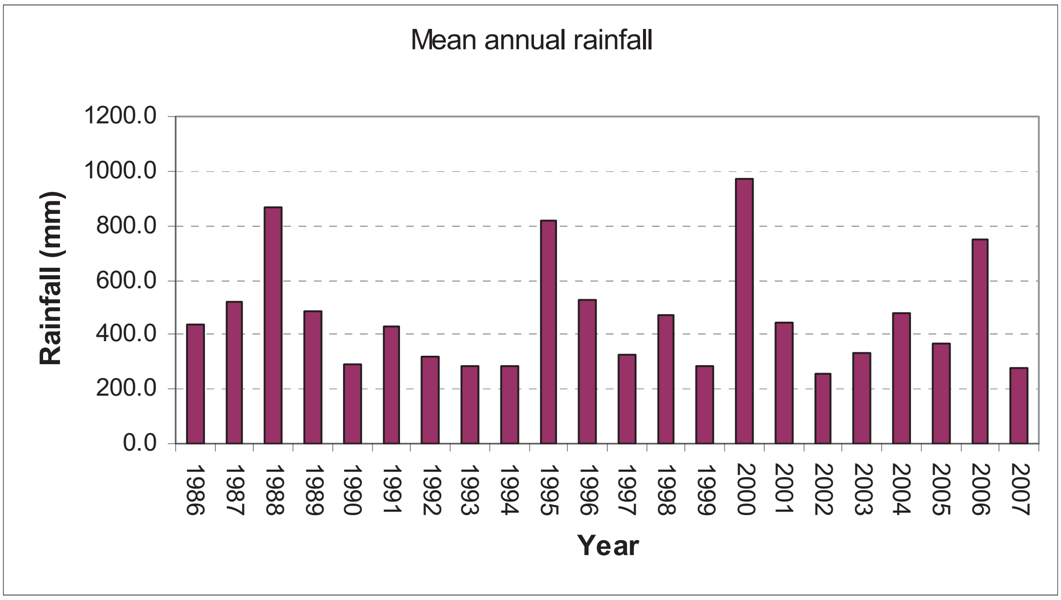


Figure 2: Serowe mean annual rainfall

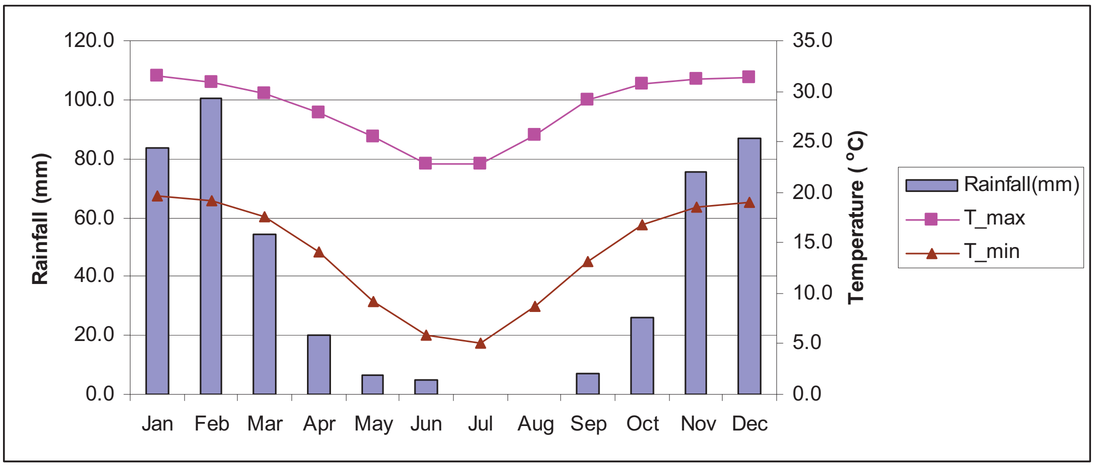


Figure 3: Rainfall and temperature pattern in Serowe

2.2.2. Temperature

It is evident from Figure 3 that the hottest months are October through April, and this coincides with the rainy season. The cool, winter months are also the dry months.

2.2.3. Relative humidity

Relative humidity data was obtained from Malebala automatic data acquisition station (ADAS). Measurements were recorded every thirty minutes since 2003 at 2m, 10m and 18m heights. Relative humidity was used as input in AWSET program for estimation of reference evapotranspiration, ETo.

2.2.4. Wind speed

The wind speed was recorded at three heights of 2m, 10m and 18m. The wind speed was needed as input in AWSET program for estimation of reference crop evapotranspiration. Wind speed at 2m height was used to estimate reference evapotranspiration, ETo.

2.3. Geology and structure

A detailed description of the geology of the study area has been described by Smith (1984) and Carney et al., (1994). Smith (1984) only gave a full description of the Karoo Supergroup rocks in Botswana, while Carney et al., (1994) studied the geology of Botswana as a whole. As has already been mentioned above (section 2.1), in most cases the hard rocks are covered by thick Kalahari beds. Consequently, drilling cores, borehole drilling chips and geophysical interpretations were used to describe the lithological units. Only a brief description of the geology of the area is presented below. Furthermore, the general stratigraphy of the study area is summarized in Table 1. Figure7 shows the geology of Serowe study area with groundwater monitoring points superimposed on it.

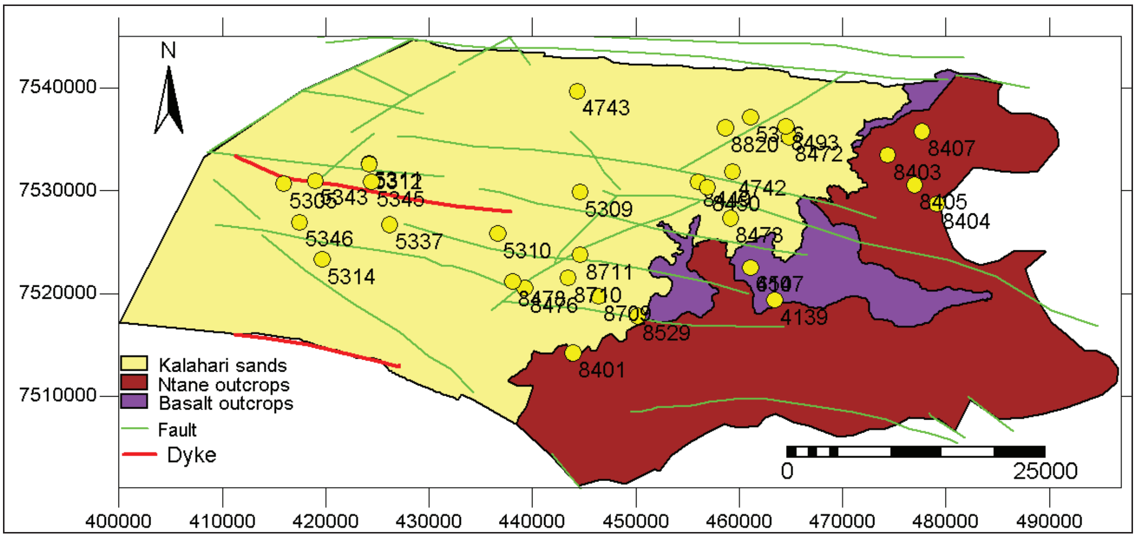


Figure 4: Geological map of Serowe (with groundwater monitoring points)

Table 1: General Straigraphy of Serowe area [After Carney et al. (1994)]

Age	Supergroup	Group	Formation	Lithological descriptions
Cainozoic		Kalahari	Kalahari beds	Sand, silt, clay and duricrust
Mesozoic	Karoo	Stormberg Lava		Crystalline, massive, amygdaloidal basalts
Upper Palaeozoic		Lebung	Ntane	Aeolian sandstone, medium to fine grained with minor mudstone intercalations, white, pink, red, brown, grey, green, yellow, partially fluviatile towards base
			Mosolotsane	Fluvial red beds, siltstones, fine grained sandstone, red mudstones
		Beaufort	Thabala	Non-carbonaceous mudstones and siltstones with minor sandstones
		Ecca	Serowe	Carbonaceous mudstones, coals, siltstones, coally carbonaceous mudstones, fine sandstone
			Morupule	Coal seams, black carbonaceous mudstone, subordinate non-carbonaceous mudstones
			Kamotaka	White, massive, coarse to medium grained sandstone, subordinate siltstones, micaceous
			Makoro	Post glacial lacustrine mudstones and siltstones marking the base of Ecca Group
		Dwyka	Dukwi	Base of Karoo sequence, tillites and shales, varved siltstones and mudstones
Proterozoic		Palapye	Shoshong	Dolerite, siltstones, shales and quartzite
Archean		Basement		Granite, gneiss and amphibolite

The pre-Karoo rocks consist of Archean Basement and Palapye Group rocks. The Archean Basement and rocks of the study area consist of an assemblage of gneisses, migmatites and other plutonic rocks with varying degrees of foliation. These rocks are not outcropping within the study area. Very few boreholes intersect these rocks at great depths. However, outcrops of the Palapye Group rocks are found east of Serowe village where they are represented by quartzites, siltstones, shales and sandstones with various shades of grey and pink colours. Pre-Karoo and post Palapye Group intrusions occur as coarse grained dolerite dykes intruded into the Palapye Group or Basement rocks (Carney et al., 1994).

Overlying the Basement rocks are the Karoo Supergroup rocks, Table 1. This assemblage is mainly comprised of sandstones, siltstones, mudstones and tholeiitic flood basalts (Carney et al., 1994; Smith, 1984). The basalts were intruded onto the sandstone, and mark the end of Karoo sedimentation. To the west of the escarpment the Karoo Supregroup rocks are overlain by varying thicknesses of Aeolian sediments of the Kalahari Group. To the east and along the escarpment the karoo basalt and sandstone are mostly exposed at the surface due to gully erosion. These are the most important rocks in terms of hydrogeological significance in the area as they host the primary aquifer (see section 3.4 below for hydrogeological description).

Widespread faulting has disrupted the pre-Kalahari rocks, with many faults aligned in the WNW-ESE direction, breaking the Karoo into a series of graben and horst structures (Swedish Geological Company, 1988). The WNW-ESE structures are of hydrogeological importance as they control the groundwater flow. Superimposed on this fault pattern are a conjugate set of NE-SW and NW-SE trending faults and fractures. Refer to figures 7 (above) and 8 (below).

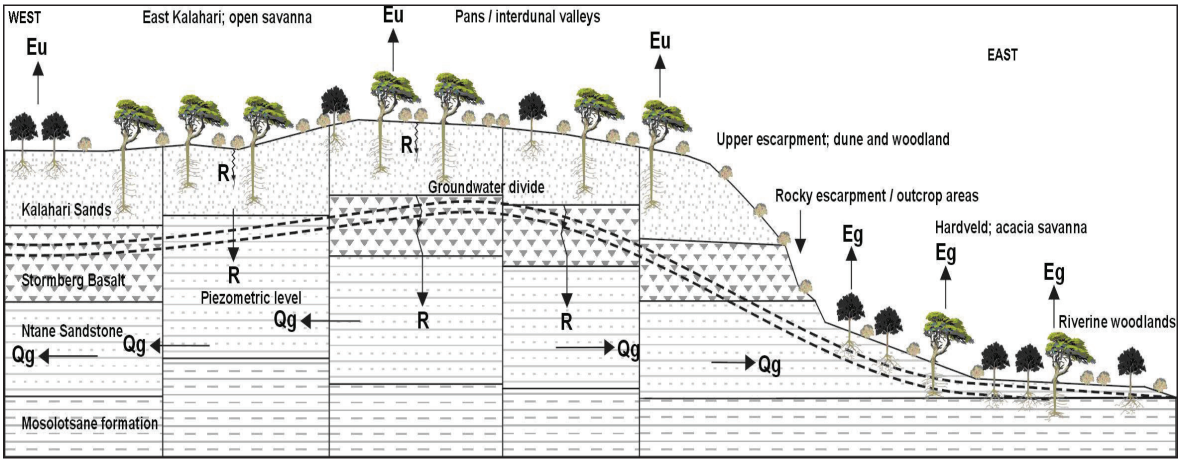


Figure 5: Hydrogeological cross-section of the study area (Lubczynski, (2000))

Where,

Eu = evapotranspiration from unsaturated zone

Eg = groundwater evapotranspiration

R = recharge

Qg = groundwater flow

2.4. Hydrogeology

A number of projects have contributed significantly to the understanding of the hydrogeology of Serowe area. These include amongst others, Swedish Geological Company (1988) and Wellfield Consulting Services (2000). The hydrogeologically important Karoo rock units in Botswana are the sandstones of Lebung and Ecca groups Table 1.

In the study area the most important aquifer is the Ntane Sandstone Formation of the Lebung Group. It has primary porosity because it is arenaceous and poorly cemented. It also has secondary porosity resulting from the fracturing and weathering of this formation. The Ecca sandstone has not been fully explored in the area probably due to its depth, and suspected poor water quality. The Ntane sandstone is thicker (more than 60m) in the sandveld portion of the study area, where also the groundwater table is very deep. It is thinner in the eastern hardveld, where the depth to the water table is shallow (Figure 9). The base of the Ntane sandstone aquifer is the Mosotshane Formation, which consists of very low permeability mudstones and siltstones. A series of faults and dykes that criss-cross this aquifer, play a significant role in defining the aquifer recharge and groundwater flow direction. Dykes and faults may act as barriers to groundwater flow or act as conduits for preferential flow depending on the fracture pattern and whether they are in-filled or not. An impermeable graben structure to the north of the study area forms the northern boundary, while an impermeable dolerite dyke is the southern boundary. The eastern and south eastern boundaries are defined by the limit of the Ntane sandstone. The aquifer is normally confined when it is overlain by thick basalt lavas. Where there is no basalt or the base of basalt is below the sandstone then unconfined conditions prevail.

