GROUNDWATER AND LAKE WATER BALANCE OF LAKE NAIVASHA USING 3-D TRANSIENT GROUNDWATER MODEL

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Thesis submitted to the Faculty of Geo-Information Science and Earth Observation of the University of Twente in partial fulfillment of the requirements for the degree of Master of Science in Geo-information Science and Earth Observation.
Specialization: Water Resource and Environmental Management

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ABSTRACT

Integrated water resources management is necessary, particularly in a system where considerable interactions exist between ground and surface water resources. Integrated study requires reliable estimation on an overall basin water budget and reliable estimates on hydrologic fluctuations between ground water and surface water resources. The objective of this study is to construct and calibrate a 3-D transient groundwater model that simulates the long term groundwater and lake water balance of the Lake Naivasha basin and that could be utilized to evaluate the effects of changes in system flux over time.

Methodological design of this study starts with a field work Geodetic-GPS survey program in order to accurately measure height of the groundwater level and surface water levels. The data analysis involved a separate characterization of both surface and subsurface parameter. Time series data including lake level, surface water-inflow, evaporation and precipitation were analyzed on a monthly basis. Pump test data was analysed for recently drilled boreholes. Recharge was estimated by relating monthly change in groundwater level and average recharge measured in the area. Water abstraction data mainly from the irrigated commercial farms was analysed based on the irrigation area-depth relationship.

The model developed using GMS software, covers an area of 1817 sq. km with two aquifer systems. The upper aquifer is unconfined and the lower aquifer is semi-confined. The upper aquifer is in hydraulic link with the lake. The model grid contains 104 rows, 120 columns with a uniformly horizontal spacing equal to 500 m. The lake bathymetry was represented by the lake package Triangular Networks (TIN) of GMS functionality. The model design spans over 79 years (1932-2010) with a total of 942 stress periods and a single time step. In the modeling process the applications of the conceptual model approach of Modflow and Lake Package LAK3 was extensively explored.

Model calibration was highly constrained by observing the measured and calculated aquifer and Lake Level. The final calibrated model, implements the application of parameter estimation tools, PEST. The model matches the observed lake level with R^2= 0.985, steady state and R^2= 0.905, transient state. Model sensitivity analysis result shows that the steady state model is highly sensitive to increasing and decreasing of recharge and highly sensitive to a decreasing than increasing in hydraulic conductivity. The transient model shows equal sensitivity with increasing and decreasing of the storativity but with a slow response.

The long term lake water balance is calculated by Modflow using the stage-volume rating curve of Lake Package LAK3. The long term average storage volume is 8.4 * 10^8 m^3/month. The long term average fluxes in to the lake are precipitation 7.72 *10^6 m^3/month, surface inflow 19.36*10^6 m^3/month and groundwater inflow (Lake seepage-in) 1.1*10^6 m^3/month. The long term average fluxes out of the lake are evaporation 21.41*10^6 m^3/month, lake water abstraction 1.92 *10^6 m^3/month (equivalent to 5*10^6 m^3/month over the past 30 years) and groundwater outflow (Lake seepage-out) 5.5*10^6 m^3/month. The lake water balances suggests that the lake is not in equilibrium with the inflow and outflow terms, a long term net lake level fall of 5.4m resulted in a lake storage loss of 6.73 * 10^8 m^3 over the period, 1932-2010.

A long term groundwater budget is calculated reflecting all water flow in to and out of the regional aquifer. The inflow components include recharge 2.8*10^6 m^3/month, river leakage-in 1.4*10^5 m^3/month and Lake Seepage-in (groundwater outflow from the lake) 5.56*10^6 m^3/month. The outflow components include well abstraction 7.5*10^5 m^3/month (equivalent to 2*10^6 m^3/month over the past 30 years), river leakage-out 2*10^4 m^3/month, Lake Seepage-out (groundwater inflow in to the lake) 1.1*10^6 m^3/month and groundwater outflow through the head dependent boundaries 6.7*10^6 m^3/month.The model water balance suggests that lake Naivasha basin is in equilibrium with a net outflow about 1% greater than the inflow over the calibrated period of time (1932-2010)

Key words: Lake Naivasha, Groundwater modeling, Transient, Water balance, Interaction Modeling

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Abbreviations	**Description**
3-D	Three-Dimensional
GPS	Global Positioning System
TIN	Triangular Networking
GMS	Groundwater Modeling System
LAK3	Lake Package 3
PEST	Parameter Estimation Tool
WRAP	Water Resources Assessment and Planning Project
GB01	Stream gauging stations name
ASTER	Advanced Spaceborne Thermal Emission and Reflection Radiometer
UTM	Universal Transverse Mercator
2GD1	Lake level measuring stations
DO	District office
GIS	Geographic Information System
SRTM	Shuttle Radar Topography Mission
USGS	United States Geological Survey
DEM	Digital Elevation Model
WDD	Water development Department
RGS	Regular gauging station
LAYCODE	Layer type for Modflow software
LPF	Layer Property Flow
BCF	Block Centred Flow
Mm^3/month	Million meter cube per month
GHB	General Head boundary
WREM	Water Resource and Environmental Management

1. INTRODUCTION

1.1. Background

Lake Naivasha has been considered as a highly significant national fresh water resource in Kenya by several authors. Its water is not only being utilized for domestic water supply and recreation but also sustains important economic activities such as flower and vegetable growing, tourism and fishing. Rapid industrial development and increase in agricultural production like in Lake Naivasha, have led to freshwater shortages in many parts of the world. In view of increasing demand of water for agricultural, industrial and domestic consumptions, a greater emphasis is being laid for sustainability and optimal utilization of water resources.

Sustainable development of water resources needs quantitative estimation of the available water resources. Quantitative estimation is necessary to maintain the groundwater reservoir in a state of dynamic equilibrium over a period of time and the water level fluctuations to be kept within a particular range over the dry and wet seasons. Water balance studies have been extensively implements to make quantitative estimates of water resources. Water balance also helps to evaluate quantitatively the contribution of individual sources of water in the system in time and space and studies the degree of variation in water regime due to changes in components of the system.

1.2. Research problem

Lake Naivasha is the only freshwater resources among many saline lakes in the Kenyan rift valley. The freshness of the lake makes it suitable for development of flower production, horticultural production, tourist industries and other human activities around the shores of the lake. In the last 10 years the industrial expansion around the lake has definitely translated into a correspondingly high increase in the demand of water use. At this time the ever increasing demand of lake and groundwater water usage for irrigation and other activity is reflected by water level decline and water quality deterioration in the study area.

In sought for solution and reflect what is going on the study area different water balance studies have been made in the area. However most of the previous study attempts were to characterize the two, the lake and the groundwater bodies as separate systems in the study area. Nevertheless, the degree interaction is less investigated; the surrounding aquifer and the lake Naivasha are beloved to be in dynamic interactions. Although the hydrogeological study of Lake Naivasha basin is expected to be very complex due to the interaction of groundwater and surface water flows, 3-D transient modelling approach is presented in this research to study the long term spatial and temporal variations in the system.

1.3. Objective

1.3.1. General objective

Construction and calibration of 3-D transient groundwater model that simulates the groundwater and lake water balance of the study area and that could be utilized to evaluate the effects of changes in system flux over time.

1.3.2. Specific objectives

The general objective will be achieving by solving the following specific objectives during this study.

- Develop and calibrate 3-D transient groundwater model to the (1932-2010) years of lake/aquifer exploitation
- Calculate long-term groundwater and lake water balance.
- Simulation of the lake-aquifer abstraction from the basin.

1.4. Research questions

- How fluxes in the aquifer/Lake system vary in space and time?
- What are the water balance components of the system?
- What is the steady state groundwater flow pattern of the area?

1.5. Research hypothesis

- The spatial and temporal variation of fluxes of the lake Naivasha basin can be simulated through 3-D transient flow model.

1.6. Relevance of the study

- Seasonal variability of groundwater-surface water exchange fluxes and its spatially and temporally variable impact on the water balance. Hence water balance analysis is important in order to quantify the linkages between the surface water and groundwater regime
- Lake Naivasha is fresh water which is currently as the center of industrial development. Hence water balance analysis is important to provide a technical basis for decisions on the quantity of water available and economic development activities on the area

1.7. Literature review

Lake Naivasha being a fresh water lake within the Kenyan rift valley with no known outflow has drawn many researchers interested in different aspects of the lake. Exploration of the Naivasha area began as early as the 1880's by European explorers. Thompson, of the Royal Geographical Society of England, during a visit at that time, he noted the freshness of the lake's waters, and attributed it to the lake being either of recent origin, or having an underground channel (LNROA, 1996)

(Ojiambo, 1996)discusses the hydrogeologic conditions around the lake. He indicates that the main subsurface outflow is from around the intersection of Oloidien Bay and the main lake with outflow fluxes ranging from 18-50 * 10^6 m^3/year.

(Ojiambo, 1996)in his thesis Characterization of Subsurface Outflow from a closed basin Freshwater Tropical Lake, Rift Valley, Kenya, he pointed out that groundwater level to the north of the lake have dropped below the lake level compared to what they were in 1972 when studied by McCann. The drop in water levels in northern wells around Manera may not be wholly explained by the drop in the lake level, but may be largely explained due to increasing pumping from the aquifer

(Mcann, 1974) In the report hydrogeologic investigation of the Rift Valley catchment, he pointed out that "in the Naivasha catchment groundwater generally flows towards the lake from the Mau and Aberdare escarpments, although it is diverted locally by the presence of faults that either form barriers or conduits

(Sikes, 1936) made the first statistical attempt to estimate monthly and annual water budget for the lake, and estimated the magnitude of the proposed underground seepage. He estimated water was seeping out of the lake at a rate of 43 * 10^6 m^3/year. (Mcann, 1974) estimated that about 34 * 10^6 m^3/year of water recharge the shallow groundwater aquifers from Lake Naivasha.

(Ase, 1986) worked on the surface hydrology of Lake Naivasha. He calculated the lakes monthly water balance for the period 1972 to 1980 based on mass balance equation. He estimated ground water outflow in the range 45-50 million cubic meters per month.

(Trottman, 1998) exercised preliminary ground water model to investigate the hydraulic interaction between Lake Naivasha and the surrounding unconfined aquifer and to study the changes in ground water storage of the aquifer in response to fluctuating lake levels. However many assumptions and generalization were made in calculating the model inputs which oversimplified the complex aquifer system of this area.

(Baher, 1999) was tried to improve the knowledge of interaction between the lake and the surrounding aquifers. He used a cross sectional model to study the interaction between the lake and groundwater. He studies ground water storage by optimizing different aquifer parameters like Transmissivity and storage coefficient, which are used to quantify the storage change. He also investigated the ground water storage behaviour of the aquifer in relation to the lake level and to quantify the contribution of ground water as a potential water resource with scarce aquifer parameters and inaccurate boundary conditions.

(Mmbui, 1999) studied the long-term water balance of the basin and calculated a groundwater outflow of $4.6 * 10^6$ m^3/month. He estimated a long-term average total combined inflow into the lake $2.26 * 10^6$ m^3/month.

(Owor, 2000) studied the long-term interaction of ground water with the lake to determine the long-term water budget for the lake and estimate water abstraction from both the surface and ground water resources. This was an integration of two previous studies: Long-term water balance of Lake Naivasha by Mmbui(1999) and groundwater flow modeling of the Lake Naivasha basin by Hernandez (1999). His study was a better approach in providing a more realistic insight into the long-term interaction of the lake and groundwater.

(Kibona, 2000) modelled the aquifers north of the lake .She modelled the lake by using a specific definition of the upper layer as a lake. She sought to understand the variation of ground water levels in space & time by setting up both transient and steady state.

For groundwater-surface water balance modeling recharge is the most important input variable. In attempt to understand the spatial and temporal distribution of recharge in the study area, (Nalugya, 2003) investigate that recharge in the study area is low and governed by Rainfall, Evapotranspiration and soil type. According to here study the natural areas around Kedong received the highest recharge (43.75mm/year), followed by Marula (33.75mm/year). Ndabibi and Three points receive the lowest, 0.69mm/year and 4.38mm/year respectively.

Geophysical surveys particularly resistivity, gravity, and magnetic were carried out in the past on the geothermal areas in the Naivasha basin. The works of Tsiboah(2002) is among the most important studies in the basin that gives more attention on subsurface characterization aquifer geometry definition. According to the study Lake Naivasha basin is made up of two aquifer system exists at a depth of 20-40m and 50-80m (Tsiboah, 2002).

(Becht & Harper, 2002) calculated the water balance of Lake Naivasha from a model based upon the long-term meteorological data of rainfall, evaporation and river inflows. The study estimated an annual abstraction rate for the period (1983-1998) as 60*106 m3 /year.

(Mohammedjemal, 2006) conducted a feasibility artificial recharge study in the horticultural area north of Lake Naivasha. He made injection and hydraulic conductivity test to investigate the infiltration capacity of the shallow aquifer in the study area

To understand the hydro-geological behaviour of the rift lakes it is essential to gain good conceptual view of the geological and palate-hydrological processes. (Nabide, 2002) develop 3-D conceptual hydrogeological model for Lake Naivasha area based on the integration of geology, hydrochemistry, isotopic analysis, and boundary conditions. This model is a good basis to construct a calibrated groundwater model and to decrease the various assumptions made in the past modeling histories. The most recent study probably the most important analysis and recommendations about groundwater-lake water interaction was made by(Yihdego, 2005). He modifies the conceptual model developed by (Nabide,

2002) and develops a steady state 3-D groundwater model using high conductivity "high-K lake" method to simulate the lake. However the "high-K lake" method is suitable for relatively simple geometries and lakes with slower and smaller fluctuations (Chui & Freyberg., 2008)

Table 1:1 Summary of prevous studies in the study area

Parameters	**Author**	**Estimated values**
Evapotranspiration	(Ase, 1986)	1865 mm/year
River inflow (Malewa River)	(Podder, 1998)	215 Mcm /year
River in flows (Gilgil River)	(Lars-Erik Ase, 1986)	24 Mcm/year
Lake water outflow	(Sikes, 1936)	43 Mcm/year
	(Mcann, 1974)	34 Mcm/year
	(Ase, 1986)	46-56 Mcm/year
	(Clarke A.C.G. D. Allen, 1990)	50 Mcm/year
	(Ojiambo, 1996)	40 Mcm/year
	(Mmbui, 1999)	57 Mcm/month
	(Baher, 1999)	55 Mcm/year
	(Mmbui, 1999)	4.54 Mcm/ month
	(Asfaque, 1999)	5.46 mm pan evpo
	(Owor, 2000)	4.76 Mcm/month
	(Becht & Harper, 2002)	60 Mcm/year
Lake water inflow (groundwater & baseflow)	(Viak, 1975)	1.8 Mcm/year
	(Graham, 1998)	60 mm/year
	(Graham, 1998)	137 mm/year
	(Owor, 2000)	0.22 Mcm/ month
Abstraction (lake & aquifer)	water bailiff's	32.7 Mcm/year
	(Goldson, 1993)	35 Mcm/year
	Domestic (Water Bailiff, 1993)	21.6 Mcm/year
	(Hermandez, 1999)	18 000 m3/day
	(Owor, 2000)	18000-25000 m3/day
	(Kibona, 2000)	14 000 m3/day
Aquifer transmisivity	(Hermandez, 1999)	1- 5000 m2/day
Aquifer storativity	(Hermandez, 1999)	0.1-0.15
storage volumes	(Owor, 2000)	6.9 Mcm/month

2. DESCRIPTION OF THE STUDY AREA

2.1. Location and accessibility

The study area is located in the Kenyan, Nakuru District, at about 100 km Northwest of Nairobi. It is located in the central rift valley of Kenya between latitudes 0^0 10'S to 1^0 00'S and longitudes 36^0 10'E to 36^0 45'E, with UTM zone 37 South and covers an area of about 3500 km^2. It is accessible by the mainline of the East African railways and a major road that services the western part of the country. There is an even distribution of all-weather roads within the area. The study area is situated in North-eastern part of the Naivasha basin at a mean altitude of 1885m above mean sea level.

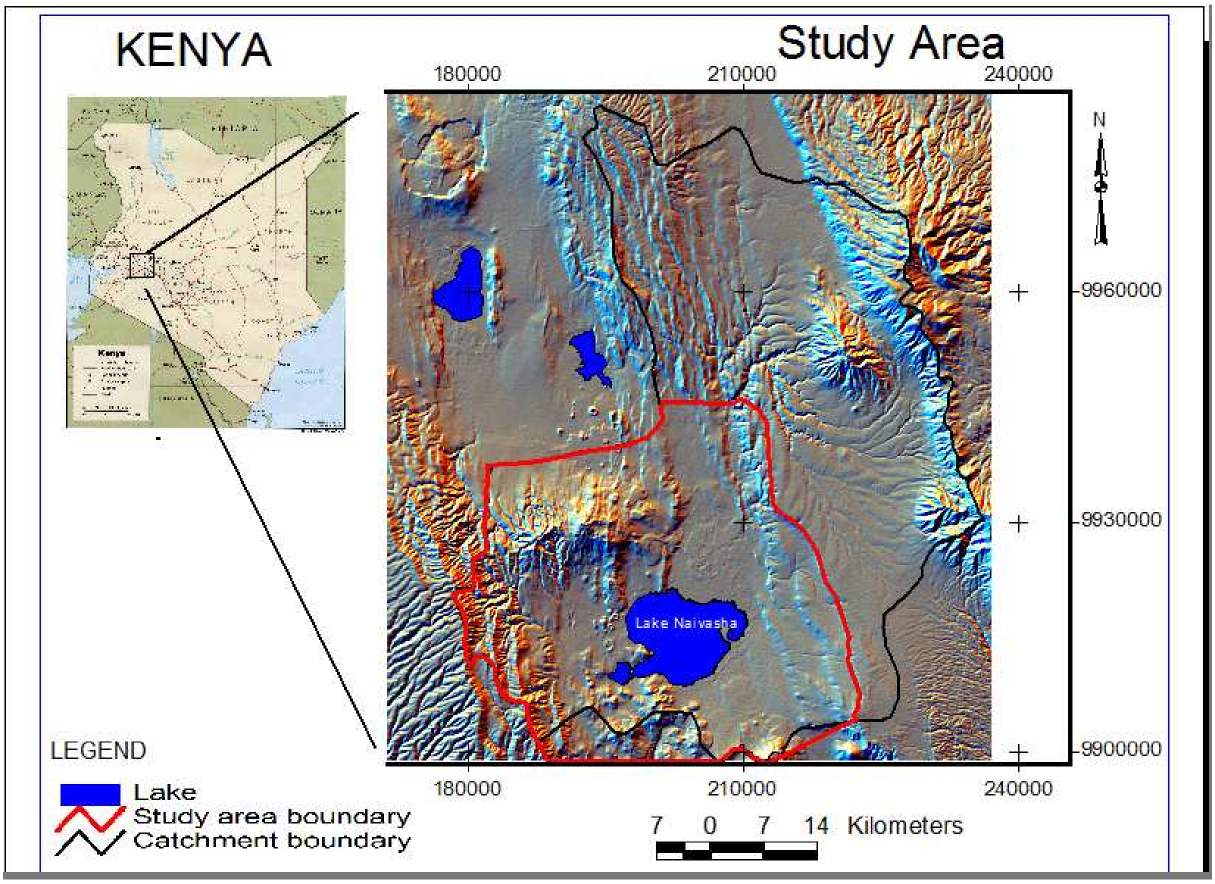


Figure 2:1 location map of the study area

2.2. Physiography and land use

Lake Naivasha dominates the central part of the Naivasha basin. It has a mean surface area of 145 km^2 at an average altitude of 1887.3 m. a.m.s.l .(Mmbui, 1999).The Mau escarpment on the western fringe rises up to a maximum of 3080 m. The escarpment is rugged and deeply incised with numerous faults and scarps that are prevalent.

The principal land use is agriculture which includes crop farming (horticulture, vegetables and fruits) around the lake and a mixing farming on the rain fed slopes of the escarpment. The Eburru hills, Mau, and Longonot escarpments are all hosts to indigenous hard wood forests that form the main water shed of the lake basin.

2.3. Climate, hydrology and drainage

The basin lies within the semi-arid belt of Kenya with average annual precipitation of 700m.The rainfall pattern is bimodal with the main rainy period in April-May and the shorter one from October-November. It is greater along the Mau and Aberare escarpments where it averages from 1250-1500mm annually and is lower in valley areas (like Lake Naivasha) where it averages about 650mm annually. There is an annual potential evaporation estimated at about 1700 mm (Mcann, 1974), monthly averaged potential evaporation on the floor of the basin exceeds rainfall by a factor of 2 to 8 for every month except April when the potential evaporation still exceeds rainfall for the wettest years. Mean daily temperatures vary between 9^0c at night to 25^0c during the day.

The major streams that drain the study area are the Malewa River and the Gilgil River. Ground-water discharge from the weathered volcanic aquifers provides base flow to the rivers. The Malewa River is one of the two main perennial rivers that drain the lake and flow in a graben at the foot of the kinanagop plateau. The Malewa and Turesha rivers have a combined drainage area of about 1,730 km^2. The Kinanagop Rivers are captured by the main Malewa River in the north east of the basin. Further downstream the Malewa River is joined by the Turash a river and the two rivers flow south wards. The Gilgil River flows in a narrow basin to the north of the basin and is the second major perennial river that drains the lake.

2.4. Lake morphology and general setting

The Lake is shallow with an average depth of 5m. It has very flat bottom with major decrease in depth only close to the shores. The deepest part of the main is located near Hippo Point on the southern western part of the lake. A nearly west-east profile of the Lake bottom shows the flatness of the main part of the Lake and the crater like morphology of the two deepest parts of the Lake, the Oloidien bay and Crescent Lake. The two deepest parts of the Lake have typical crater shaped morphology indicating volcanic origin of formation.

2.5. Geologic setting

The geology of the area is generally made of volcanic rocks and lacustrine deposits (sedimentary rocks). In the basin are complex geological structures, which have been subjected to several tectonic processes leading to varying structural features.

2.5.1. Sedimentary unit

The lake sediment comprises alluvial, lacustrine air fall (wind deposits), reworked volcanic. It is a heterogeneous mixture lakebed deposits and fluvial deposits.

