MODELLING GROUNDWATER STORAGE CHANGE

IN RESPONSE TO

FLUCTUATING LEVELS OF LAKE NAIVASHA KENYA

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FLUCTUATING LEVELS OF LAKE NAIVASHA, KENYA

by

Damali K. Trottman

This thesis is submitted in partial fulfillment of the requirements for the award of a Master of Science Degree in Water Resources Survey, with emphasis on Groundwater Resources Management, to the Water Resources Department , International Institute for Aerospace Survey and Earth Sciences (ITC), Enschede, The Netherlands.

Board of Examiners:

Prof. Dr. Ir. J. C. Van Dam (Chairman)
Delft University of Technology

Prof. Dr. A. M. J. Meijerink (Head, Water Resources Dept.)
ITC, Enschede

Drs. R. Becht (Supervisor)
ITC, Enschede

Drs. D. Kovacs (Director of Studies)
ITC, Enschede

Drs. M. Lubczynski
ITC, Enschede

INTERNATIONAL INSTITUTE FOR AEROSPACE SURVEY AND EARTH SCIENCES, ITC
ENSCHEDE, THE NETHERLANDS

ABSTRACT

Groundwater models are tools used to simulate the behaviour of groundwater in response to natural or artificially imposed stresses. In this study, a groundwater model is created to investigate the hydraulic interaction between Lake Naivasha (Kenya) and the surrounding unconfined aquifer, and to study the change in groundwater storage of the aquifer in response to fluctuating lake levels. Modelling is done using PMWIN, a finite difference groundwater modelling package. The relationship between the lake and the aquifer is found to be dynamic, with the lake both losing water to, and gaining water from the aquifer, depending on the relative head difference between the two. A lake water balance is computed for three climatic conditions, a wet year, a dry year, and an average year. The effect of lake changes on groundwater storage volumes are modelled under varying combinations of these climatic conditions. Groundwater storage is found to represent a significant percentage of the yearly lake balance, and creates a buffering effect against dry weather conditions.

ACKNOWLEDGEMENTS

As with any major undertaking, there are a number of people without whom this would not have been possible, or, without whose help, this would have been made more difficult than it should have been.

First of all, I express my gratitude to the Netherlands Government for awarding me a scholarship to come to ITC to study, and the Jamaican Government (through the National Water Commission) for allowing me to make use of this opportunity.

Thanks to all my lecturers for their willingness to patiently impart their knowledge, and the friendly attitudes they displayed. A special thanks to my Director of Studies, Mr. Dennis Kovacs, who was always there and always willing and eager to help in whichever way was needed.

Thanks to my supervisor, Mr. Robert Becht for his patience, persistence, and guidance, and to Professor Meijerink and Mr. Lubczyinski, who were willing to help whenever asked.

Mr. Polman and Mr. Mulder, my computer cluster managers, you made sure my computer was always up and running. You were very important people. Thanks as well to Benno and Job for all the photocopying and scanning and supplying of equipment.

There are some key people in Kenya whom I must thank for their assistance and cooperation: Bas van Helvoort (Kenya-Netherlands Wetland Training Programme), the WRAP team (Water Resources Assessment Project), Sara Higgins (lake Naivaisha Riperian Owners Association) and the people from the KWS Training Institute.

Of course I must say thanks to my classmates, the 'Naivasha Group'. We have worked together, complained together, shared data, and encouraged each other.

To my family, I am so grateful for your support, prayers and understanding over these two years.

Lord, without YOU, all of this would have been totally impossible! Thanks for looking out for me…

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1.0 INTRODUCTION

'Water is a limiting resource throughout most of Africa, but within East Africa it is most crucial in Kenya, both for the needs of the environment, as well as for food production, industry and increasing urbanization. Thus wherever water occurs, there is pressure upon it'.

1.1 Research Justification

Freshwater occurs in Lake Naivasha. Lake Naivasha is the highest of a series of rift valley lakes in Kenya, lying at an elevation of approximately 1883 masl. The lake is geologically very young, and is the remnant of a once large lake (12000-9200 B.P) which in former times expanded to include the neighbouring alkaline lakes, Nakuru and Elmenteita. The freshness of Lake Naivasha's waters is unexpected as the lake lies in a closed drainage basin and has no visible outlet.

Being the only freshwater lake in a region of otherwise alkaline lakes has made Lake Naivasha of great importance. The lake supports a wide range of economic activities in the Naivasha area. The continuous development of Naivasha Town is undoubtedly inseparably linked to the activities taking place around the lake.

Official recording of lake levels began in 1908. They have since fluctuated widely , by as much as 11.3 m (Tetley, 1948). The general trend is towards a decrease in levels, area and volume. This is cause for concern at both the local and national level as not only is the economy of Naivasha threatened, but Lake Naivasha is Kenya's second Ramsar Site and hence the Kenyan Government is under international obligation to properly protect and manage the lake and its resources. Proper management of the lake water resources is however not a simple task. Integral to any sort of management plan or process is an awareness of the quantity of the resource available and how this varies in time. Changes in the lake level and volume can be observed, but very little is known on the changes which occur in groundwater levels, and the contribution of groundwater to the available resources. An understanding of the role of groundwater can be achieved through groundwater modelling.

1.2 Research Objectives

The main objectives of this study are as follows:

- to study the hydraulic interaction between Lake Naivasha and the surrounding aquifer
- to use groundwater modelling techniques to investigate the groundwater storage behaviour of the aquifer as it relates to changing lake levels
- to quantify the contribution of groundwater as a potential water resource

1.3 Literature Review

Although numerous studies have been done on Lake Naivasha, none of these (as far as is known) have concentrated on observing the response of groundwater storage to changes in lake level, which is the focus of this study. The following literature review thus looks briefly at work done on the hydrology of the lake, and the lake water balance as these play an important role in influencing lake levels.

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, described Lake Naivasha as being '12 x 9 miles in size (19.3 x 14.5 km, i.e 280 km^2), shallow, fringed with papyrus, containing no fish, but lots of duck and hippopotami'. 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, 1993). Since then, various researchers have attempted to find a reason for the lake's freshness.

- Gregory (1921) suggested that the lake's freshness was due to an undiscovered underground outlet
- Nilsson (1938) proposed the lake's freshness was a result of water both entering and leaving the lake via underground seepage
- Sikes (1936) made the first statistical attempt to estimate a monthly and annual water budget for the lake, and estimated the magnitude of the proposed underground seepage. It is uncertain which methods he used, but he estimated water was seeping out of the lake at a rate of 43×10^6 m^3/yr (Darling et. al, 1990).

- McCann (1974), also created a lake water budget. His results are shown below. He did not include groundwater inflow in the budget. Evaporation rates were calculated using evaporation pan data, with a pan coefficient of one (i.e pan evaporation was assumed equal to evapotranspiration). His estimate of the rate of groundwater outflow was 34 x 10^6 m^3/yr.

INPUT (mill. cu. m/yr)	
Precipitation	
- over lake	90
- over swamp	42
Streamflow (gauged)	248
Total	380

a)

OUTPUT (mill. cu. m/yr)	
Lake evaporation	237
Swamp evapotranspiration	109
Groundwater outflow	34
Total	380

b)

Table 1.1 - Water budget (McCann, 1974)

The lake level fell during the 1980's prompting further research into its hydrogeology.

- Ase (1986) carried out a study of the lake hydrology which supported the underground seepage theory. He calculated that groundwater outflow from the lake ranged between 46-56 x 10^6 m^3/yr (Ase et al., 1986). This is comparable with the results obtained by Sikes.

Because the lake is located on the highest point of the rift floor, potential exists for groundwater seepage to occur not just to the south as is generally agreed upon, but also to the north.

- Darling et al. (1990) were able to indirectly determine (using stable isotope analysis and a water mixing model) the directions of subsurface outflow from the lake. They concluded that there is considerable outflow to the south (50-90% of lake outflow) and significantly less outflow to the north. (Their research suggests that the northerly outflow is confined to the area between Eburru and Gilgil, while the southerly outflow is between Olkaria and Longonot. See fig.1.1) This concurs with the work of Allen et al. (1989) who previously came to the same conclusion that most of the lake outflow ends up between Olkaria and Longonot.

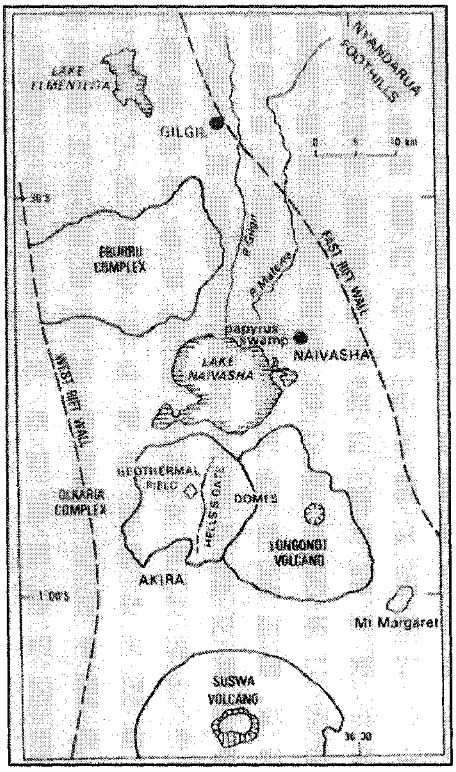


Figure 1.1 - Physical Features of the Lake Naivasha Section of the Kenya Rift Valley

1.4 Research Methodology

The approach taken in carrying out this study is summarized in the flow chart below.

DATA COLLECTION

↓

DATA PROCESSING

↓

GROUNDWATER MODELING

↓

DISCUSSION OF RESULTS

Figure 1.2 - Flow chart showing research methodology

A three week field visit was made to the study area (Oct. 6-27, 1997) to collect data. Much of the time was spent trying to install seepage meters in Lake Naivasha (to measure groundwater in/outflows), and conduct pumping tests. Sites were visited to take groundwater level measurements, and data downloaded from loggers.

The data collected was later processed, and the necessary maps and databases created, and the calculations of model inputs made.

PMWIN (Processing Modflow for Windows) was the groundwater modelling package used. A set of hypothetical situations were run in the model in order to achieve the objectives of the study.

Results of the modelling were discussed, and conclusions and recommendations made.

2.0 THE STUDY AREA

2.1 Location of Study Area

The area of study, Lake Naivasha, is located just south of the equator in the Kenyan Rift Valley, SW Kenya, between latitudes 36°15′ - 36°25′ E and longitudes 0°37′- 0°50 S. The lake has a surface area of approximately 127 km^2, and lies about 70 km NW of Nairobi.

Administratively, the Naivasha area falls within the much larger Nakuru District (7921 km^2), Rift Valley Province. At an elevation of roughly 1883 masl, Lake Naivasha is the highest of a series of lakes in the Kenyan Rift Valley, with lakes Nakuru and Elmenteita lying in close proximity.

There are a number of important locations to which reference is frequently made in this report. These are shown in fig. 2.2.