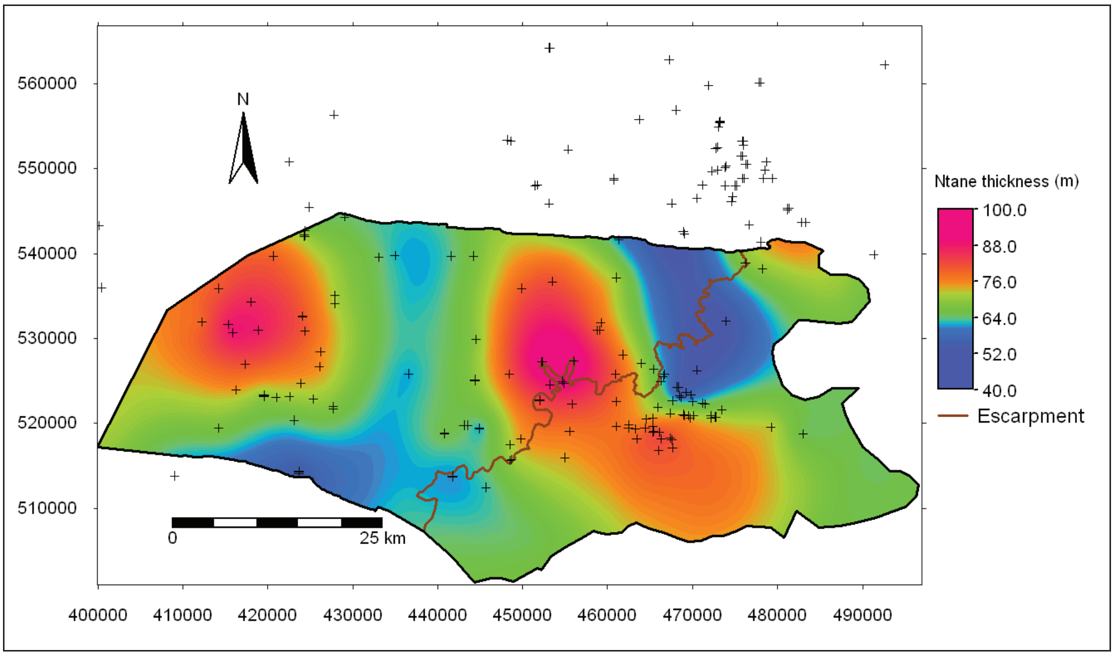


Figure 6: Ntane sandstone thickness

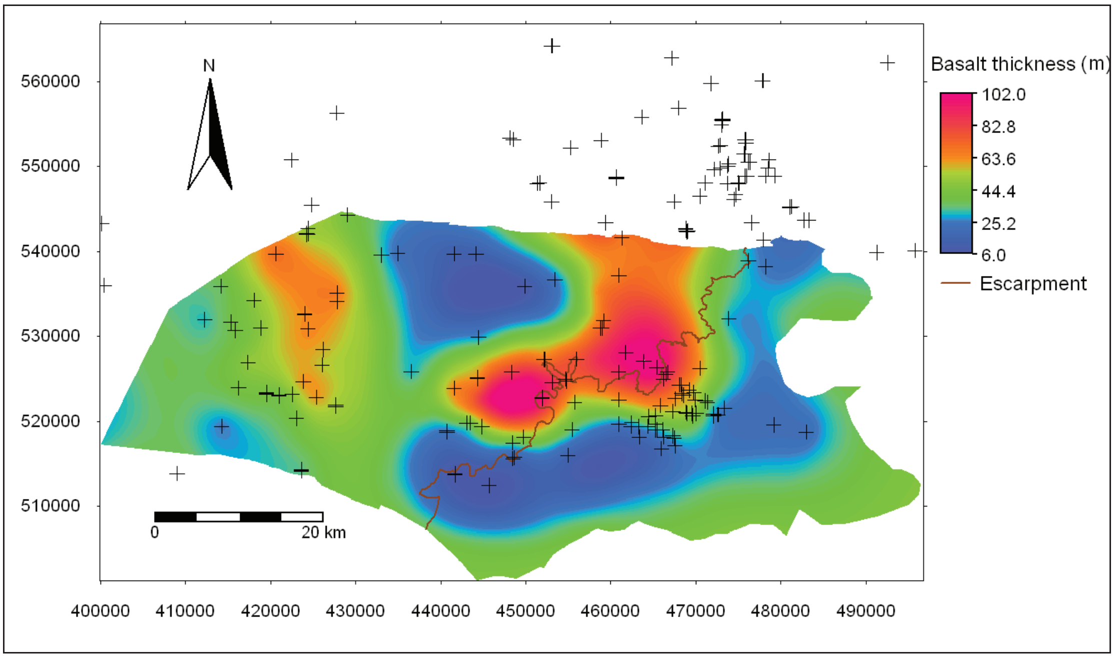


Figure 7: Basalt thickness map

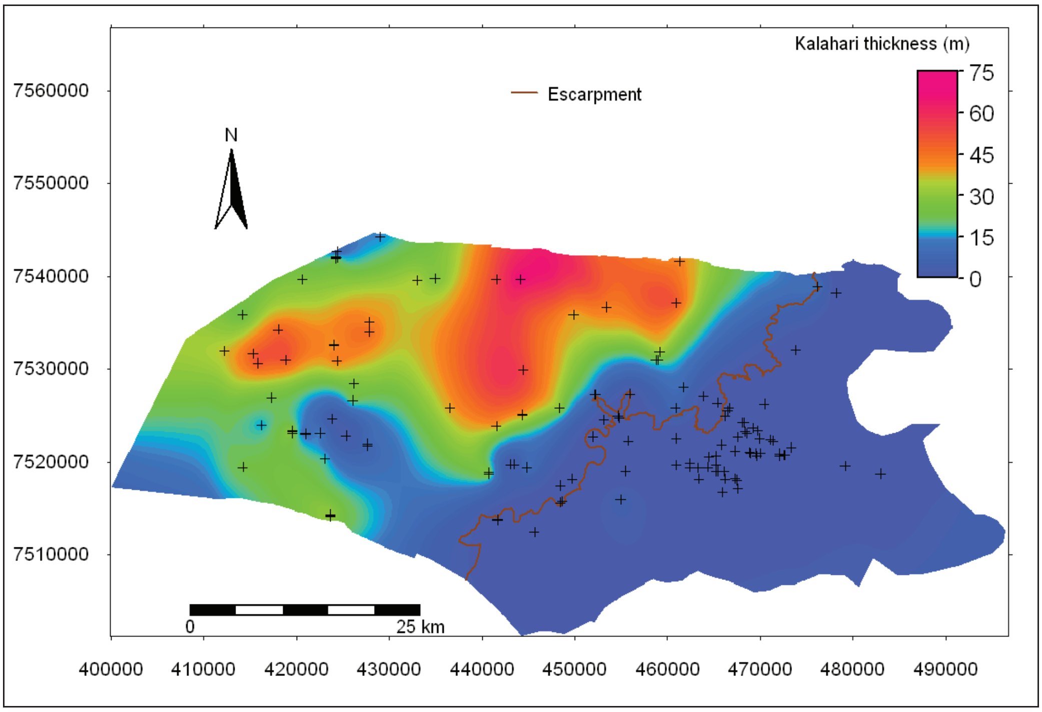


Figure 8: Kalahari thickness map

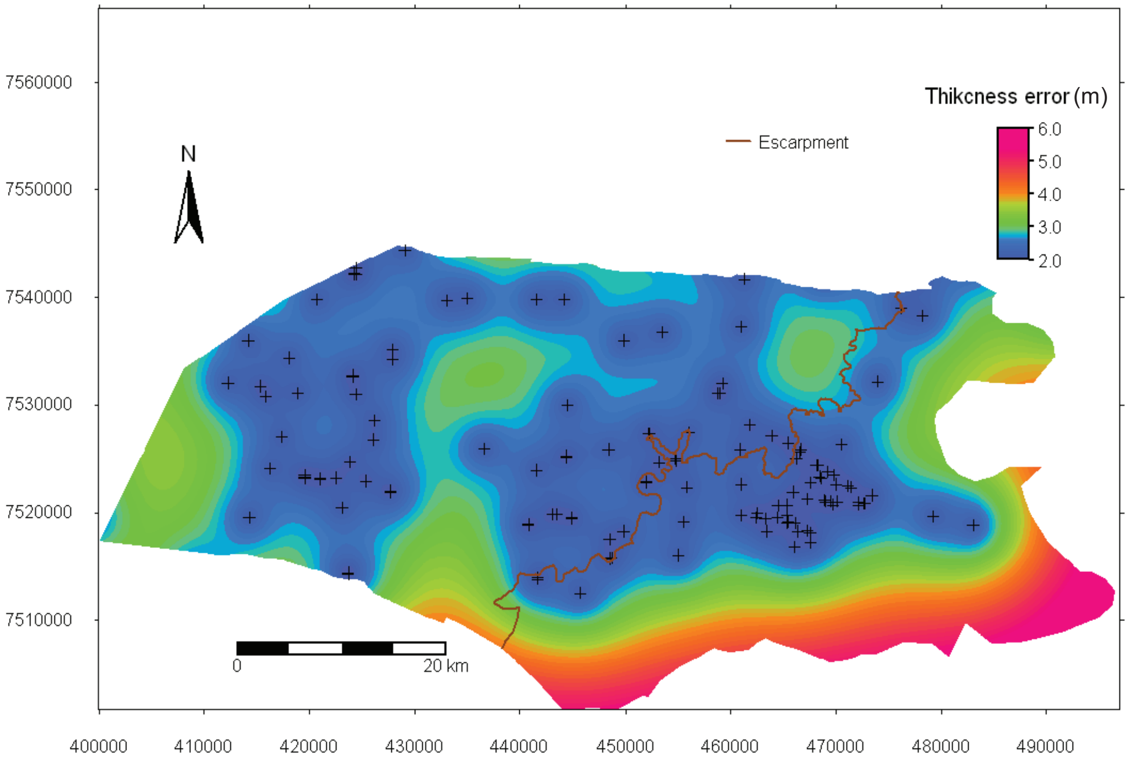


Figure 9: Kalahari thickness error map

3. Previous studies

A lot of work related to groundwater resources evaluation and development has been done in and around the study area (Swedish Geological Company, 1988). Some academic researches that have been undertaken in the study area are: (Gieske, 1992; Lubczynski, 2006; Magombedze, 2002; Magombedze et al., 2004; Obakeng, 2000; Obakeng, 2007; Timmermans and Meijerink, 1999; Zziwa, 2003). None of these researches developed a fully-transient ground model of the area which is complex and requires network of monitoring schemes (Lubczynski, 2006). A lot of data has been collected over the past eight years (since 2001) which allows fully-transient calibration with spatio-temporally variable fluxes such as recharge and groundwater evapotranspiration.

The available numerical groundwater MODFLOW model of the Serowe aquifer was developed in 1998 and only calibrated in steady state and in partially transient mode (Lubczynski, 2000; Lubczynski, 2006). This study helped a lot in understanding the groundwater balance of the study area; as well as highlighting the importance of temporal hydrological data acquisition. The study recommended installation of automatic data acquisition systems (ADAS) to use in spatial recharge and evapotranspiration modelling, and interfacing with numerical groundwater model.

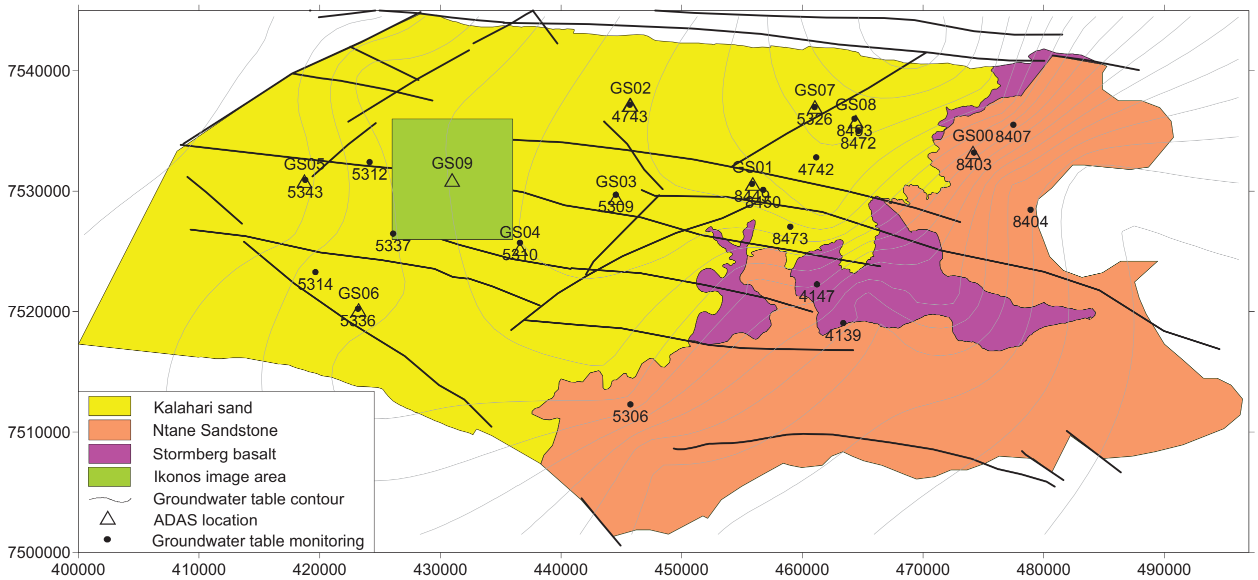


Figure 10: Monitoring network in Serowe area

Table 2: Data collected

Archive Data	Field Data
Borehole / well coordinates, elevations	Water level data – 45 boreholes
Pumping test records – tested yield, rest water levels, drawdowns	Water chemistry data – 19 samples; pH, temperature, and EC.
Meteorological data – rainfall, temperature (1971-2007)	Geophysics surveying (resistivity imaging) pseudosection
Borehole drilling records – borehole number, drilled depth, water strike, yield, lithological logs (rock types and thicknesses)	
Well abstraction records, chloride content and water levels	

A monitoring network was established in the study area in order to acquire hydrogeological parameters of the aquifer as well as meteorological variables of the Serowe area (Figure 10). Most of these were installed as part of PhD research work aiming at studying the soil moisture dynamics and evapotranspiration in the Botswana Kalahari (Obakeng, 2007).

The data processing by Obakeng (2007) terminates in March 2005. However, data monitoring/acquisition was continued beyond March 2005. Hence the aim of this study is to organise and process additional data that has been acquired since 2005.

4. Groundwater Monitoring

This chapter discusses the groundwater level monitoring within the Serowe aquifer. Hydrographs of monitoring wells are presented and discussed. Available water level abstraction records are also presented and discussed here. The chloride records are presented and analysed in chapter 6 (Recharge assessment).

Monitoring of groundwater levels is very crucial for proper management of an aquifer or well field because this can give an indication on water level trends so that appropriate remedial action can be taken to safeguard the aquifer. A good groundwater monitoring program should include water level records, borehole abstraction records, as well as groundwater quality monitoring.

Monitoring of the Serowe aquifer started in the earlier 1980s. Fairly good records of groundwater monitoring are available for the study area. However, the monitoring data is not always in a format that can be readily used by someone who does not know the area well or who is not experienced enough. Data is available both in digital and hard copies. Well hydrographs give a quick indication of the water level trends in the wells, and can be analysed to estimate the renewability of the groundwater resource by the well hydrograph method which assumes that a rise in groundwater table is due to rainfall recharging the aquifer. However this is not always the case as there are other factors that can cause a water table fluctuation, such as atmospheric tides, lateral flow and borehole abstraction (Obakeng, 2007). It should be noted here that this method is not readily applicable in the locations where the recharge is equal to discharge as in those instances there is no change in water level. This method is applicable only where there is a marked change in water table.