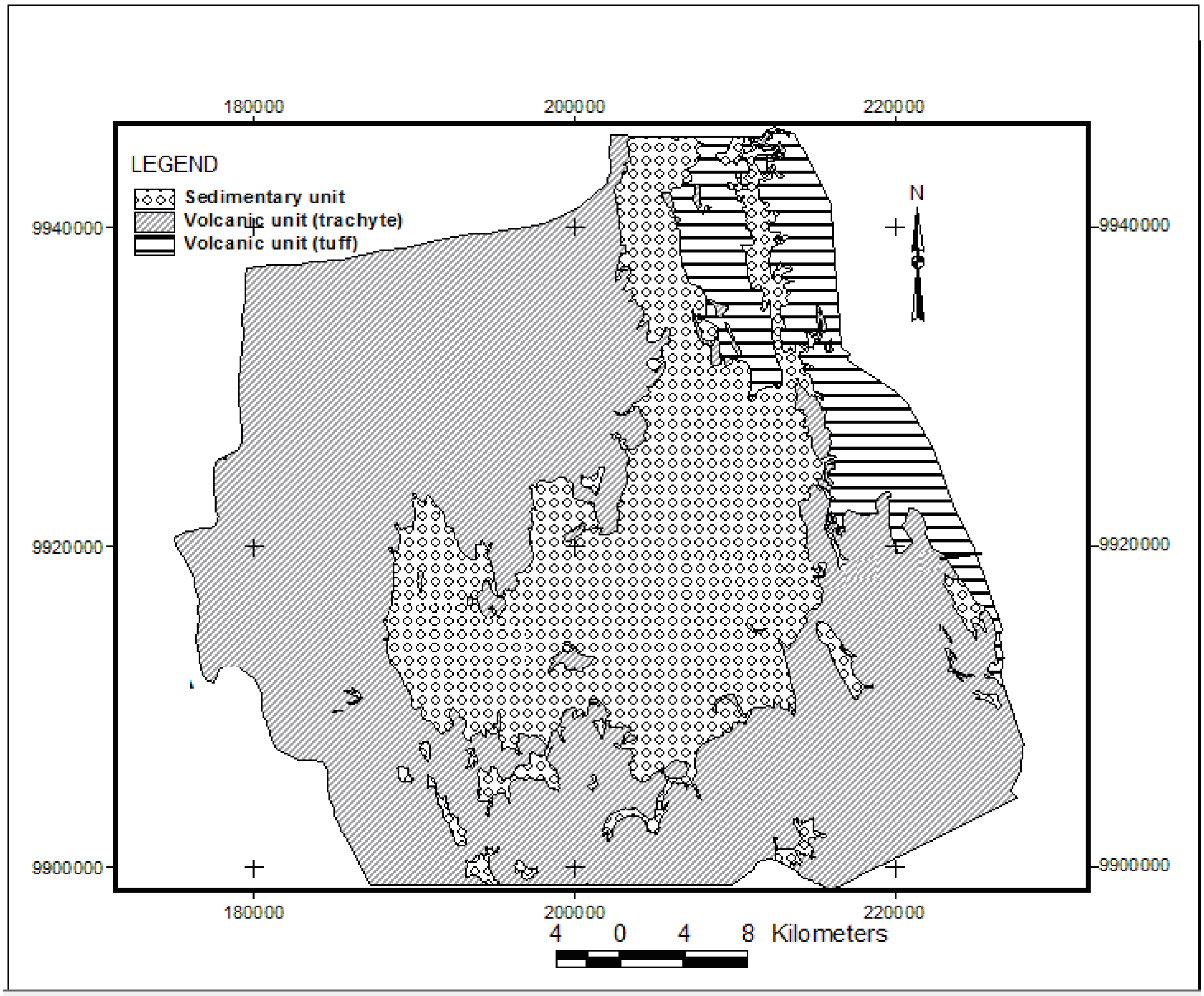


Figure 2:2 Generalized geological map of the study area

2.5.2. Volcanic unit

The volcanic rocks consist of (trachyte unit and tuff unit). The volcanic stratigraphy is very complex and has been the subject of numerous studies. Because volcanic stratigraphy and its physical features are genetically related to the location of the units with respect to particular structural blocks and volcanic centres, the hydro-geologic units were defined on the basis of stratigraphic position and lithologic properties from (Nabide, 2002).

2.6. Hydrogeological setting

The hydrogeology of Lake Naivasha is complex (Clarke M. C. G., 1990). Hydrogeology is greatly influence by the geology, topography and climatic factors that pertain in the area.

2.6.1. Groundwater occurrence

Groundwater occurrence is greatly determined by the geological conditions as well as the available water for storage. Fresh volcanic rocks tend to be compact with no primary porosity although secondary porosity may be well developed. These rocks underlying the rift valley therefore have low permeability.

In the localized highland areas, there exist deep groundwater tables as well as steep groundwater gradient. The high hydraulic gradient accounts for the substantial outflow of groundwater from the lake to the south as well as some outflow towards the north. Structural features such as faults often optimise storage, Transmissivity and recharge with the significant of these occurring in places that are adjacent to or within a surface drainage system. Shallow groundwater table, low precipitation and low values of recharge characterize the valleys.

2.6.2. Aquifer systems and aquifer properties

(Clarke M. C. G., 1990) noted that aquifers are normally found in fractured volcanics, or along weathered contacts between different lithological units. These aquifers are often confined or semi-confined and storage coefficients are likely to be low. The main aquifer is found in sediments covering parts of the rift floor. These, aquifers usually have relatively high permeability and are often unconfined with high specific yield. On the rift escarpments, the estimated hydraulic conductivities range from 0.1 m/d for the Kinangop Tuff and 1.1 m/d for the Limuru Trachyte.

2.6.3. Piezometer and groundwater flow:

(Clarke M. C. G., 1990) write that the area has a complex hydrogeology, because while it is lower than the Rift escarpments it is at the culmination of the Rift floor. Groundwater certainly flows away from Lake Naivasha because the lake water is fresh, even though the lake has no outlet and lies in an area of high evaporation. Northerly flow may occur both via Gilgil and under Eburru. Southerly flow must also occur, following the hydraulic gradient.

Pizometric plots and isotopic studies show that underground movement of water is occurring both axially along the rift and laterally from the bordering highlands into the rift. Analysis of Pizometric map and aquifer properties of the rocks in the area show that much of the subsurface outflow from the Naivasha catchment is to the south, via Olkaria-Longonot towards Suswa

The structure of the Rift Valley and in particular major marginal Rift faults and the system of grid faulting and the Rift floor undoubtedly have substantial effect on the groundwater flow systems of the area. In general faults are considered to have two effects on fluid flow. They may facilitate flow by providing channels of high permeability, or they may prove to be barriers to flow by offsetting zones of relatively high permeability. In the Rift Valley the main direction of faulting is along the axis of the Rift, and this has a significant effect on the flows across the Rift. It is apparent from the high hydraulic gradients that are developed across the Rift escarpments that the effect of the major faults is to act as zones of low permeability. The effect of faulting is to cause groundwater flows from the sides of the Rift towards the centre to flow longer paths reaching greater depths, and to align flows within the Rift along its axis.

3. METHODOLOGY

Given that the objectives of research, the methodological design of this study involves separate characterization of both surface and subsurface parameter before put in to the model. The research contains three phase with the main stages being data preparation includes (pre-fieldwork and fieldwork) data processing and modeling: A schematic representation of the breakdown and sequence of the study process is shown in Figure: 3.1.

3.1. Data preparation

3.1.1. Pre field work

In the preliminary stages of the study a literature review and preparation for fieldwork was carried out. The existing well database was updated and reorganised. Available data were screened and pre-processed, field survey points mapped out, mapping units delineated and appropriate field materials and tools identified. The following materials were used

Topographic Maps
Geologic Maps: Geological Map of Longonot Volcano, the Greater Olkaria and Eburru Volcanic Complexes and adjacent areas, 1988, 1:100 000, Ministry of Energy Geothermal Section (Kenya)
Satellite Images: Landsat TM images (bands 741)
Groundwater well records: Pumping tests data, drilling completion records, groundwater level data
References: Research papers, previous MSc thesis, drilling reports and journal articles from previous Works in the study area were used (see references).
Equipment: Geodetic GPS for levelling of the wells, water level transducers
Geological equipment: Geological compass, geological hammer, magnifier, Laptop and camera.

3.1.2. Field work

A 3-week fieldwork was carried out from the third week of Sep, 2010 to the first week of October, after the required data has been identified and preparations were made. The following activities were carried out in the field

Hydrogeological observation

Hydro geological observations were taken with the help of Aster satellite images, geological maps and cross sections and previous studies of the area. The primary observation sites were selected along river Malewa, river karate and previously drilled borehole site and geological cross section shafts. For comparison with borehole drilling findings, the thickness and location of the different geological units were recorded.

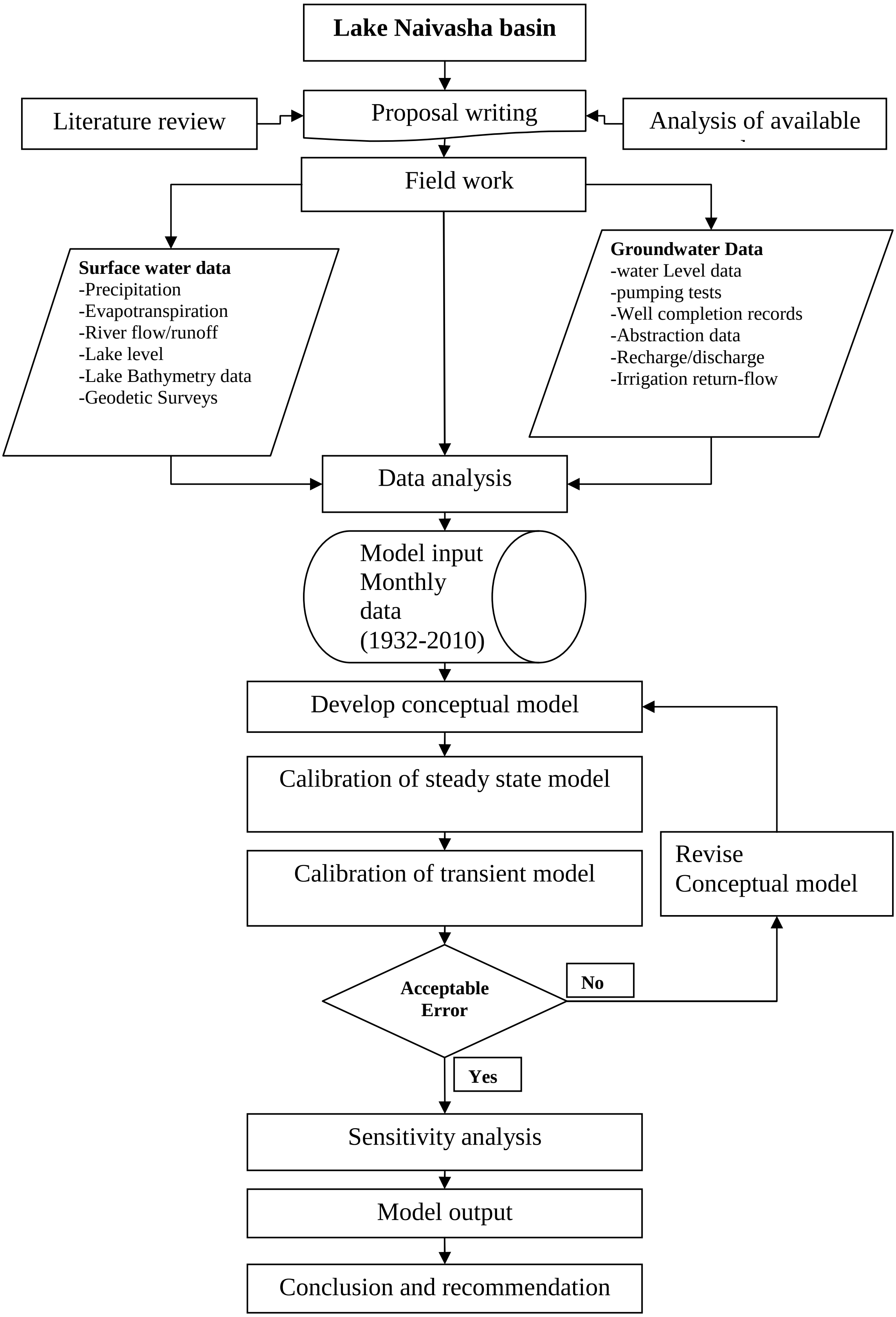


Figure 3:1 Schematic representation of the breakdown and sequence of the study process

Geodetic survey

A Geodetic GPS survey program was made during the field work. The main objective of this program is order to accurately measure the groundwater level and surface level at each well, river stage and lake level. For this reason a Geodetic GPS of the Leica GPS Receiver type (tripod-mounted) was preferred than the normal Garmin GPS.

The method of data acquisition in the field work requires a method of relative (differential) positioning system where two Leica GPS operating at the same time. One GPS as base station mounted on a tripod at a reference point and the second GPS as Rover on a pole which can be moved to the measuring site. The concept of the relative (differential) positioning uses one stationary antenna as a reference point. The stationary antenna's receiver then tracks at least the same satellites (preferably all visible satellites) as the moving receiver does.

Before start measuring the reference station was set at previously Geodetic levelled borehole named Menera farm BH7 with known coordinates 211231E, 9924924N and elevation 1903.32m This point was take as bench mark for the whole survey period. The antenna was centered above the station on a tripod and the height to the antenna phase centre was measured. All the cables were connected and the receivers were initialized so that visible satellites were acquired. After the R-time mode is set to "Reference" and the appropriate antenna is selected survey began. The configuration of the second GPS (Rover) also follow a similar set up except in this case the R-time mode is set to "Rover"

When the tracking had begun, it was ensured that the receiving device was functioning properly and that both measurements and broadcast ephemeris were recorded. The tracking performance was then monitored by watching receiver signal quality indicators. In this well levelling program an accuracy maximum 23 cm and minimum 0.3 cm is registered

The office processing was done using software called Leica Geo Office. In the processing the reference benchmark was used to compute co-ordinates of the other user-defined stations using baseline processing and single point processing where required. After the whole data had been processed, the interpolation method was used to relate and transform the co-ordinates from the Cartesian Clark 1880 system to the local coordinate system. The two co-ordinate systems were then matched using common tie points to obtain the transformation parameters. The transformation parameters were necessary to transform the coordinates from one datum to another. The transformed coordinates was therefore obtained from Projection: UTM, Spheroid: Clark 1880 and Datum: Arc 1960 in the local system coordinates system.
In this field work a total of 39 wells were geodetically levelled appendix: 1

Figure 3:2 Locating the reference station (left) and well top elevation survey (right)

Water level

Water level measurements of number boreholes with access openings were carried out by lowering a probe attached to an electric cable. Water level was measured relative to the top of the access hole. Where there was not access well, the measurement was taken relative to the top of the concrete slab. In few cases the wells are found without access hole or electric pipes are installed in the wrong direction, in this case the water level history of the well was traced from the drilling history record. Water level for geodetically levelled wells was taken in centimetre accuracy.

Pump test

Pump test data for the most recently drilled deep borehole is collected from different drilling companies. In addition to pump test result from all previous works is collected to interpret the results based on the present hydrogeological knowledge of the basin.

River flow data

Time series river inflow data for 12 gauging stations is collected from the Naivasha district office (Dc'S). The data are gauge height data collected on daily basis by stuff, weir and automatic data recorder. Most of the Gauge stations are located outside the study area from which about 90% of the river inflow is supposed to drain from. The main river is Malewa River (2GB 01) which has the longest record history in the basin (1931-2007) and the smallest is Karati River (2GD02) with no data record. During the field work the data for each station was checked for anomalies and missing gaps.

Lake level

The water level of Lake Naivasha has been monitored since 1908. Stations 2GD1, 2GD4 and 2GD6 have been the historical lake level measuring stations in the area (Mmbui, 1999). Recently other two stations have been installed at Yacht club weather station and Oserian metrological station. During the field work this station was not functioning due to maintenance problem. The Oserian metrological station is found on the south-western side of the lake. The station has recording history since 1991.

Rainfall

Rainfall data was one of the important input parameter to compute the amount of direct precipitation into Lake Naivasha. Naivasha District officer (DO) rainfall station data was selected for its location relative to the lake and data quality and availability. The station had data for the duration 1910 to 2010. For missed data filling purpose, rainfall data from other two nearby stations (Naivasha Kongoni Farm and Kenya nut rainfall station) were also collected

Evaporation

Being inside the rift valley, Evaporation is an important component to calculate the water balance of the lake. Evaporation data was obtained from Oserian metrological station. The station has recording history since 1991 on daily basis. Evaporation data from (1932-1997) was documented by Mmbui (1999).

Water abstraction data

Water abstraction data mainly from the irrigated commercial farms was collected along with the irrigation techniques. In addition to time series satellite image processing was also made to calculate the irrigation land cover change occurred since 2008. The development of irrigation lands, irrigation types as well as the main crops grown per farm before 2008 was documented by Musota (2008)

3.1.3. Post field work

Post fieldwork was the main process of this work. It includes data analysis and modeling with their scientific approach and application for this research.

4. DATA ANALYSIS

4.1. Study area set up

Construction and delineations of the study area involves the application of GIS and remote sensing in representing the areal image of the lake Naivasha basin. The general topographical and hydrological background was described from available records, maps, aerial photos and satellite imagery. The Landsat images band composite 543 was processed using ILWIS software. The study area boundary includes the simulation of subsurface hydrogeological boundaries and influence of major structures in addition to the catchment boundary.

4.1.1. Surface elevation model

In modeling of lake- aquifer interaction, a key step forward has been the incorporation of the Lake bottom bathymetry in to the digital surface model in order to the map the aquifer out crops on the bottom of lakes and to simulate the aquifer-lake interaction in detail. In this procedure the top surface elevation and the lake bottom bathometry are processed differently and overlaid together to form the top surface elevation of the study area. The accuracy assessment is made based on the geodetically levelled well head points.

4.1.2. Surface elevation map

The surface elevation map is created from SRTM data available from the USGS server. The source data has 90 m spatial resolution with absolute horizontal accuracy of 60m and vertical absolute accuracy of 16m. After the data is downloaded, imported and registered in ILWIS software, it was resampled to 50m special resolution. In order to calibrate the DEM with the actual surface elevation of the area, elevation points from the levelled wells and the row DEM are compared continually until the elevation match. In Figure 4.1 (left), the DEM is directly compared with the measured elevation values and the correlation value was 0.89. Next based on the constant reference objects like the lake surface, the DEM is made to rise by a constant value 3m. Through successive calibration on the DEM and on the measured values, the final result with a correlation of 0.98 Figure: 4.1(right) was used as the surface elevation map for the model input.

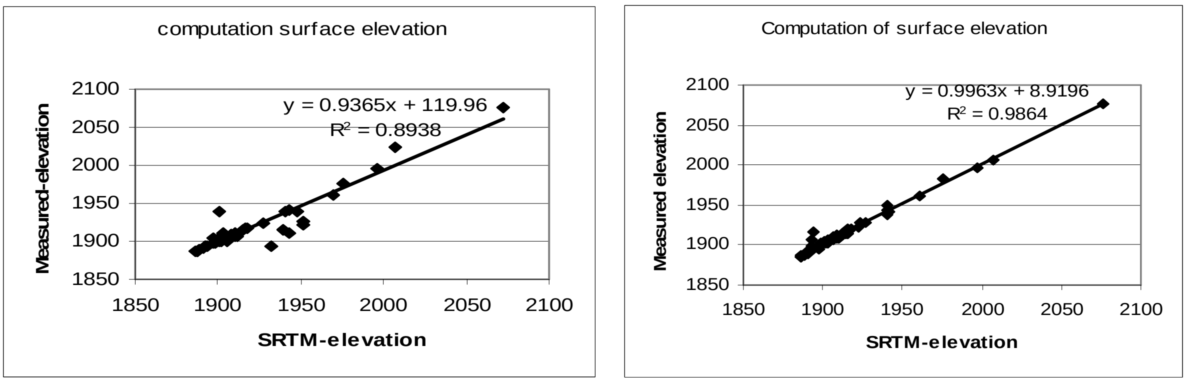


Figure 4:1 Computation of surface elevation processing (left) and after processing (right)

The base system of the lake bathymetry was originally surveyed in 1957 by the Ministry of water Works (Kenya). The contours of a map of the Naivasha area that was mapped by Viak (1975) were also raised by 2m to the level of the lake bathymetry. The digitization and interpolation of the contour maps were made by Owor (2000). Lake Bathymetry was incorporated in to the surface elevation map using map calculation function in ILWIS software. The lake bathymetry is an input parameter for the lake package in order to calculate the lake water balance of the model, Figure: 4.2

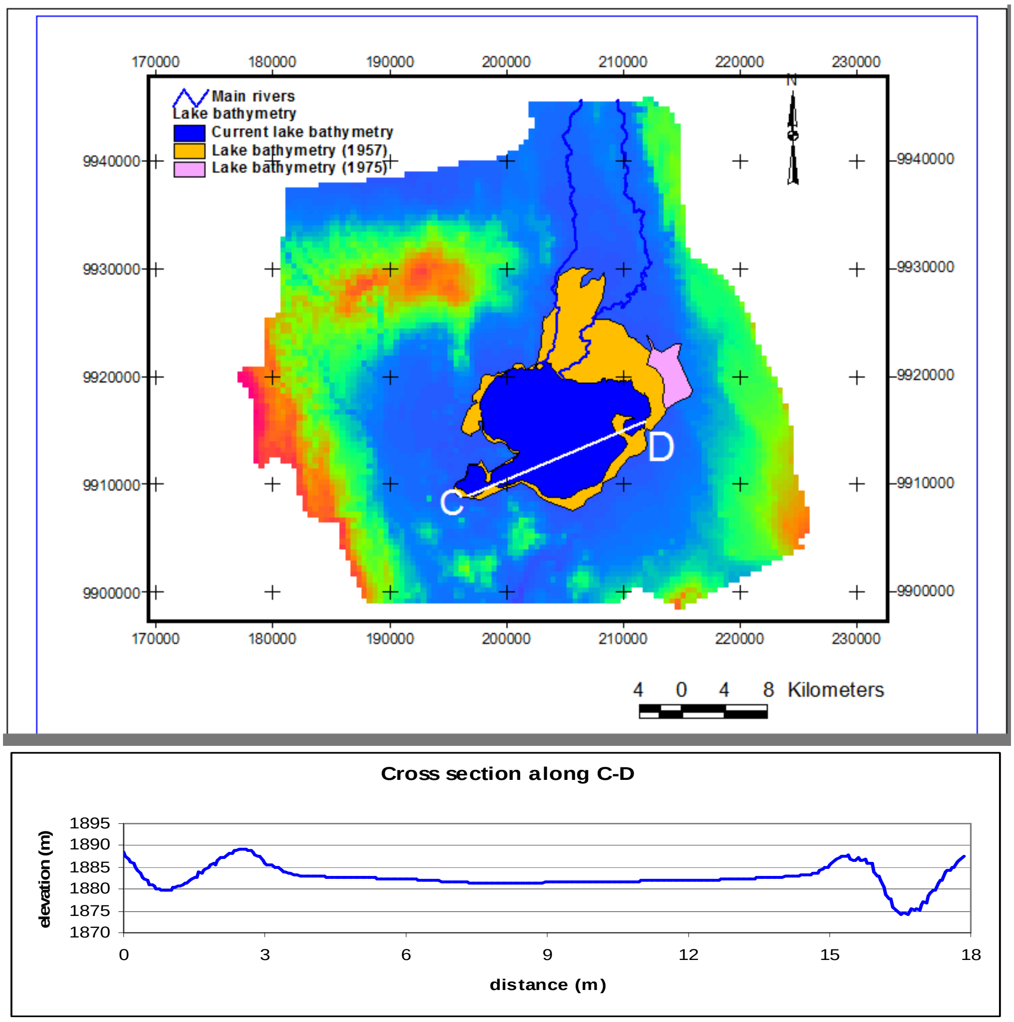


Figure 4:2 Profile of Lake Naivasha WRAP (1998)

Rating curve for the lake Naivasha were derived from the bathymetric surveys carried out by WRAP, 1998, the survey were include informations (curves) for calculating the Lake stage- volume and Lake stage -area relationships.