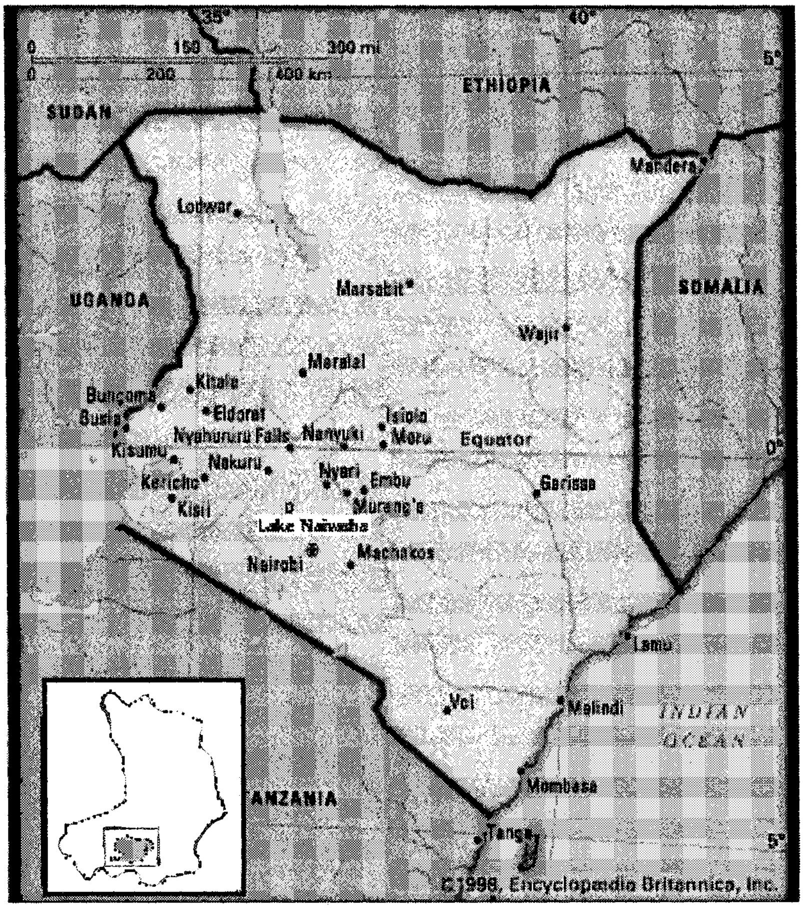


Figure 2.1 - Location of study area

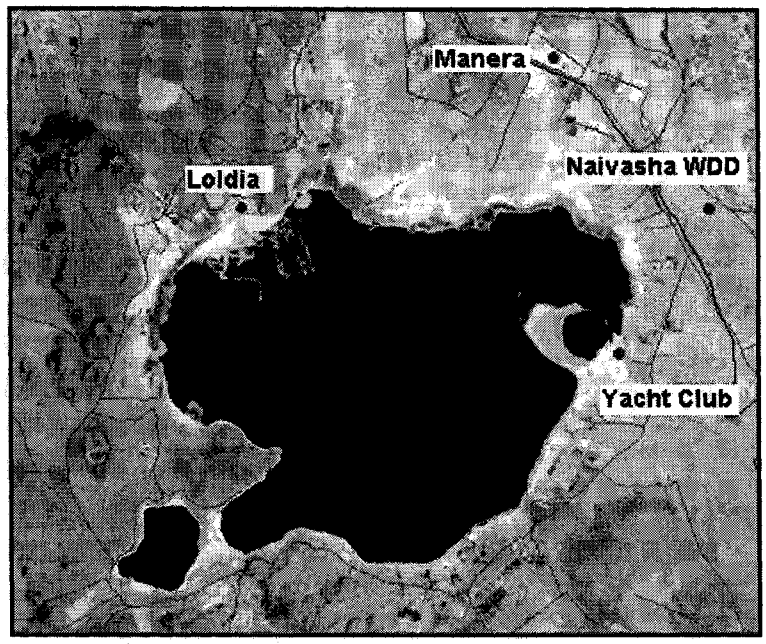


Figure 2.2 - Important locations

2.2 Geology & Geomorphology

The study region may be divided into three main geomorphologic units: the Mau Escarpment to the west, the Kinangop Plateau to the east, and between these two highlands, the Rift Valley plains (see fig.2.3). The region is geologically very young, with all rocks and structures having formed during the past 4 Ma (Min. of Energy, 1990). There were four major periods of volcanic activity (V1-V4) and faulting (F1-F4) which led to the present day situation (Baker et al, 1988).

EPISODE		ACTIVITY	AGE RANGE
V4		Late quaternary to recent salic volcanoes	0.4-0 Ma
	F4	Extensive minor faulting of rift floor	0.8-0.4 Ma
V3		Quaternary flood lavas of rift floor	1.65-0.9 Ma
	F3	Renewed faulting of rift margins	1.7 Ma
V2		Early quaternary flood trachytes	2.0-1.8 Ma
	F2	Formation of step faults (narrowing of graben	3-2 Ma
V1		Pliocene ash flows	3.7-3.4 Ma
	F1	Major faulting of eastern rift margin	4-3 Ma

Table 2.1 - Major Volcanic and Deformation Episodes (from Min. of Energy, 1988)

2.2.1 Mau Escarpment

The Mau Escarpment forms the western margin of the Rift Valley in the study region. Its height reaches over 3000 masl, and it has a N to NNW orientation. The escarpment is composed largely of soft, porous volcanic ashes and tuffs, with rare outcrops of agglomerates and lavas (Thompson et al, 1963). Downfaulted platforms with fault scarps up to 300 m separate the escarpment from the Rift Valley (Min. of Energy, 1990). Faults and scarps are difficult to trace either due to their being eroded, or covered with new material (McCann, 1974).

Unlike the Kinangop Plateau, the Mau Escarpment is not flat-topped, but rugged and deeply incised.

The main river draining the escarpment is the Marmonet. It fails to reach Lake Naivasha, instead recharging the alluvium of the Ndabibi Plain (see fig. 2.10). There is no drainage from the escarpment reaching the lake via surface water courses.

2.2.2 Kinangop Plateau

The westernmost part of the Kinangop Plateau forms the eastern margin of the study area. The plateau is a broad, flat plain which within the study area reaches a maximum elevation of about 2740 m. It is deeply incised by tributaries of the Malewa river, the largest river directly recharging Lake Naivasha. Like the Mau Escarpment, the plateau is downfaulted in a series of steps, and is separated from the rift floor by a fault scarp, the NNW trending South Kinangop fault scarp (100-20 m high) (Min. of Energy, 1990). The plateau is composed of soft volcanic rocks.

2.2.3 Rift Floor Plains

The rift floor plains in the Naivasha basin (40-50 km wide) form part of the Gregory Rift Valley, which contains numerous volcanic cones and craters, scarps and lakes (see fig. 1.1). Longonot volcano lies to the northeast of the lake, and Eburru volcanic complex to the southwest.

The Naivasha basin is largely covered with sediments derived from erosion of the surrounding volcanic rocks of the rift margins. These sediments were deposited in a lacustrine environment during the Gamblian stage of the Pleistocene period, and are commonly referred to as Gamblian lake sediments. Despite their extensive distribution, the lake sediments are not thick, and rarely exceed 30 m (Thompson et al, 1963).

The lake sediments are composed of sand and pebble beds, and quite commonly, gravel comprised of rounded pumice clasts (Studdard et. al, 1995). Volcanic rocks underly the lake sediments. Quaternary alluvial sediments also form a minor part of the Naivasha basin cover.