In the study area, water level monitoring is done both manually and with automatic water level recorders. There are some gaps in the data due to various reasons, ranging from logistical to mechanical. Water level graphs of some boreholes in the area are presented in Figure 11 below, all at the same scale.
The graph shows that water level changes in the Serowe are in the order of a few centimetres, considering the period 2001 to 2008. The highest hydraulic head change is about 2m at borehole BH8450. However, the water levels are showing a gentle downward trend over along period, with some sporadic water level rises in some years. This clearly shows that net recharge of the aquifer is both spatially and temporally variable. It is also seen that where groundwater levels rise, they generally start rising about one to three months after the rainfall, indicating delayed aquifer response to recharge.

In order to get a better picture of the water level trends at individual boreholes, a few hydrographs of the selected wells (see Figure 10 for locations) are presented individually below each having its own scale.

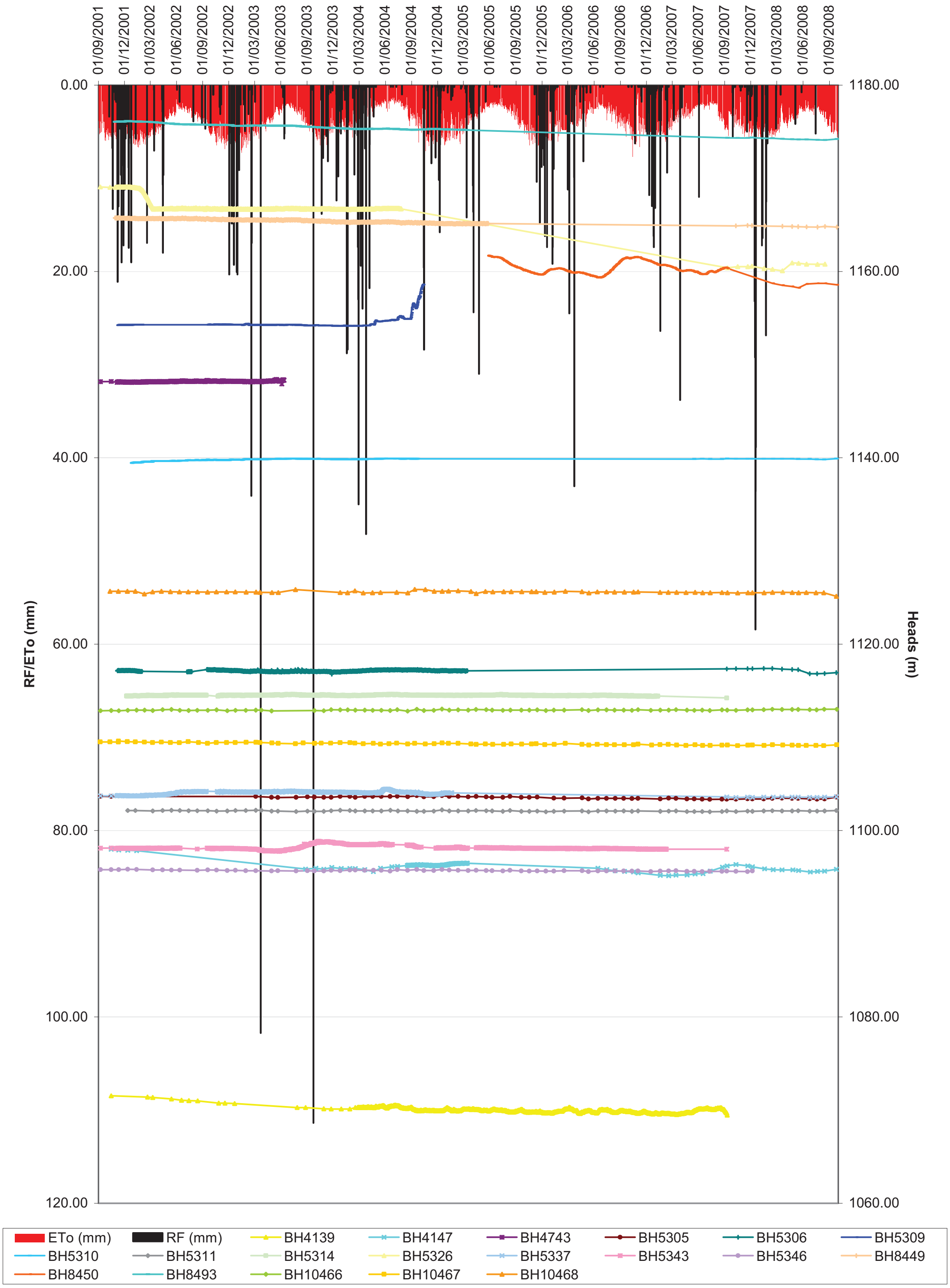


Figure 11: Serowe borehole hydrographs

BH4147 Hydrograph (Hardveld)

This borehole is within Serowe well field in the hardveld just below the escarpment. The aquifer here is overlain by the Stormberg Basalt, and has a maximum change in head of about 1m. However, the general trend of the groundwater level is downward. High water level fluctuations could be due to well abstraction effects.

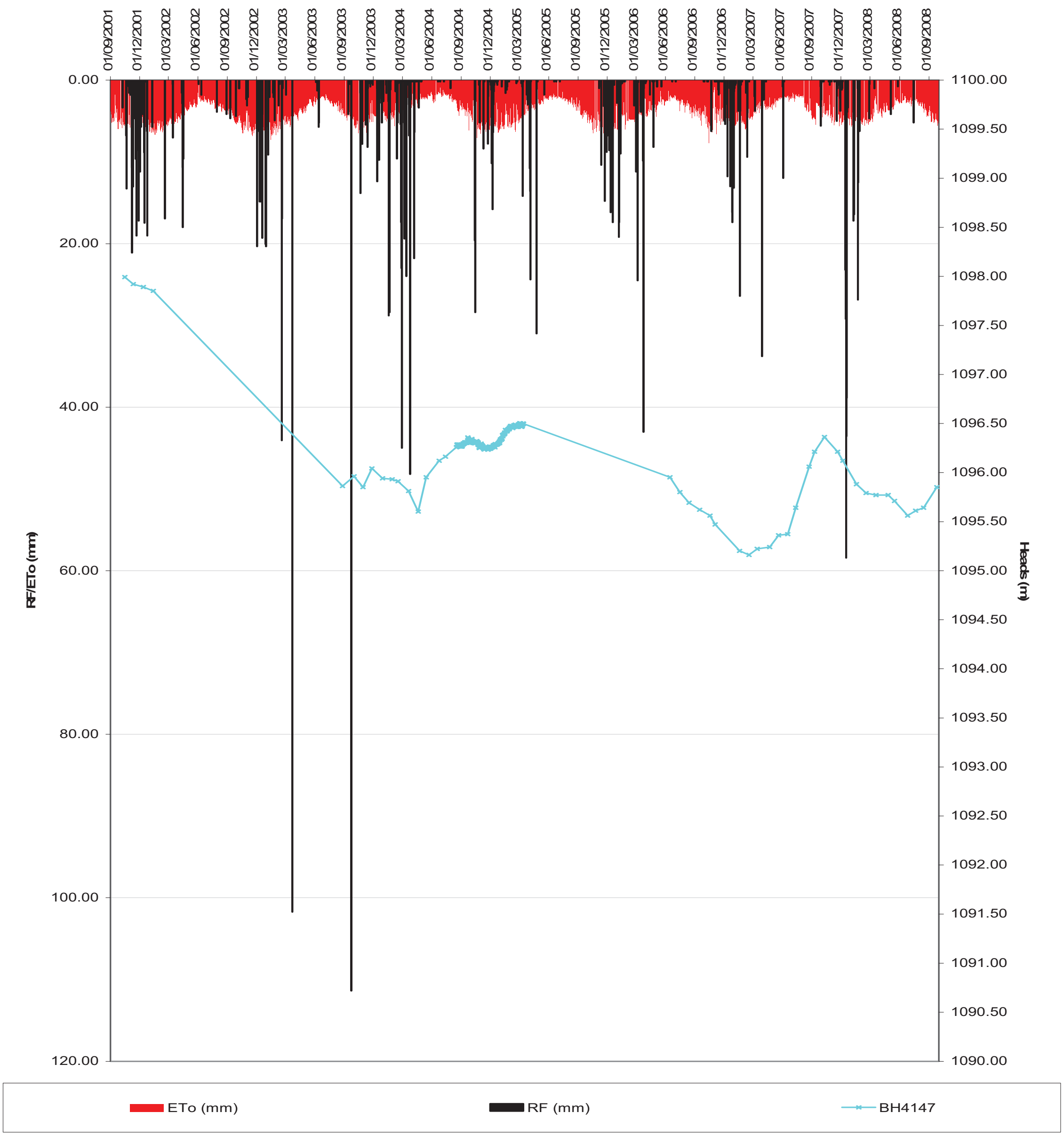


Figure 12: BH4147 water level hydrograph vs rainfall (hardveld borehole)

BH10466 Hydrograph (sandveld)

This borehole is to the western end of the project area in the sandveld. The change in water level hydrograph is in the order of a few centimetres only. This borehole is situated away from pumping boreholes. Therefore, the water level response here better approximates the actual groundwater recharge of the aquifer as it is not disturbed by pumping. It is close to the head constant boundary of the transient model of the area. A critical observation of the hydrograph shows that the water level is rising at this particular borehole, though at a very low rate.

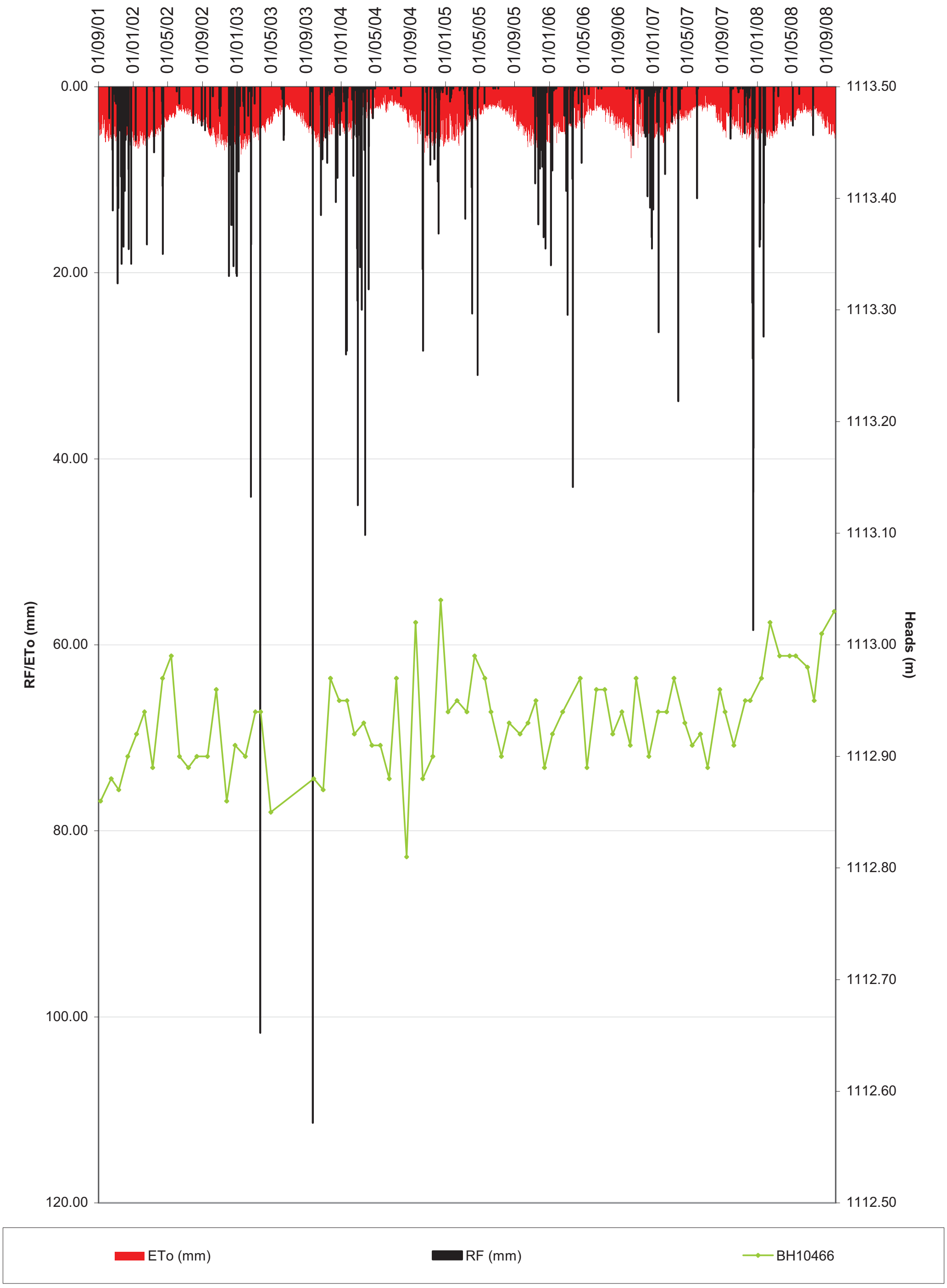


Figure 13: BH 10466 water level hydrograph vs rainfall/ETo

BH8493 Hydrograph (Sandveld, water divide)

This borehole is within Serowe production well field, above the escarpment, and close to the groundwater mound. The hydrograph shows a downward decline with time, which could be attributed to well field abstraction.