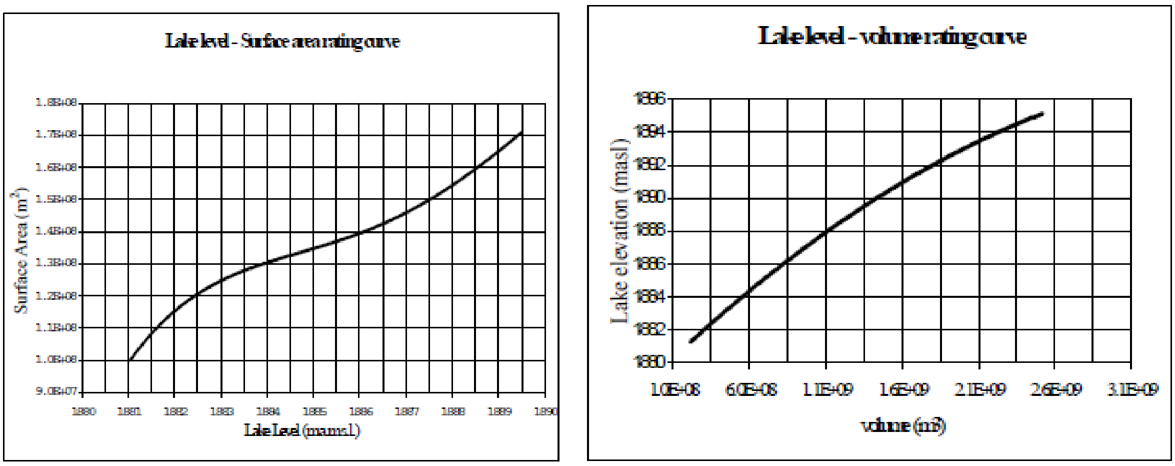


Figure 4:3 Surface area-Stage rating curve (left) and Volume-Stage rating curve (right)

4.2. Surface water data analysis

4.2.1. Precipitation

Rainfall data was collected from Naivasha district office meteorological station found at the center of the study area. The station had data record for the duration 1910 to 2010. After filtering for anomalies, the daily rainfall records were aggregated in to a monthly basis, Figure 4.4. The rainfall data considered in this station will be applied as a direct precipitation into the lake and for the general model area that has been used for the runoff and recharge estimates.

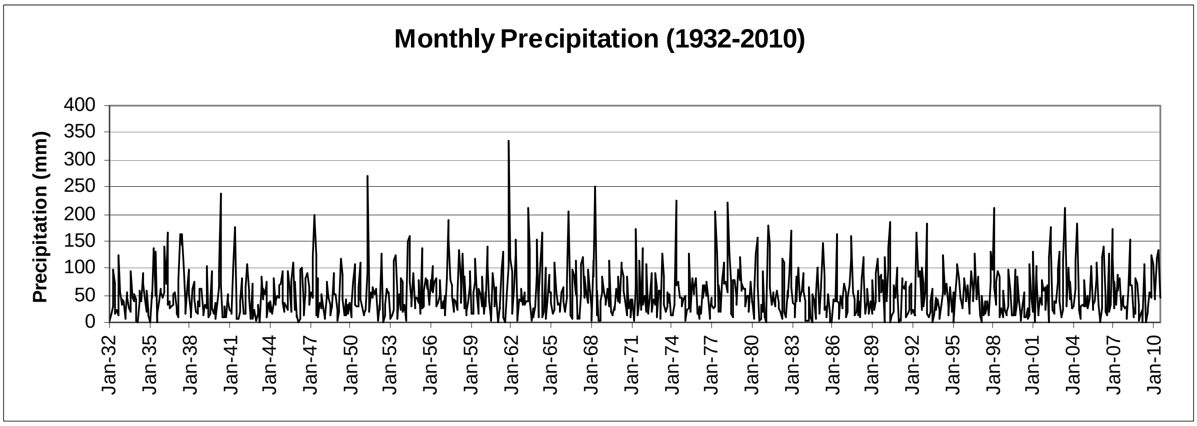


Figure 4:4 Monthly precipitations

4.2.2. Evaporation

Evaporation data was collected from Naivasha Water development Department (WDD) and from Oserian meteorological station. The WDD Evaporation data covers 1959-1990. The data was screened for typing errors and outliers using scatter plots and data gaps for missing data were in filled using linear regression. In order to backdate the evaporation data from 1959 to 1932, long term monthly averages were computed (Mmbui, 1999). These long-term averages were used to infill months with no recorded data.

The Oserian station is a private station found on the western side of the lake. The station has been record evaporation on daily basis since 1991. The data from the main farm was pre-processed for anomalies and data gaps. After aggregating from daily to monthly the record was used to extend the record series to 2010, Figure: 4.5.

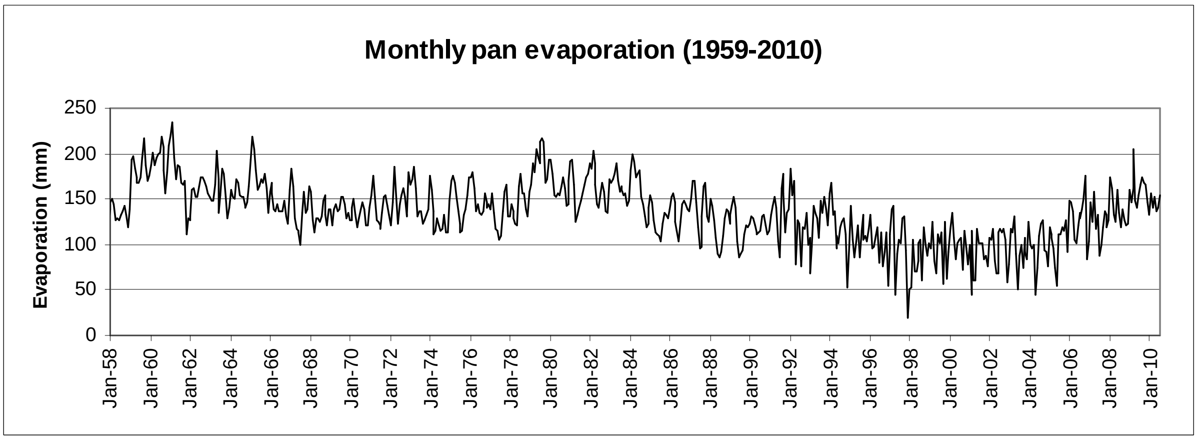
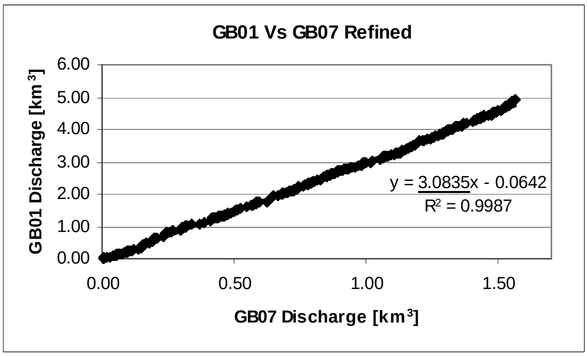


Figure 4:5 Monthly evaporation

4.2.3. Surface/River inflow

To quantify the time series river inflow in to lake Naivasha, The river discharge stations 2GB1and 2GB7 (Rivers Malewa), 2GA5 (on Gilgil) and 2GD2 on (Karati) were used. They were chosen for their proximity to the lake and discharge relationship among the stations. The amount of monthly discharge in to the lake from 1932-1997 was extensively documented by different researchers, (Becht & Harper, 2002), and (Mmbui, 1999). They were using linear interpolation, simple and multi-linear regression with the neighbouring stations, to fill the data gaps (Mmbui, 1999).

To extend the data series to the present, (2010) an assessment of the status of the gauge stations was made, In this study, data from station "2GB7" has been used to update the model to March 2010. This station has water level records from 1959 to 1994 and from 2000 to 2010. The position of 2GB1 in relation to 2GB7 suggest that the flows recorded at 2GB1 will be more than that of GB7 (i.e. GB7 is upstream of GB1). In order to use data from 2GB7, therefore, (1) the relationship between the stations had to be established, (2) In order to convert the water level to discharges; the flow rating curve for 2GB7 had to be established. After different statistical pre-processing (aggregation, data missing in fill and correction for anomalies), a multiplication factor of 3.08 was obtained at a correlation coefficient of 0.9987 and RMSE of 0.064 between the stations. Figure: 4.6 (left). With this factor, aggregated monthly discharges at 2GB7 from 1959 to 1994 were all transformed into their GB01. The flow rating curve Figure: 4.6 (right) were obtained from the historical data for 2GB07 the historical data has information on discharge, velocity and gauge height readings



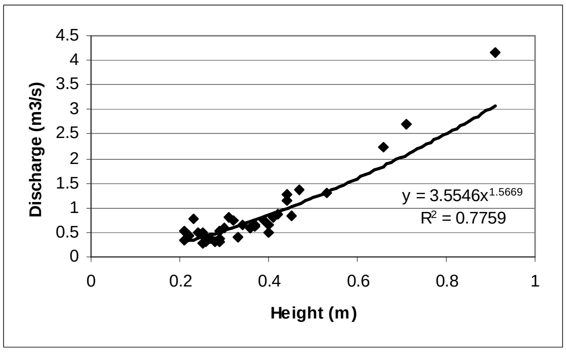


Figure 4:6 Monthly discharge GB01 vs 2GB07 (left) and Flow rating curve 2GB7 (right)

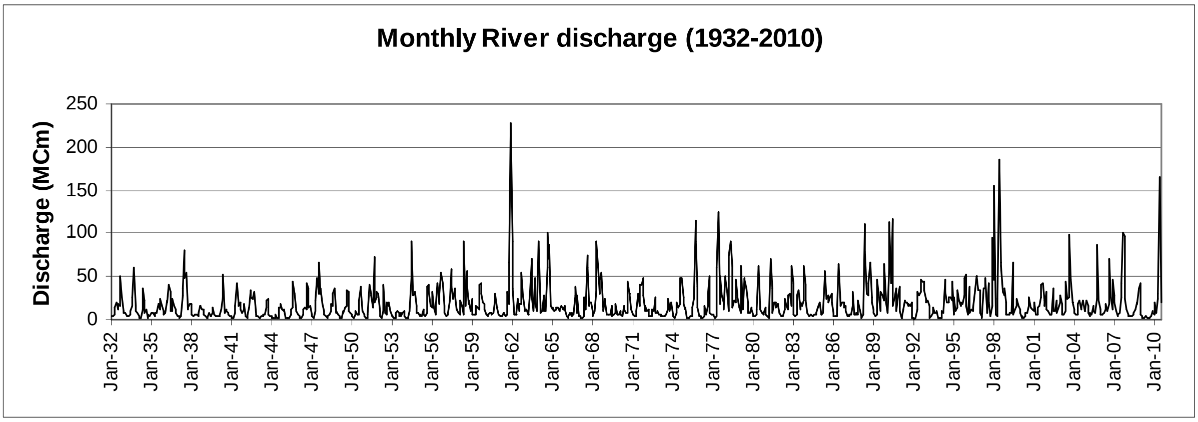


Figure 4:7 Monthly surface inflow

4.2.4. Lake level

Lake level series from 1932- 1997 was documented by (Mmbui, 1999). He was used records from the three main lake level stations namely: 2GD4, 2GD6 and 2GD1. Also He was included private monitoring stations at the Sulmac and Vaughan farms to fill data gaps in the series.
For this study the series had to extend to the year 2010, therefore the above station was visited again and data was collected and reprocessed. For this purpose data from 2GD6, which is a national regular gauging station (RGS) was primarily considered. However the station has data missing from Jan 2000-Dec 2003. Oserian farming company is one of the companies which have been practicing irrigated farming using water from Lake Naivasha. The farm is situated at the western part of lake. The company did its own daily monitoring of the lake levels. According to the information from the company officer, Oserian station carried out Levelling on the 17th March 2009 using Kengen's reference point of 1889.373m, and Oserian graduated staff stationed at the Lake was higher by 0.84 m. So, previous values were reduced by 0.84 m

length. The corrected daily record is available from 1991 to the present. After checking for type and some anomalies, the data from this station was used to fill the data gap for 2GD6. Finally the daily records were averaged to a monthly basis using excel software and the complete picture of the Lake Naivasha water level was constructed. Figure: 4.8

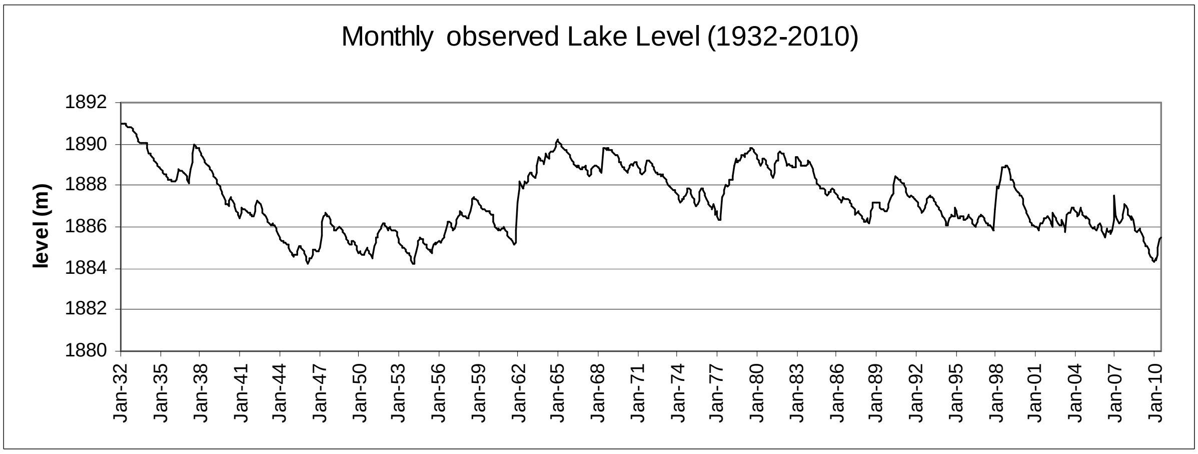


Figure 4:8 Monthly lake-level

4.3. Groundwater data analysis

4.3.1. Groundwater Level

The Pizometric surface of the wells was calculated as depth to water surface from the corrected Surface elevation map section 4.1 The wells with available groundwater levels in the area were collected from Naivasha District office (Kenya), GB drilling company (Nakuru, Kenya) and from previous works (ITC data base). Adjustments have been effected to reflect corrections based on the knowledge attained from recently drilled wells, newly levelled wells and the wells in which the water levels were measured during fieldwork.

4.3.2. Pump test

Pump test analysis was mad for 10 recently drilled boreholes in the study area. The analysis was made using two methods, Jacob straight-line method and Aquifer test software. During the analysis using the Jacob method, a plot of the drawdown versus time was constructed for the boreholes on a semi-log paper and an approximately straight-line graph were obtained from the test. When the best fit is obtained the Transmissivity of the boreholes is calculated from the pumping rate and the slope of the time-drawdown graph using the following relationship.

$$\mathrm{T} = \frac{2.3Q}{4\Pi \Delta S} \qquad \text{Equation: 1}$$

Where; Q = pumping rate in m3

ΔS = slope of the time-drawdown graph (change in drawdown per log cycle), and

T = Transmissivity

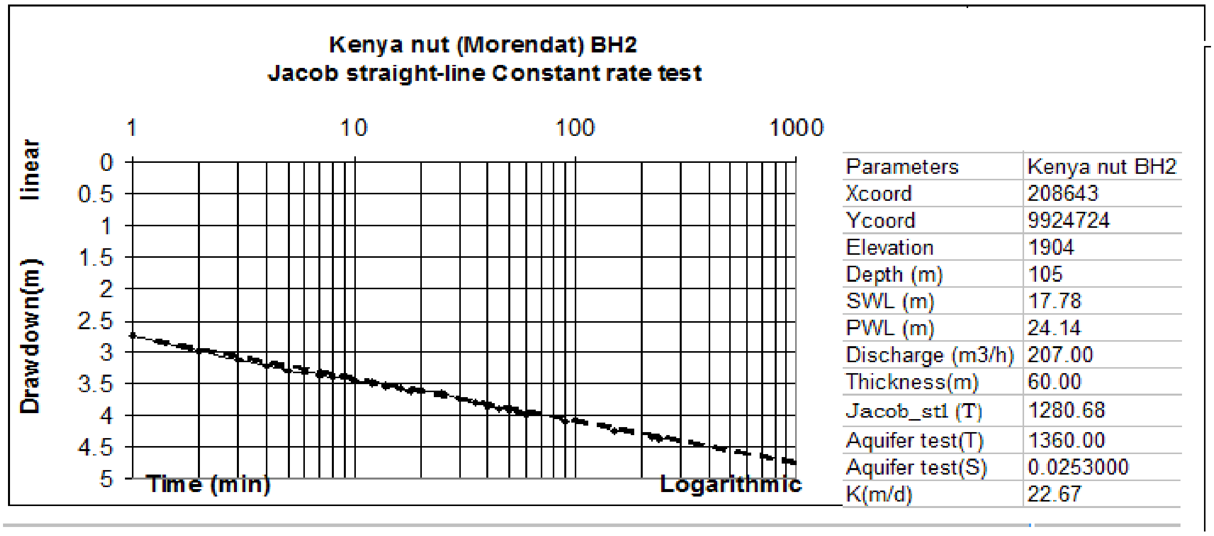


Figure 4:9 Pump test analysis result using Jacob straight-line method

The some data set was calculated using Theis analysis in aquifer test considering a single well pump test analysis assumption Figure 4.10. The single well analysis method gives the opportunity to estimate the aquifer parameters of a single well. For this pumping test, there is only one well used for both pumping and for recording drawdown measurements. The Transmissivity value calculated using both methods is presented in appendix: 2. (a) Borhole information and (b) pumptest result

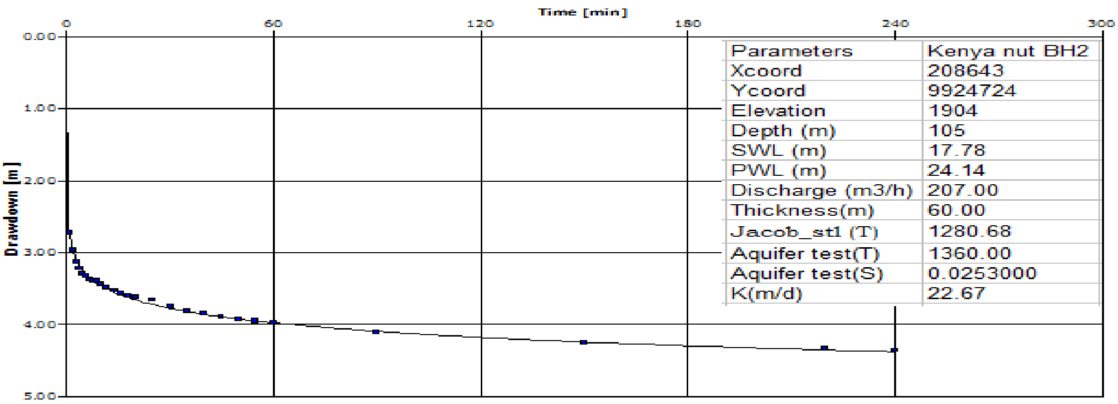


Figure 4:10 Pump test analysis result using aquifer test method

Previous estimate aquifer Transmissivity by Hermandez(1999) found values varying from less than 1 m^2/day to more than 5000 m^2/day. He estimated storativity values between 0.01. (Kibona, 2000) For well BH C at Three Point Ostrich farm within the lake sediments she estimates using the Hantush method a transmissivity of 1150 m^2/d and a storativity of 3.95×10^{-3}. With the Cooper-Jacob yields a transmissivity of 462 m^2/d and storativity of 1.46×10^{-3}.

Table 4:1 Transmissivity calculated by Clarke et al (1990) for the whole of Lake Naivasha

Area	Lithology	Transmissivity m^2/day		Average depth (m)	Hydraulic conductivity (m/d)
North east	Sediment & volcanics	1170	307	60	19
South east	Sediment & volcanics	3082	502	60	51
South west	Sediment & volcanics	940	297	60	15
North west	Sediment & volcanics	5308	1601	60	88

An analysis of shallow aquifer that yields water in the study area was done using well data kept by Ojiambo (1996). According to his analysis, transmissivity value in the area ranges from 3-1200 m^2/day. The corresponding hydraulic conductivity calculated from transmissivity values range from 14 to 750 m/day. Summary of previous estimate of aquifer parameters if found in appendix: 3

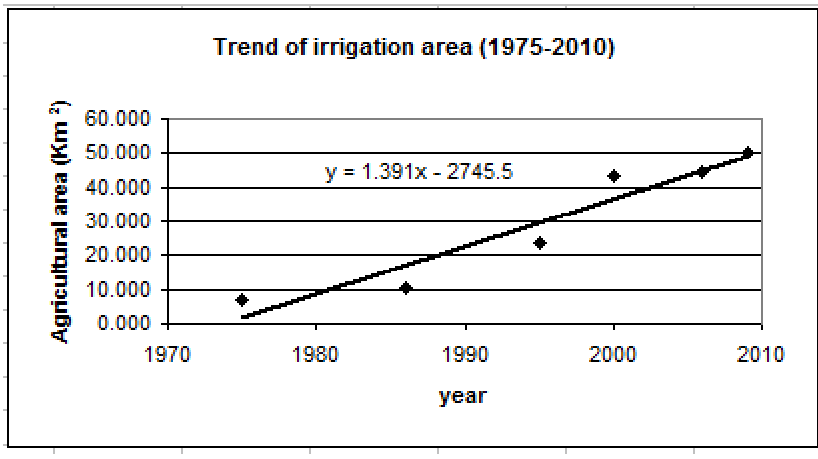
4.4. Abstraction

Abstraction is the most undocumented input data in the study area. Water abstraction in the study area occurs from irrigation and domestic wells. Abstraction for irrigation use is much larger than for domestic use. Irrigation abstraction from the lake sediment aquifer is especially important in the commercial farms.

Water abstraction data mainly from the irrigated commercial farms was collected along with the irrigation techniques and the area of the farms. Quantification of abstraction from different water bodies such as Surface water (river, lake) and groundwater (boreholes) was made based on the abstraction record data sheet obtained from the Naivasha District officers. The abstraction data sheet contains records such as the total irrigated area, location, owner, abstraction permit (from surface and groundwater) and actual abstraction per day for each irrigation farms. However it is often thought that the irrigation is supplied more than the estimated irrigation demand. Assessment of mount of abstraction was also made by considering the relation between abstraction and area of irrigation lands

Abstraction rate (m^3/day) = area of irrigated land * depth of irrigation   Equation: 2

Estimation of historical development of irrigation areas in the basin was made by processing time series satellite images. Estimation of depth of irrigation was made based on previous estimations different statistical relationships. (Yihdego, 2005) estimation was 1000mm/year.