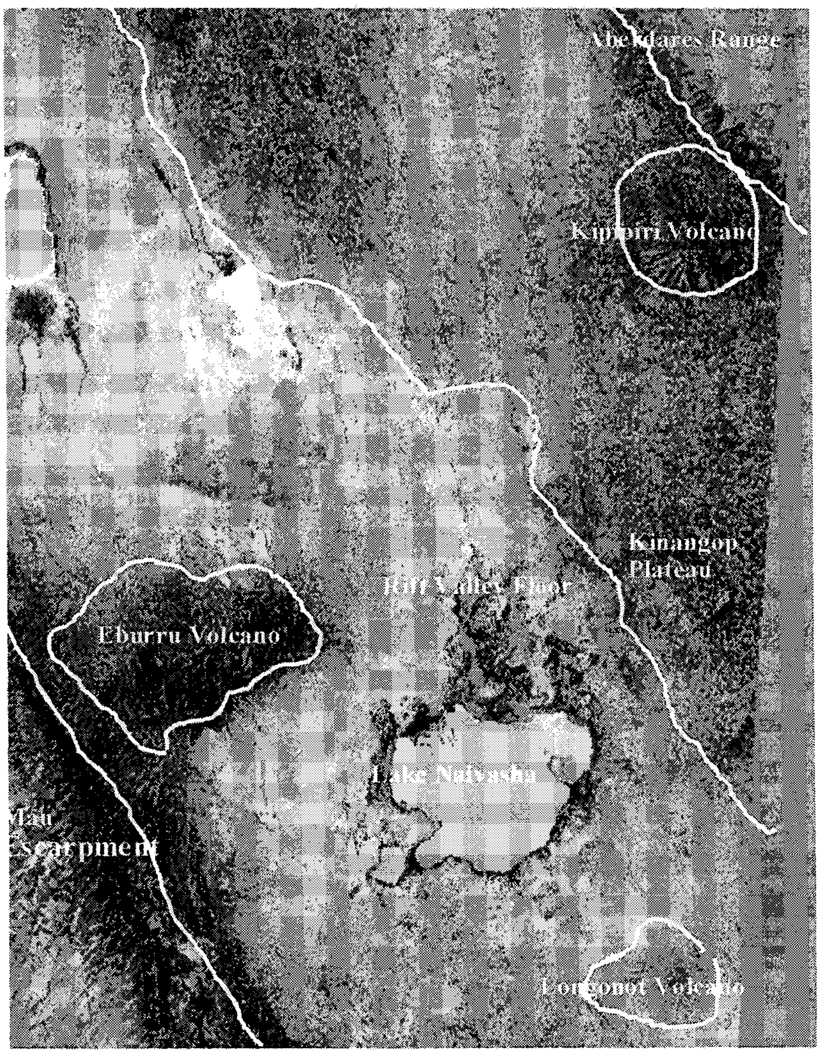


Figure 2.3 - Main geomorphological units in the study area

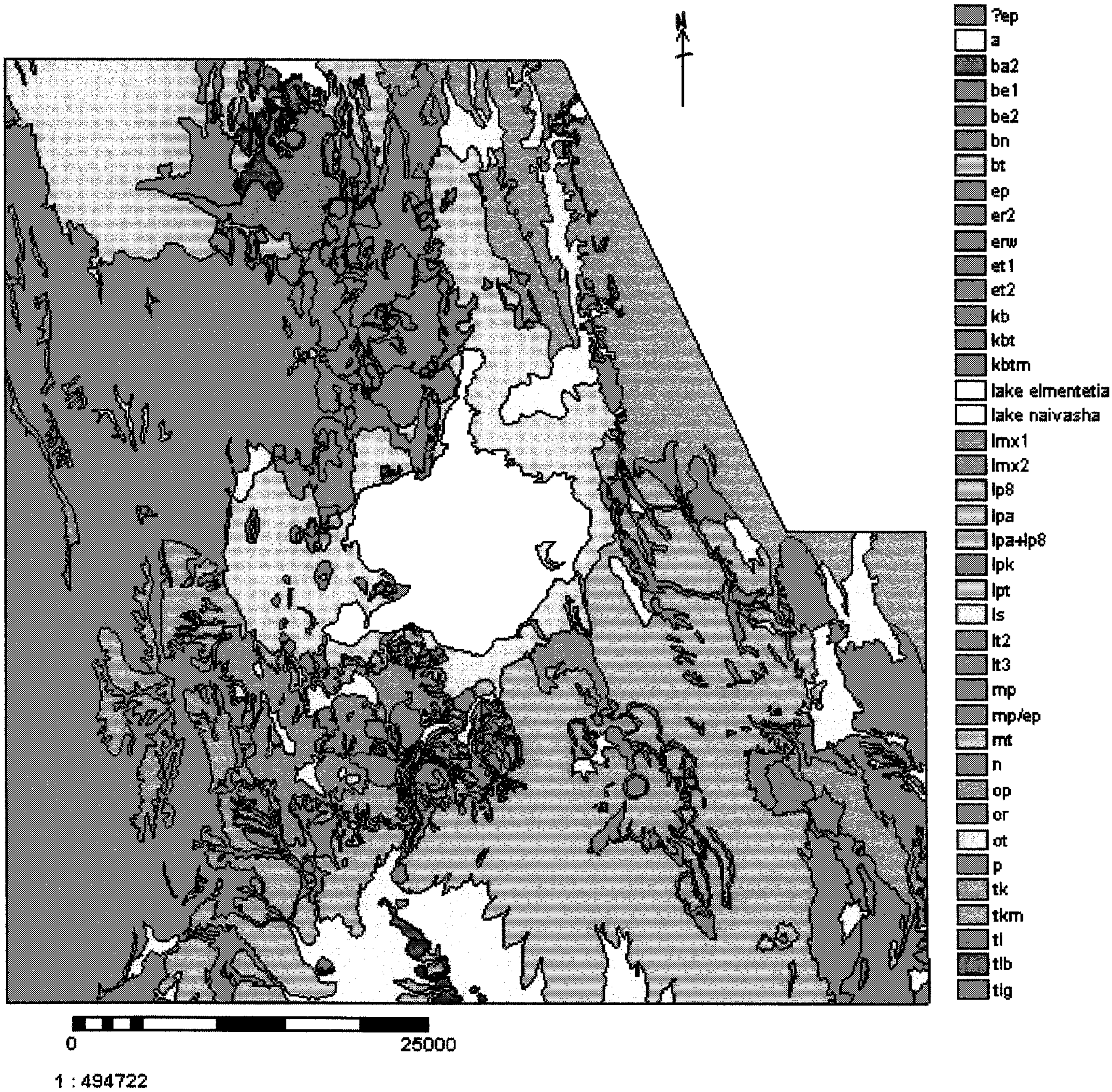


Figure 2.4 - Geological map of the Study Area

Legend for the Geology Map

Unit	Description
?ep	Eburru pumice; pantellerite, trachyte pumice, ash fall deposits
a	Alluvial deposits
ba2	Alkaria basalt; basalt and hawaiite, lava flows, pyroclastic cones
be1	Older elementetia basalt, hawaiite lava flows, pyroclastic cones
be2	Younger elementetia basalt; basalt, hawaiite and mugearite/benmoreite lava flows, pyroclastic cones
bn	Ndabibi basalt, hawaiite lava flows, pyroclastic cones
bt	Surtseyan/strombolian ash cones
ep	Eburru pumice; pantellerite and trachyte pumice, ash fall deposits
er2	Eastern eburru pantellerite; lava flows, pyroclastic cones
erw	Waterloo ridge pantellerite; welded and unwelded pyroclastics
et1	Older eburru trachyte; lava flows and pyroclastic cones
et2	Younger eburru trachyte; lava flows and pyroclastic cones
kb	Kijabe hill basalt
kbt	Surtseyan tuff cones
kbtm	Surtseyan tuff cones with laterally equivalent fall tuffs
lmx1	Lower longonot mixed basalt/trachyte lava flows and pyroclastic cones
lmx2	Upper longonot mixed basalt/trachyte lava flows, pyroclastic cones
lp8	Longonot ash
lpa	Longonot akaria pumice
lpa+lp8	Longonot ash and akaria pumice
lpk	Kedong valley tuff; trachyte ingimbrites and associated fall deposits
lpt	Longonot volcanice; pre-caldera welded pyroclastics, lava flows
ls	Lacustrine sediments
lt2	Lower longonot trachyte; lava flows and pyroclastic cones
lt3	Upper longonot trachyte; lava flows and pyroclastic cones
mp	Maiella pumice; trachyte, pantellerite pumice, ash fall deposits
mp/ep	Maiella pumice/ ebrru pumice
mt	Magaret trachyte; unwelded and welded pyroclastics
n	Ndabibi comendite lava flows, domes, pyroclastic cones
op	Olkaria comendite; pyroclastics (include pre-lpk lacustrine sediments, reworked pyroclasitcs in ol Njorowa gorge)
or	Olkaria comendite; lava flows and domes (includes ol njorowa pantellerite lava and welded Pyroclastics)
ot	Olkaria trachyte; lava flows
p	Ndabibi pantellerite lava flows
tk	Kinangop tuff (eartern rift margin)
tkm	Mau tuff (western rift margin)
tl	Limuru trachyte
tlb	Karati and ol mogogo basalts
tlg	Gilgil trachyte

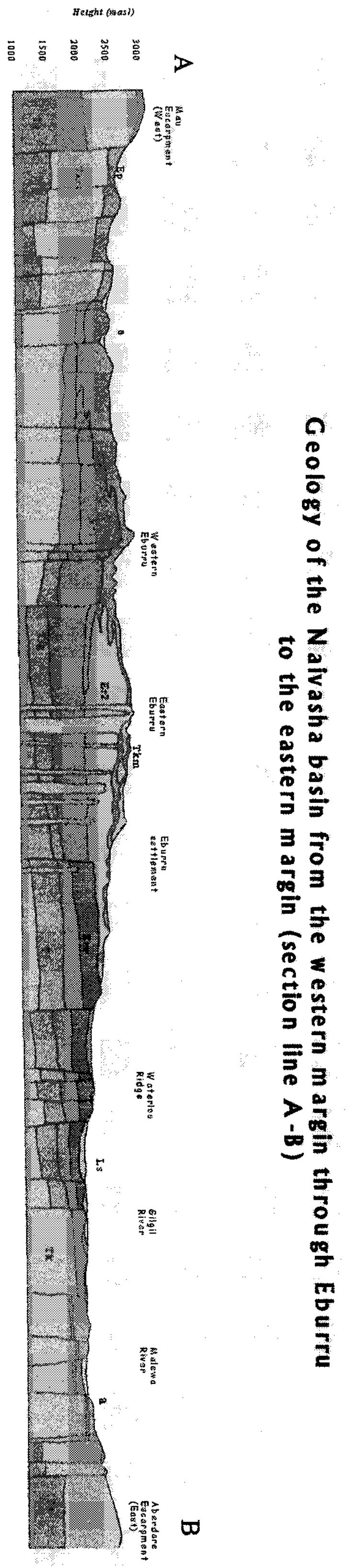


Figure 2.5 - Geological cross section across section line A-B

2.3 Climate

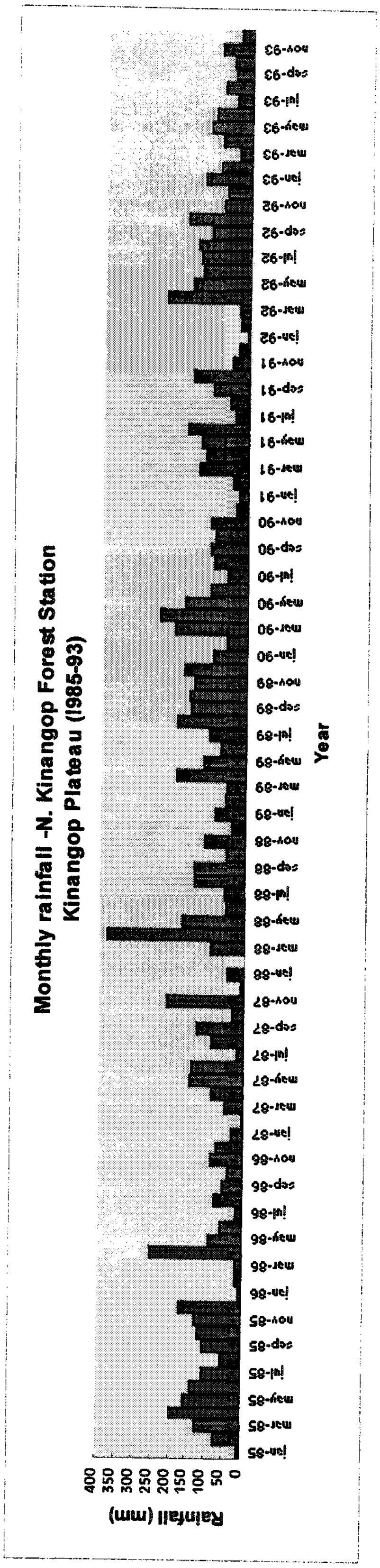
2.3.1 Rainfall

Of the many factors influencing groundwater availability, the most crucial and basic is that of precipitation. High rainfall alone however is not enough to guarantee good groundwater recharge as significant precipitation inevitably is lost to run-off, evapotranspiration, and in satisfying a soil moisture deficit. Nonetheless, recognizing the rainfall pattern is essential.

According to Vincent et al. (1979), 'Much of East Africa experiences two major rainfall seasons (March-May and October-December) with a third season during July-September which brings rain to western Kenya and Uganda'. The rainfall pattern within the study area is however bimodal. The rainy seasons are typically from April to May (sometimes June), and November to January. The April-May rainy season is the main rainy period, known as the 'long rains', and the November-January period, the 'short rains'.

The rainfall pattern displays a high temporal and spatial variability related to altitude. For example, North Kinangop Forest Station on the Kinangop Plateau to the east (elevation 2630 masl) has a 9-year average rainfall of 1146 mm (1985-93), while for the same period, the station of the Naivasha Water Development Department (WDD - on the rift floor - 1935 masl) had a significantly lower annual average value of 637 mm. This low rainfall is typical of the Naivasha area as much of the rainfall directed toward the central rift valley from the east is intercepted by the surrounding highlands (Verschuren, 1996).

From the rainfall data recorded at the Naivasha WDD station, 1989 can be considered a wet year, 1993 a dry year, and the years 1985-88 and 1991, average years (see fig. 2.7). Based on this data, a wet year is considered to have a total rainfall of 875 mm, an average year a rainfall total of 596 mm, and a dry year a rainfall total of 490 mm. (These figures are later used in the water balance in section 3.3). In a wet year, potential evapotranspiration is estimated to be 1350 mm. In an average year, it is estimated to be 1472 mm, and in a dry year, 1600 mm (see section 3.2). Figure shows comparisons between monthly rainfall and potential evapotranspiration for wet, dry and average year conditions. It is only in April of a wet year that rainfall exceeds potential evapotranspiration. In all other months of the year, under the varying climatic conditions described, potential evapotranspiration around Lake Naivasha exceeds rainfall.(Using monthly rainfall averages from 1985-93, the percentage of rainfall occurring in each month was calculated. These percentages were applied to the total rainfall values for the wet, dry and average years to get the monthly rainfall amounts under these conditions.)



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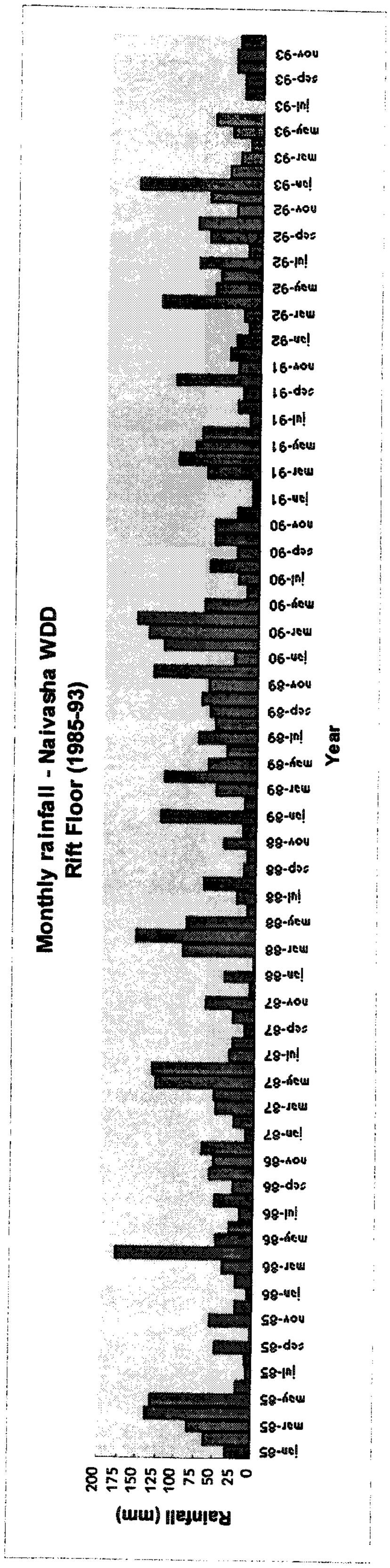


b)