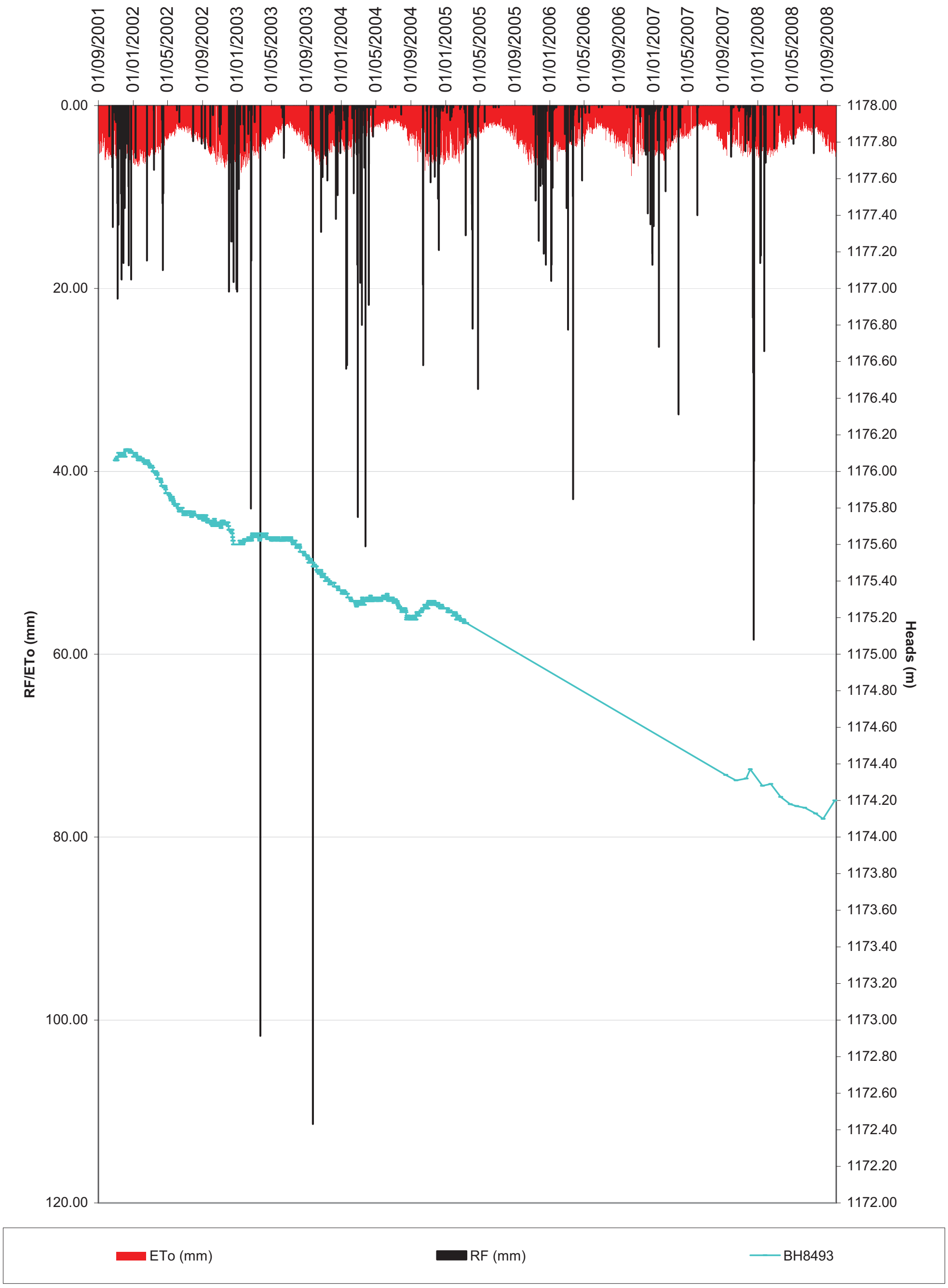


Figure 14: Bh8493 Hydrograph

BH5346 Hydrograph (in sandveld)

The borehole is located to the west of the study in the sandveld. The water level fluctuation is equal or less than 10cm. Even though the water level rise is so small, it can be seen that recharge to the aquifer does take place. There is a delay of about one month to three months between the time of precipitation and water table response.

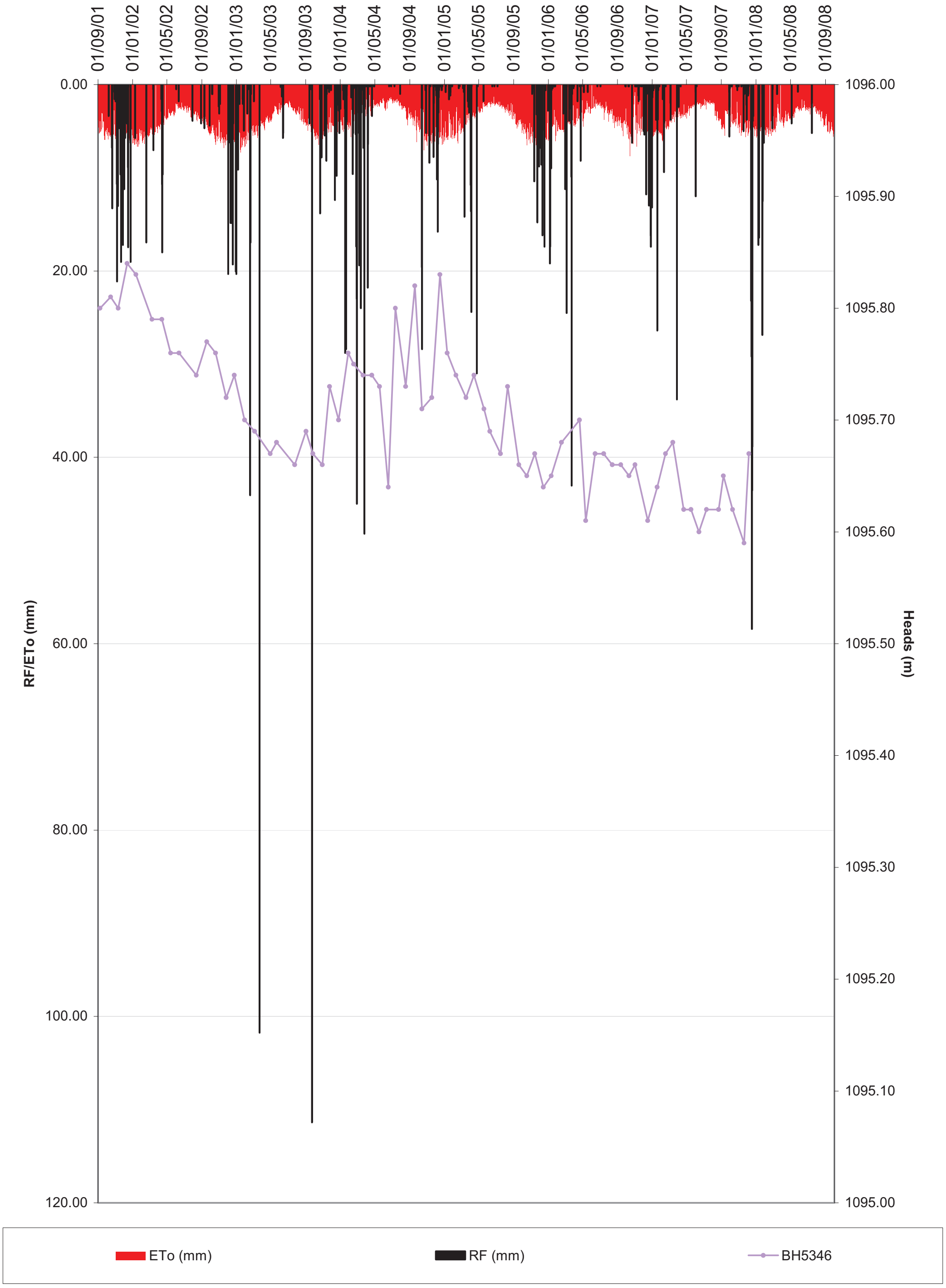


Figure 15: BH5346 Hydrograph

5. Evapotranspiration

Studies have shown that potential evapotranspiration rate and reference FAO evapotranspiration (FAO No.56, 1998) in Serowe, and for Botswana as a whole, are much higher than the annual rainfall. The evapotranspiration that would occur when there is no shortage of water is referred to as potential evapotranspiration (PET). The evapotranspiration that actually happens is referred to as actual evapotranspiration, Ea, and is governed by the availability of water. The evaporation that occurs from a reference surface that is not short of water is referred to as reference evapotranspiration, ETo. This reference surface is defined as a "hypothetical reference crop with an assumed crop height of 0.12m, a fixed surface resistance of 70sm-1 and an albedo of 0.23".

The recharge assessment with EARTH algorithm (van der Lee and Gehrels, 1990) presented in this study requires continuous daily ET_0 (or daily PET) and daily rainfall as input. In this study daily ET_0 was used.

ETo was estimated from time step microclimatological data (relative humidity, wind speed, rainfall and solar radiation) in AWSET program which uses in-built Penman-Monteith (P-M) equation. The standard Penman-Monteith equation (FAO No.56, 1998) for actual evapotranspiration is given by:

$$\lambda ET = \frac{\Delta(R_n - G) + \rho_a c_p \dfrac{(e_s - e_a)}{r_a}}{\Delta + \gamma\left(1 + \dfrac{r_s}{r_a}\right)} \qquad \textbf{5-1}$$

where, λET is the latent heat flux (actual evapotranspiration) in [MJ m^{-2} day^{-1}]

Rn is net radiation at the crop surface [MJ m^{-2} day^{-1}]

G is soil heat flux [MJ m^{-2} day^{-1}]

(e_s-e_a) is air vapour pressure deficit [kPa]

p_a is the mean air density at constant pressure [Kg m^{-3}]

c_p is the specific heat of air [kJ Kg^{-1} $^oC^{-1}$]

Δ is slope vapour pressure curve [kPa $^oC^{-1}$]

γ is the psychrometric constant [kPa $^oC^{-1}$].

r_s is the bulk surface resistance [s m^{-1}]

r_a is the aerodynamic resistance [s m^{-1}]

The FAO Penman-Monteith reference evapotranspiration (ET_o) can be calculated by equation 4.3 as below:

$$ET_o = \frac{0.408\Delta(R_n - G) + \gamma \frac{900}{T+273} u_2 (e_s - e_a)}{\Delta + \gamma(1 + 0.34 u_2)} \quad \textbf{5-2}$$

where ETo is reference evapotranspiration [mm day-1],
Rn is net radiation at the crop surface [MJ m-2 day-1],
G is soil heat flux density [MJ m-2 day-1],
T is mean daily air temperature at 2m height [oC],
u2 is wind speed at 2m height [m s-1],
es is saturation vapour pressure [kPa],
ea is actual vapour pressure [kPa],
es-ea is saturation vapour pressure deficit [kPa],
Δ is slope vapour pressure curve [kPa oC-1],
γ is the psychrometric constant [kPa oC-1].

ET_0 for the study area was calculated by Obakeng (2007) using Penman-Monteith method (FAO No.56, 1998) with meteorological data acquired from automatic data acquisition systems (ADAS) in the area at the stations GS00 – GS10 (see Figure 10), but only up to March 2005. Obakeng concluded that PET in the area is less spatially variable but highly temporally variable. He found that the correlation of meteorological data for all the ADAS stations was high. An earlier report by Wellfield Consulting Services (2000) also showed that there was good correlation between GS00 and Mahalapye station. This is why Malebala ADAS station (GS10), the only functioning after 2005, was considered as representative for the entire study area; so the calculated ET_0 could be used to calculate recharge in that area.

As Malebala (GS10) station has data gaps, these were filled by correlating them with the nearest, long-term, continuous daily data of the Mahalapye microclimatic station located about 100km south-east of Malebala (GS10) station, at latitude 23.08^{o}S and longitude 26.8^{o}E. The Mahalapye data was obtained from the website of NOAA's National Climatic Data Centre (National Climatic Data Centre (NCDC)). This data was available for the period 09/11/2003 to 05/12/2009.

Unfortunately the web NCDC data did not include the solar radiation indispensable for calculation of the FAO ET_0 according to equation 4-3. Instead therefore, the Hargreaves formula that does not require solar radiation but gives reliable estimates in arid and semi-arid conditions (Allen et al., 1998; Allen et al., 2001; FAO No.56, 1998) was used to extrapolate incomplete Mahalaphye P-M ET_0. The Hargraeves ET_0 was calculated according to:

$$ET_o = 0.0023(T_{mean} + 17.8)(T_{\max} - T_{\min})^{0.5} R_a \quad \textbf{5-3}$$

where ET_o is reference evapotranspiration [mm d^{-1}]
R_a extraterrestrial radiation [mm d^{-1}]
T_{max}, T_{min} and T_{mean} are daliy maximum, minimum and mean air temperature [oC], and
T_{mean} is calculated as average of T_{max} and $T_{min.}$

The correlation between Mahalapye Hargreaves ETo and Malebala FAO ET_0 gave a correlation coefficient of 78% (Table 3), and a regression function of 99% (Figure 16). Such moderate correlation between is due to the large distance (~100km) between the two measurement locations and also because of the different calculation algorithms used i.e. FAO and Hargreaves. However, the eventual ET_0 extrapolation uncertainty seems to be of minor importance in the recharge calculation (objective of this study) because the simulated in this study recharge is not highly sensitive to temporal variability of ET_0.

Table 3. Correlation of Mahalapye (NCDC data) and Serowe ADAS stations

	GS00	GS01	GS02	GS03	GS04	GS05	GS06	GS07	GS08	GS10	Mahalapye
GS00	1										
GS01	0.96	1									
GS02	0.95	0.98	1								
GS03	0.95	0.99	1.00	1							
GS04	0.94	0.98	0.96	0.96	1						
GS05	0.96	1.00	0.98	0.98	0.98	1					
GS06	0.96	1.00	0.99	0.99	0.98	1.00	1				
GS07	0.96	1.00	0.99	0.99	0.98	1.00	1.00	1			
GS08	0.96	1.00	0.98	0.98	0.98	1.00	1.00	1.00	1		
GS10	0.80	0.82	0.75	0.77	0.77	0.82	0.81	0.80	0.81	1	
Mahalapye	0.63	0.67	0.61	0.63	0.60	0.67	0.66	0.64	0.67	0.78	1

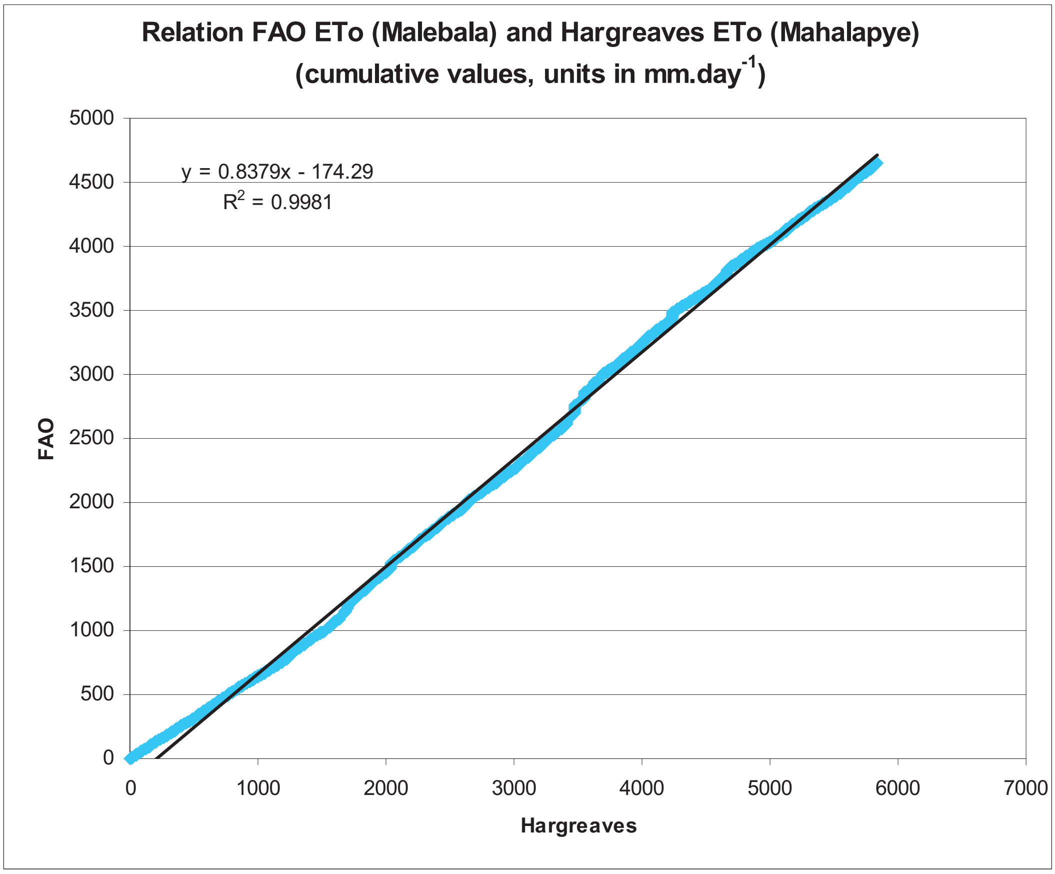


Figure 16: Regression ET_o equation for the Malebala and Mahalapye stations

A graph of the estimated reference evapotranspiration (ET_0) for the study area is presented below in Figure 17. From this graph it is clear that reference evapotranspiration obtained by using the Hargreaves method (equation 4-3) is higher than that obtained by the FAO Penman-Monteith method. This is in line with results from earlier works from other areas such as Gavilan et al.,(2006). who did a calibration of the Hargreaves equation in Spain.