Year	Agricultural area (km^2)
1975	7.147
1986	10.336
1995	23.383
2000	43.146
2006	44.672
2009	50.23

Figure 4:11 Trend of irrigation areas (1980-2010)

Water abstraction data mainly from the irrigated commercial farms was collected along with the irrigation techniques and the area of the farms. Quantification of abstraction from different water bodies such as Surface water (river, lake) and groundwater (boreholes) was made based on the abstraction record data sheet obtained from the Naivasha District officers. The abstraction data sheet contains records such as the total irrigated area, location, owner, abstraction permit (from surface and groundwater) and actual abstraction per day for each irrigation farms.

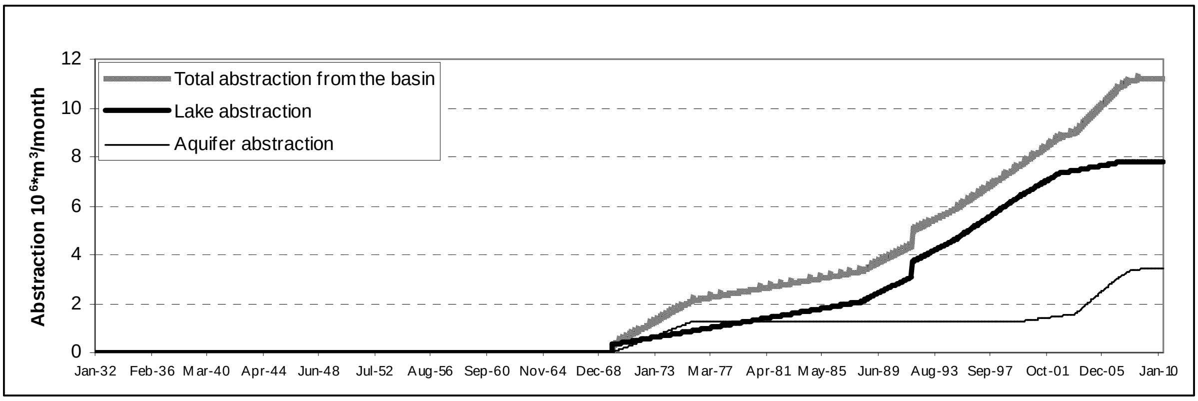


Figure 4:12 Analysis result of abstraction from different water bodies

4.5. Recharge

The mechanisms of groundwater recharge in the study area are; direct recharge by direct percolation through the videos zone and indirect recharge from water bodies and lateral inflow from adjacent aquifers. The most factors contributing to recharge in the study area are rainfall, evapotranspiration rates and soil types. The area experiences an annual rainfall of 640mm and annual evapotranspiration of 1700mm which is characteristic of a semi-arid climate. Since the potential evapotranspiration of the area was estimated to be higher than rainfall, direct recharge is not a permanent process in the area, but a process which occurs during rain seasons and, only when there is high intensity (Nalugya, 2003)

Table 4:2 Direct recharge estimated from SWAP model (1991-1998) by Nalugya (2003)

Local name	Location		Recharge(mm/day)		Average recharge(mm/year)
	UTM_X	UTM_Y	Min	Max	
Kedong	209691	9908544	0	7	43.75
Ndabibi	194490	9914863	0	0.27	0.69
TPF	213403	9924948	0	0.1	4.38
Marula	208444	9930840	0	5.5	33.75

Table 4:3 Direct recharge estimates from 1-D mixing model by Nalugya (2003)

Type of recharge	Stable isotope	mmday-1	mm year-1	Average (mmyear-1)
Mixed recharge; Lake: rain,	18-Oxygen	1x10-5 – 7x10-5	3.65 – 25.55	14.6
	Deuterium	3x10-5 – 7x10-5	10.95 – 32.85	21.9

The reliability of recharge estimates using different techniques is variable, the techniques based on Surface-water and unsaturated zone data provide estimates of potential recharge whereas those based On groundwater data generally provide estimates of actual recharge (Scanlon, 2002)

Estimated groundwater recharge from a basin (Mcann, 1974)

$$R = A * \sum_{i=1}^{n} \Delta hi * Sy \qquad \text{Equation: 3}$$

Where

R=Total ground water recharge (m^3/d),

A=area (m2)

Δhi = seasonal increase in ground water level (m)

Sy =Specific yield of water yielding materials

In this study the approach of Equation (3) is applied. The implementation follows the following procedure
1.The study made by Nalugya (2003) was taken as the steady state average recharge for the basin.
2. The seasonal increase in ground water level is assumed to be directly proportional to the lake level fluctuation. And can be replaced by monthly lake level increase.
3. The specific yield, I called it "the relating factor" is estimated from the relationship in equation: 3

$$Sy = \frac{avR}{\sum_{i=1}^{n} \Delta hi}$$ Equation: 4

Where

Sy= "Relating factor"
avR=average recharge in (m) of the area estimated by Nulugya(2003)
Δhi = seasonal increase in lake level

4. The transient recharge was calculated using the relation of equation: 3 and equation: 4

$$Ri = A * \Delta hi * Sy$$ Equation: 5

Where

Sy= "Relating factor" constant
Δhi = seasonal increase in lake level at time t=i, indicates the stress period i=1-942
Ri= groundwater recharge at time t=i

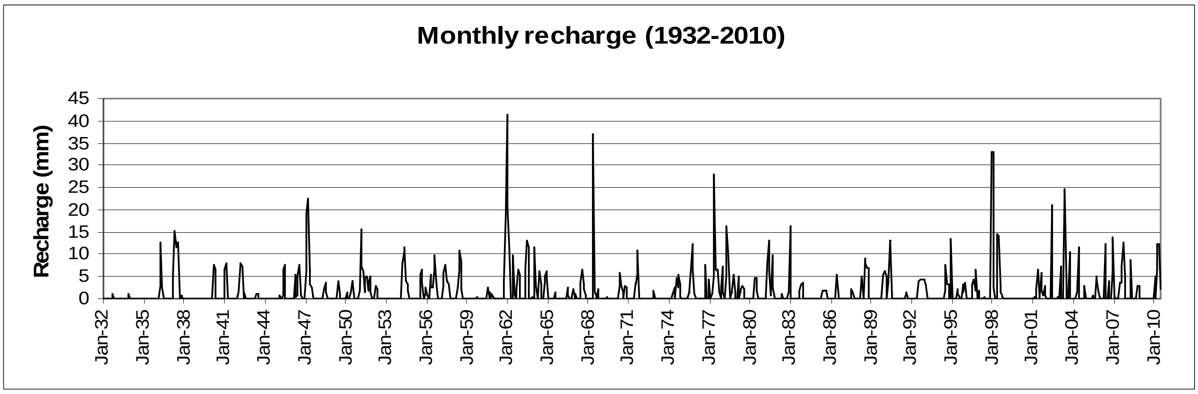


Figure 4:13 Monthly recharge

4.6. Irrigation return flow

The process of re-entry of a part of the water used for irrigation is called return flow. Percolation from applied irrigation water, derived both from surface water and groundwater sources, constitutes one of the major components of groundwater recharge.

Assessment of the irrigation return-flow can be made in different methods one of which is the estimation of the amount of Irrigation return water based on the Abstraction-Return flow relationship.

Return flow = C * Abstraction Equation: 6

Where, C is the return flow coefficient ($0 < C < 1$).

The irrigation return flow depends on the soil characteristics, hydraulic properties of the aquifer, depth to water table, irrigation practice and type of crop. A study was made in the study area by (Mohammedjemal, 2006) targeting to assess the feasibility of artificial ground water recharge using the runoff harvested from the greenhouses. During the study, Injection and pumping test was carried out to determine the intake capacity and hydraulic property of the aquifer. In addition to grain size and water quality analysis was made. The field experiment results revealed the texture variation within the top 30 to 32m unsaturated zone. Besides the injection test results in very low intake rate potential and the hydraulic conductivity is in the order of 0.01 to 0.2mday-1. Based on The above findings he concludes the infeasibility of shallow infiltration in the study area.

Based on this argument and a similar argument by Hagos (2008) who also made another artificial recharge test in the study area, the contribution of irrigation return flow in the long term water balance of the basin is assumed to be insignificant and will not be considered in this study.

4.7. Evapotranspairation

The study area is characterized by an annual potential evaporation about 1700 mm (McCann, 1974). Monthly averaged potential evaporation on the floor of the basin exceeds rainfall by a factor of 2 to 8 for every month except April when the potential evaporation still exceeds rainfall save for the wettest years (Owor, 2000), (Yihdego, 2005).

The natural vegetation surrounding the lake is mainly papyrus swamp vegetation. Natural vegetation outside of the lake surroundings are shrub, cactus, savannah and acacia (to the north). Acacia trees can be attributed to the shallower ground water table. The depth of root zone in the study area is assumed to be 3 to 12 metres (Asfaque, 1999). But little is known about the contribution of the Acacia trees for groundwater evapotranspiration. Generally due to the very nature of the study area, like; very high evaporation over precipitation, flat topography and rift valley environment, evapotranspiration from the land surface is assumed have insignificant contribution in the water balance of the basin. On the other hand, direct rainfall and evaporation from the lake surface includes estimates for swamp evapotranspiration on the lakeshores.

5. MODELING

5.1. Model setup

5.1.1. Modeling protocol

Modeling protocol is intended to provide guidance that will promote consistency in the application of groundwater modeling in the study area. According to (Anderson & Woessner, 1992), modeling protocol includes code selection, model design, calibration, Sensitivity analysis and prediction. The basic modeling protocol implemented in this study is designed as

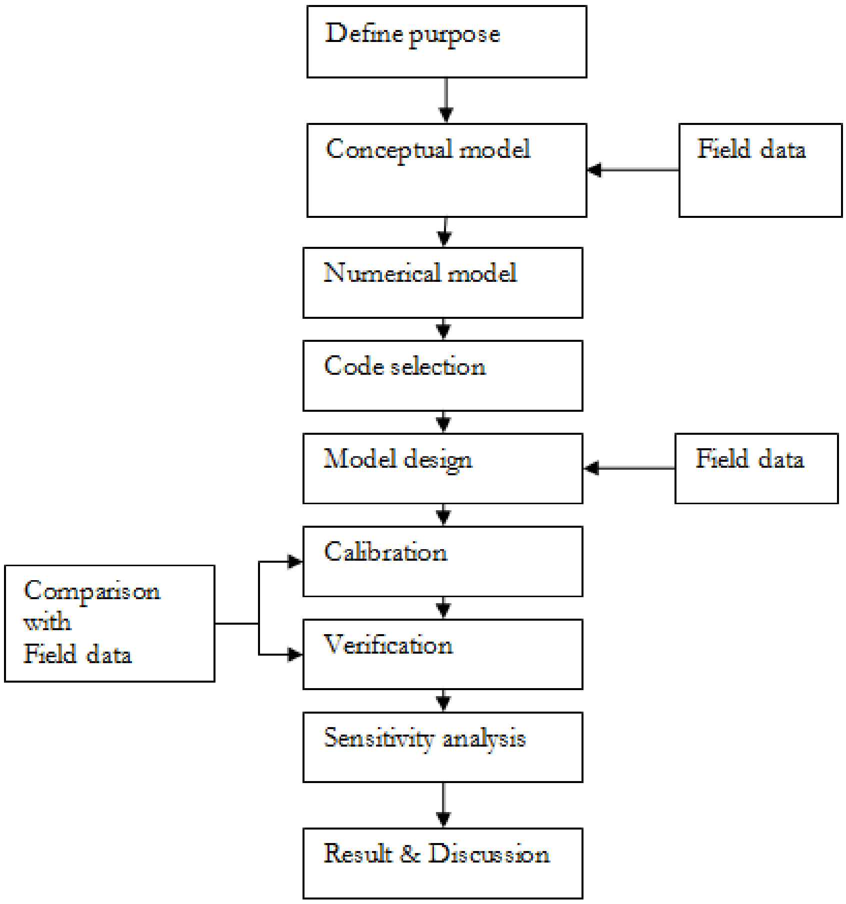


Figure 5:1 Steps in modeling protocol adopted from Anderson & Woessner (1992)

5.1.2. The groundwater flow equation

The partial-differential equation of ground-water flow used in modflow is (Harbaugh & McDonald, 2000)

$$\frac{\partial}{\partial x}\left(K_{xx}\frac{\partial h}{\partial x}\right)+\frac{\partial}{\partial y}\left(K_{yy}\frac{\partial h}{\partial y}\right)+\frac{\partial}{\partial z}\left(Kzz\frac{\partial h}{\partial z}\right)+W = S_s\frac{\partial h}{\partial t} \qquad \text{Equation: 7}$$

Where

Kxx, Kyy, and Kzz are values of hydraulic conductivity (L/T) along the x, y, and z coordinate axes

h= is the potentiometric head (L);

W= is a volumetric flux per unit volume representing sources and/or sinks of water,

W<0.0 for flow out of the ground-water system, and W>0.0 for flow in

SS= is the specific storage of the porous material

T= is time (T).

5.1.3. Modflow lake package (LAK3)

In the Lake Package described in this study, the lake is represented as a volume of space within the model grid which consists of inactive cells extending downward from the upper surface of the grid. Active model grid cells bordering the space and adjacent aquifer, exchange water with the lake at a rate determined by the relative heads and by conductance that are based on grid cell dimensions, hydraulic conductivities of the aquifer material, and user-specified leakance distributions that represent the resistance to flow through the material of the lakebed.
Implementation of a lake water budget includes input of parameters those representing the rate of lake atmospheric recharge and evaporation, overland runoff, and the rate of any direct withdrawal from, or augmentation of, the lake volume. The lake/aquifer interaction can be simulated in both transient and steady-state flow conditions.

5.1.4. Seepage between lake and aquifer

The direction and magnitude of seepage between a lake and the adjacent aquifer system depends on the relation between the lake stage and the hydraulic head in the ground-water system, both of which can vary substantially in time and space. Seepage from a lake into the surficial aquifer that surrounds it, where the lake acts as a source of recharge to the aquifer, occurs when and where the lake stage is higher than the altitude of the water table in the adjacent part of the aquifer. Quantification of the rate of seepage between the lake and the aquifer is made by an application of Darcy's Law equation: 8. The volumetric flux (L3/T) is expressed by integrating the specific discharge over some cross section of area A (L2) in a plane perpendicular to the direction of flow equation: 9.

$$q = K \frac{h_l - h_a}{\Delta l} \qquad \text{Equation: 8}$$

Where
q = seepage rate (L/T);
K = hydraulic conductivity (L/T)
hl = is the stage of the lake (L); and ha= is the aquifer head (L);
Δl = is the distance (L) hl and ha;

$$Q = qA = \frac{KA}{\Delta l}(h_l - h_a) = c(h_l - h_a) \qquad \text{Equation: 9}$$

Where
c = KA/Δl is the conductance (L2/T).
K/Δl = is the leakance (T -1).

5.1.5. Lake water budget

The interaction between the lake and the surficial aquifer is represented in this Lake Package by updating at the end of each time step a water budget for the lake that is independent of the ground-water budget represented by the solution for heads in the aquifer. Inherent in the calculation of a lake water budget is the computation of current values of lake volume and stage. The lake stage is crucial in making the estimates of ground-water seepage to and from the lake that are used by Modflow.

The implementation of a separate water budget for the lake that accounts for seepage losses to and seepage gains from the aquifer provides the capability to use the model to make a separate estimate of the stage of the lake and its relation to the water table. Updating a lake water budget also requires that estimates be made of gains and losses of water from the lake other than by seepage, such as (1) gains from rainfall, overland runoff, and inflowing streams, (2) losses to evaporation and out flowing streams, and (3) anthropogenic gains and losses (withdrawals for water supply or augmentation with water from another source.
The water budget procedure incorporated in the Lake Package is implied by the equation used to update the lake stage. The explicit form of this equation is:

$$h_1^n = h_1^{n-1} + \Delta t \frac{P - E + SRO - W - SP + QI - QO}{A}$$ Equation: 10

Where

h the lake stages (L)

Δ t time step length (T)

P rate of precipitation (L3/T)

E rate of evaporation (L3/T)

SRO rate of surface runoff to the lake (L3/T)

W rate of water withdrawal from the lake (L3/T)

QI rate of stream inflow (L3/T)

QO rate of stream outflow (L3/T)

A surface area of the lake (L2)

SP net rate of seepage between the lake and the aquifer (L3/T)
(Positive Value indicates seepage from the lake into the aquifer)

5.2. Conceptual model

A conceptual model is the pictorial representation of the groundwater flow system. (Anderson & Woessner, 1992), Hydro-geologically the area is complex due to rift floor geometry and tectonics. Therefore the purpose of the construction this conceptual model is to simplify the field problems and organize the associated field data so that the system can be analysed more readily. Simplification is necessary because a complete reconstruction of the field system is not feasible (Anderson & Woessner, 1992).

The basic conceptual model assumes two non-coinciding aquifer systems. The upper aquifer is unconfined aquifer: lake and fluvial/lacustrine sediment interbeded with clay silt and volcanic materials. The lower aquifer is semi- confined aquifer: reworked and weathered/fractured volcanic materials. The upper aquifer is in hydraulic link with the lake. It is characterized by variable (heterogeneous) horizontal and even (homogenous) vertical aquifer properties. The lower aquifer is characterized by homogenous vertical and horizontal aquifer properties. The lower aquifer is also in hydraulic connection to the upper aquifer through a leakage terms.

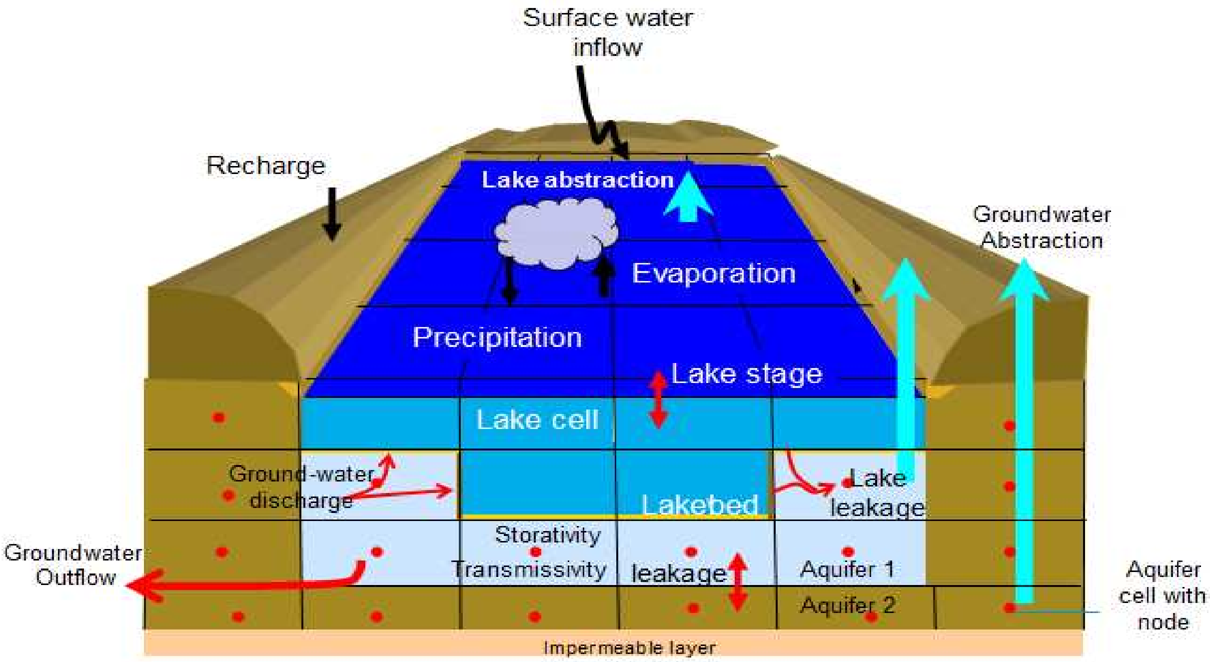


Figure 5:2 conceptual model reflects the study area

5.2.1. Hydrostratigraphic units

The simplified Hydro Stratigraphic units of the area as discussed also by Nabide (2002) and Yihdego(2005) consists of three units: tuff, trachyte and sedimentary. The sedimentary unit is composed of fluvial and lacustrine sediment. The unit is laterally variable. The volcanic unit is composed of reworked and weathered volcanic. The unit is laterally and vertically continuous. The tuff hydro-geological unit is considered as impermeable base by assigning at least a two order of magnitude contrast in hydraulic conductivity.

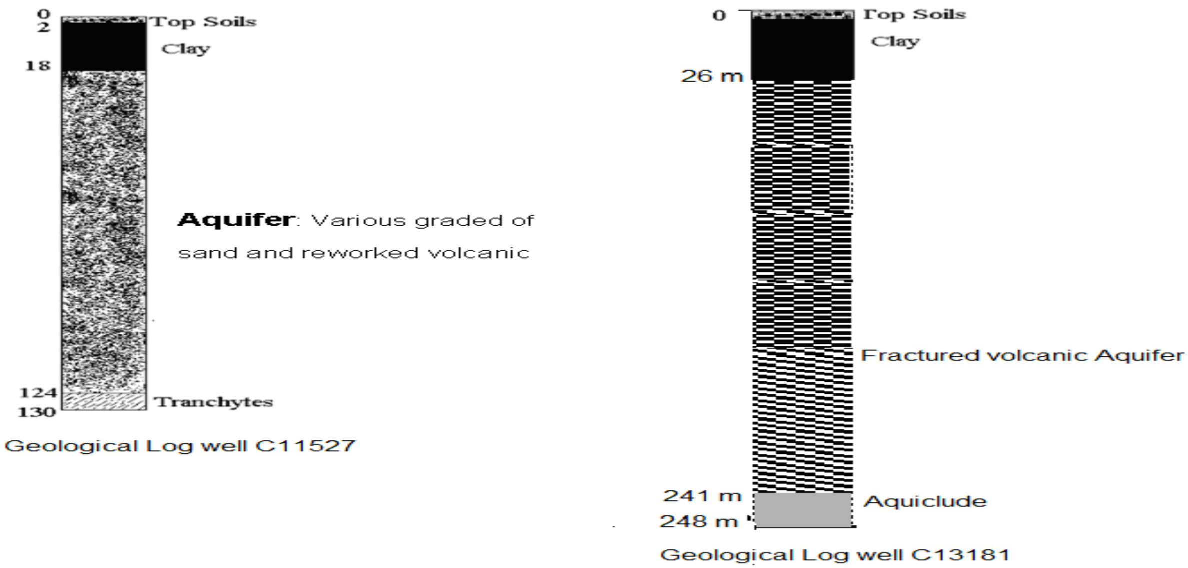


Figure 5:3 Geological Log of well C11527 (sedimentary unit) and well C13181 (volcanic unit)

5.2.2. Groundwater flow pattern

Groundwater flow system pattern and direction was analysed on the bases of the available heads. The natural groundwater flow of the area was inferred from Piezometer heads before the abstraction stress on the local aquifer was happened, roughly around 1980's. As indicated in Figure: 5.4 below, the ground water flow has two flow patterns: lateral flow pattern and axial flow pattern. The first component of lateral flow into the basin was from the western escarpment that dispersed into three directions. North east wards, South east wards and directly East wards that recharged the lake sediments. The second lateral flow was from the eastern escarpments that dispersed again into three arms: North West wards, East wards to the Lake and Southwest wards. Axially, the bulk of the flow from the lake was to the south through the Olkaria-Longonot volcanic complexes and North and North West towards the lake Elementa. The Northward flow from the lake towards and through the Eburru hills and Elmenteita Lake basin was discussed by (Clarke A.C.G. D. Allen, 1990)

Construction of the conceptual model

Two approaches can be used to construct a Modflow simulation: the grid approach and the conceptual model approach. The grid approach involves working directly with the 3-D grid and applying sources/sinks and other model parameters on a cell-by-cell basis. The conceptual model approach involves application the GIS tools in the Map module to develop a conceptual model of the site being modelled. The location of sources/sinks, layer parameters such as hydraulic conductivity, model boundaries, and all other data necessary for the simulation can be defined at the conceptual model level. Once this model is complete, the grid is generated and the conceptual model is converted to the grid model and all of the cell-by-cell assignments are performed automatically. The conceptual model approach is the most efficient approach for building realistic, complex models is the conceptual model approach... For this modeling the conceptual model approach was used. The GIS input maps were pre-processed using Arc GIS and ILWIS software.