Figure 2.6 - Comparison between rainfall at different altitudes

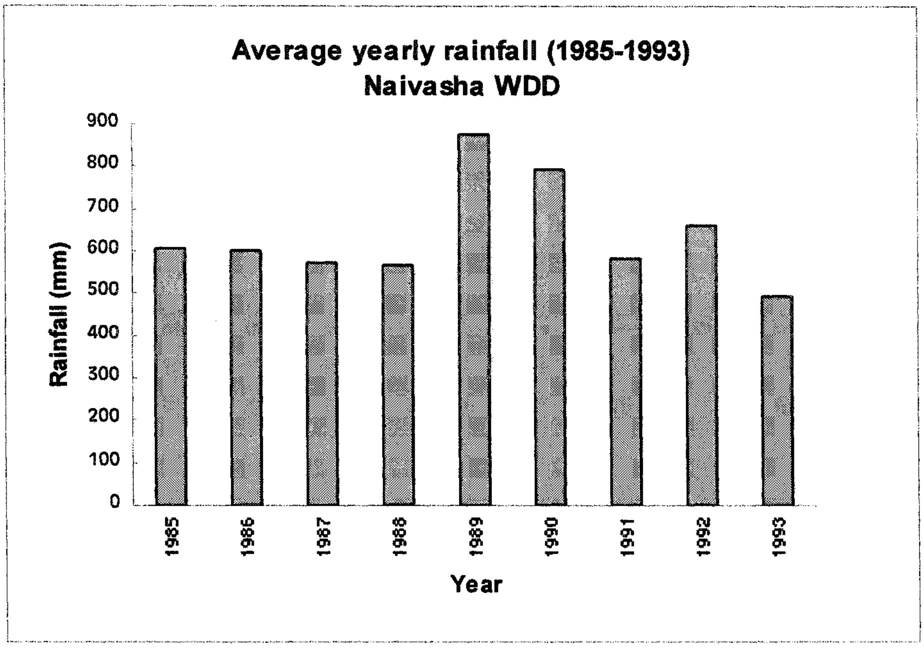
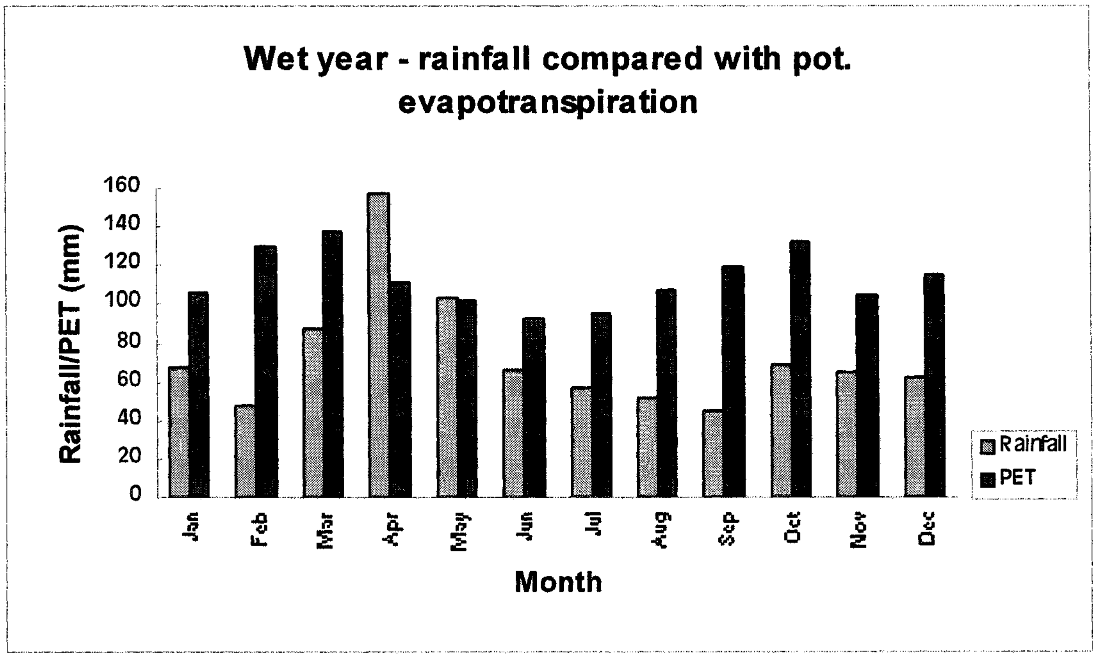
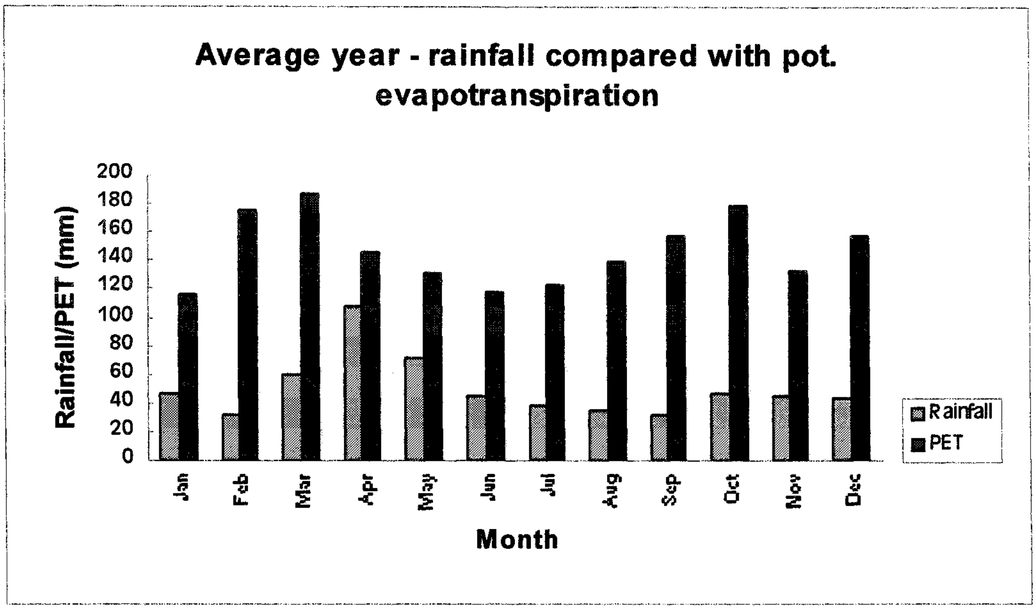


Figure 2.7 - Average yearly rainfall (1985-1993)



a)



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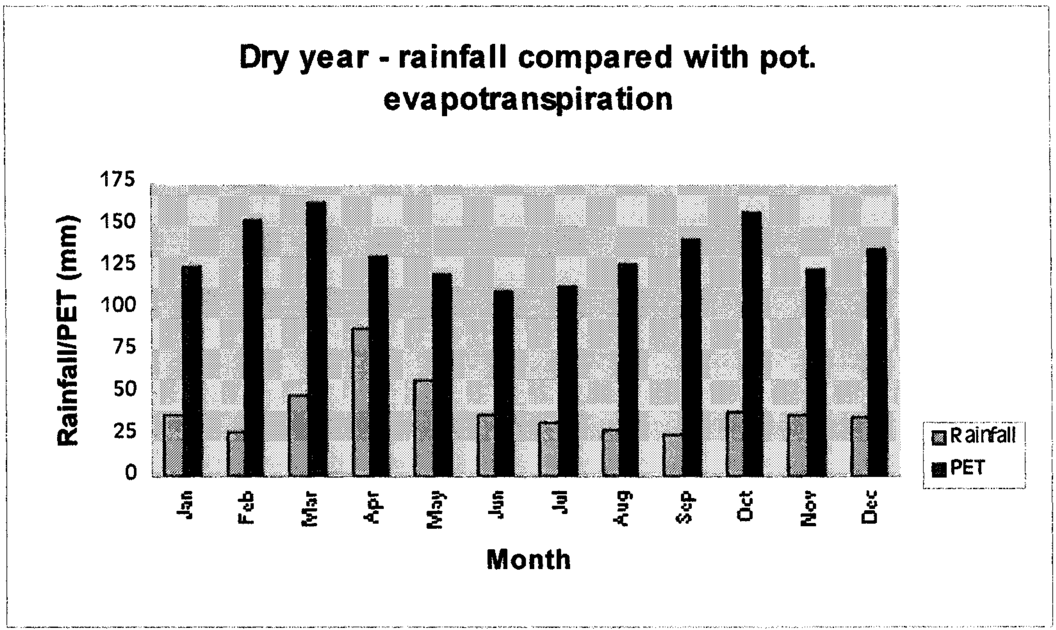


c)

Figure 2.8 - Rainfall compared with potential evapotranspiration: wet, dry, average years

2.3.2 Temperature

Temperatures in the Naivasha area are highest in January and February and lowest in July and August. Mean monthly maximum temperatures range between 24.6°C to 28.3°C, and mean monthly minimum temperatures between 6.8°C and 8.0°C in. The average monthly temperature ranges between 15.9°C and 17.8°C.

2.3.3 Evapotranspiration

In Naivasha, low relative humidity and an average daily temperature of 24 °C combine to cause an annual potential evapotranspiration of 1500-1900 mm/yr (Ase et al., 1986). This is far in excess of rainfall resulting in a strongly negative hydrological balance near the lake (see fig.2.8). Potential evapotranspiration is higher during dry seasons and lower during rainy periods.

'It is acknowledged that pan evaporation rates may contain errors and that there is disagreement about pan coefficients which should be applied to records to obtain comparative 'open water' rates...' (McCann, 1974; Sikes, 1936). A 'class A' evaporation pan has however been set up alongside the lake shore at the Yacht Club to monitor lake evaporation. This is a recent development and so data exists only for the period Feb-Apr and Aug-Oct, 1997 (see.2.9 - pan factor not applied).

Monthly 'class A' pan evaporation measurements from 1966-80 were available (table 2.2). Average yearly evaporation based on these measurements is 1787 mm. The pan was located along the eastern shore of Lake Naivasha. This data is later used in calculating a monthly water budget for the lake (section 3.2).

Month	J	F	M	A	M	J	J	A	S	O	N	D
Evaporation (mm)	118	178	190	149	132	120	125	142	158	183	134	158

Table 2.2 - Average monthly evaporation (1966-80)

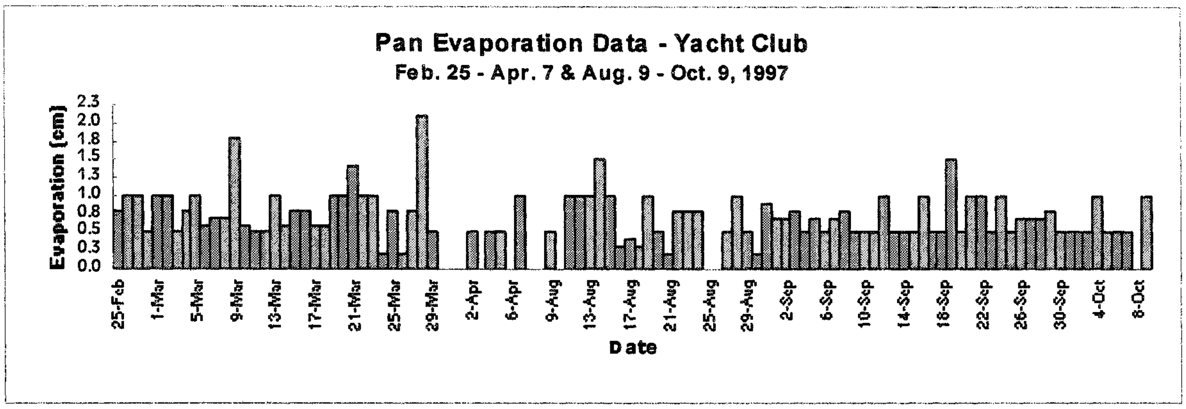
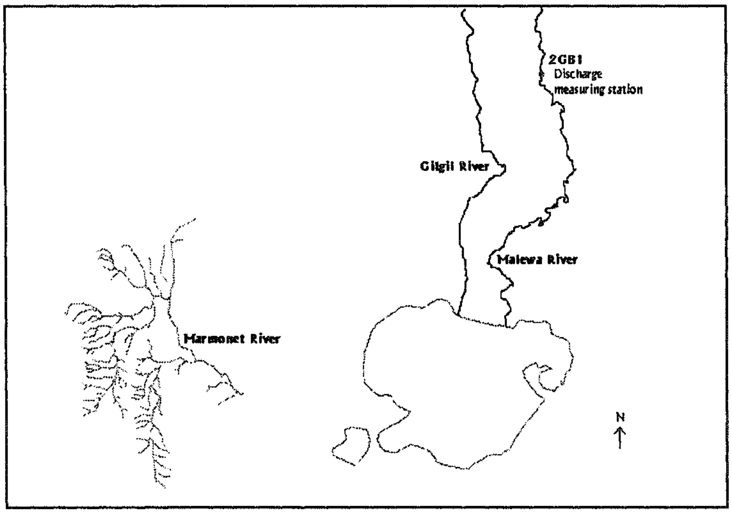


Figure 2.9 - Daily pan evaporation data

2.4 Hydrology

The main rivers in the study area are the Marmonet river on the west side of the lake draining the Mau Escarpment, and to the N-NE of the lake, the Gilgil, Malewa and Karati rivers (see fig. 2.10). The Karati river is perennial, and contributes very little inflow to the lake. The Gilgil and Malewa rivers collect runoff from the Aberdare mountains and their foothills to the NE of the lake, and discharge into the papyrus swamp forming part of the northern lake shore (Darling et al., 1990). The Marmonet river, although it flows towards the lake, fails to reach it.