Also from Figure 17 it can be seen that ETo has maximum peak during the local summer season (October/March) which is the rainy season and temperatures are at a maximum. ETo is low around April to March which is the dry winter period, characterised by low temperatures; meaning less energy and water for evapotranspiration as opposed to the summer period.

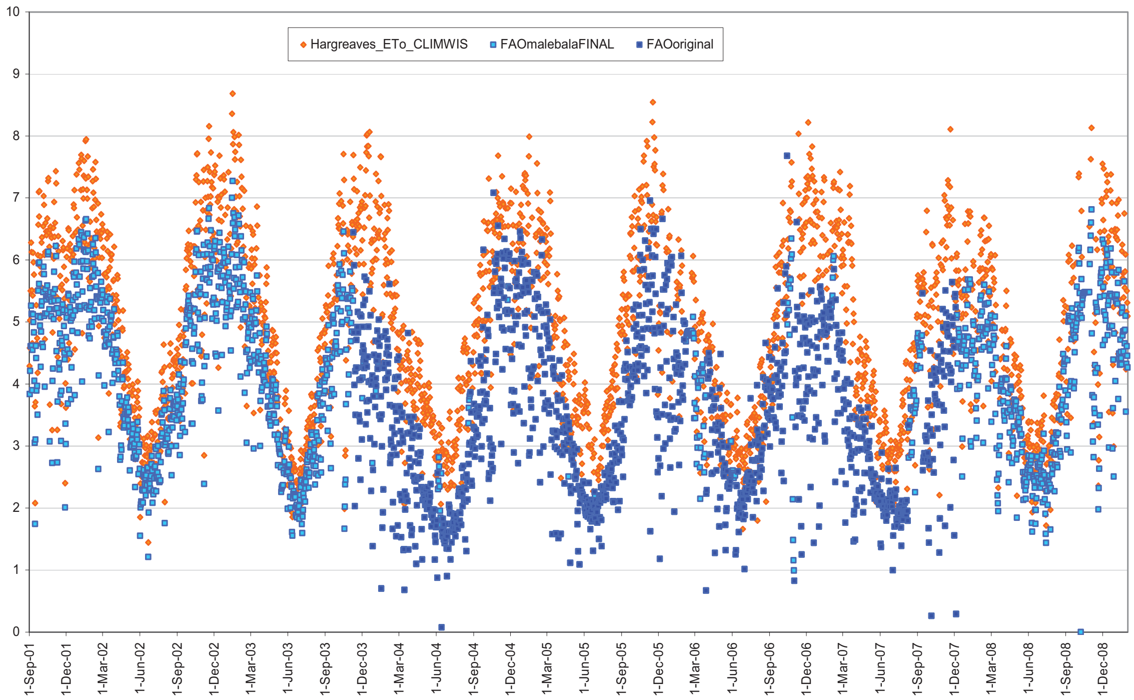


Figure 17: Daily reference evapotranspiration for Mahalapye and Malebala

6. ASSESSMENT OF NET GROUNDWATER RECHARGE

6.1. General overview

Groundwater recharge (R) is the rate at which water in an aquifer is replenished. A number of methods are available for quantifying groundwater recharge. Each of the method has its own limitations in terms of applicability and reliability. The objective of the recharge study should be known prior to selection of the appropriate method for quantifying groundwater recharge as this may dictate the required space and time scales of the recharge estimates (Xu and Beekman, 2003). Water resource evaluations for instance would require information on recharge at large spatial and temporal scales whereas assessments of aquifer vulnerability to pollution would require more detailed information at local and shorter time scales.

In semi-arid areas like the Serowe area, it is difficult to estimate evapotranspiration (ET) since recharge in those areas is usually very low (few millimetres per year). It is therefore not suitable to estimate recharge by conventional methods of subtracting ET from rainfall as recharge is usually smaller than the errors in rainfall and evapotranspiration (Obakeng, 2007). Another important aspect is to distinguish between recharge (R) and net recharge (Rn), also referred to as effective recharge (Re). The net recharge is defined as the water entering a groundwater body after infiltration and percolation through the unsaturated zone minus the amount that is being extracted from the aquifer by groundwater evapotranspiration, ETg (equation 5-1).

$$Rn = R - ETg \qquad 6\text{-}1$$

Groundwater evapotranspiration (ETg) consists of two components (Lubczynski 2009): groundwater evaporation (Tg) and groundwater transpiration (Tg).

$$ETg = Eg + Tg$$

In the Kalahari hardveld (Fig.10) the water table is close to the surface and there are lots of trees so in this part of the study area during the dry seasons, water is lost by evaporation from the groundwater (Eg), as well as by groundwater transpiration (Tg) from the shallow water table so the Rn is equal to:

$$Rn = R - (Eg + Tg)$$

In the Kalahari sandveld (Fig.10) groundwater table is very deep >60m groundwater table, which is too deep for the evaporation from water table (Eg = 0) but still accessible by certain tree root systems like *Boscia albitrunca* or *Acacia erioloba* as stated by Obakeng (2007). Therefore in the sandveld area the Rn is equal to

$$Rn = R - Tg \qquad \textbf{6-2}$$

6-3

The spatial and temporal groundwater recharge of the Serowe study area is determined by using two methods: chloride mass balance method and EARTH 1-D modelling.

6.2. Chloride Mass Balance Method

6.2.1. Theoretical Background

Chloride mass balance (CMB) is an indirect method of recharge estimation based on the law of mass conservation and steady state chloride mass balance at the water table interface (Selaolo, 1998).
Chloride is the most important environmental isotope used for regional groundwater recharge estimation (Simmers et al., 1997). The CMB method can be used under a wide range of climatic, soil and geologic conditions. The basic equation for recharge calculation with the chloride mass balance method is given by equation 5-4 below:

$$R_T = \frac{(P \times Cl_p) + D}{\overline{Cl_{gw}}} \qquad \textbf{6-4}$$

where: R_T is areal recharge (mm/yr)
P is average annual precipitation (mm/yr)
Cl_p is chloride content in precipitation (mg/l)
D is dry deposition of chloride measured during the dry season ($mgm^{-2}yr^{-1}$)
$\overline{Cl_{gw}}$ is harmonic mean of chloride concentrations in groundwater (mg/l), and should be calculated as shown by equation 6.5 below :

$$\overline{Cl_{gw}} = \frac{N}{\sum_{i=1}^{N} \frac{1}{Cl_{igw}}}$$ **Equation 6-5**

The numerator of the above equation (i.e. P x Clp + D) is termed the total chloride deposition rate (TD), which assumes steady-state surface chloride flux (Selaolo, 1998).
The chloride mass balance method makes the following assumptions:

- Chloride is conservative, and as such will not react with the aquifer material, and will also not be leached or taken up by vegetation,
- The source of chloride in groundwater is from precipitation only.

This technique is excellent in providing minimum estimates of recharge but addition of solutes from various non-atmospheric sources may take place, limiting its usefulness. CMB may be especially useful in areas where groundwater levels do not fluctuate or data on groundwater levels are lacking (Xu and Beekman, 2003). Despite these shortcomings, the CMB method is highly recommended for semi-arid regions as well as in fractured rock systems. It is a relatively simple and least expensive method. The method is most attractive in areas with high evapotranspiration, where the infiltrating water gets highly concentrated (Simmers et al., 1997). The CMB method should not be applied in areas underlain by evaporites or areas where up-coning or mixing of saline (ground) waters occurs. The method should be applied with great caution in areas close to the sea where rainfall chloride contents are highly variable.

This study did not have rainfall samples to measure chloride concentration in rain water; and also no data on dry deposition of chloride. Consequently, TD as determined by previous studies has been used to calculate areal recharge of Serowe aquifer. Gieske (1992) recommended TD values of between 400 and 500mgm^{-2}yr^{-1} after doing a research on the dynamics of groundwater recharge in eastern Botswana.

Table 4. Total chloride deposition rates from rain gauge and totaliser networks(values from (Selaolo, 1998))

Station	Period (Sept–Sept)	Raingauge (mgm^{-2}yr^{-1})	Totaliser Network (mgm^{-2}yr^{-1})
Serowe	1986 to1993	442 ± 124	629

(

6.2.2. CMB Results

A total of 19 boreholes were sampled during the fieldwork campaign. But due to the vastness of the investigation area these samples were combined with existing chloride records of the area to get a fairer distribution of groundwater chloride of Serowe area. This was done after analysing temporal variability of groundwater chloride concentrations in the area which was found to be low (**Figure 18**). From this figure it is evident that the chloride concentrations of most boreholes from 1989 to 2008 do not show any significant variability with time. Hence measurements used for spatial interpolation of chloride are average chloride concentrations at each borehole for 126 boreholes in the study area as they are more representative of the area and not a particular year.

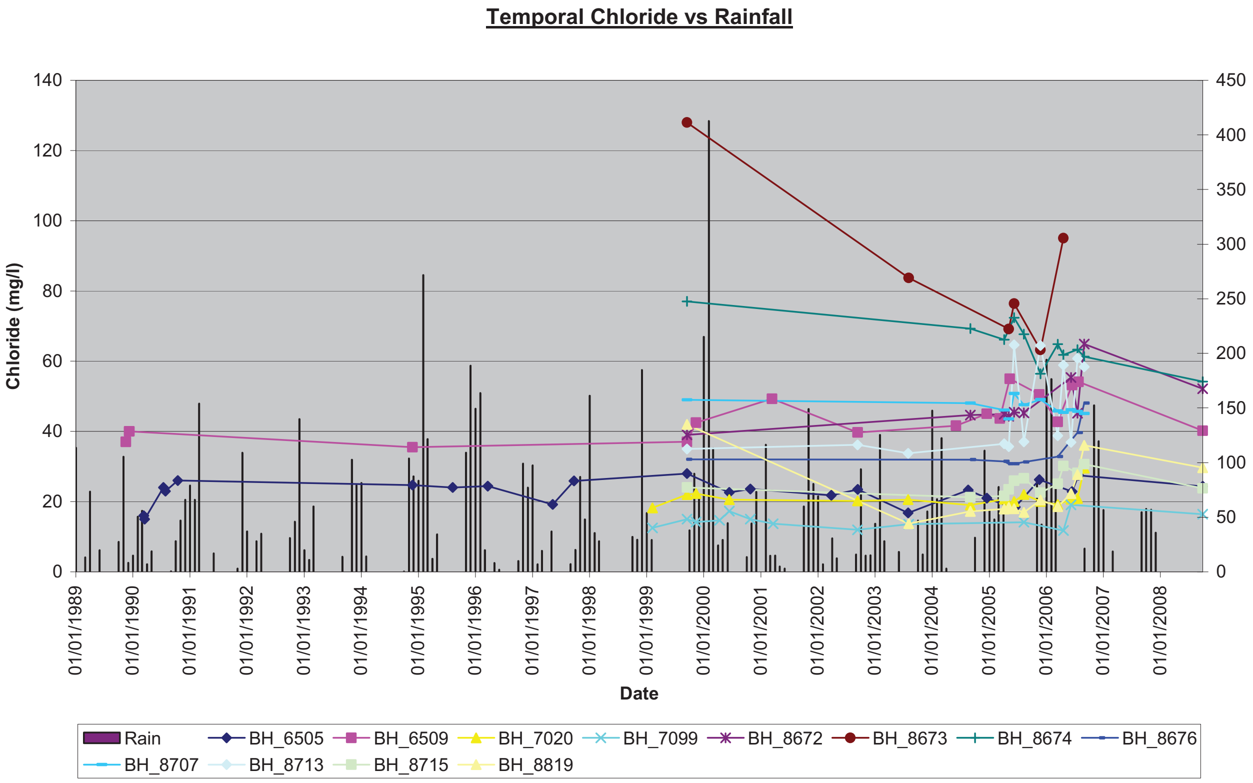
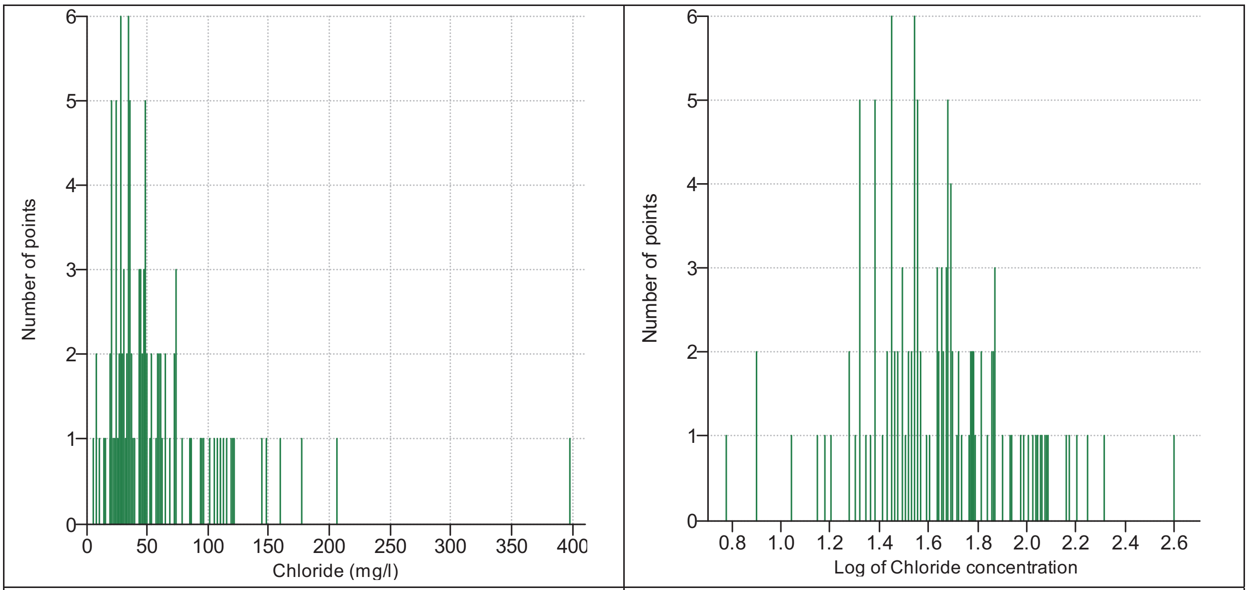


Figure 18: Temporal variability of chloride, Serowe aquifer.