5.3. Numerical model

5.3.1. Code selection

One of the necessary criteria for code selection is weather the code includes a water balance computation. A water balance calculation should be part of every modeling exercise (Anderson & Woessner, 1992). This is because the water balance involves computation of flows across boundaries, to and from sources and sinks and storage.

The Groundwater Modeling System (GMS) is a complete graphical user environment for performing groundwater simulations. The entire GMS system consists of a graphical user interface (the GMS program) and a number of analysis codes like Modflow. The GMS interface is developed by Aquaveo, LLC in Provo, Utah

Modflow is a 3-D, cell-centered, finite difference, saturated flow model developed by the United States Geological Survey (McDonald & Harbaugh, 1988). Modflow can perform both steady state and transient simulations and has a wide variety of boundary conditions and input options. GMS supports Modflow as a pre- and post-processor. The input data for Modflow are generated by GMS and saved to a set of files. These files are read by Modflow when Modflow is launched from the GMS menu. The output from Modflow is then imported for post-processing in GMS.

For this study GMS V.7 including Modflow 2000 was used for post and pre-processing. GMS v7 incorporates the lake package LAK3 which was utilized to estimate the water budget of the lake Naivasha.

5.3.2. Type and number of layers

The modelled domain covers an area of 1817sq km with two non-conceding aquifers. Layer one is unconfined with a thickness that varies from few meters to 100m. Layer two is semi confined aquifer with thickness varies from 30 to 220m. The first layer contains the lake. For simplification purpose the two layers are simulated as a horizontal continuous layer with an average thickness 60m for the first layer and 100m for the second layer. In order to reflect the non-conceding nature of the layers a low hydraulic conductivity was assigned to the boundaries of the first layer. A contrast of two orders of magnitude in hydraulic conductivity (Neuman & Witherspoon., 1969) between the aquifer and a non-permeable unit at the boundaries result in ignoring the horizontal flow in the layer (Anderson & Woessner, 1992).

To simulate the semi confined nature of the second aquifer a quasi-three dimensional modeling approach was assumed. In quasi three dimensional modeling approaches, the confining layer to the second aquifer is simulated by means of leakage terms. The leakage term is represents vertical flow between the two aquifers. Modflow 2000 now has a package called the layer property flow package (LPF). The LPF package relative to the BCF package is that there are now only two layer types: confined and convertible. A convertible layer is similar to the layer LAYCODE = 2 and layer LAYCODE = 3 types in the BCF package. The layer can be confined or unconfined depending on the elevation of the calculated water table. In this modeling conceptualization the above two layer were simulated as convertible layer.

5.3.3. Grid design

The model grid contains 104 rows, 120 columns and two layers. The horizontal spacing is uniformly equal to 500 meters. With a cells origin at 172251 m easting and 9896300 m northing, a total number of 24960 cells designed as 14562 active cells and 10398 in active cells.

5.3.4. The boundary conditions

The boundary condutions are represented if Figure: 5.4

Western part: The watershed boundary that peaks at the Mau scarp was taken to be a no flow boundary.
North-western part: The north western boundary spanned by the Eburru hills beneath which are acknowledged to be outflow to the Elmenteita Lake basin (Darling et al., 1996). In order to show the conceptual implication of the outflow through this boundary, the boundary condition is located behind the Eburru hills. This hydraulic boundary is simulated by General Head boundary with head elevation specified at 1830m
Eastern part: The South Kinangop fault trending due NNW is considered to impede most of the inflow from the Kinangop plateau. Most of the flux is considered to take place in the deeper horizons. Minimal in flow through this area has been considered negligible. The boundary is specified as no flow boundary.
Southern part: Flow through the Olkaria and Longonot volcanic complexes has been considered to be the conduit for most of the lake outflow from the basin most of which percolates into the deeper geothermal systems. To the southeast of the Longonot volcano, some considerable outflow from the basin has been considered to account for the fluxes from the Kinangop plateau that gets into the basin in a south-westward direction. These boundaries are specified as General Head boundaries fixed at 1800m (southern) and 1850m (south-eastern).
The basement: The bottom of the system has been considered to be composed of undifferentiated volcanic materials that have a very low ($K_{zz} \ll K_{xx} \approx K_{yy}$).
The Surface: The lake surface at the centre of the domain is a time variant boundary whose boundary has been defined within the Lake Package.
The two main rivers, Malewa and Gilgel, are time variant boundaries whose boundaries have been defined within the River Package.

5.3.5. The initial conditions

The initial conditions have been considered to be the hydrologic stresses (lake levels, River flows, evaporation, and precipitation) at the 1932 period. The initial groundwater levels have been derived as a long-term average value from 1932 to 1979 interpolated within the model to obtain the initial Pizometric surface.

5.3.6. Representation of the lake

The current coverage of the lake is approximately 110-120 sq. km (from ASTER image of 2009). However the initial (historical lake surface area was about 181 sq. km (from historical data at the year 1932). Therefore for model conceptualization purpose the lake surface area was set to the initial coverage as surveyed in 1957 by the Ministry of water Works (Kenya). The initial lake bathymetry was represented by the lake package Triangular networks (TIN) at the center of the model. The lake represented with a minimum elevation 1874m (at the crescent lake) and maximum 1896m (maximum stage that can be achieved during the simulation. In the steady state simulating the lake was assumed to have an initial stage of 1891(stage at the initial condition of the year 1932) minimum stage 1874m and a maximum possible stage 1896m

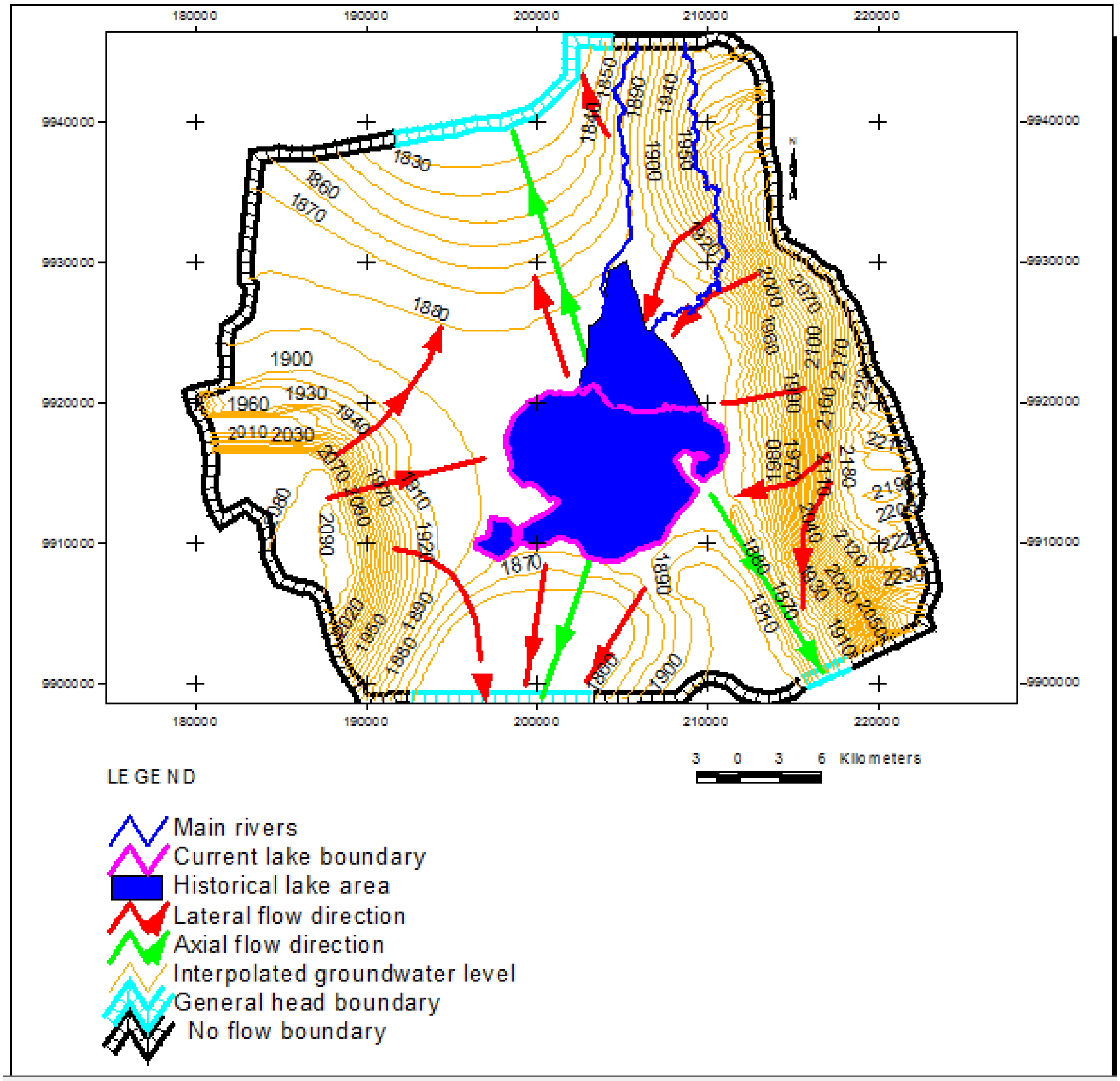


Figure 5:4 Representations of boundary conditions and the lake surface area

5.4. Steady state calibration

An important part of any groundwater modeling process is the model calibration process. In order for a groundwater model to be implemented in any type of management role, it must be established that the model can successfully simulate observed aquifer behaviour. Calibration is a process where a certain parameters of the model such as recharge and hydraulic conductivity are changed in a systematic fashion and the model is repeatedly run until the computed solution matches field-observed values within an acceptable level of accuracy.

5.4.1. Initial model execution

As discussed earlier in this modeling process the modeling approach followed is the conceptual modeling approach. In this approach data was entered through the conceptual model. After synthesizing and screening for errors, it converted/mapped in to Modflow layer and run. This process is a continuous process throughout the modeling process.

The initial hydrologic stresses (lake levels, River flows, evaporation, and precipitation) at the 1932 period were used for the steady state calibration. The initial groundwater levels calculated as a long-term average value from 1932 to 1979 was interpolated within the model to obtain the initial Pizometric surface. This was so done because (1) levels within this duration correspond to the natural stresses that were acting in the system then, (2) lack of enough data to adequately describe the Pizometric surface at the start of the simulation period. Depending on the problem and modeling objectivity, it may be appropriate to assume that water level measured during a certain period of time represents quasi steady state condition under stress that prevail during that period (Anderson & Woessner, 1992).

5.4.2. Steady state observation Data

Two types of observation data were used in the calibration process: water table elevations from observation wells and observed lake level in the year 1932. To consider the distribution and the density of observation wells is an important step in steady state calibration process. For example a key to success at using parameter estimation tools is to have a greater number of observations than parameters being estimated.

5.4.3. Steady state calibration procedure

During steady state calibration, the model parameters of aquifer hydraulic conductivity and riverbed conductance and Lake Leakance were estimated. The steady state calibration was accomplished using 41 observations and 3 adjustable parameters: hydraulic parameters layer 1 and layer 2 recharge layer 1, River bed conductance (river Malewa and Gilgel) and Lake Leakance (Lake Naivasha). The steady state calibration was accomplished by minimizing the difference between model-predicted steady state aquifer water levels and measured groundwater level and lake stage during the time of simulation.

5.4.4. Calibration techniques

There are two calibration techniques commonly used in groundwater modeling environment. Trial and error calibration and automated calibration. A trial and error method can be used to iteratively adjust model parameters until the model computed values match the field observed values to an acceptable level of agreement. In this approach parameter values are adjuster by the modeler in sequential model run to much simulated heads to the calibration target. In this modeling process a trial and error adjustment was made to some parameters known with a high degree of certainty and therefore requires only slightly adjustment. However because trial and error calibration influenced by the modeler's experience and biases, the method is less efficient to quantify the statistical uncertainty and reliability of the results. Anderson & Woessner(1992) express to this techniques as "unquantifiable" method to indicate the subjective of the method to the good judgment of the modeler.

Better Calibration can be achieved using an inverse modeling. An inverse model is a tool that automates the parameter estimation process. The inverse model systematically adjusts a user-defined set of input parameters until the difference between the computed and observed values is minimized. GMS contains an interface to an inverse model called PEST. The following sections describe the parameter estimation tools used for the model Calibration.

5.4.5. Parameter Estimation Tools (PEST)

PEST, a nonlinear, least-squares inverse modeling program developed by (Doherty, 1998) was used to calibrate the model. During calibration, PEST runs the Modflow model thousands of times, comparing model-predicted results with observations. After each model run, the objective function is analyzed to determine whether the model run was an improvement over the previous run. After each model run, PEST evaluates each adjusted parameter to determine the next best adjustment to that parameter. PEST then prepares the input data set for the next model run with the adjusted parameters, runs the model and re-evaluates the output. The goal is a weighted, least-squares optimization of the fit between the model-predicted values and the observations.

5.4.6. Model parameterization

Values of the Transmissivity/hydraulic conductivity characteristics of the aquifer systems by previous authors and recently calculated in this field work are listed in appendix: 2. Most of the aquifer parameters collected from pump test result that were basically derived from the geologic map and pump test result of the area, were in transmissivity values however the model requires hydraulic conductivity values. For this reason the field transmissivity values were converted in to hydraulic valued assuming an average thickness of 80m.These values were used to map the initial estimates for the zones. In the estimate for recharge an initial value of 20 mm/year (Nalugya, 2003)equivalent to about 4% of average precipitation within the rift area was assumed for the lake sediments. The western scarp area of Mau stretching north to the Eburru area was considered to have a higher recharge (Owor, 2000) Previous work by (Wiberg, 1976) estimates recharge at 50 mm/year equivalent to 7% of the average precipitation. (Mmbui, 1999) noted that the Pleistocene pyroclastic that flank the Mau and Aberdare escarpments appear relatively absorptive and doubtless transmit infiltrating precipitation and runoff to the underlying fracture and fissure systems of less absorptive and permeable rocks.

5.4.7. Defining the parameter zones

Parameter zonation was made for recharge (Layer: 1) and hydraulic conductivity (layer: 1) Figure: 5.5 and 5.6.Recharge was assigned only to layer one. Hydraulic conductivity for layer-2 was estimated as 0.1 m/day as initial values and was assumed to be uniform in the model area. The vertical hydraulic conductivity of the formations is not well known. It is often safe to assume 10% of the horizontal hydraulic conductivity.

For the delineation of parameter zones the zonation made by Owor(2000) was used as initial estimation for the distribution of both hydraulic conductivity and recharge. But later the distribution was adjusted to reflect the latest finding; recharge by Nulgya(2003) and 3-D conceptual model by Yihdego (2005) and Nabide (2002). In the implementation of the parameter zonation with PEST, the number of zones is a function of the number of observation points. The number of zones or parameters to be estimated should be less than the number of observation points. On the other hand the number of zones defined for each parameter was also an important consideration for running PEST. The more zones there were, the longer it took to perform optimization iteration.

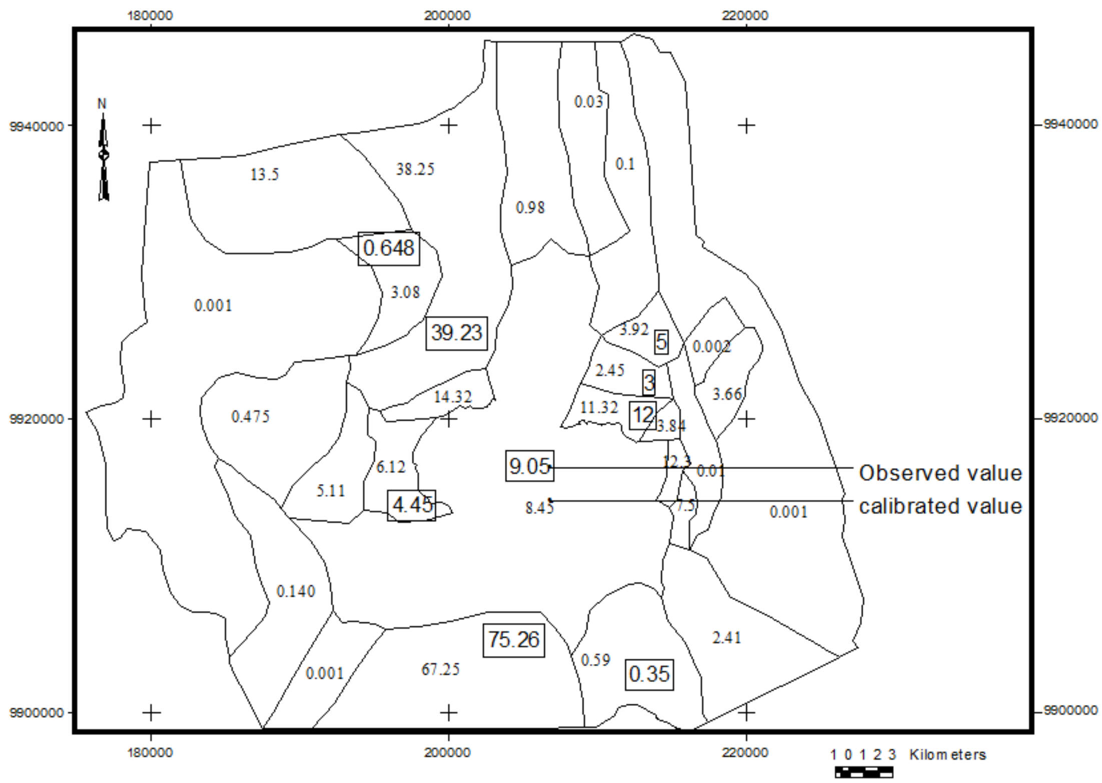


Figure 5:5 Parameter zones for Hydraulic conductivity (layer: 1)

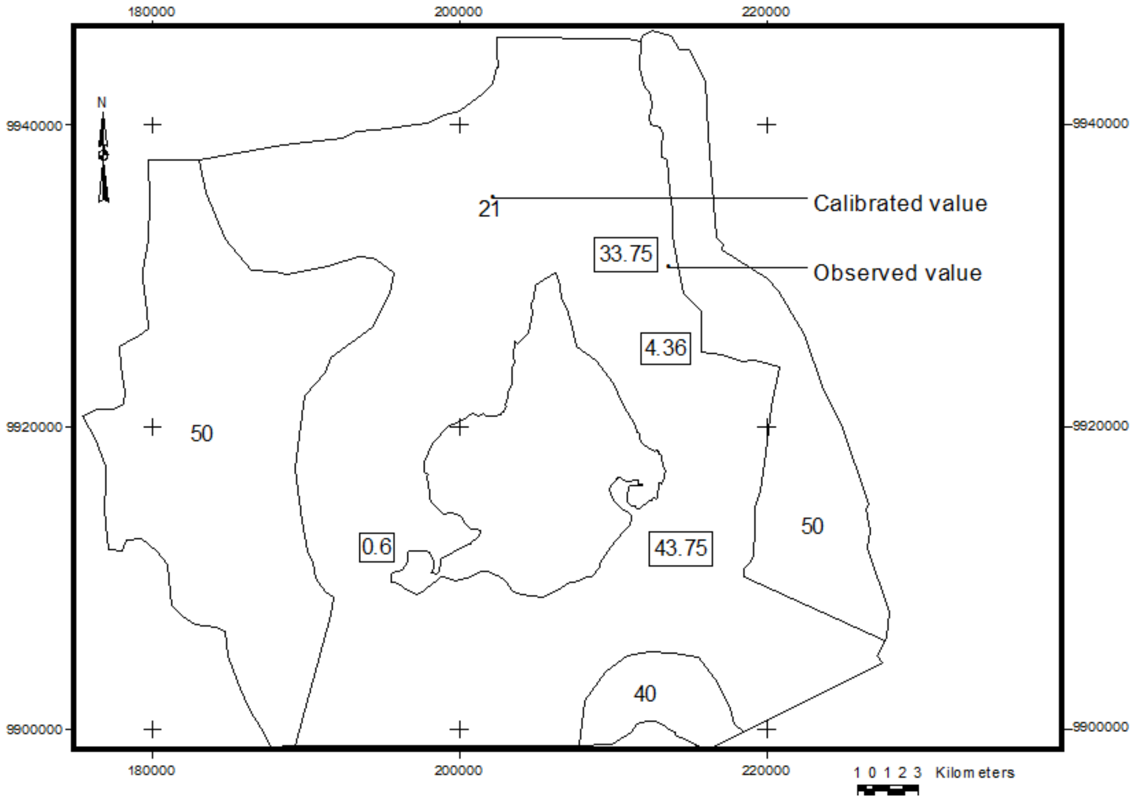


Figure 5:6 Parameter zones for recharge (layer: 1)

5.5. Transient model calibration

In transient problem the groundwater heads are a function of time. A transient simulation typically begins with a steady state initial condition and ends before or when a new steady state is reached (Anderson & Woessner, 1992). The simulated steady state aquifer heads were used as initial aquifer heads for the transient state simulation. Measure aquifer heads and the transient lake level were the important calibration targets for this run. The stress period step-size was increased to a monthly basis to match the available lake data. Mainly due to lake of date and few is known about it, groundwater evaporation from the saturated zone was not considered in this study.

5.5.1. Storage parameters

The transient simulation requires initial estimates for the storage coefficients of the aquifer. In the modeling conceptualization above the two layers were simulated as convertible layer, implies that the layer can be either confined or unconfined depending on the elevation of the computed water table. Storage parameter for convertible layers demands the assignment of both the specific storage (Ss) And Specific yield (Sy) for individual layers. However this parameter was the least known parameter in the study area.

In this study the values that were initially considered by previous studies were taken as initial run for the model. These values include: Specific yield estimated by (Wiberg, 1976) (0.0015), (Ojiambo, 1996) (0.0044), (Trottman, 1998) (0.12 for the unconfined aquifer, and 0.0001 for the confined aquifer), (Hermandez, 1999) (0.01-0.15) and (Kibona, 2000) (0.00146-0.00395 for the lake sediments)

5.5.2. Initial condition

The initial conditions refer to the head distribution everywhere in the system at the beginning of the simulation and thus are boundary conditions in time. The initial condition for the transient simulation was the steady state head solution generated by the calibrated steady state. Use of model generated head values ensures that the initial head data and the model hydrologic inputs and parameters are consistent (Franke, 1987).