The Malewa river is the most important contributor to Lake Naivasha's river inflow, providing an estimated 80% of the total river inflow (Vincent et. al., 1979). The average daily discharge of the Malewa is 6.3 m^3/sec (1960-85)(SEE Appendix 2).



Gauged rivers	Catchment area (sq. km)
Malewa	1588
Gilgil	534
Karati	151
Ungauged	841

Figure 2.10 - Sketch showing main rivers in the study area

Table 2.3 - Size of catchment areas

2.5 Hydrogeology

'In general, rock permeability in the Rift Valley is low, although local variations are common. Aquifers are normally found in fractured volcanic rocks or along weathered contacts between different lithological units. These aquifers are often confined or semi-confined and storage coefficients are likely to be low. However, aquifers with relatively high permeability are found in sediments covering parts of the rift floor (particularly around Lake Naivasha). These aquifers are often unconfined and will have relatively high specific yields' (Stuttard, 1995).

Borehole data exists for 69 boreholes in the study area (see Appendix 3). These boreholes were drilled between 1939 and 1997. Although water level measurements were taken in 1997, the boreholes are not leveled. This presents a problem when trying to construct a piezometric map, as most of the boreholes are located around the lake where the water table is relatively flat. Given these constraints, it was possible to construct only a very general piezometric map (heavily based upon topography) which shows that groundwater flows laterally toward the lake from the Mau escarpment (west) and Kinangop plateau (east). There were very few groundwater level measurements available south of the lake, however all reports consulted make reference to outflow along the south shore (Ase et al, 1986; Gaudet and Melack, 1981).

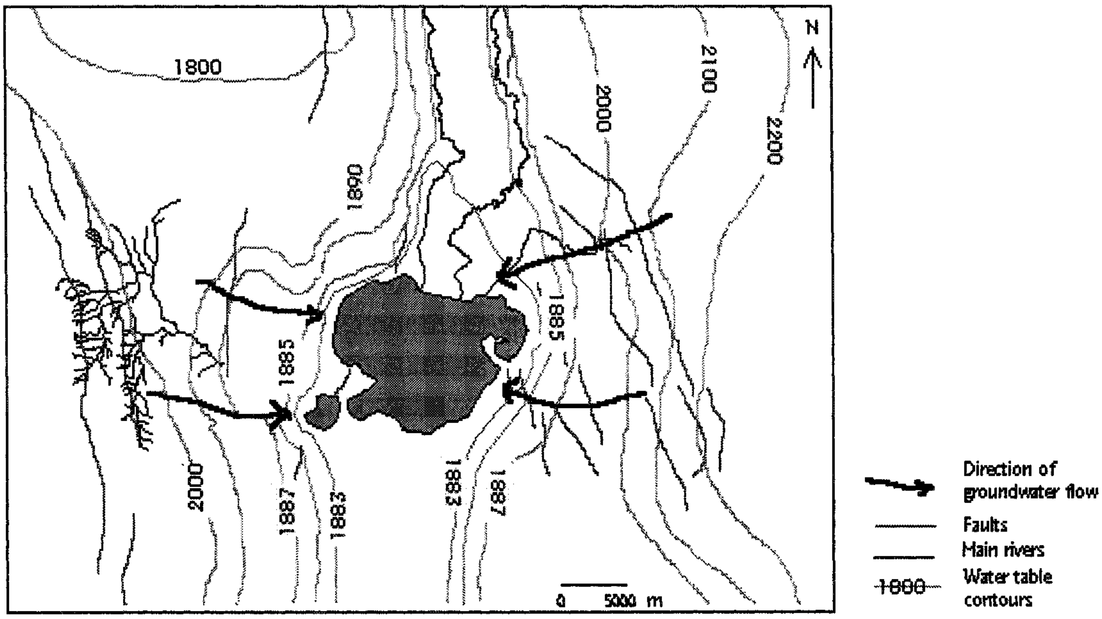


Figure 2.11 - Piezometric map showing main groundwater flow directions

2.6 Groundwater Level Fluctuation

Groundwater level variation has been automatically recorded using absolute loggers since July, 1997 at select locales. Absolute loggers measure the absolute pressure above the top of the logger. When the logger is placed below the water table in a borehole, it records the pressure of both the column of water above it, and the pressure of the atmosphere.

absolute pressure = atmospheric pressure + pressure head

As the groundwater table rises and falls, the absolute pressure will increase and decrease. Changes in atmospheric pressure will also cause the absolute pressure to change, so a reference logger has to be placed nearby (in a house for example) to record atmospheric pressure only. When the logger in the borehole is removed and the data downloaded, it has to be corrected for atmospheric pressure. This is done by subtracting the atmospheric pressure recorded by the reference logger from the absolute pressure recorded by the logger in the borehole (over the corresponding time period).

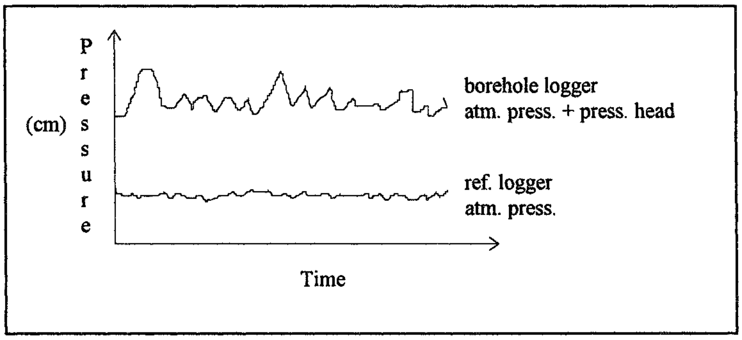


Figure 2.12 - Sketch showing graphs of logger data from borehole and reference loggers

To link the resulting pressure measurements to groundwater levels, when the logger is removed, the depth to groundwater must be measured, and the time noted. By knowing that a certain pressure reading at a particular time corresponds to a particular groundwater level, groundwater levels at other times can be extrapolated.

Processing of the logger data was carried out using the TDVX software. The resulting groundwater level fluctuations are shown in fig. 2.13 below.

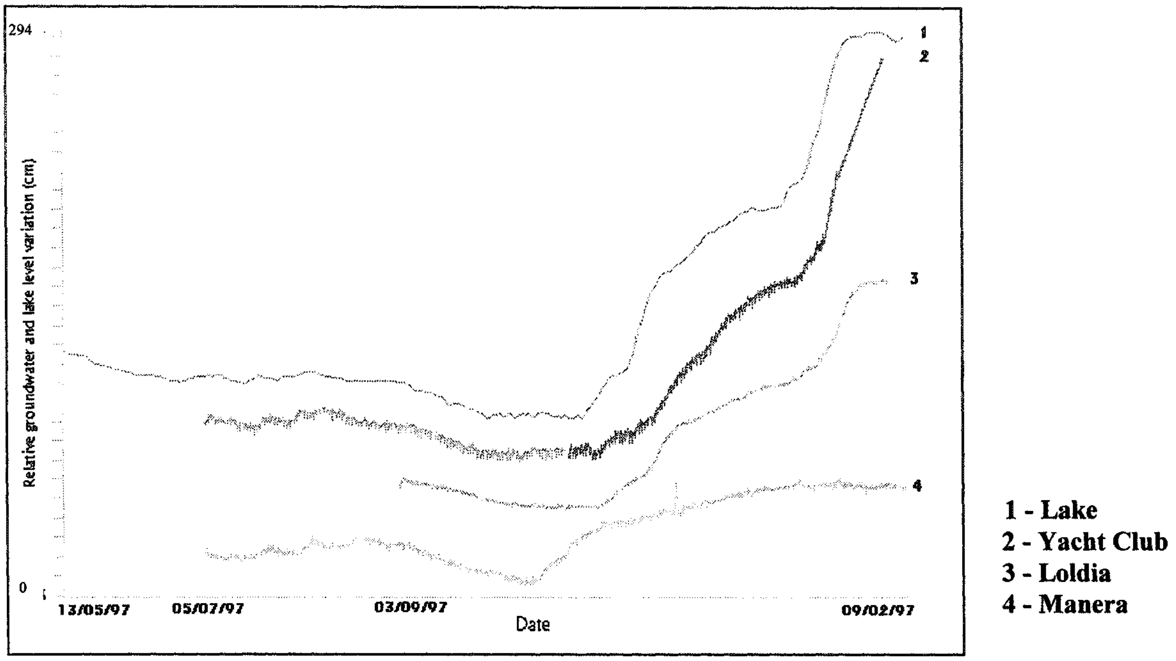


Figure 2.13 - Relative lake level and groundwater table fluctuations

The graphs are arranged in order of increasing distance from the lake. Graph 1 shows the lake level fluctuation between the period May 13, '97 - Feb. 9, '98. The lake experiences an almost consistent decline in level between May to October. However, from November to the end of the observation period, a time length of 101 days, the lake level rises rapidly by 2.07 m due to heavy rainfall.

As one would expect, Yacht Club, being closest to the lake (135 m), and connected to it by a highly transmissive zone, mimics the lake fluctuation pattern very closely without any significant time lag. The groundwater level variation graph for the Yacht Club logger is compiled using data from two boreholes. These boreholes are only roughly 100 m apart in what appears to be a homogenous area, and so it is safely assumed that they would undergo similar groundwater fluctuations. It was necessary to do this as when the lake level rose, the borehole closer to the lake was submerged. The logger in the other borehole was also affected by the rise in lake level. The groundwater level rise was so much that it exceeded the active range of the logger. The end section of graph 2 thus was linearly interpolated using the observer measured depth to groundwater at the end of the period when the logger was removed.

At 213 m from the lake, Loldia exhibits a delayed response to the lake level rise. For a 1.92 m rise in lake level over 92 days (Nov.1 - Feb. 1), the groundwater rises by only 1.16 m for the corresponding period.

The Manera logger (graph 4) is located approximately 3800 m from the lake. From inspection of the graph, it seems as if the groundwater level at this distance is affected by some other influence in addition to the lake fluctuations. The first half of the graph exhibits the same downward trend as the lake, however, while the lake level is still falling, the groundwater table at Manera begins to rise. There must therefore be some other driving force causing this rise in groundwater level. The rise could be due to the fact that there are a number of irrigation wells nearby. Discontinued pumping of these due to the heavy rains falling at this time could cause the groundwater table in the vicinity of the Manera borehole to rise.

2.7 The Lake

Lake Naivasha can be divided into 4 sections which are all hydrologically connected either at the surface or by groundwater flow.

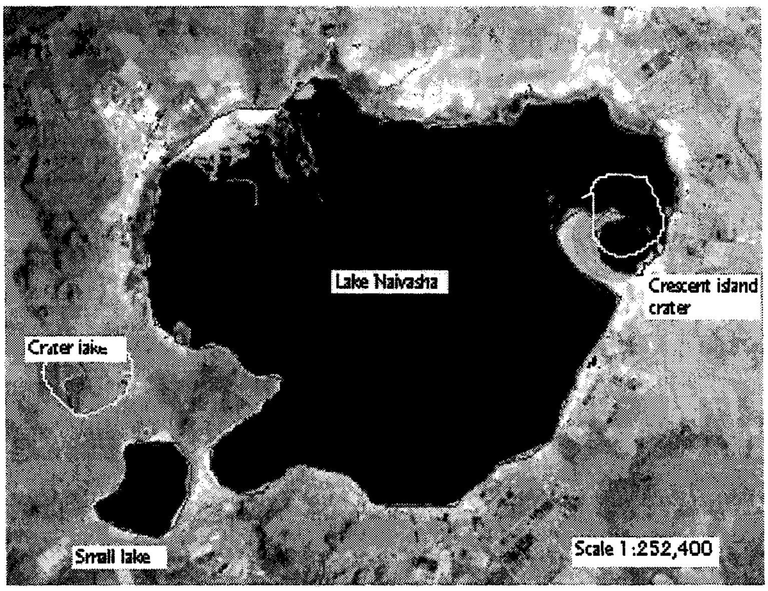


Figure 2.14 - Sections of Lake Naivasha

2.7.1 The Main Lake (Lake Naivasha)

The main lake is fairly large, but shallow. The surface area (derived from a 1995 Landsat TM image) is approximately 127 km^2. The depth averaged 7 m in the period 1963-1993 (Verschuren, 1996), however depth measurements made during the field visit result in an average depth of 4 m. The drainage basin is topographically closed, but the lake itself is hydrologically open, with groundwater flowing into the lake from the north, east and west, and outflowing mainly in the south-southeast (Gaudet and Melack, 1981; Allen et al.,1989). This inflow and outflow of water keeps the lake waters fresh. Conductivity measurements taken during the site visit in October 1997 averaged 500 µS/cm.