A spatial distribution of chloride was obtained by geostatstical interpolation using kriging method. In kriging, a semi-variogram model has to be chosen for as long as there is normal distribution of points, otherwise a logarithm of these points has to be used for determining the semi-variogram model. A histogram plot of the interpolation points shows that the chloride values are not normally distributed but right-skewed, Figure 19 (a). However, a logarithm of the chloride values gives a normally distributed histogram Figure 21 (b). Therefore a log of the chloride values was used to generate a semi-variogram model.

The interpolated groundwater chloride distribution map with points used for interpolation superimposed on it is shown in Figure 21.



(a) Right-skewed | (b) Normal distribution

Figure 19: Histogram of chloride concentration

An experimental semi-variogram was obtained by using a lag spacing of 2000m. Then a spherical model with a nugget of 0.045, sill of 0.09 and range of 20000m was fitted into the experimental semi-variogram as shown in

Figure 20 below.

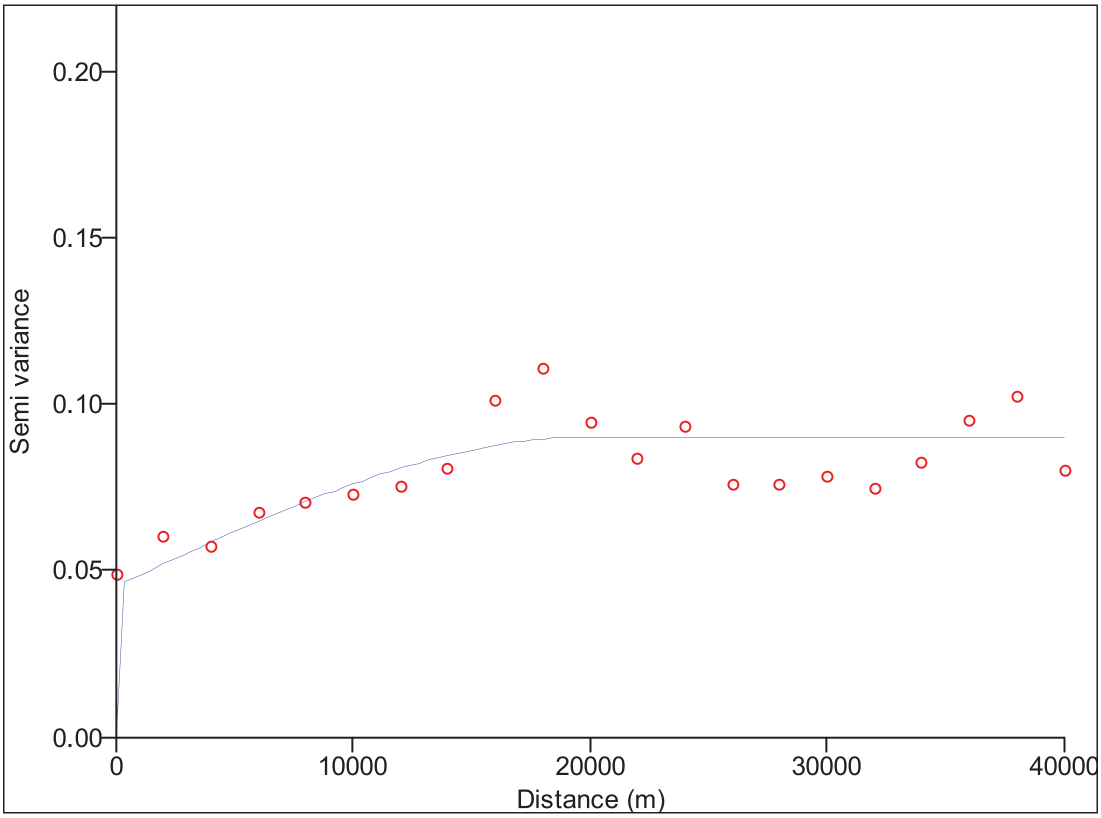


Figure 20: Semi-variogram of chloride distribution in groundwater

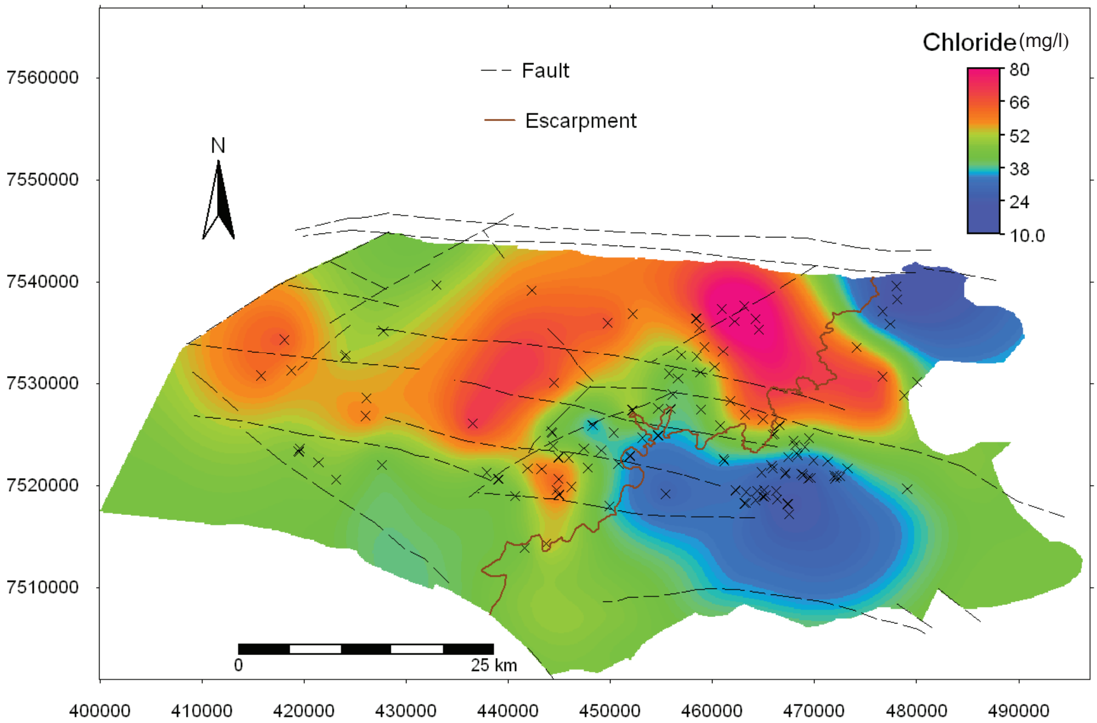


Figure 21: Spatial distribution of chloride concentration in groundwater

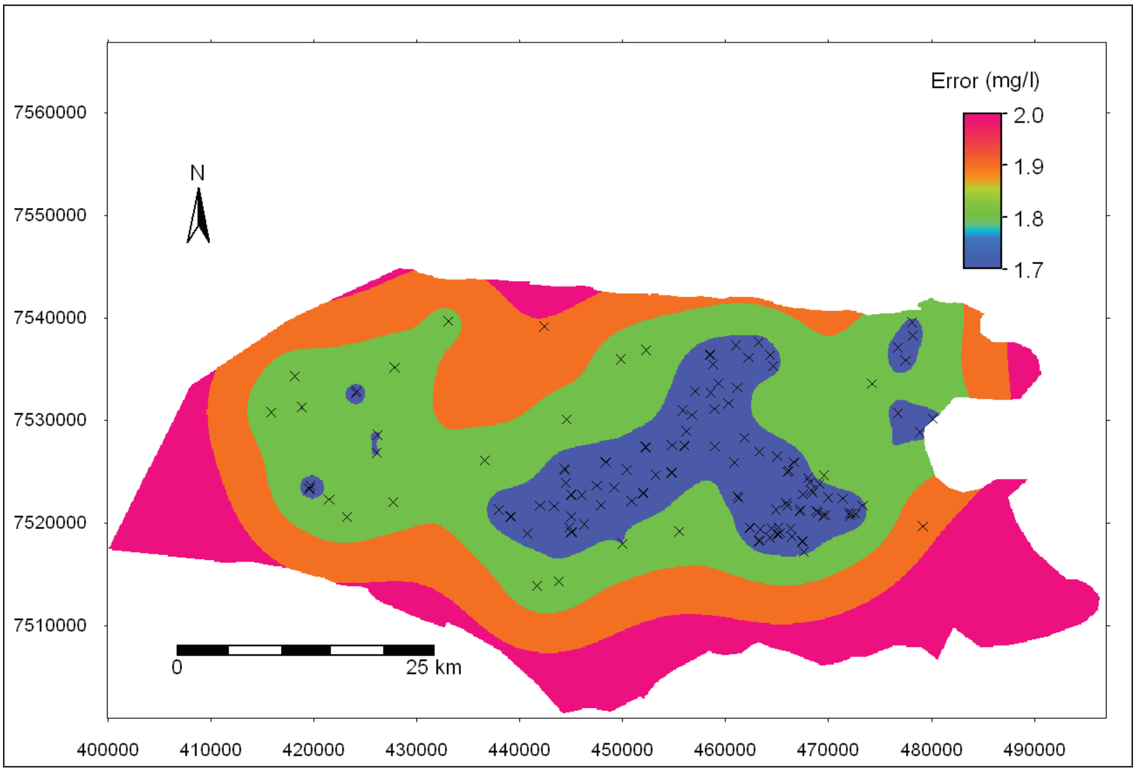


Figure 22: Error map of distribution of chloride concentration in groundwater

From Figure 21 it is evident that chloride values are generally low in the central part of the study area around the escarpment and eastern hardveld. This coincides with thick, highly faulted/fractured basalt confining the sandstone aquifer. The aquifer is probably being recharged by preferential flow of water through these fractures and faults. Lower chloride concentrations towards the east of the escarpment could be indicating active recharge due to absence of Kalahari sand cover, thin or no basalt and outcropping of the aquifer. There is no data farther east and south of the study area, hence the high error as shown by the interpolation error map Figure 22.

High chloride concentrations in the west, central and northern parts could be attributed to thick Kalahari sand cover as the area has thin to no basalt cover. Hence very little recharge reaches the aquifer as the erratic precipitation is lost to evapotranspiration. Groundwater in this portion of the area is relatively older than that in the east.

The point recharge values obtained by the chloride mass balance equation above were interpolated by kriging using ILWIS software application. The calculated recharge values are presented in Table --- (Appendix----), whereas Figure 23 below shows the interpolated spatial groundwater recharge of the area. It is evident that recharge is generally lower where the sandstone aquifer is confined by basalt and/or covered by thick Kalahari sand. As such recharge is generally low in the sandveld due to thick Kalahari sand cover. It seems that faults have no significant contribution to groundwater recharge in this part of the study area as they are deep-seated. In the eastern hardveld part of the study area,, the recharge values (Figure 23) are relatively higher since the aquifer is mostly outcropping. Low recharge in the north-east could be linked to high evapotranspiration from the shallow sandstone aquifer and/or confinement due to unfractured and unfaulted basalt.

Using the minimum and maximum TD of 442 and 629mg.m^{-2}.yr^{-1} gives minimum and maximum average spatial net recharge of Serowe aquifer as 12.6mm/yr and 17.9mm/yr. The long term mean annual rainfall of Serowe is 437mm/yr, (Obakeng, 2007), therefore groundwater recharge accounts for about 2.9% to 4.1% of the mean annual rainfall.