5.5.3. Stress periods and time steps

Selection of the simulation time step is a critical step in transient model design because the value of the space and time discretization strongly influences the numerical results(Anderson & Woessner, 1992). Stress periods in Modflow are the blocks of time of variable length used in simulation of each time step. Stress periods and time steps should not be too large to miss important changes neither should they too small to take a long time to make unnecessary detail calculations in the system (Magombedze, 2002). As discussed in the analysis part of this study, data was available from 1932 to 2010 spanning over 79 years that included monthly lake levels, stream flow, and evaporation and precipitation data over the whole duration. The transient recharge was calculated based on the previous study by (Nalugya, 2003) to much the existing data in transient simulation. Transient water abstraction was estimated based on the irrigation area-depth relationship. To much the observed lake level, monthly stress periods were chosen with one time step each. Therefore the model was design with a total of 942 stress periods. And with a single time step uniform for each stress period. A trade-off had to be made with stress period and the time-step size with in each stress periods. More stress periods and more temporal variability in the calibration process, allow for a better fit between calculated and measured heads, but also make the calibration task more complicated and more time consuming because more stress periods and time steps need more input data and require therefore more processing time (W. Lubczynski & Gurwin, 2005).

5.5.4. Transient observation data

The lake level has been measured on monthly basis for the whole simulation time (1932-2010). Aquifer groundwater level was measured during this field work (Sep, 2010). A total of 14 water level data and the lake stage were .assumed to constraint the model solution. The well location and descriptions of the observation data is given in appendix: 4

5.5.5. Transient calibration target

Since the lake stage is a reliable record to use as a base station, the main calibration target considers the observed lake level as a base for the other well records. The groundwater level were measured in Sep, 2010, the last simulation time of the transient model was Jun, 2010. There is a time gap of two months between the end of the simulation time and the water level observation time. Although the time gap is considered, the water level data is also as second important calibration target for the model.

5.5.6. Transient calibration procedure

During the transient state calibration, the steady state model parameters, boundary conditions, aquifer hydraulic conductivities, riverbed conductance and Lake Leakance were kept constant. The steady groundwater outflow and inflow which are indication of the groundwater - lake interaction are calculated as a function of the head difference between the lake and the aquifer. The storage parameters, specific storage and specific yields were adjusted by try and error in each calibration attempts.

The transient model required a warm-up period of about five stress periods because observations during the initial stress period (1932) is partly dependent on events that occurred years prior to 1932. By using the ending steady state heads as the transient starting heads, the transient state were run as a steady state for five stress periods. This helps the model to introduced to and adjust itself with the successive stress periods in the simulation times.

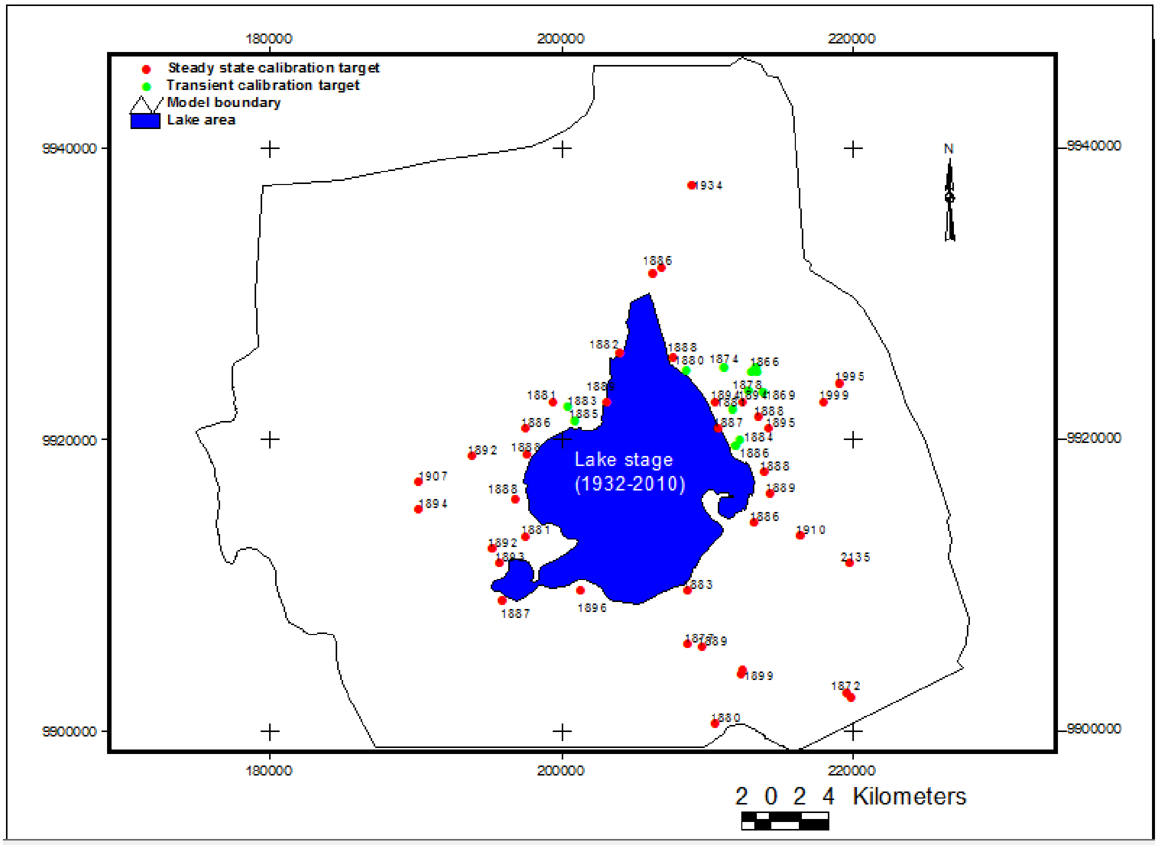


Figure 5:7 Location map of calibration wells

6. RESULT AND DESCASION

6.1. Evaluation of steady state calibration

The result of the calibration should be evaluated both qualitatively and quantitatively (Anderson & Woessner, 1992). The measured and simulated heads with their difference are listed in appendix 4. Model residuals (the difference between model-calculated and observed values) are generated by the PEST software, providing an indication of how well the model calculated values match the observed values. Model statistics for the steady state calibration indicate an overall R^2 =0.985 between measured and modeled aquifer water levels. The standard error for the aquifer water level estimates is 0.738m indicating that about 95 percent of the modeled aquifer water levels are within about -0.2m and 2.6m of observed values.

6.1.1. Steady state model calibration errors

The three ways of expressing the average difference between simulated heads (hs) and measured heads (hm) are: the (ME), Mean absolute error (MAE) and the root mean square error (RMS). The objective of the calibration is to minimize these errors.

The mean error (ME): is the mean difference between measured head (hm) and simulated heads

$$ME = \frac{1}{n}\sum_{i=1}^{n}(h_c - h_o)_i \qquad \text{Equation: 11}$$

The mean absolute error (MAE) is the mean of the absolute value of the difference in measured and simulated heads

$$MAE = \frac{1}{n}\sum_{i=1}^{n}|(h_c - h_o)_i| \qquad \text{Equation: 12}$$

The root mean square error (RMS) or the standard deviation is the average of the square difference in measured and simulated heads

$$RMS = \left[1/n\sum_{i=1}^{n}(h_m - h_s)i^2\right]^{0.5} \qquad \text{Equation: 13}$$

The R^2 value

In statistics, a value is often required to determine how closely a certain function fits a particular set of experimental data. R^2 values range from 0 to 1, with 1 representing a perfect fit between the observed and simulated aquifer heads, and 0 representing no statistical correlation between the data. The R^2 value (often referred to as the goodness of fit) is computed as follows

$$R^2 = \frac{\sum(\overline{m} - m)^2 - \sum(s - m)^2}{\sum(\overline{m} - m)^2} \qquad \text{Equation: 14}$$

Where

$\overline{m}$ = mean of observed lake level, m= observed lake level, s= simulated lake level

Table 6:1 Steady state error summary

Evaluation Criteria	Error value
Mean error	1.239
Mean absolute error	4.736
Root mean square error	5.729
R^2	0.985

6.1.2. Steady state model scatter plot of observed and simulated heads

Scatter plot of measured against simulated heads was used to show the calibrated fit. Figure: 6.1. The scatter plot are visually examined whether points in a plot show deviation from the straight line in a random distribution or have systematic deviation, where systematic deviation of the plots can indicate systematic error in adjusting the parameter values with in parameter zone. The scatter plot shows a R^2 = 0.985.

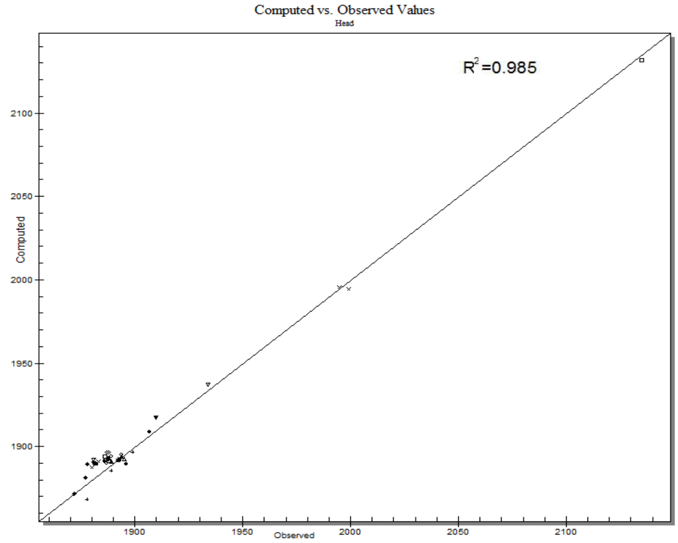


Figure 6:1 Scatter plot of computed Vs observed values

6.1.3. Steady state model water budget results

The steady state groundwater seepage from the lake amounts to 4.49 Mm^3/month and seepage into the lake is calculated as 0.22 Mm^3/month Table: 6.2. Significant inputs to the flow system of the area include recharge and lake seepage (sustained by the excess of the Surface inflow and precipitation over evapotranspiration).The outflow to the south and north of the basin provides the main outflow from the area.

Table 6:2 Steady state water budget for the Lake

Flow components	Inflow(Mm^3/month)	Outflow(Mm^3/month)
Precipitation	8.788	0.000
Evaporation	0.000	22.138
River/surface inflow	17.624	0.000
Groundwater inflow (Lake seepage in)	0.224	0.000
Groundwater outflow (Lake seepage out)	0.000	4.498
Total	26.636	26.636
Inflow-Outflow		0.000

Table 6:3 Steady state water budget for entire model

Flow components	Inflow(Mm^3/month)	Outflow(Mm^3/month)
storage	0.000	0.000
Head dependent boundary	0.000	7.125
River leakage	0.114	1.192
Recharge	4.014	0.000
Lake seepage	4.498	0.224
Total	8.625	8.541
Inflow-Outflow		0.084
Percent of discrepancy		0.010

6.1.4. Steady state model contour map of simulated heads

The contour map of simulated aquifer heads is shown in, Figure 6.2. Both layers have a similar spatial Pizometric configuration. The Pizometric heads around the vicinity of the lake mapped exactly as the observed values. Away from the lake horizons, the aquifer hydraulic properties have been affected by different factors like heterogeneity and lithostratigraphy effects. The head dependent boundary specified at the North, south and south eastern were calibrated as 1830m, 1800m, and 1850m respectively.

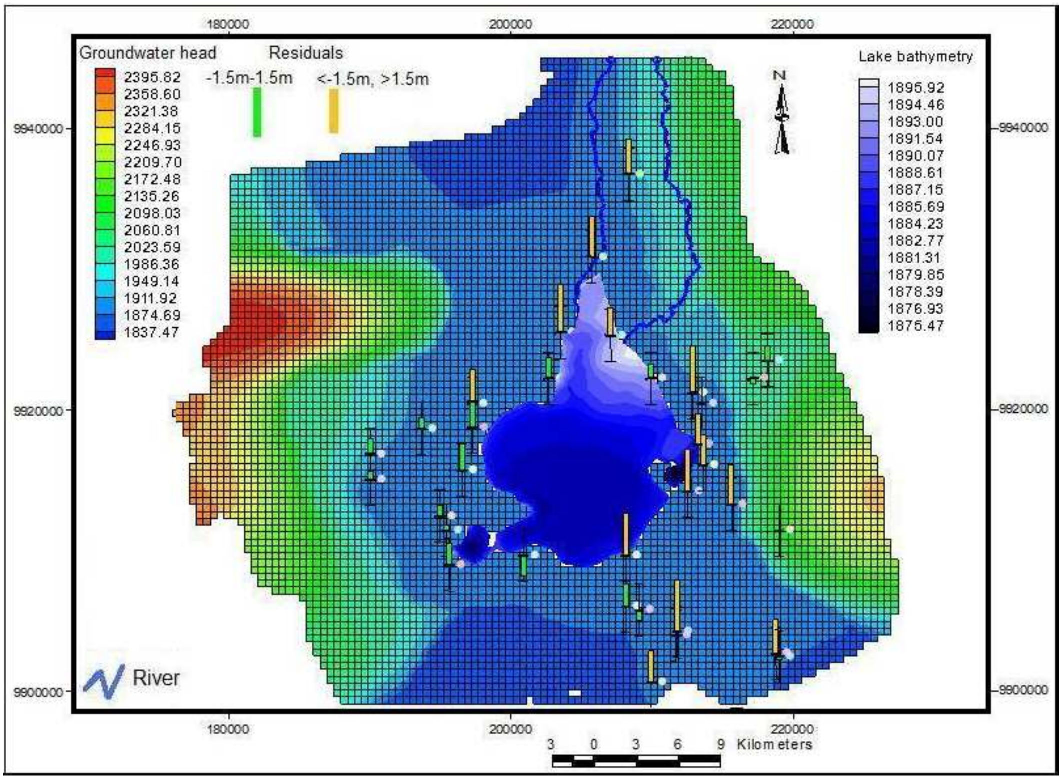


Figure 6:2 Contour map of simulated aquifer heads

6.2. Evaluation of the calibrated transient state mode

6.2.1. Transient model calibration Errors

Table 6:4 Summary of transient calibration error

Evaluation Criteria	Lake level	Aquifer heads
Mean error	-0.161	-0.932
Mean absolute error	0.387	1.54
Root mean square error	0.478	2.449
R2	0.905	0.732

6.2.2. Transient model time series plot of observed and calculated lake level

The transient model simulation results were evaluated by observing the measured and calculated aquifer and Lake Level. The transient simulation in this case includes 14 transient observation wells and the lake stage. Calculated aquifer head is presented in appendix: 5

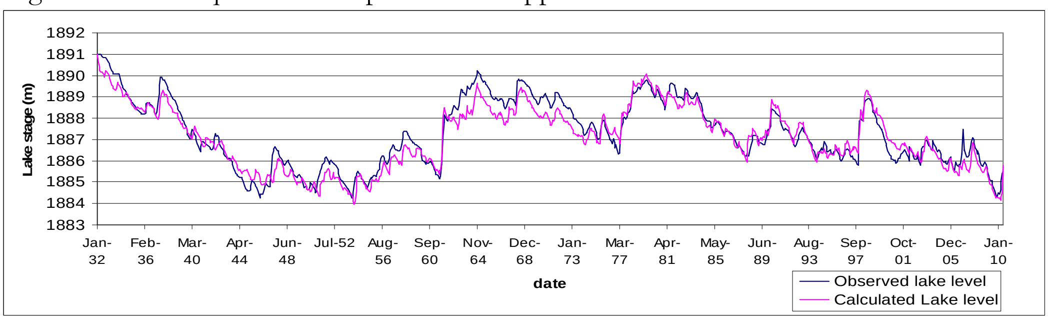


Figure 6:3 Time series plot of observed and calculated lake levels (1932 to 2010)

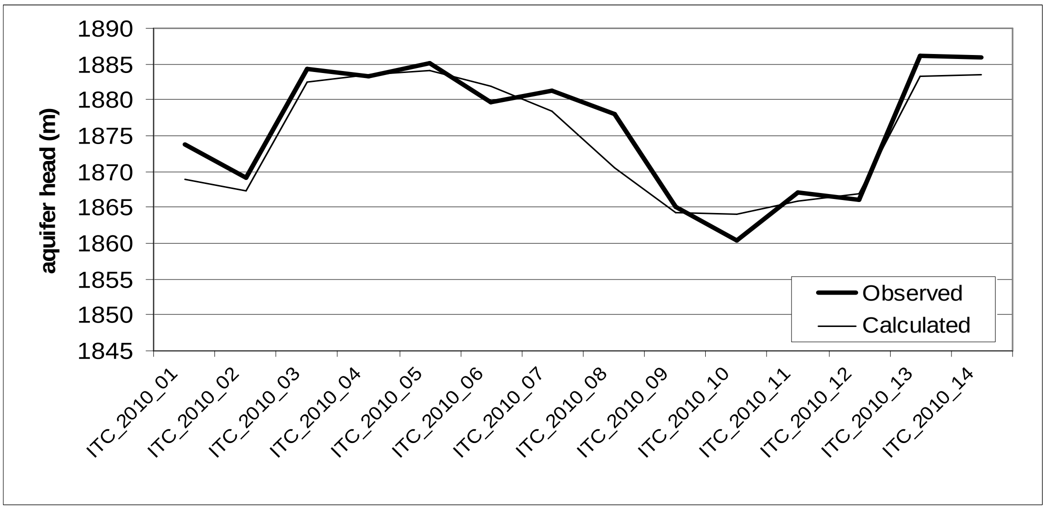


Figure 6:4 Observed and calculated aquifer head (Observation, Jun, 2010)

6.2.3. Transient model scatter plot of observed and calculated level

Transient scatter plot of measured against simulated aquifer heads and lake stage was used to evaluate the transient model. The scatter plot are visually examined whether points in a plot show deviation from the straight line in a random distribution or have systematic deviation, where systematic deviation of the plots can indicate systematic error in adjusting the parameter values with in a parameter zone.

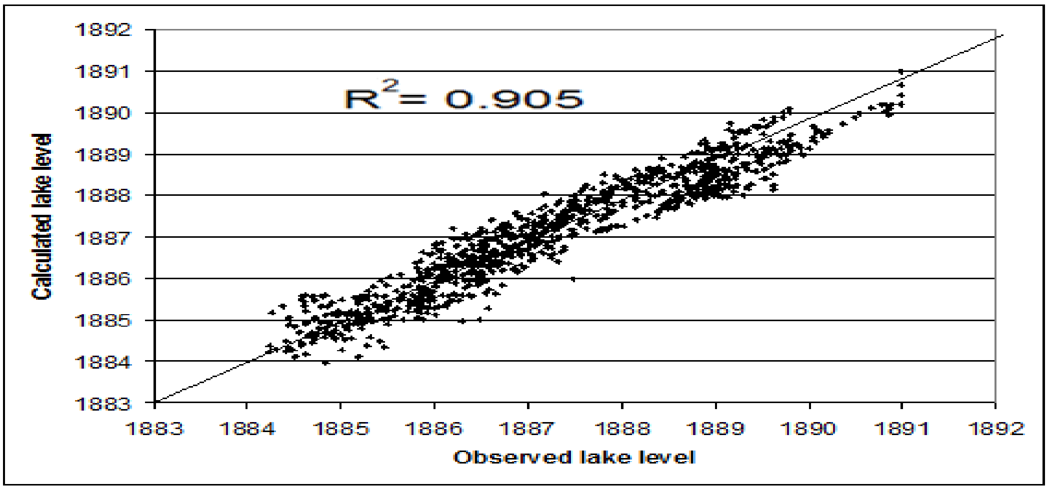


Figure 6:5 Scatter plot of observed and calculated lake levels (1932 to 2010)

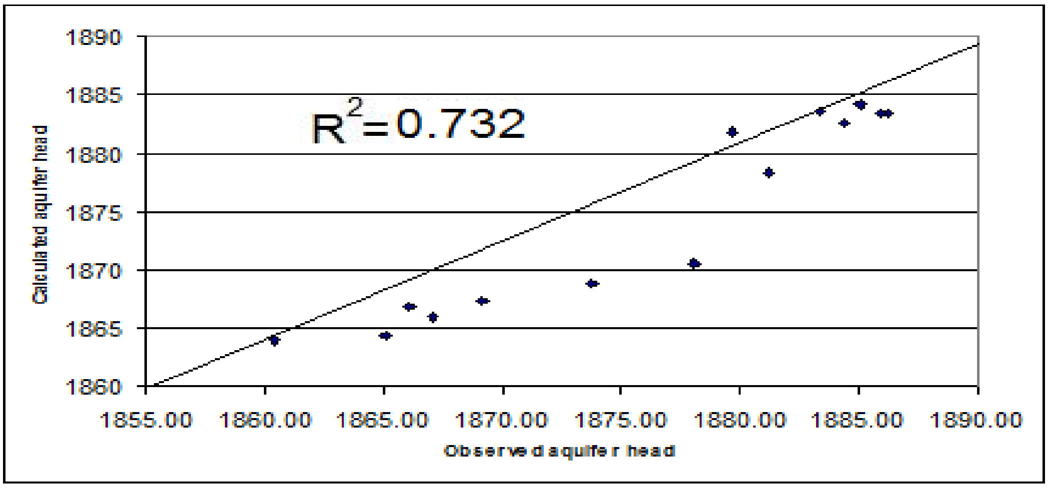


Figure 6:6 scatter plot of observed and calculated aquifer heads (Jun, 2010)

6.2.4. Sensitivity analysis

Sensitivity analysis is a measure of uncertainty in the calibrated model caused by uncertainty in the aquifer parameters and boundary conditions. The main objectify of the sensitivity analysis is to understand the influence of various model parameters and hydrogeological stresses on the aquifer system and to identify the most sensitive parameter(s) which will need a spatial attention in the feature studies. Sensitivity analysis was performed by systematically changing the calibrated value conditions (Anderson & Woessner, 1992). In similar approach, the sensitivity analysis for this particular study was performed by systematically changing aquifer and hydraulic parameters from calibrated values and evaluating the change on observed aquifer heads. Groundwater recharge and hydraulic conductivity (for the steady state model) and the storage parameter (for transient model) were each varied separately. In this procedure, the calibrated hydraulic conductivity and recharge values (steady state) and aquifer specific yield (transient state) were increased and decreased by a magnitude equivalent to 20%, 40% and 60% of the calibrated values. Root mean squire error (RMSE) was used as a statistical evaluation criteria.

The analysis result is presented below. The steady state model is sensitive to both recharge and hydraulic conductivity. The model is highly sensitive to increasing and decreasing of recharge. The model is also show strong sensitivity to a decreasing than increasing in hydraulic conductivity. On the other hand the transient model shows equal sensitivity to increasing and decreasing of the specific yield but with a slow response.