2.7.2 Crescent Island Crater

This is a small (2.1 km^2), but deep depression along the eastern shore of Lake Naivasha (Verschuren, 1996). It is formed from a partly submerged volcano, the rim of which lies above water forming the island. Based on 1983 lake levels, the depth of this section of the lake averaged 18 m (Stuttard,1995). Like the main basin, the waters in this section of the lake are fresh (Verschuren, 1996).

2.7.3 Small Lake (Lake Oloidien)

The 'small lake' (5.5 km^2), located at the western end of Lake Naivasha, is hydrologically closed. According to Gaudet and Melack (1981), groundwater output is negligible and water losses are due to evaporation. Electrical conductivity values of this lake section are very high (3423 µS/cm). Although 'small lake' is hydrologically closed, it is fed by the main lake as when water levels in the main lake reach a certain level, water seeps through the swampy separation between the two (Verschuren, 1996). When the lake level rises above 1885.5 masl, Lake Oloidien becomes joined with the main lake (Verschuren, 1996).

2.7.4 Crater Lake (Lake Sonachi)

Although physically separated from the main lake by a distance of approximately 2700 m, Crater Lake (0.14 km^2)may be considered a part of the Lake Naivasha system. Crater lake, as the name implies, is a lake which formed in an old volcanic crater. Its waters are very saline (conductivity measurements range between 3000-11,550 µS/cm, Verschuren, 1996). Given the high rate of evaporation, and low rainfall in the area, the lake's existence must be maintained by groundwater inflow. It has been suggested that the lake is in hydraulic connection with Lake Naivasha. This connection is supported by direct observation of synchronous lake level changes in Crater Lake and Lake Naivasha (MacIntyre and Melack, 1982) (Verschuren, 1996). The height of Crater Lake relative to Lake Naivasha was not determined in the field.

2.8 Lake Levels

The level of Lake Naivasha has fluctuated widely over the past 113 years, from a minimum of 1881.5 masl in 1975 to a maximum of 1896.5 masl in 1895.

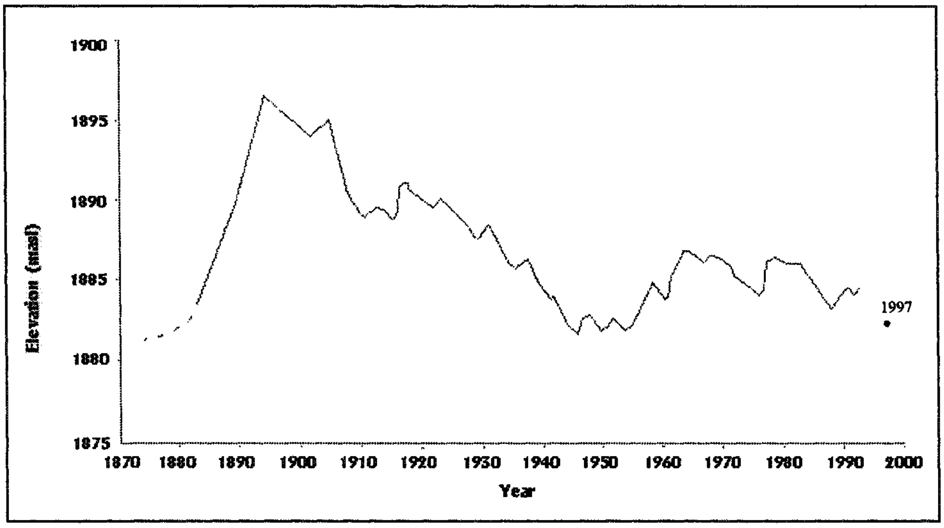
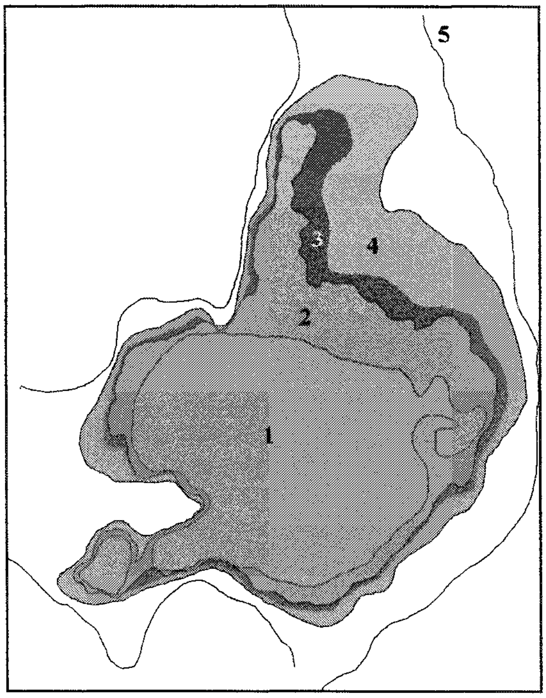


Figure 2.15 - Historic lake level fluctuations

This fluctuating character of the lake is likely to have given rise to its name. The word Naivasha is a corruption of the Maasai word *E-na-iposha* which means 'heaving waters' (Internet source), an apt name for the lake as its levels have constantly risen and fallen over the last century. Lake Naivasha is actually the remnant of a former lake which in Late Pleistocene to Early Holocene times (Gamblian period) extended northwards to include lakes Elmenteita and Nakuru, approximately 30 and 45 km away respectively.

Because the Naivasha lake basin is shallow (see section 2.7.1), relatively small changes in lake level result in large changes in open-water surface area. Lake surface area and storage changes were calculated in ILWIS using a digital elevation model (DEM) of the area, which included the lake bathymetry. The DEM was created by interpolation of digitized contour lines (40 m contour interval). The lake bathymetry was derived based on depth transects made during the field visit, and adjustment of a 1955 bathymetric map. At the time of visit (October, 1997), based on the DEM, the lake surface area was approximately 125 km^2, and the storage 403 million m^3.

Figure 2.16 shows how the lake surface area has varied in the past.



AREA	YEAR	LAKE LEVEL masl	SURFACE AREA sq. km
1	1956	1882.5	128
2	1973	1884.0	189
3	1930	1887.5	217
4	1895	1896.5	282
5	10-12,000 BP		

Figure 2.16 - Former Lake Naivasha shorelines

(adapted from Vincent et. al, 1979)

At the time of visit, the lake level stood on average at 1882.4 masl. In the following three months, heavy rains resulted in a rise of 1.7 m in the lake level. Based on the DEM, the lake surface area increased by approximately 60 km^2 in the three month period. This difference in surface area is similar to the difference between the 1956 and 1973 shorelines in fig. 2.16. A better appreciation of the tremendous change in surface area of the lake can be gotten by observing plate 3. In the photo, all the grassland seen was submerged as the lake rose. This has important implications for groundwater storage as is discussed in chapter 4.

3.0 THE LAKE WATER BALANCE

The lake water balance/budget refers to the balance between the input of water to, and output of water from the lake. This balance is of importance as firstly, it determines how much lake water is available for development, and secondly, it influences the quantity of groundwater stored in the aquifer surrounding the lake. In this study, the lake water balance is considered for three climatic conditions - wet conditions, dry conditions and average conditions. How the lake inputs and outputs are determined under these conditions is discussed in the following sections, and the resulting definition of wet, dry and average conditions summarized in tables 3.6 and 3.7.

The inflows into the lake are derived from precipitation, river discharge and groundwater inflow, whereas the outflows occur via evaporation, evapotranspiration, groundwater outflow and pumped abstractions. Differences between inflow and outflow volumes result in a change in storage/volume of the lake.

3.1 Lake Inflows

Precipitation

The rainfall data used was data recorded at the Naivasha Water Development Department (WDD) from 1985-93 (see Appendix 5). When this data is averaged on a yearly basis, it can be seen that 1989 has the most rainfall, 1993 the least rainfall, and the remaining years (except 1990), an average amount.

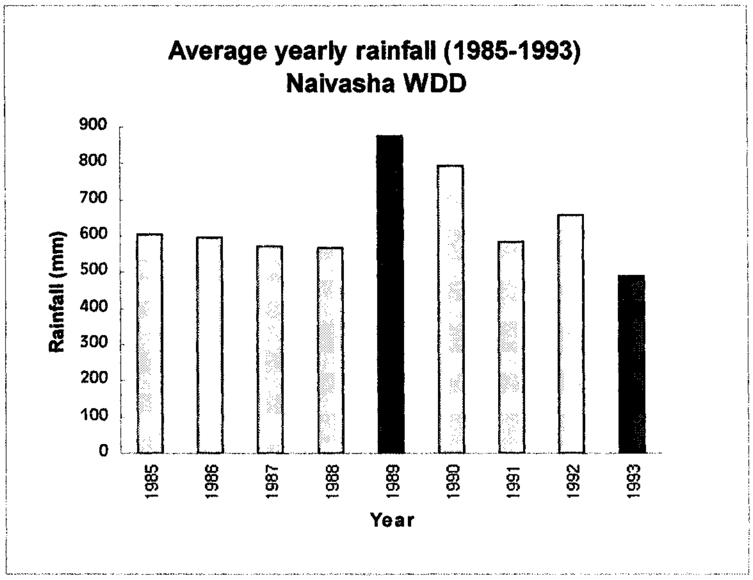


Figure 3.1- Average yearly rainfall (1985-93)

Based on this rainfall pattern, 1989 was taken to be representative of a wet year, 1993 was taken to be representative of a dry year, and the mean rainfall of the remaining years (except 1990) was taken to be representative of an average year. The amount of precipitation over the lake is therefore 875 mm in a wet year, 490 mm in a dry year, and 596 mm under average conditions.

Discharge

Three rivers flow into Lake Naivasha, the Malewa, Gilgil and Karati rivers (see section 2.4). The Karati river is perennial. England and Robertson (1969) calculated that when it flowed for at least half the year, its discharge was less than 2.5×10^6 m^3 (0.02 m). Due to this low discharge, in this study, the contribution of the Karati river discharge to the lake water inflows is considered negligible.

The flow in a stream or river is called runoff (Fetter, 1994). There is a relationship between the amount of rainfall in a catchment and the amount of runoff (flow) out of the catchment. This relationship is described by the runoff coefficient. The runoff coefficient depends on characteristics of the catchment such as slope, soils, vegetation, evaporation, catchment morphology. Ase et al. (1986) calculated the runoff coefficient for the Malewa catchment is 0.1, and for the Gilgil catchment, 0.18. Based on aerial photo interpretation, the terrain of the ungauged areas is similar to parts of both the Malewa and Gilgil catchments, so the runoff coefficients of the two were averaged and used for the ungauged areas. By knowing these runoff coefficients, and by calculating the mean rainfall in the catchments under wet, dry and average year rainfall conditions, it was possible to determine river discharge for wet, dry and average years.

$$\mathbf{Q = P * c}$$

discharge = catchment precipitation x runoff coefficient

In order to calculate the mean rainfall in the catchments, a Theissen polygon map was created (see fig. 3.2). The Theissen method gives an <u>estimate</u> of the rainfall as the Theissen method of calculating rainfall assumes that rainfall is equally distributed within each polygon. Furthermore, in mountainous areas, orographic effects can create microclimates over small distances. Significant precipitation can therefore fall on one side of a ridge but little on the other (Fetter, 1994).