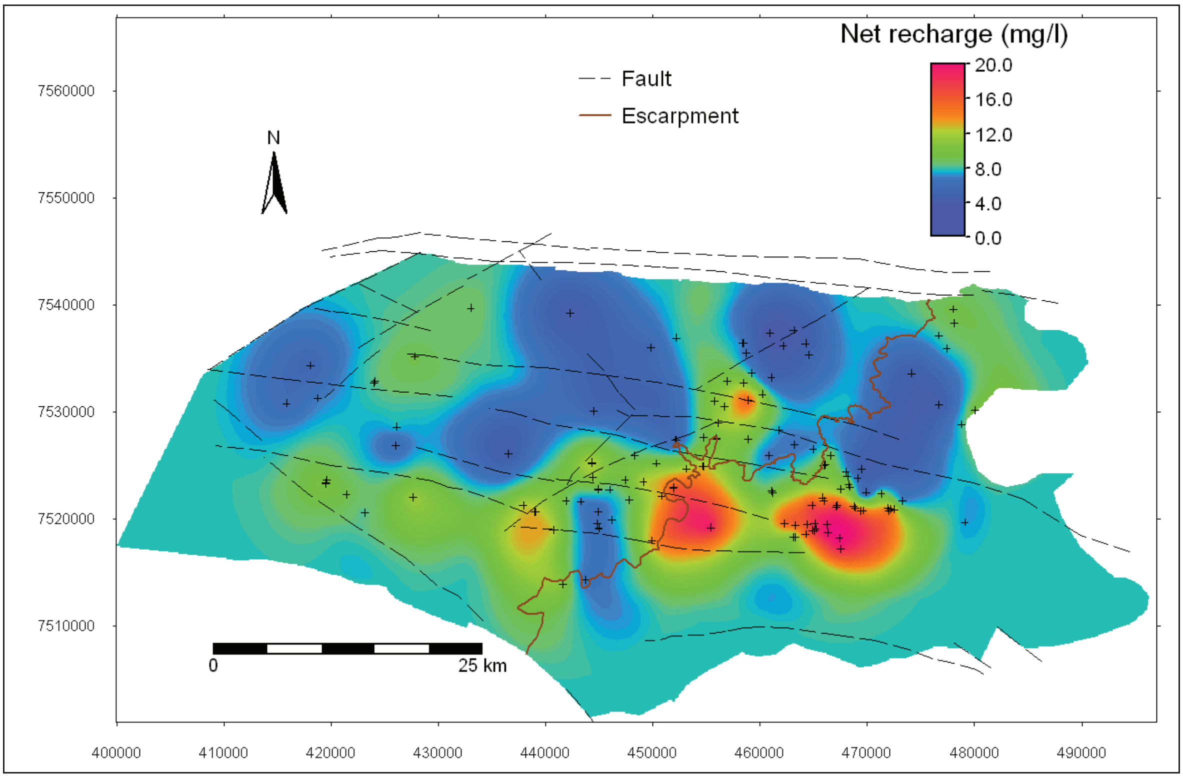


Figure 23: Recharge distribution from kriging of point recharge values

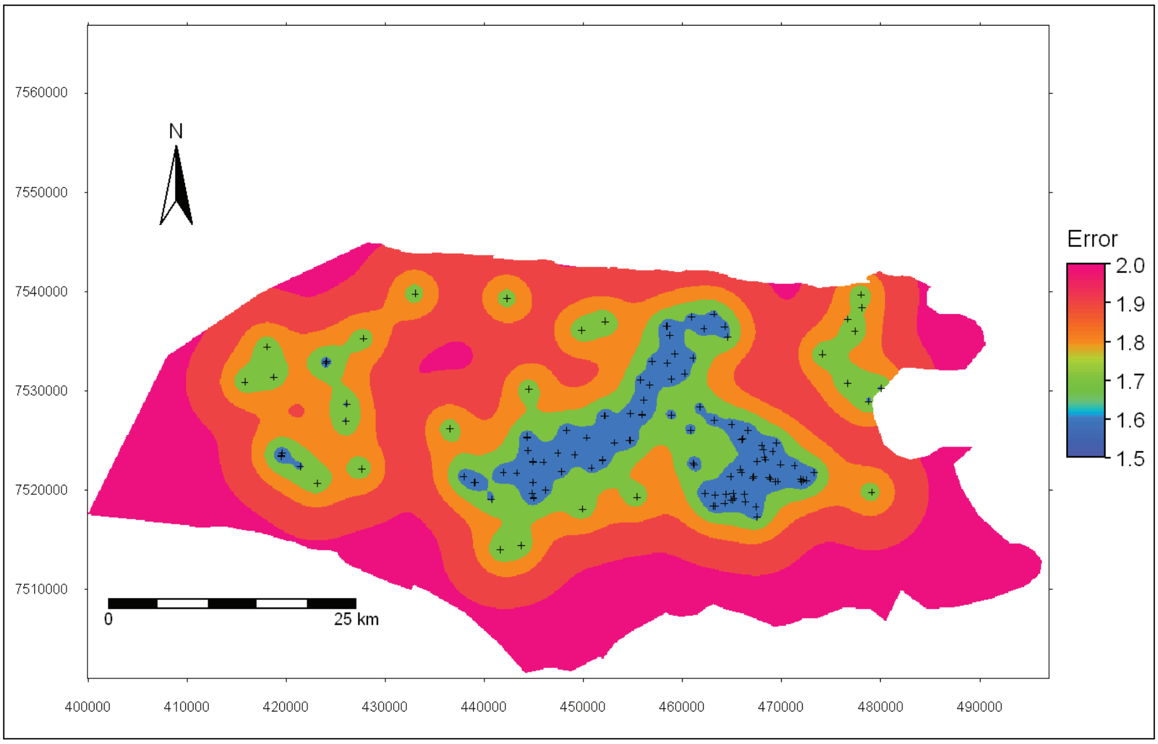


Figure 24: Error of net groundwater recharge

6.3. EARTH 1D Model

6.3.1. Theoretical Background

The Extended model for Aquifer Recharge and soil moisture Transport through the unsaturated Hard rock (EARTH) is a 1D lumped parameter model for the simulation of recharge and groundwater level fluctuations (van der Lee and Gehrels, 1990). The model uses both the direct and indirect method of recharge calculation. The model has five different modules Figure 25. The first three modules, MAXIL, SOMOS and LINRES estimate recharge by direct methods. The indirect part of the model is the SATFLOW module, which calculates groundwater level using recharge estimated from the direct part. The model was originally developed for semi-arid climate of Botswana. The input data for the model are daily rainfall, daily potential evapotranspiration, and 11 other soil and groundwater related parameters for the simulation of net recharge and groundwater level fluctuations. The observed groundwater levels and soil moisture measurements are used to calibrate the model. The output of the model is daily groundwater levels, net recharge, actual evapotranspiration, percolation, ponding and surface runoff.

The individual modules of EARTH 1-D model are presented and described below. For a very detailed description of the modules, the reader is referred to van der Lee and Gehrels, (1990).

MAXIL (MAXimum Interception Loss): Here the precipitation excess (Pe) or effective rainfall, which is the amount of water that actually infiltrates, is estimated by taking into account the fraction of the original precipitation that is retained by leaves and stems of the vegetation or as depression storage. The effective rainfall (P_e) is then calculated by the formula:

$$P_e = P - MAXIL - E_0$$

where P is precipitation (mm)

$MAXIL$ is the intercepted fraction of P (mm)

E_0 is evaporation from ponding surface storage (mm).

SOMOS (SOil MOisture Storage): The change in soil moisture storage is estimated by distributing the precipitation excess into actual evapotranspiration, percolation, ponding and/or runoff

This change in soil moisture storage is given by the mass balance equation () below:

$$\frac{dS}{dt} = P_e - ET_a - R_p - (SUST + Q_s)$$

where $\frac{dS}{dt}$ is change in soil moisture storage [mm]

ET_a is actual evapotranspiration [mm],

R_p is the percolation [mm]

$SUST$ is ponding water [mm]

Q_s is runoff [mm]

The product of the volumetriv soil moisture conten and the thickness of the zone where soil moisture changes occur is termed the soil moisture S [mm].

ET_a is determined by assuming linear relations between S and potential evapotranspiration (PET), actual voulmetric soil moisture (θ), porosity (ϕ) and the permanent wilting point (θpwp).ETa is given by:

$$ET_a = PET \frac{(\theta - \theta_{pwp})}{(\phi - \theta_{pwp})}$$

Percolation, Rp, is derived by the Darcy equation as:

$$R_p = K\left|\frac{dh_p}{dz} + 1\right| \approx K_s\left|\frac{\theta - \theta f_c}{\phi - \theta fc}\right|$$

where K is unsaturated hydraulic conductivity [mm d^{-1}]

$\frac{dh_p}{dz}$ is the hydraulic head gradient,

Ks is saturated hydraulic conductivity [mm/day]

θ is the actual volumetric soil moisture

θf_c is the soil moisture at field capacity

ϕ is the porosity Below the root zone, soil water movement in unsaturated zone is mainly determined by gravity and capillary gradient is less influenced i.e. term $K.dh_p/dz$ is negligible compared to gravitational component. Therefore it is assumed the pressure head is constant with depth. The percolation in equation (5.4) reduces and equal to unsaturated hydraulic conductivity. Ks is saturated hydraulic conductivity, θfc is soil moisture at field capacity and other terms are defined as above. The water fraction leaves from SOMOS is assumed to equal of recharge amount. But this groundwater recharge is delayed due to transfer through VADOS zone.

If infiltration rate exceeds the percolation when soil in SOMOS is at saturation, the surface ponding is appears. Surface runoff occurs when ponding water increases beyond the threshold of $SUST_{max}$. The equations for these conditions are

$$\frac{d(SUST)}{dt} = P_e - ET_p - R_p - E_o$$

$$Qs = SUST - SUST_{\max}$$

Where E_0 is open water evaporation (mm), $SUST_{max}$ is maximum surface storage capacity and Qs is runoff.

SUST (SUrface STorage): This module calculates the amount of water that accumulates (ponding) at the surface when maximum soil moisture is reached and the percolation rate is lower than infiltration rate such that surface runoff takes place. The equation below then applies:

$$\frac{d(SUST)}{dt} = P_e - ET_a - R_p - E_o$$

Where SUST is surface ponding [mm]

E_o is open water evaporation [mm]

When ponding water exceeds maximum surface storage ($SUST_{max}$) then runoff (Q_s) occurs. This is represented by :

$$Q_s = SUST - SUST_{\max}$$

LINRES (LINear REServoir routing): Percolation (Rp) from SOMOS is a function of time and depth.to groundwater table. The time that the percolating water takes to reach the groundwater table is calculated in this module by the equation:

$$R = Y_n = \frac{f}{(1+f)} \sum_{i=0}^{n} (1+f)^{-i} \; Y^*{}_{n-i}$$

$$Y_o = \frac{(1 + f)}{f} R_p$$

where, R is recharge [mm d-1],
f is unsaturated recession constant,
n is the number of reservoirs,
Y^* is the result from previous time step,
R_p is percolation [mm]
Y_0 is upper boundary condition.

SATFLOW (SATurated FLOW): This module simulates groundwater level rise by using the recharge calculated from the LINRES module. When there is no recharge the groundwater level is simulated from an exponential drop in water table by using the recession constantThe equation used is:

$$\frac{dh}{dt} = \frac{R}{STO} - \frac{h}{RC}$$

Where dh/dt is change in groundwater level over time [m d^{-1}]
R is recharge
STO is storage coefficient
h is groundwater level (m) above local base level
RC is the saturated recession constant (days

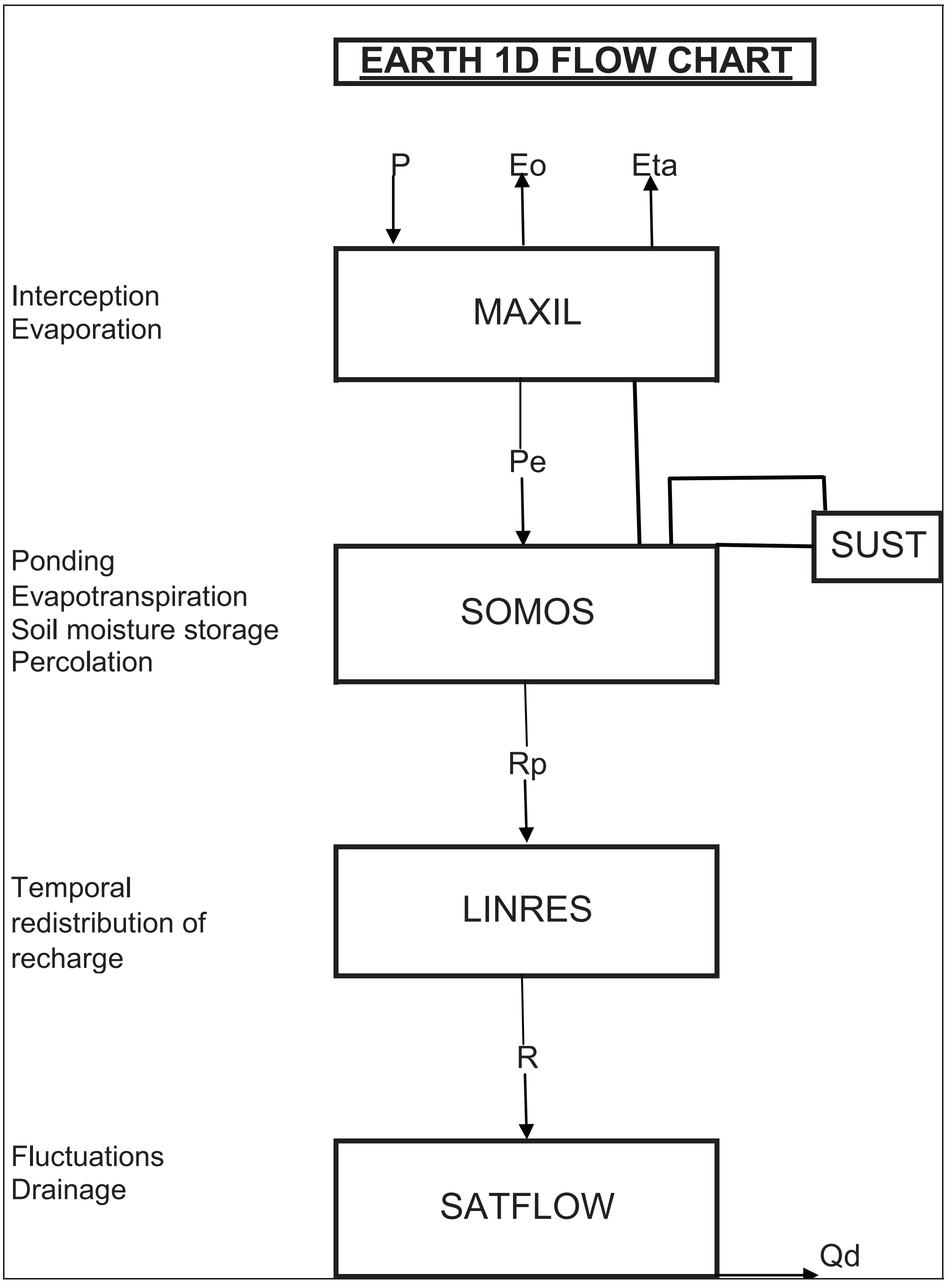


Figure 25: Flow chart of EARTH 1D model. Source: (van der Lee and Gehrels, 1990)

6.3.2. Modelling results

Model calibration aims at obtaining the best fit between measured and predicted variables. This is achieved by reducing the calibration error as low as it is practically possible. A well calibrated model can then be used to predict or simulate reality where there is lack of data. Different state variables may be used to calibrate a model. In this study groundwater levels were used to calibrate EARTH model.

The model calibration covered the period 1st September 2001 to 1st October 2008 using groundwater levels from four boreholes spread across the study area as shown in Figure 10. The boreholes used for model calibration are BH4139, BH5305, BH5306 and BH8493. Model calibration was done by trial and error method, changing one parameter at a time, beginning by paying particular attention to those parameters that cause greater deviations.

The groundwater levels that were calibrated were acquired as part of the continuous groundwater monitoring programme of the Serowe aquifer. Apart from groundwater levels, the other important input data needed for EARTH modelling are daily rainfall and potential evapotranspiration. In this study reference evapotranspiration was derived from the FAO Penman- Monteith (chapter 5). Rainfall and other meteorological data needed for calculation of reference evapotranspiration was obtained from Malebala (GS10) automatic data acquisition system (ADAS) as explained in chapter 5.

The main parameters that were found to affect the model calibration are soil moisture at field capacity, saturated and unsaturated recession constant, and the number of reservoirs. The local base level and initial base level were determined from groundwater level hydrographs. The rest of the model parameters were fixed at the beginning but later fine-tuned during model calibration as and when necessary. The model parameters used for the calibration are presented in Table 5. EARTH 1-D calibration input data below, whereas the calibrated EARTH model recharge of the boreholes is presented in Table 1.