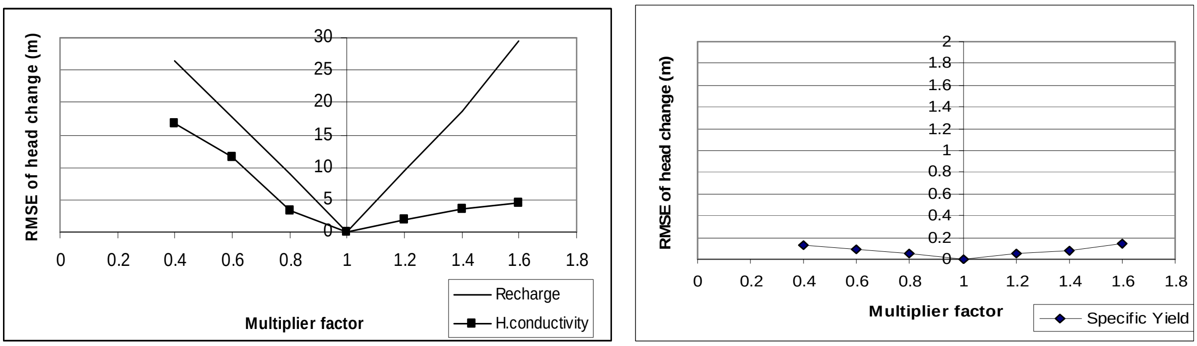


Figure 6:7 Sensitivity of the steady state model (left) and transient model (right)

6.2.5. Calibrated model parameters

Hydraulic conductivity

Steady state simulation results indicate a wide range in horizontal hydraulic conductivities from 0.1 to 75 m/day for the lacustrine sediments and 0.001-0.1m/day for the volcanic aquifer. The steady state model was calibrated to the initial states of the lake (lake stage =1891m) this means that the initial aquifer water level around the lake had to be higher in order to have the some head as the lake itself.

But later during the transient simulation, when the lake stage starts to decrease, the hydraulic head in the surrounding aquifer was not immediately respond to the stress on the lake. In other word, change in the lake flux does not bring the some effect on the lake shore aquifers. This means that the preliminary steady state hydraulic conductivity at the immediate vicinity of the lake had to be scaled up in order to allow lower hydraulic head equivalent to the lake stage during the transient simulation. For the some reason hydraulic conductivities along major outflow directions were also increased by 15-50% of the steady state calibration values. The adjustment does not bring a significant effect on the expected model solutions. Considering the sensitivity hydraulic conductivity in the model, Figure 6.8 the result indicates that the model is less sensitivity to an increasing in hydraulic conductivity.

Aquifer storativity

Initial values of aquifer storativity were estimated based on previous works. A values of specific yield in the range of (0.01-0.1) for the lacustrine sediments and (0.01-0.001) for the volcanic aquifer was calibrated. The corresponding specific storage values were in the range of (0.001-0.0001) adjusted during calibration. Storativity is often assumed to be uniform with in an aquifer or confined bed (Anderson & Woessner, 1992).The storage coefficient values in this study have been generally lumped into two main zones to the lake sediments/alluvial materials around the vicinity of the lake, and the reworked volcanic materials.

General Head boundary (GHB)

(Anderson & Woessner, 1992)Warns that Caution must be used in using the GHB to insure unrealistic flows do not develop. One of the effects of this package may the introducing of unlimited deeper recharge of groundwater over the model boundaries. In transient simulation Time-Variant Specified-Head package is rather preferred than the GHB to simulate specified head boundaries that can change within or between stress periods. However the GHB is found simple to implement, because it minimizes the model constrain in the process of calibration. One of the problems in the study area is the presence of boundary aquifer heads set below the elevation of the upper aquifer. The problem can result the development of unsaturated zone around the boundary heads which can be difficult to be simulated by Time-Variant Specified-Heads package.

In this model the application of GHB package is implemented along the entire outflow boundaries. The model was calibrated by introducing a constraint mechanism that allowing only groundwater outflow from the model. The calibration method gives interesting results where no un-necessary recharge was introduced in to the model Table: 6.3 and Table: 6.4. The result is also logical in the context of topographic set up of the lake. The lake elevation (1887m .a.m.s.l) is exceptionally higher relative to the physical boundaries which are set at elevation (1800m .a.m.s.l).

6.3. Simulation of lake-aquifer abstraction from the basin

As an initial input to the model, amount of abstraction was estimated by considering the relation between abstraction and area of irrigation lands. Area of irrigation was calculated by calculating time series Landsat images and depth of irrigation was estimated using different statistical analysis (see section 4.4 for the detail). Simulation of abstraction in this study aims at estimation of combined abstraction (abstraction from the aquifer and withdrawal from the lake). The simulation was evaluated according to two main scenarios, "with abstraction" and "without abstraction". The effect of the stress was evaluated based on the observed aquifer heads and lake stage at the end of the simulation time.

6.3.1. Scenario one: Without abstraction

Agricultural activities in the basin before 1980 were negligible. Owing to this reason all water loses from the lake during this period, save for loses due to evapotranspiration and groundwater outflow. This scenario assumes that the there was no abstraction at all before or after 1980. The model was made to run from 1932-2010 and the result was evaluated. The lake stage graph result shows a divergence around the 1980 Figure: 6.9. The effect is also evidence in the aquifer water levels where the calculated heads are higher Figure: 6.8. The difference between observed and calculated aquifer and lake level after 1980 was taken as important indication of the effect of abstractions from the basin for agricultural, industrial and domestic purposes

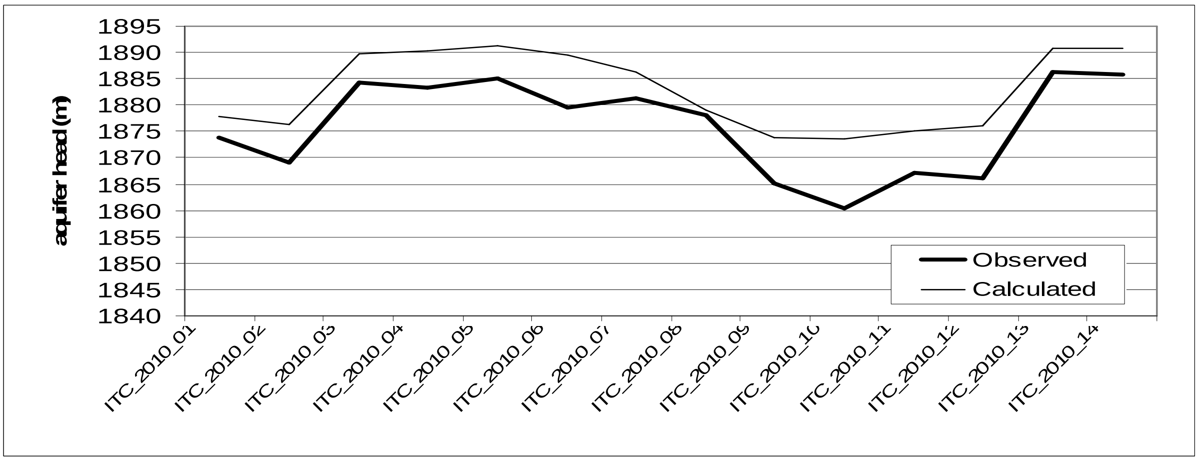


Figure 6:8 Observed and calculated aquifer head (abstraction not included)

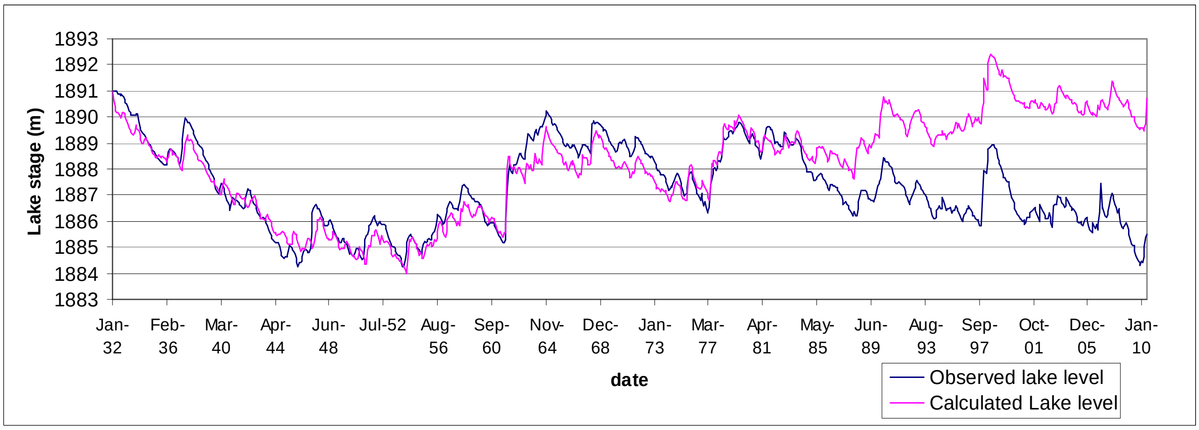


Figure 6:9 Time series plot of observed and calculated lake level (abstraction not included)

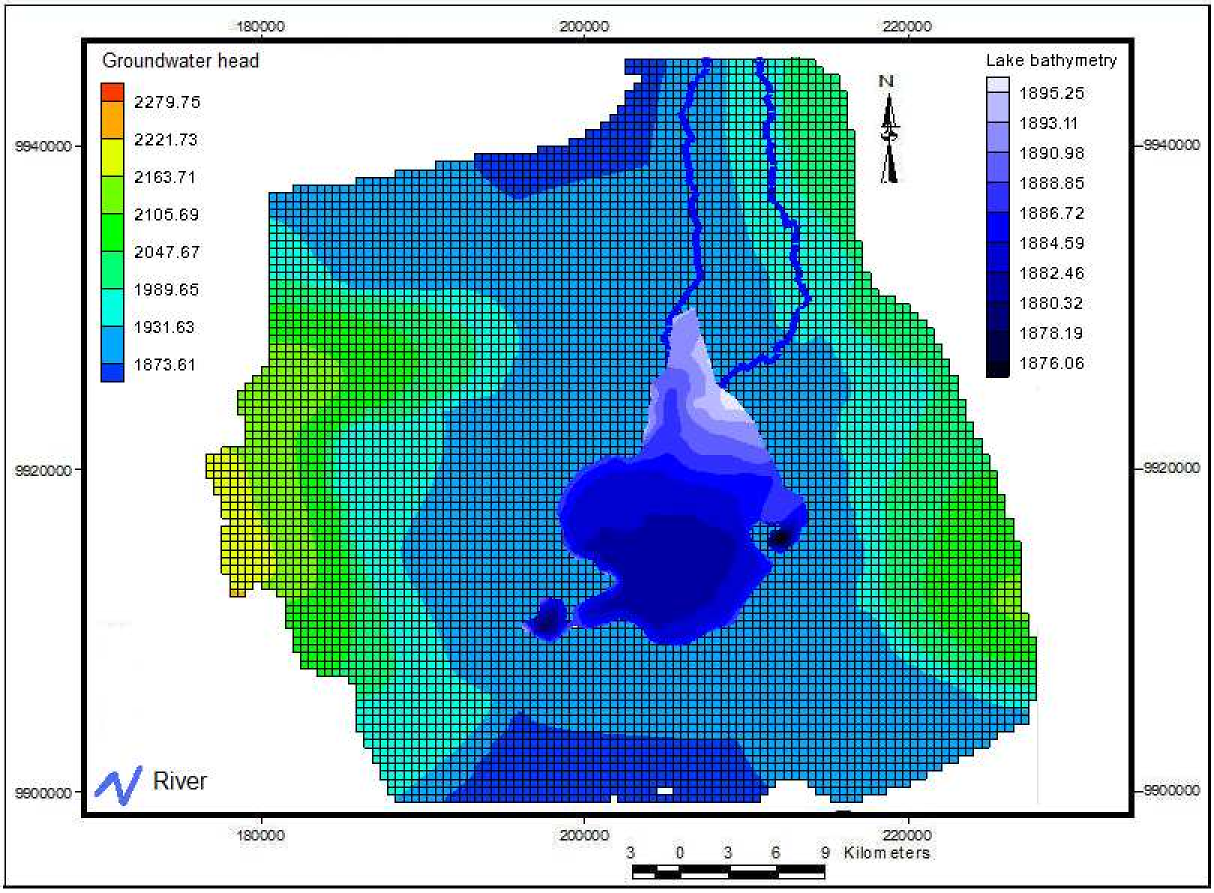


Figure 6:10 Groundwater contour map of simulated heads (abstraction not included)

6.3.2. Scenario two: With abstraction

According to the analysis on Section: 4.4, average abstraction from the basin estimated is about 6.25 *M m^3/month with a long term trend of abstraction ratio is 30% (groundwater) and 70% (lake water). In order to observe the simulation of abstraction on the lake the model application of this scenario was break down in to three.

1. All the abstraction was from the lake
2. All the abstraction was from the groundwater (aquifer)
3. Abstraction was in the ratio of 30% (groundwater) and 70% (lake water).

Lake abstraction from the lake is implemented by applied the abstraction stress directly on the lake. Well abstraction in the area is everywhere. However not all the single wells have measurable stress on the aquifer of the area. The main wellfield is found on the Northeastern part of the lake. The site is highly affected by groundwater abstraction since the history of agriculture activities in the basin.

In this scenario, the total stress of abstraction was applied to the aquifer. Implementation of the scenarios is made on seven wells. The wells are indicated on the ground by the main Pivot wells found in three well fields Figure: 6.12. Abstraction stress applied to all wells was the same, the total abstraction divided by the seven wells.

Wellfield A: well field: A includes (Three Point, Manera and Delamer site) the wellfield is the most exploited site in the area. The well field is represented by four wells.
Wellfield B: well field B includes all wells around Marula sites. The well field is represented by two wells.
Well field: C. Wellfield C includes all wells very near to the lake. The wellfield was represented by one well (assuming that there could be an option to pump water directly from the lake).

Response of the lake stage to the different abstraction schema is presented in Figure: 6.11.The deviation of simulated from observed water levels in the past three decades is distinct and indicates the magnitude of industrial abstraction. The deviation between the different curves, as a result the different abstraction stress in the area indicates the degree of interaction between the surface water bodies and the groundwater.

Abstraction estimation was needed to optimize in order to match the observed values. Optimization was made by adjusting the initial abstraction input values until the simulated aquifer and lake stage matches the observed values. The final optimization result allows estimate of an average combined abstraction from the basin to be about 7*Mm^3/month equivalent to 84 Mm^3/year since 1980. Previous estimation, 57*106 m3/year (Mmbui, 1999), and 60*10^6m3/year (Becht & Harper, 2002).

The result is in very good agreement to the estimates calculated from abstraction analysis Section: 4.4. The little increase in the current estimate supports the current condition that the groundwater level at the wellfield and the lake level are currently in the state of decreasing while abstraction is continuously increasing. The calculated abstraction has resulted in a lake which might have been 4.8m higher than was observed during the current field work, (Jun, 2010).

6.3.3. The contour map of simulated heads

The final simulation contour map of aquifer heads and lake stage shown in, Figure: 6.12. The long term groundwater level fluctuation of the area was between minimum of 1800.233m and maximum of 2337.773m. The long term lake stage fluctuation was between minimum 1885.8m and a maximum 1891m.The final lake stage is calibrated as 1885.8 the result was very close to the observed value which was 1885.45m. Low water level at the Northern, southern and southeastern limit shows the level specified at the boundary condition. The boundary heads specified at the boundaries were 1830m, 1800m, and 1850m respectively.

The most interesting part of this simulation is that the development of low groundwater level anomalies in the well field: A. while no enhanced abstraction effect was observed in the other two well fields. The second interesting result is that the cone of depression in this well field is remains confined to that specific location. During the simulation dry cells were also observed only in this particular wellfield. The result indicates that the wellfield is not in direct hydraulic connection to the main recharging water body, the lake. The dry cells during simulation indicates the applied abstraction stress is the maximum stress that the will field could carry.

The reason why a similar development of cone of depression is not enhanced in the other well field could have several implications. One of this could the fact that these well-fields are located relatively very near to the main recharging zones. The Boreholes in the high abstraction area are suspected of another source of recharge namely from rivers. According to previous studies the isotopic composition of Marula (well-field: B) signifies that these boreholes have their source of recharge from precipitation and river Malewa (Oppong-Boateng, 2001).This is also confirmed from the isotopic composition of unsaturated zone of Marula (d18O =-2.80 o/oo, d2H = -23.6 o/oo) which is a mixed of river Malewa and rain (Naulgay 2003).

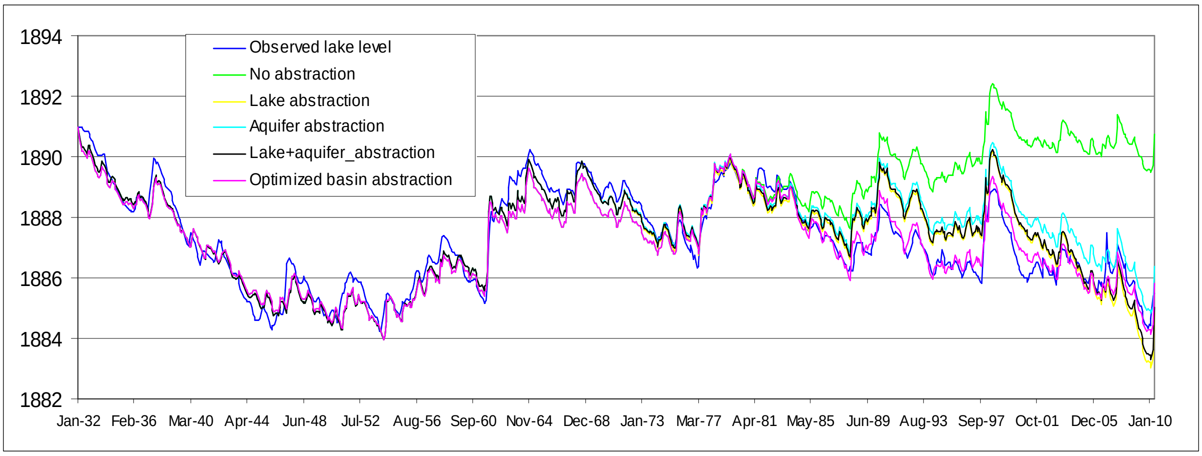


Figure 6:11 Response of the lake stage to the different abstraction schema

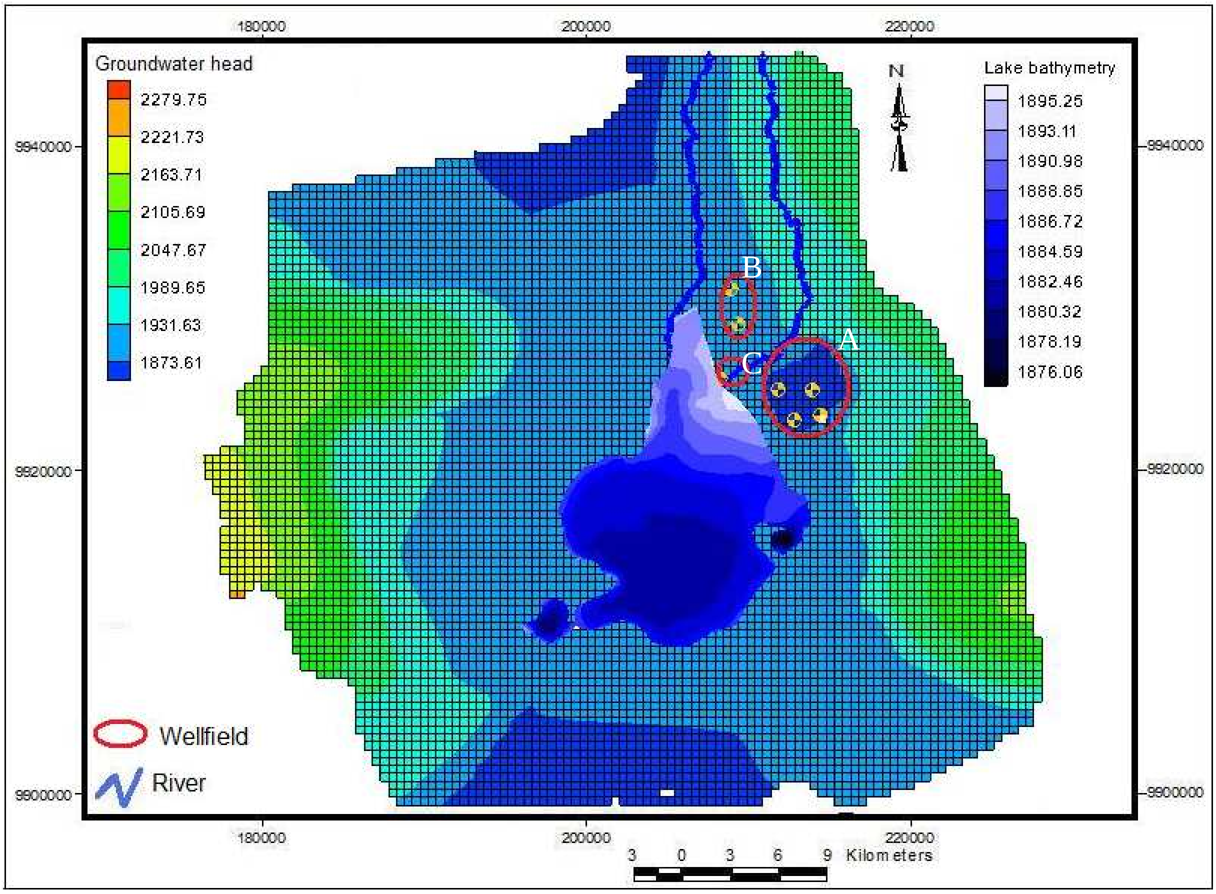


Figure 6:12 Groundwater contour map of simulated heads (abstraction included)

6.4. The long term water budget

6.4.1. Long term lake storage

The lake storage was calculated by Modflow using the stage-volume rating curve of Lake Package LAK3. Stage-volume rating curve was generated using the DTM-derived Lake bathymetry within the limitations of the DTM. The lake storage volume imitates the temporal lake level fluctuations Figure: 6.13 the average transient storage volume calculated is, 8.4 * 10^8 m^3/ month.

Figure 6:13 Simulated lake storage volume -stage-surface area relationship

6.4.2. Long term aquifer-lake interaction

There is a temporal fluctuation of the total amount of water seeping in to and from the lake to the aquifer. The maximum quantities of seepage to the groundwater are during those periods of consistent lake stage rise Figure: 6.14. The long term lake seepage out (groundwater outflow from the lake) was calculated as 5.56 *10^6 m^3/ month equivalent to 66.7 *10^6 m^3/ year.

The lake levels are sustained during those periods when the total inflow (stream inflow, runoff and rainfall) exceeds the total outflow (lake seepage abstraction and evapotranspiration). These periods are linked to consistent inflows into the lake possibly during the high rains after the dry spells. Similarly the long term lake seepage in (groundwater inflow in to the lake) was calculated as 1.1 *10^6 m^3/ month equivalent to 13 *10^6 m^3/ year.

The result is quite comparable to the result obtained by previous researchers. The previous results for the outflow term was 50 *10^6 m^3/year (Clarke M. C. G., 1990); 57 *10^6 m^3/year (Owor, 2000);
56 *10^6 m^3/ year (Becht & Harper, 2002) The relatively increase in the current estimation indicates the increase of lake water outflow as a result of intense abstraction in the study area.