The calculated average catchment rainfall and discharge is shown in the table below. For each Theissen polygon, the polygon area was multiplied by the rainfall recorded at the station representing that polygon to get the total rainfall in that polygon area. For each catchment, the rainfall values were added together to get the total catchment rainfall for the year (see Appendix 5). The total catchment rainfall was then multiplied by the catchment runoff coefficient and catchment area to get the yearly discharge from that particular catchment. The Malewa catchment has an area of 1588 km^2, Gilgil 0.18 km^2 and the ungauged area 841 km^2.

Catchment	Runoff coeff.	Climatic condition	Year	Mean yearly catchment rainfall (mm)	Total yearly discharge (mill. cu. m)
Malewa	0.10	Wet	1989	1078	171
		Avge	avged yrs.	896	142
		Dry	1993	799	127
Gilgil	0.18	Wet	1989	627	60
		Avge	avged yrs.	524	50
		Dry	1993	479	46
Ungauged areas	0.14	Wet	1989	216	69
		Avge	avged yrs.	171	55
		Dry	1993	162	52

Table 3.1 - Average catchment rainfall (Theissen method)

Previous researchers in the study area calculated different river discharge volumes from the ones calculated above. Their results were averaged with the results above to arrive at a final value to use in the water balance.

Researcher	Catchment	Condition	Method	Yearly disch. (Q) (mill cu. m)	Yearly disch calc. above (mill. cu.m)	FINAL yearly Q mill.cu.m)
Ase et.al (1986)	Malewa	Wet	rating curve	367	171	269
		Avge	1960-85 av.	198	142	170
Ase et.al (1986)		Dry	rating curve	51	127	89
Ase et.al (1986)	Gilgil	Wet	rating curve	71	60	66
Tetley (1948)		Avge		15	50	33
Ase et.al (1986)		Dry	rating curve	3	46	25

Table 3.2 - Yearly river discharge under varying climatic conditions

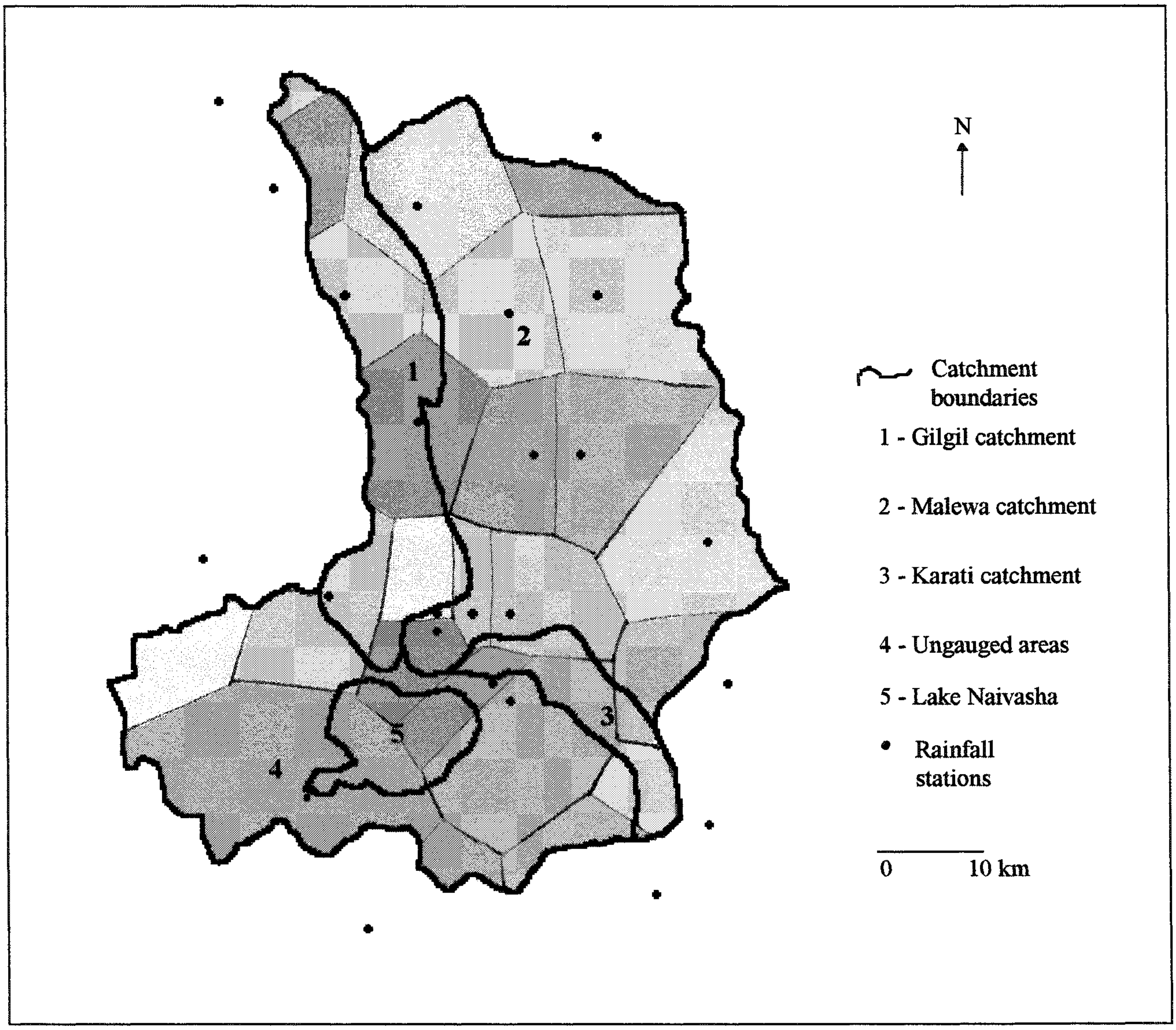


Figure 3.2 - Theissen polygon map

Groundwater Inflow

Groundwater inflow rates used were based on those estimated by Gaudet and Melack (1981). They estimated groundwater inflow into the lake ranged between 40-60 x 10^6 m^3/yr.

3.2 Lake Outflows

Evaporation & Evapotranspiration

Evaporation and evapotranspiration are very important in the lake water balance, accounting for more than 50% of the annual lake water losses. A number of methods can be used to calculate lake evaporation. These include the water balance method the and evaporation pan method.

Evaporation pan measurement is the only direct measurement of evaporation available (Ven te Chow, 1964). Monthly average pan measurements (1966-80) were obtained from a type 'class A'. The pan is located along the east shore of the lake. Depending on the pan environment (relative humidity, openness of the pan site), a correction factor has to be applied to the pan measurements to get the actual evaporation. Knowing what pan factor to apply is difficult. Brind and Robertson (1959) in their evaporation study of the lake suggested a pan factor of between 0.84-1.04 be used. The average of this was taken (0.94) and applied to the measured pan data to give the lake evaporation (Evap.). Lake evaporation (Evap) is derived from pan evaporation measurements using the formula:

$$\text{Evap} = \text{pan factor} \times \text{pan evaporation}$$

	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec	Total
Epan (mm)	118	178	190	149	132	120	125	142	158	183	134	158	1787
Evap. (mm) factor .94	111	167	179	140	124	113	118	133	149	172	126	149	1680

Table 3.3 - Average monthly lake evaporation (1966-80)

Reference crop evapotranspiration (Eto) is defined as 'the rate of evapotranspiration from an extensive surface of 8-15 cm tall green grass cover of uniform height, actively growing, completely shading the ground and not short of water' (Doorenbos and Pruitt, 1984). The reference crop evapotranspiration (Eto) can be obtained from:

$$\text{Eto} = \text{pan factor} \times \text{pan evaporation}$$

This is the same formula used to calculate lake evaporation. It can therefore be assumed that Eto and lake evaporation are numerically equivalent.

Monthly Eto measurements calculated using the Penman method were available (FAO's Cropwat) (see Appendix 4). These measurements were averaged with the pan measurements to obtain a final monthly lake evaporation value for average year conditions. Evaporation under dry and wet conditions were estimated based on the average year conditions (see table)

	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec	Total
Eto (mm) (Penman)	118	115	121	102	96	90	90	99	108	118	99	102	1258
Evap (mm) (Pan)	111	167	179	140	124	113	117	133	149	172	126	149	1680
FINAL evap. (mm)	115	141	150	121	110	102	104	116	129	145	113	126	1472

Table 3.4 - Monthly lake evaporation (average of Penman and pan)

Monthly Eto values were also calculated using the Thornthwaite method and the Langbein-Turc method. The results obtained by these methods were not in keeping with expected amounts. The Thornthwaite method gave a yearly evapotranspiration of 768 mm under average year conditions, and the Langbein-Turc method a yearly evapotranspiration of 443 mm (see Appendix 4).

Using the reference crop evapotranspiration (Eto), the actual evapotranspiration of a particular crop or vegetation type (Eta) can be derived by applying a crop factor (Kc). The crop Kc for the swamp surrounding the lake was estimated to be 1.15. The resulting values for the swamp evapotranspiration under average year conditions is shown in table 3.5 below. The Kc factor was applied to the lake evaporaton under wet and dry conditions to get the swamp evapotranspiration under these conditions (see table).

	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec	Total
Eto (mm)	115	141	150	121	110	102	104	116	129	145	113	126	1472
Eta (mm) Kc = 1.15	132	162	172	140	127	117	120	134	148	167	130	145	1693

Table 3.5 - Monthly swamp evapotranspiration

Groundwater Outflow

Groundwater outflow was estimated by Ase et al. (1996) to vary between 46 -56 million m^3 per year. The upper limit was used in the lake balance (table 3.7) as the groundwater outflow under wet conditions, and the lower limit, the outflow under dry conditions. Outflow under average conditions was estimated to be 50 million m^3 per year.

Abstraction

Abstraction data was taken from a report of the LNROA (1993). According to this report, "where information on water abstraction was unreliable, estimates based on electric power consumption used in pumping water were made based on the Kenya Power and Lighting and Ministry of Water office records in Naivasha".

Lake Area and Storage

The area of the lake and the lake storage volumes corresponding to the lake levels in table 3.6, were calculated from a digital elevation model (DEM). The DEM was made by digitizing then interpolating contour lines from 1:50000 scale topographic maps of the study area. A 40 m contour interval was used, except in key areas such as near rivers and on the tops of hills. In such places, a smaller interval was used. Depth transects were made in the lake at the time of field visit, and these were used to update an existing lake bathymetry map made in the 1950's.

3.3 Lake Water Balance

The final yearly lake water balance arrived at is shown in table 3.7. The balance was created for wet, dry and average climatic conditions. Variables used in the budget are shown in the table below. The yearly evaporation and evapotranspiration depths are multiplied by the lake and swamp areas respectively to obtain the volumetric amounts of evaporation and evapotranspiration. Similarly, the depth of rainfall over the lake is multiplied by the lake area to obtain the total volume of rainfall added yearly to the lake. The budget is separated into lake inputs and outputs.

VARIABLES	WET CONDITIONS	MEAN CONDITIONS	DRY CONDITIONS
Direct rainfall on the lake (mm)	875	596	490
Open water evaporation (mm)	1350	1472	1600
Evapotranspiration from swamps (mm)	1553	1693	1840
Area of swamps (sq. km)	18	12	9
Area of lake (sq. km)	190	164	125
Lake storage (millions of cu. m)	1000	670	380
Average lake level (masl)	1886	1884	1882

Table 3.6 - Variables used in the water budget

Table 3.7 - Estimated annual water budget for Lake Naivasaha (million cu. m)

INPUTS	WET CONDITIONS	MEAN CONDITIONS	DRY CONDITIONS
Recharge by direct rainfall	166	98	61
Malewa river	269	170	89
Gilgil river	65	33	25
Unguaged areas	69	55	52
Groundwater inflow	50	45	40
TOTAL INPUT	619	401	267

a) Lake inputs

OUTPUTS	WET CONDITIONS	MEAN CONDITIONS	DRY CONDITIONS
Evapotranspiration	28	20	17
Open water evaporation	257	241	200
Groundwater outflow	56	50	46
Abstraction	34	45	53
TOTAL OUTPUT	374	356	316
YEARLY BALANCE	**+ 245**	**+ 44**	**- 49**

b) Lake outputs

4.0 THE MODEL

4.1 Groundwater Models

A model is a simplified representation of reality. It can be physical, electrical analogue or mathematical (Abbott and Refsgaard, 1996).