Table 5. EARTH 1-D calibration input data

Name in EARTH	Parameter name in pyEARTH	Calibrated value			
		BH4139	BH5305	BH5306	BH8493
MAXIL	Maximum interception loss	2.4	2.4	2.4	2.4
Smax	Max soil moisture	28.0%	27.5%	28.5%	28.0%
Sfc	Soil moisture at field capacity	27.0%	26.5%	27.0%	27.0%
Sr	Residual soil moisture	16.0%	16.0%	16.0%	16.0%
Si	Initial soil moisture	27.5%	20.0%	28.0%	27.5%
D	Depth root zone	1000	1000	1000	1000
Ks	Saturated conductivity	500	500	500	500
SUSTm	Maximum surface storage	0	0	0	0
n	Number of reservoirs	2	2	2	2
f	Unsaturated recession constant	20	20	20	30
RC	Saturated recession constant	700	400	350	2500
STO	Storage coefficient	0.05	0.05	0.05	0.05
hi	Initial GWL	2.25	0.35	0.15	3.65
h0	Local base level	1069.40	1103.35	1116.90	1172.50
----	Piezometer Elevation (m)	1099.66	1174.90	1148.06	1240.20

Table 6. EARTH model recharge results

BH	R total (% of rainfall)	R total (mm)	R daily (mm/day)	RF total (mm)	RMSE (mm)
BH4139	2.0%	47	0.018	2322.8	0.217
BH5305	2.3%	53.1	0.021		0.098
BH5306	3.0%	70.7	0.027		0.1
BH8493	2.0%	47	0.018		0.077
Simulations between 2001-09-01 and 2008-10-01					

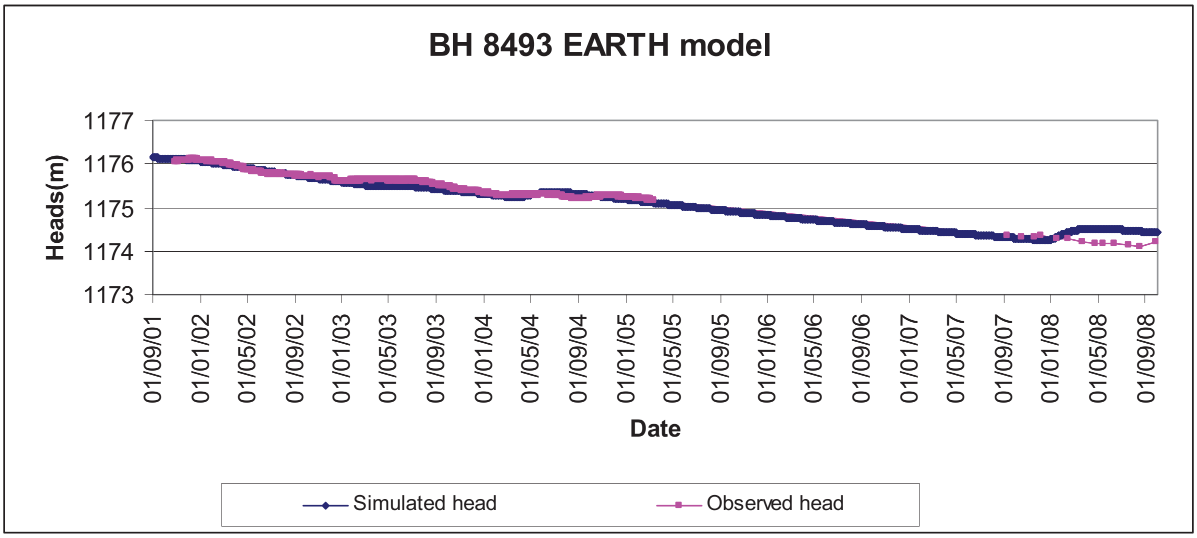


Figure 26: BH8493 EARTH model simulation

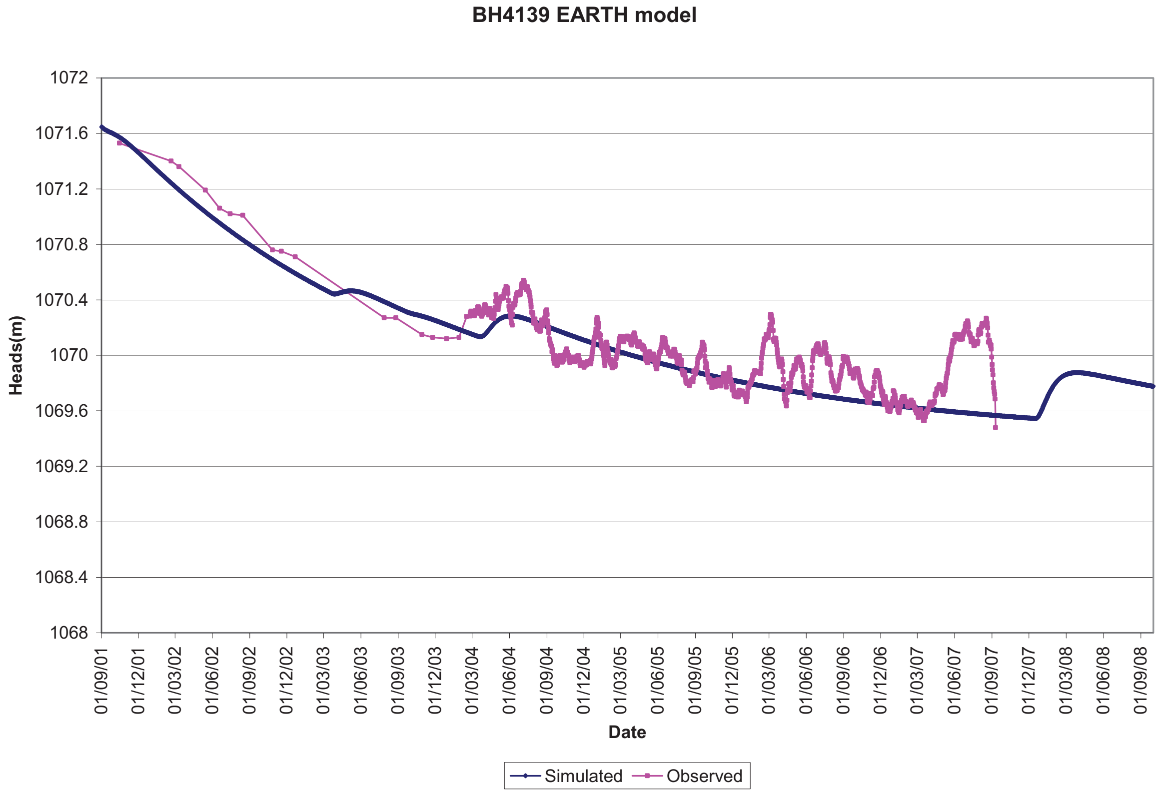


Figure 27: BH4139 EARTH model simulation

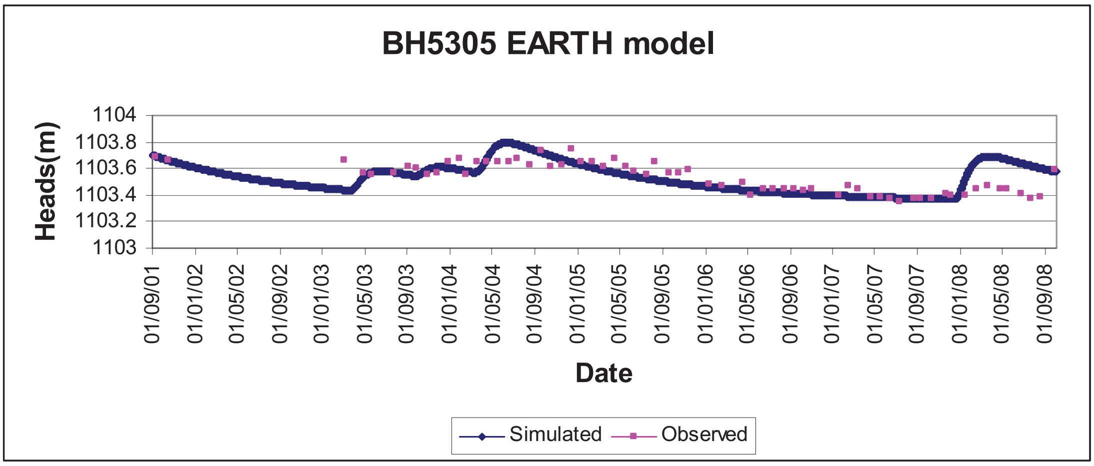


Figure 28: BH5305 EARTH model simulation

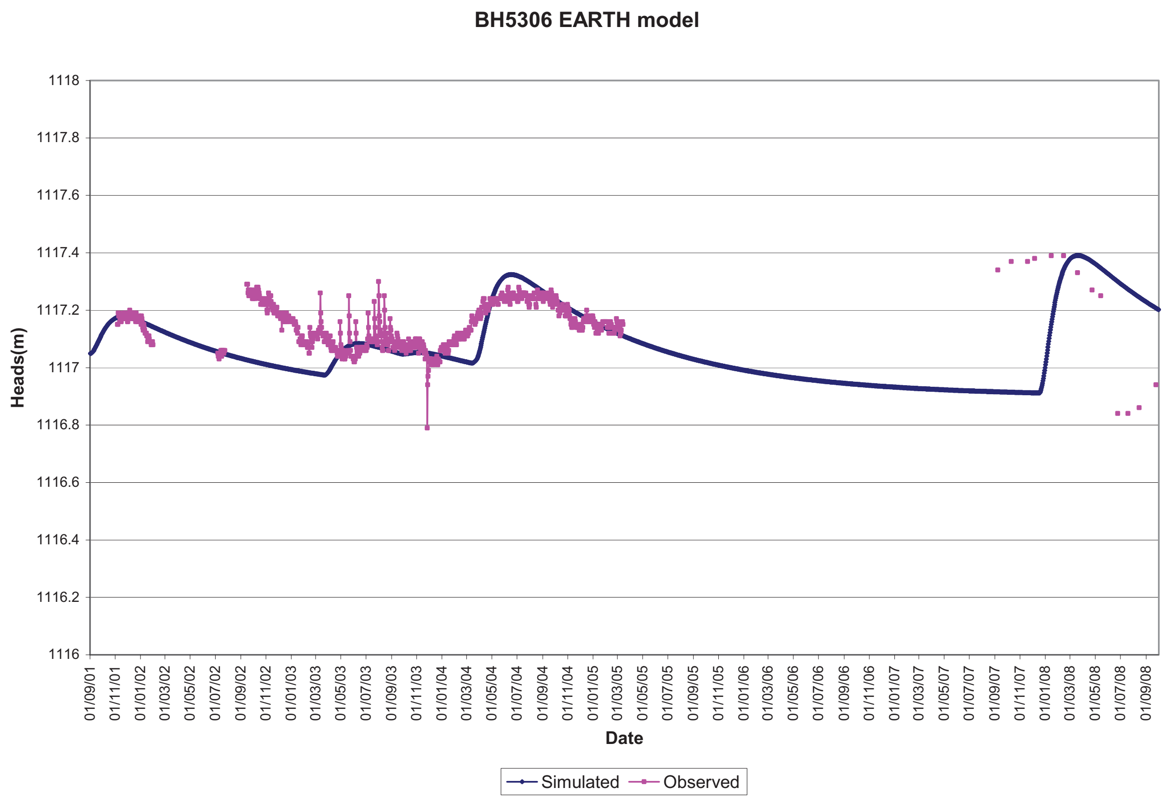


Figure 29: BH5306 EARTH model simulation

7. DISCUSSION AND CONCLUSION

The objective of assessing groundwater recharge of the Serowe aquifer was achieved by using two recharge assessment methods, namely chloride mass balance (CMB) and EARTH 1-D modelling. These methods were selected basing on their applicability to the semi-arid climate of the study area, data availability and the aim of the research work.

The net groundwater recharge of Serowe aquifer as calculated by the CMB method gives an average value of 12.6mm y^{-1}, when a total chloride deposition rate of 442mg m^{-2} yr^{-1} is applied.Using the minimum and maximum TD of 442 and 629mg.m^{-2}.yr^{-1} gives minimum and maximum average spatial net recharge of Serowe aquifer as 12.6mm/yr and 17.9mm/yr. The long term mean annual rainfall of Serowe is 437mm/yr. Therefore, net groundwater recharge of Serowe aquifer is 2.9% of mean annual rainfall when using total chloride deposition rate of 442 mgm^{-2} yr. A high recharge value of 4.1% is obtained when using total deposition rate of 629 mg $^{-2}yr^{-1}$.

EARTH 1-D modelling showed that groundwater recharge does not only vary in space, but also in time. Some rainfall events do not translate into recharge to the aquifer. Recharge occurs only in times of exceptionally high rainfall and does not result in immediate rise in groundwater levels. Recharge is normally delayed by some weeks to months with respect to rainfall and it ranges from 2 to 3% of annual rainfall as calculated from the four boreholes calibrated.

The CMB method showed that chloride concentration varies in space. Several factors can be attributed to the variability of recharge in space. These are geology and structure, topography, soil type, vegetation and land use. The data analysis shows that net groundwater recharge in the study area is high where the sandstone aquifer is at shallow depth and unconfined. High recharge is also evident at topographically high areas around the escarpment and in areas characterised by high faulting/fracturing which induces preferential recharge.

Low recharge values generally occur in the areas where the Kalahari sand is thicker and/or the aquifer is overlain by fresh Stormberg basalt layer.

Comparison of recharge by CMB and that obtained from EARTH 1-D shows that the CMB gives slightly higher recvhagre than EARTH model. However, their differences are not so big. Recharge ranges from 2 to 3% by EARTH while it is 2.9% to 4.1% by CMB. Therefore, the two methods can be used together to compliment each other.

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8. Recommendations

The current monitoring programme of the aquifer is doing fine, but there is still room for improvement. It is recommended that there be a proper and a well co-ordinated monitoring of aquifer parameters such as water levels, abstraction records and water chemistry. These are very critical for well field and aquifer management as their proper recording is very useful for recharge estimation and wellfield management.

Currently there are a number of automatic water level recorders installed in the area, but most of them are currently not operational due to lack of proper maintenance and service. At the time of writing, all the microclimtic weather stations are not working, and all but one have not been working since 2005. Therefore there is need to review the current operating and maintenance schedule. This could save the government lots of money in the long run.

The available records show that there is a lot of data that has been collected in the area that can be used for doing more reliable numerical groundwater models. Unfortunately this data is not well archived, and the instruments are not well looked after. Moreover, this data needs to be under the responsibility of someone who understands the usefulness of collecting and archiving monitoring data such as a hydrogeologist.

Effort must be made to retrieve data from the many malfunctioning data loggers. These could be shipped back to the manufacturer who might with some luck manage to retrieve such very important and crucial data.

There should be further detailed study of vegetation and soils of the area to see what their contribution is to groundwater recharge of Serowe.
Monitoring of water chemistry in the study area needs improvement as temporal chloride concentrations have shown that

Finally, every effort must be made to do a fully-transient state numerical model of the aquifer. With a bit of extra effort a fully transient model could be done as there is already some data to build on.

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