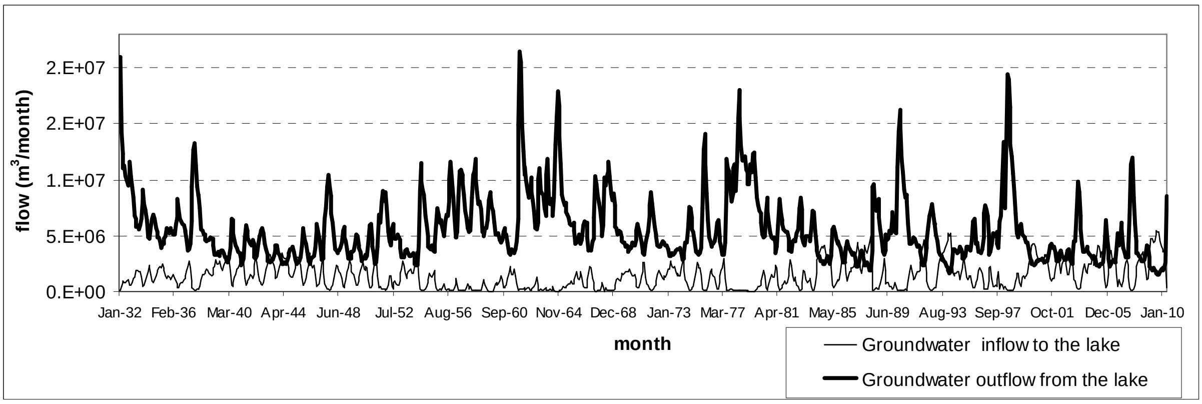


Figure 6:14 Graph showing long term ground inflow and outflow from the lake

6.4.3. The long term lake water budget

The long-term (1932 to 2010) lake water balances obtained from the transient model is shown in the Table: 6.5 below. The result indicates a long term net lake level fall of 5.4m resulted in a lake storage loss of 6.73 * 10^8 m^3 over the simulated period, 1932-2010. The result indicates that the lake is in equilibrium with a long term average precipitation 7.72 *10^6 m^3/month, evaporation 21.41*10^6 m^3/month, surface water inflow 19.36 *10^6 m^3/month, lake water abstraction 1.92 *10^6 m^3/month equivalent to 5.02*10^6 m^3/month, Lake seepage inflow 5.56 *10^6 m^3/month and Lake seepage outflow 1.1*10^6 m^3/month. At the end of the simulation period the final status of the lake (stage, volume, and surface area) is calculated as stage 1885.5m, volume 6*10^8 m^3 and Surface area 1.04 *10^2 km^2.

Table 6:5 Water budget for the Lake over the period 1932-2010

Flow components	Inflow(Mm3/month)	Outflow(Mm3/month)
Precipitation	7.722	0.000
Evaporation	0.000	21.414
Surface inflow	19.366	0.000
Lake water withdrawal	0.000	1.921
Groundwater inflow (Lake seepage in)	1.137	0.000
Groundwater outflow (Lake seepage out)	0.000	5.563
Total	28.225	28.898
In-Out		-0.673

6.4.4. The long term groundwater and lake water budget

A long term groundwater budget is prepared reflecting all water flow in to and out of the regional aquifer. Table: 6.6 show the overall water budget of the study area. The inflow components include recharge 2.8*10^6 m^3/month, river leakage-in 1.4*10^5 m^3/month and Lake Seepage-in (groundwater outflow from the lake) 5.56*10^6 m^3/month. The outflow components include well abstraction 7.5*10^5 m^3/month (equivalent to 2*10^6 m^3/month over the past 30 years), river leakage-out 2*10^4 m^3/month, Lake Seepage -out (groundwater inflow in to the lake) 1.1*10^6 m^3/month and groundwater outflow through the head dependent boundaries 6.7*10^6 m^3/month.

The long term average river water inflow in to the groundwater is 1.4*10^5 m^3/month. (Baher, 1999) calculate a similar amount, 1.5*10^5 m^3/month, of river seepage in to the groundwater using Darcy's equation. The river water balance shows that the river water inflow in to the groundwater is 75% greater than groundwater inflow in to the river. This result indicate that the two main rivers, Malewa and Gilgil, are dominantly loosing river and are recharging zones for the groundwater

Groundwater moves out of the model through the Head dependent boundaries with a long term average rate 6.7*10^6 m^3/month. The outflow to the south and north of the basin provides the main outflow from the area. The boundary conditions are simulated using general head boundary package. Inflow in to the model through head dependent boundaries is zero. The result indicated that the model (the Lake Naivasha basin) is ever loosing water to the surrounding aquifers.

The total long term average inflow in to the model is calculated as 9.8 *10^6 m^3/month. Similarly, the total long term average outflow from the model is calculated as 1.0 *10^7 m^3/month. The model water balance suggests that Lake Naivasha basin is in equilibrium with outflows about 1% greater than the inflows over the calibrated period of time (1932-2010).

Table 6:6 Water budget for the entire model over the period (1932-2010)

Flow components	Inflow(Mm3/month)	Outflow(Mm3/month)
Storage	1.295	1.359
Head dependent boundaries	0.000	6.770
wells	0.000	0.755
River leakage	0.141	0.020
Recharge	2.803	0.000
Lake seepage	5.563	1.137
Total	9.802	10.040
In-Out		-0.238
Percent of discrepancy		-0.013

6.4.5. Comparison with other water balance models

The interaction between the lake and the surficial aquifer in Modflow is represented in the Lake Package LAK3. By updating at the end of each time step, a water budget is calculated for the lake that is independent of the ground-water budget. The implementation of a separate water budget for the lake provides the capability to use the model to make a separate estimate of the budget of the lake and its relation to the aquifer through the seepage in and out from the lake.

The spreadsheet water balance model was used for comparison of the lake water balance. The model was used for evaluation of the lake water balance by several previous researchers, Mmbui(1999) and Becht & Harper (2002). The model uses mass conservation equation to sums up the net flow to the lake (from the precipitation, evaporation and stream flow and surface runoff terms) and obtains the change in storage from the previous month's storage. A new surface area is then derived and from which, using the rating curve, a new lake level is computed.

The spreadsheet cannot directly simulate abstraction stress applied on the aquifer. The model was made to run using a similar metrological time series input data as the Modflow lake package but of course excluding the abstraction information. The graphs Figure: 6.15 show a close match with the modflow lake package result indicating that there is no significant differences in the way the two models evaluate the stages from the given time step.

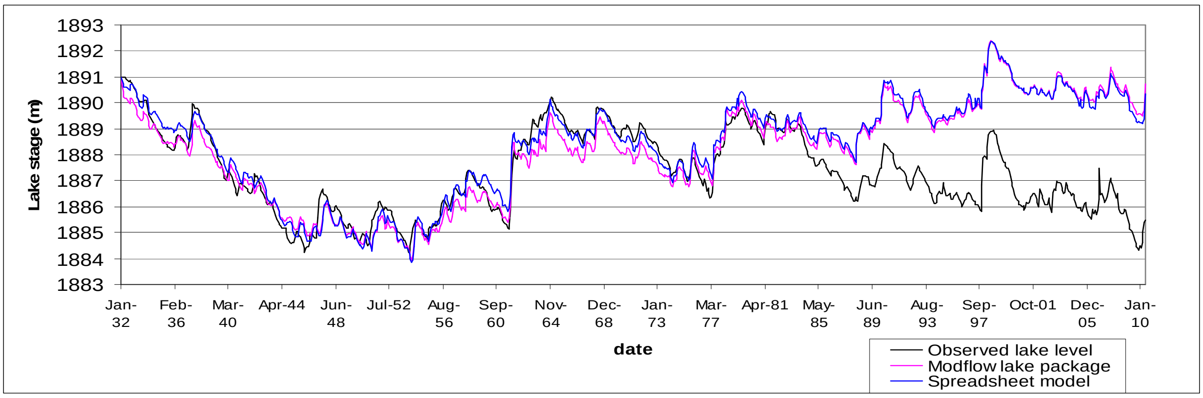


Figure 6:15 Compassion of results from Spread sheet and Modflow lake package models.

7. CONCLUSION AND RECOMMENDATION

7.1. Conclusion

The objective of calibrating of the steady state and 3-D transient groundwater model of the study area is achieved. Evaluation of calibration of transient model was highly constrained by observing the measured and calculated aquifer and Lake Level. The final calibrated model, implements the application of parameter estimation tools, PEST. The model matches the observed lake level with R^2= 0.985, steady state and R^2= 0.905, transient simulation. The transient model covers 79 years of lake-aquifer data (1932-2010) discretized in 942 stress periods, this is the longest ever calibrated ranges of years in the study of the basin.

The long term lake water balance was calculated by Modflow using the stage-volume rating curve of Lake Package LAK3. The long term average storage volume is 8.4 * 10^8 m^3/ month. The long term average fluxes in to the lake are precipitation 7.72 *10^6 m^3/month, surface inflow 19.36*10^6 m^3/month and groundwater inflow (Lake seepage in) 1.1*10^6 m^3/month. The long term average fluxes out of the lake: are evaporation 21.41*10^6 m^3/month, lake water abstraction 1.92 *10^6 m^3/month and groundwater outflow (Lake seepage out) 5.5*10^6 m^3/month. The lake water balances suggests that the lake is not in equilibrium with the inflow and outflow terms. The result indicates that a long term net lake level fall of 5.4m resulted in a lake storage loss of 6.73 * 10^8 m3 over the simulated period, 1932-2010.

There is a temporal lake-aquifer interaction in the study area. The long term lake seepage-out (groundwater outflow from the lake) was calculated as 5.5 *10^6 m^3/ month and the long term lake seepage-in (groundwater inflow in to the lake) was calculated as 1.1 *10^6 m^3/ month.

Using field abstraction data analysis and model simulation, the combined volume of lake-groundwater used for industrial abstraction since the last three decades was estimated. This requires an average abstraction amount 7.0 *10^6 m^3/month with a long term trend of abstraction ratio 30% (groundwater) and 70% (lake water) since 1980. The amount resulted in a lake which might have been 4.8m higher than was observed in the last stress period (2010).

A long term groundwater budget is calculated reflecting all water flow in to and out of the regional aquifer. The inflow components include recharge 2.8*10^6 m^3/month, river leakage-in 1.4*10^5 m^3/month and Lake Seepage-in (groundwater outflow from the lake) 5.56*10^6 m^3/month. The outflow components include well abstraction 7.5*10^5 m^3/month (equivalent to 2*10^6 m^3/month over the past 30 years), river leakage-out 2*10^4 m^3/month, Lake Seepage-out (groundwater inflow in to the lake) 1.1*10^6 m^3/month and groundwater outflow through the head dependent boundaries 6.7*10^6 m^3/month.The model water balance suggests that lake Naivasha basin is in equilibrium with a net outflow about 1% greater than the inflow over the calibrated period of time (1932-2010).

Model sensitivity analysis was made for aquifer hydraulic conductivity and recharge (steady state) and aquifer storativity (transient model). The result shows that the model is highly sensitive to increasing and decreasing of recharge. The model is also highly sensitive to decreasing in hydraulic conductivity but less sensitive to increasing in hydraulic conductivity. On the other hand the transient model shows increasing sensitivity to increasing and decreasing of the storativity values but with a slow response.

At the end of the work a comparison was made between Modflow lake-package water balance result and a spreadsheet water balance model result. The spreadsheet water balance model cannot simulate abstraction stress directly applied on the aquifer. Excluding the abstraction information, there was no significant differences in the way the two models evaluate the stages from the given time step.

7.2. Recommendation

Generally it is better to assume that the model is a regional groundwater model. For this reason, the model is best used for broad-scale predictions. At this point, the user should avoid the temptation to model localized impacts. A primary objective of the model development and calibration was the characterization of the interaction between the aquifer and the lake. The model can be used to provide a general sense of groundwater to surface water and groundwater to groundwater impacts in the basin. However, the model is best used for prediction of impacts to groundwater abstraction at a regional scale due to the uprising groundwater lake water usage in the basin.

The model is calibrated with coarse grid spacing which is fairly enough for a regional groundwater models. For detail modeling around the lake, further refinement to the model grid using the regional-to-local conversion method can give better accuracy to the model solutions.

The model water balance suggests that the basin is in equilibrium, with outflows greater than inflows over the simulated period (1932-2010). Extraction from water wells and withdrawal directly from the lake for agricultural purpose shows an increasing trend since the last three decades. Basin wise water resource management strategy can be design by integrating the regional groundwater model with other water evaluating soft wares. Ex. Water Evaluation and Planning (WEAP).

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Appendix: 1 Geodetic surveyed wells (2010)

ITC_ID	Northing	Easting	Elevation (m)	accuracy (m)	SWL (m)	Description
ITC_2010_01	9924923.294	211233.2822	1897.9849	0.082	24.2	BH7, reference station
ITC_2010_02	9923252.304	213852.3951	1909.666	0.003	40.5	Delamer 10A
ITC_2010_03	9919980.587	212251.547	1889.3676	0.0025	5	Test Bh1
ITC_2010_04	9922228.934	200458.1637	1923.8582	0.3982	40.5	Lodia limited
ITC_2010_05	9921265.358	200916.8954	1894.079	0.0082	9	Lodia limited
ITC_2010_06	9924723.226	208641.65	1897.4728	0.0022	17.78	Kenya nut(pivot 105)
ITC_2010_07	9922045.093	211751.5971	1892.3977	0.0134	11.2	Manera Bh4
ITC_2010_08	9923300.054	212932.3438	1890.035	0.0064	12	C4161
ITC_2010_09	9924862.17	213392.9933	1908.8915	0.1239	43.78	Panda BH3(BHD)
ITC_2010_10	9924910.378	213399.8137	1908.385	0.0102	48	Panda BHC
ITC_2010_11	9924562.84	213531.7768	1907.0672	0.1264	40	C11527
ITC_2010_12	9924645.053	213078.6487	1906.154	0.0965	40.05	Delamer BHM
ITC_2010_13	9919571.109	212036.3711	1886.5406	0.14	0.31	Slug test1
ITC_2010_14	9919564.315	212034.5959	1886.7504	0.1861	0.86	Test Bh3
ITC_2010_15	9926448.192	214880.0144	1961.104	0.1048	45	New BH1
ITC_2010_16	9924454.035	215439.0949	1940.6366	0.0848	40.63	Creative Bh2
ITC_2010_17	9924547.767	215406.3941	1941.6053	0.0954	40.63	Creative Bh1
ITC_2010_18	9923235.888	213891.3037	1909.9927	0.0027	36.7	Delemer 11B
ITC_2010_19	9919423.528	211825.1219	1886.5343	0.1129	0	Lake
ITC_2010_20	9921981.51	213411.7433	1898.0275	0.0019	17.4	Test Bh2
ITC_2010_21	9919014.073	216073.4475	1996.6692	0.0031	118.75	Naivash Water supply
ITC_2010_22	9921869.19	213698.2665	1899.5502	0.0026	19.5	BH John stone
ITC_2010_23	9924646.993	211662.4998	1904.0702	0.1088	24	Manera farm (BH6?)
ITC_2010_24	9924467.903	211874.8561	1904.5513	0.2239	dry	Old test bh
ITC_2010_25	9921482.386	202695.2235	1893.4861	0.2284	9.45	dug well
ITC_2010_26	9924790.561	209330.3976	1903.4394	0.0941	17.31	Kenya nut(95)
ITC_2010_27	9924894.645	213356.8879	1908.3756	0.0849	56	Panda BH4
ITC_2010_28	9924973.806	213328.9421	1908.894	0.1211	56.6	Panda BH5
ITC_2010_29	9925523.058	213722.6344	1915.3876	0.0946	34	Panda BHA
ITC_2010_30	9925011.115	213425.3737	1909.0467	0.1085	52	Panda BHB
ITC_2010_31	9921263.524	212983.1007	1893.0404	0.1419	11.5	dug well
ITC_2010_32	9925091.13	211934.9971	1911.6885	0.0045	35.9	Delamer BHO
ITC_2010_33	9919555.996	212044.0241	1886.4404	0.0026	0	Lake
ITC_2010_34	9927913.457	212868.1013	1915.1835	0.0019	30.7	Dray Training Institute
ITC_2010_35	9928025.108	212723.0743	1905.9599	0.2348		melewa river height
ITC_2010_36	9926841.27	212843.1553	1920.4787	0.0956	dry	Old Hund dug
ITC_2010_37	9920012.954	209934.5925	1888.0712	0.154	4	Test Bh4

Appendix: 2(a) Borhole information where Pumptest analysis was conducted

Borehole_No	Easting	Northing	Elev	Depth (m)	SWL (m)	PWL (m)	Discharge (m3/h)	Aquifer thickness (m)
Delamer BHO	213083	9924646	1906	81.00	35.50	64.81	104	30
Kreative BH1	215408	9924546	1946	133.00	40.63	48.78	21.00	60
(Morendat) BH1	209340	9924782	1929	95.00	15.78	16.50	24	36
(Morendat) BH2	208643	9924724	1904	105.00	17.78	24.14	207	60
Riftvalley BH1	212870	9927912	1920	130.00	40.90	88.35	30	54
Riftvalley BH2	212870	9927912	1920	120.00	30.70	48.64	70	42
Riftvalley BH3	212870	9927912	1920	120.00	30.20	102.76	68	42
Malewa BH	202214	9925931	2046	146.00	115.96	116.83	24	24
Sunshine BH	212613	9921328	_	110.00	18.00	28.00	48	36
Upendo village	218185	9915646	2100	220.00	157.60	167.72	6	30

Appendix: 2(b) Pumptest analysis result

Borehole_No	Cooper -JacopeStr line Transmissivity (m2/d)	Aquifer test Transmissivity (m2/d)	Aquifer test Storativity	Hydraulic conductivity(m/d)
Delamer BHO	183	131	0.0000115	4.37
Kreative BH1	77	73	0.0021	1.21
(Morendat) BH1	557	557		15.47
(Morendat) BH2	1281	1360	0.0253	22.67
Riftvalley BH1	5	9		0.18
Riftvalley BH2	59	79		1.87
Riftvalley BH3	14	14		0.34
Malewa BH	527	959	0.000122	39.96
Sunshine BH	141	267		7.42
Upendo village	16	13	0.0242	0.43

Appendix: 3 Summary of previous estimate of aquifer parameters

BH_No	Xcoord	Ycoord	Transmissivity
C1482	214316	9917024	1330
BH	207698	9925728	220
BH 1	212921	9923339	233.28
BH 3	212995	9923310	224.64
BH 4	212936	9923318	198.72
BH 9	211434	9921380	670
BH A	213712	9925550	1020
BH C	213459	9924929	1150
C1063	197600	9929926	38.9
C2071	202800	9909500	155
C2534	209050	9910000	166
C2557	195300	9912500	696
C2638	210050	9911100	166
C2657	193901	9913327	307
C2660	196950	9911950	166
C2701	195760	9909300	261
C2997	209900	9899950	21
C3924	205100	9908100	377
C4397	204900	9908300	1055
C4420	204800	9908250	671
C4500	198300	9914500	309
C4501	196100	9913900	267
C4989	208800	9909260	1382
C575	203050	9905900	6019
C579	201332	9911484	292
C630	197700	9906200	127
C630D	197700	9906200	3
KCC	209037	9925717	75
LB	214151	9920906	1000
UBH	203950	9909450	10660
PT1	210645	9911550	10640
PT2	210645	9911550	10640

Appendix: 4 Steady state calibration, Observed vs calculated heads

Field BH_code	Esthing	Northing	Observed head	Computed head	Residual
ITC136	219659	9902553	1872	1871.496	0.504
BLOCK INVEST. ESTATE	208754	9905957	1877	1880.084	-3.084
ITC102	199473	9909635	1878	1886.414	-2.414
AKIRA RANCH LTD.	219894	9902280	1878	1868.302	1.698
MARURA ESTATES	206889	9931768	1879	1887.329	-3.329
AKIRA RANCH	210617	9900426	1880	1887.796	-7.796
YUANMI	197610	9913329	1881	1888.471	-7.471
LOLDIA ESTATE	199465	9922547	1881	1886.267	-5.267
N24	204040	9925879	1882	1882.797	-0.797
LOLDIA LTD	205035	9920703	1882	1888.356	-6.356
ITC133	208751	9909641	1883	1887.84	-4.84
ITC042	207165	9925364	1886	1887.682	-1.682
ITC157	213271	9914310	1886	1890.517	-4.517
ITC082	206306	9931350	1886	1885.66	0.34
CHELINDA ESTATE	197605	9920698	1886	1887.699	-1.699
ITC043	210769	9920726	1887	1888.321	-1.321
ITC159	195974	9908951	1887	1887.399	-0.399
LONGONOT FARM/7473	212475	9904112	1887	1896.045	-9.045
ITC027	207680	9925645	1888	1888.173	-0.173
ITC074	213600	9921500	1888	1893.187	-5.187
ITC156	214009	9917763	1888	1889.42	-1.42
ITC160	196851	9915861	1888	1888.619	-0.619
ITC161	197660	9918954	1888	1888.336	-0.336
C1926	209700	9905700	1889	1884.065	4.935
ITC055	214375	9916225	1889	1890.648	-1.648
LOLDIA LTD	203174	9922549	1889	1886.648	2.352
C2557	195300	9912500	1892	1889.355	2.645
TARA	193897	9918859	1892	1889.265	2.735
ITC097	195762	9911480	1893	1889.073	3.927
C1404	190190	9915161	1894	1892.075	1.925
ITC076	212463	9922555	1894	1903.934	-4.934
MIN.OF LIV.VET.FARM	210603	9922554	1894	1894.302	-0.302
ITC084	214313	9920708	1895	1888.754	6.246
KAMERE ESTATE LTD	201333	9909636	1896	1886.424	3.576
ITC107	212412	9903826	1899	1896.011	2.989
C0466	190189	9917009	1907	1907.077	-0.077
N40	216441	9913361	1910	1914.542	-4.542
ITC047	208988	9937384	1934	1935.83	-1.83
N57	219112	9923847	1995	1999.914	-4.914
ITC058	218032	9922558	1999	1999.266	-0.266
ITC092	219888	9911496	2135	2131.418	3.582

Appendix: 5 Transient calibration, Observed vs calculated heads

Field BH_code	Easting	Northing	Observed head	Computed head	Residual	Description
ITC_2010_01	211233	9924923	1873.75	1868.828	4.922	BH7, refer station
ITC_2010_02	213852	9923252	1869.17	1867.264	1.906	Delamer 10A
ITC_2010_03	212252	9919981	1884.37	1882.576	1.794	Test Bh1
ITC_2010_04	200458	9922229	1883.36	1883.608	-0.248	Lodia limited
ITC_2010_05	200917	9921265	1885.08	1884.221	0.859	Lodia limited
ITC_2010_06	208642	9924723	1879.69	1881.815	-2.125	Kenya nut(pivot 105)
ITC_2010_07	211752	9922045	1881.2	1878.364	2.836	Manera Bh4
ITC_2010_08	212932	9923300	1878.11	1870.478	7.632	C4161, water supply
ITC_2010_09	213393	9924862	1865.11	1864.301	0.809	Panda BH3(BHD)
ITC_2010_10	213400	9924910	1860.39	1863.953	-3.563	Panda BHC
ITC_2010_11	213532	9924563	1867.07	1865.904	1.166	C11527
ITC_2010_12	213079	9924645	1866.1	1866.85	-0.75	Delamer BHM
ITC_2010_13	212036	9919571	1886.23	1883.407	2.823	Slug test1
ITC_2010_14	212035	9919564	1885.89	1883.417	2.473	Test Bh3