Groundwater models are mathematical models. In groundwater modelling, the modellers understanding of the physical processes involved are expressed by mathematical equations. A governing equation is used to represent the physical processes that occur in the system while other equations describe heads or flows along the boundaries of the model (boundary conditions) (Anderson, 1992). The governing equation for groundwater flow is derived by combining Darcy's law with the continuity equation (see fig. 3.1). The resulting differential equation (for transient flow in a confined aquifer)

$$\frac{d^2h}{dx^2} + \frac{d^2h}{dy^2} = \frac{S}{T}\frac{dh}{dt}$$

h(x,y,t) - hydraulic head at a point in time (L)
S - storage coefficient
T - transmissivity (L^2/T)

can be solved either analytically or numerically. The analytical solution gives an exact result as the equation is solved with specified limits (boundary conditions). The numerical solution gives an approximation as the differential equation is truncated after the first two expansions of the series. Numerical solutions can be solved using the finite difference or finite element method. The finite difference method solves the equation through differentiation, while the finite element method approaches the solution through integration.

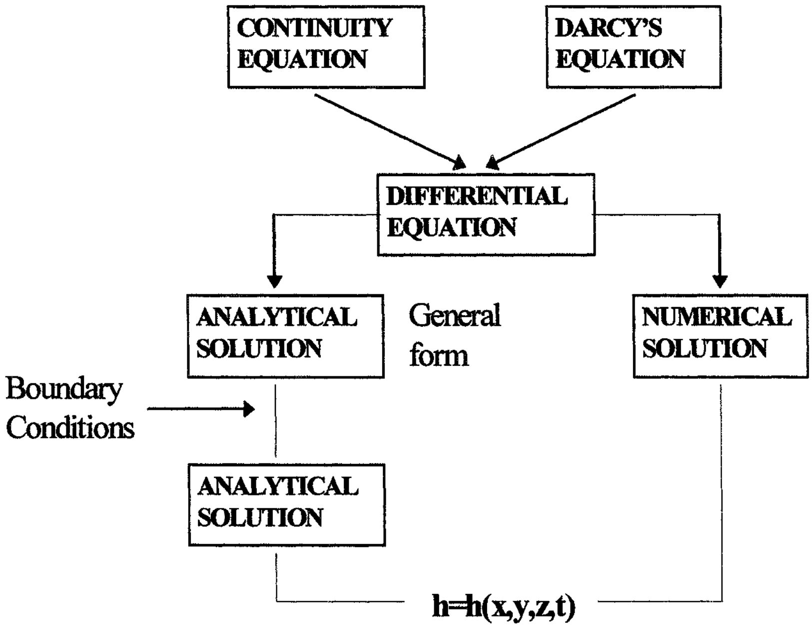


Figure 4.1- Schematic representation of mathematical solutions

4.2 Modelling Objective & General Approach

Modelling is commonly carried out for predictive purposes. However in this research, the modelling effort is primarily geared toward testing ideas about how the system behaves. The objective of the modelling is to study the effect of water level changes in Lake Naivasha (due to river inflow and evaporation) on the water levels of and groundwater storage in the aquifer surrounding the lake.

The objectives will be carried out by using PMWIN to model a set of scenarios. A basic model grid was designed (section 4.6 - Regional Model), and this is used for modelling all the scenarios. First, the actual field situation is modeled. This is followed by eight scenarios. The first three model the change in lake level and resulting groundwater level fluctuations and storage response during a wet year, a dry year, and an average year (see section 3.3). The next five scenarios model 10-year series of different hypothetical climatic conditions to see the effect on groundwater levels and storage. Whereas modelling of the lake fluctuations is done using PMWIN, calculation of the resulting storage changes is done using ILWIS. This is later explained (section 4.7). Observation boreholes are placed at varying distances from the lake, and groundwater table profiles drawn in order to clearly see the propagation of groundwater heads with time as the lake level fluctuates.

4.3 Programs Used

The modelling software used is PMWIN (Processing Modflow for Windows), a simulation system for modelling groundwater flow processes with the modular 3-D finite-difference groundwater model MODFLOW. The ILWIS (Integrated Land and Water Information Systems) GIS package is used to calculate storage changes.

In keeping with current technological trends, in some locations groundwater level fluctuations were monitored automatically using pressure loggers. The TDVX software was used for processing the logger data. This program proved to be invaluable as it allows for easy manipulation, filtering and representation of the logger derived data.

4.4 Methodology

A site visit was made to the study area for the twofold purpose of collecting data, and gaining a firsthand appreciation of and 'feel' for the area. Upon organizing the data, it was realized that much vital data is lacking. This includes storage coefficient/specific yield data, hydraulic conductivity data, proper lithological data, and groundwater level measurements taken from leveled boreholes. Leveling is extremely important as the groundwater table around the lake is flat and so flow direction is not very obvious. In the model, many assumed parameter values have been used, based on general data in existing reports, and personal judgement.

A somewhat standard approach to groundwater modelling has developed. This approach, with minor modifications, has been followed in this study and is summarized in the flow chart below. Each step in the process is of equal importance, as it impacts upon the steps to follow.

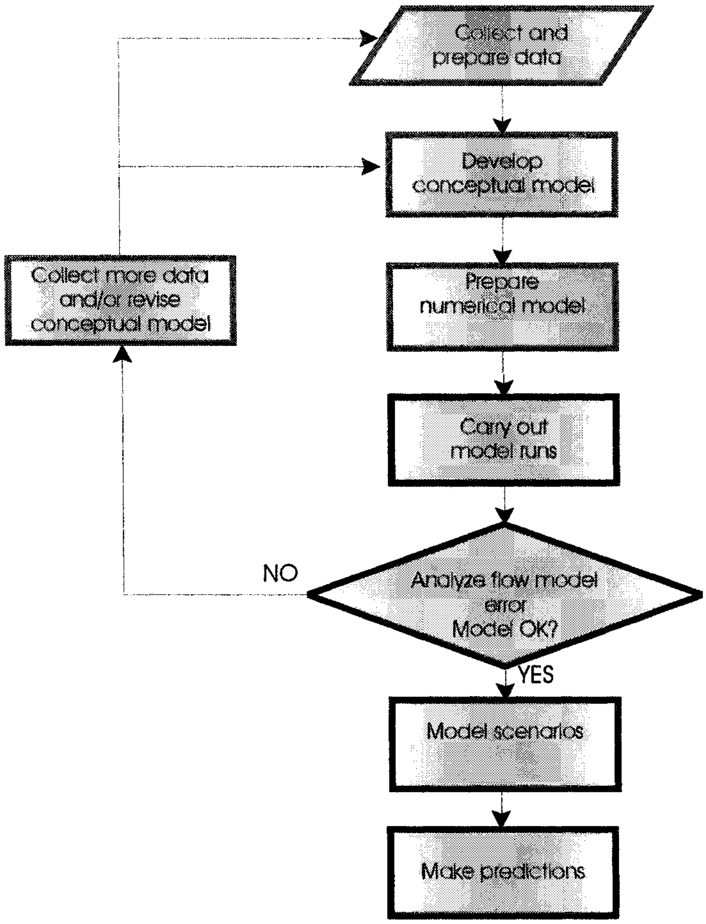


Figure 4.2 - Steps followed in carrying out the modelling exercise

Development of a conceptual model is one of the initial steps. A conceptual model is a graphical/pictorial representation of current understanding of site conditions and how the groundwater flow system operates. It includes the important features of the flow system, while incorporating simplifying assumptions. These simplifying assumptions are necessary as a complex system is difficult to model as it requires more input data and a higher level and range of skill on the part of the modeler. Additionally, increasing complexity may lead to greater output uncertainty if the input data is unavailable or inaccurate.

The purposes of developing a conceptual model include the following:
(Spitz and Morens, 1996)

- to develop a better understanding of site conditions, and be able to communicate this understanding

- to define the groundwater problem for development of a numerical model

- to aid in selecting a suitable numerical model

The conceptual model developed for the model area is shown in fig. 3.3. This conceptual model is consistent with the main features of the natural flow system. There is regional groundwater inflow to the lake from the highlands on either side as well as from the north, whereas in the south, there is groundwater outflow (see section 2.5). Depending on the lake level relative to the groundwater levels in the aquifer immediately surrounding it, there may be local flow either into or out of the lake. This is a constantly changing situation.

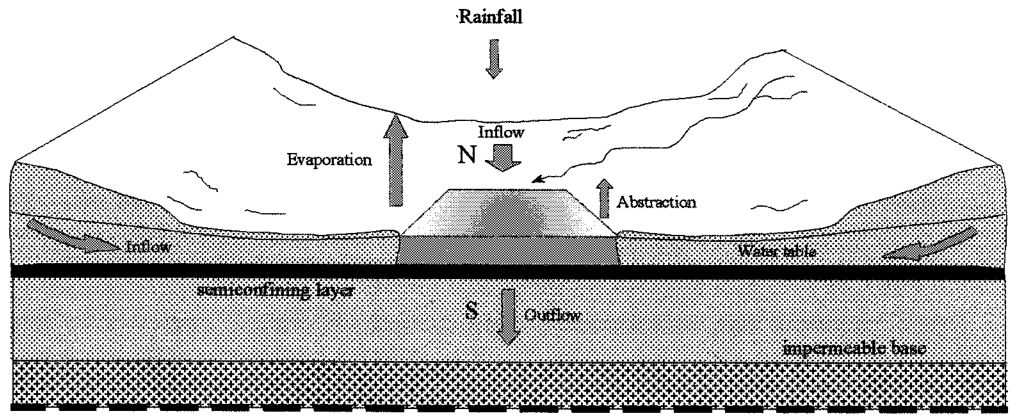


Figure 4.3 - Conceptual model

The aquifer is composed of lake sediments not exceeding a thickness of 30 m. These lake sediments are derived mainly from erosion of the surrounding volcanic rocks, and consist of volcanic sands and pebble beds, and gravels composed of pumice (see photo). Underlying the aquifer is a thick sequence of volcanic rocks, which in the model is considered an impermeable base. There is not much recharge by precipitation in the modelled area as the low rainfall rates coupled with high evaporation rates do not permit much infiltrating rain water to reach the water table (see fig. 2.8).

4.5 Submodels

When modelling an area, it is often useful to begin the exercise using submodels. Submodels are local models which are extremely simplified and unrefined. The advantages of using submodels are that they run quickly and are relatively easy to calibrate, and so can be used to test the validity of the conceptual model. Once the concept has been proven valid, the submodel can then be projected into a regional model.

In this study, initially submodels were created for three locations, the Yacht Club, Loldia Farm, and Manera Farm. At these locations there are boreholes containing pressure loggers, which continuously record groundwater level variations at six hour intervals for the Yacht Club and Loldia stations, and a half an hour interval for the Manera station. For each submodel, an attempt was made to determine aquifer transmissivity (T) and storage coefficient (S) values (or range of possible values) using the detailed logger data to calibrate the model. This exercise however proved inconclusive as when varying combinations of T and S were used, they gave the same result. In otherwords, the solution appeared to be non-unique. One parameter was kept fixed and the other varied, but even then the solution was non-unique. Additionally, the submodels could not be well calibrated, even when PEST (Parameter Estimation - an automatic calibration package) was used. The parameter values used in the model thus had to be estimated. This is discussed in the following section.

4.6 The Regional Model

Setting up a regional groundwater model is time-consuming, however once a basic design is in place, the task of improving and modifying is greatly simplified.

The basic regional model is composed of 64 rows, 56 columns, and 2 layers, both modelled as confined layers. In reality, the upper layer is unconfined but is modelled in PMWIN as a confined layer. Running the model in the confined mode influences the way PMWIN calculates for transient simulations. In the confined mode, the transmissivity of each cell is kept constant throughout the simulation and the confined storage coefficient is used to calculate the rate of change in storage. By assigning a typical unconfined aquifer storage coefficient value to the confined model, it reacts like an unconfined aquifer with constant transmissivity. Given the small ratio between variations in saturated thickness and aquifer thickness (1:15), this simplification is thought to be justifiable (Ven te Chow, 1964).

The model cells are all of the same dimension, 500 x 500 m. Because the lake and its immediate surroundings is the focus of the study, the model domain covers only this small section of the catchment (27% of the total catchment area), and the entire grid area is considered active. No flow boundary conditions exist at all the boundaries of the model. The Malewa river discharges into the lake, and is included in the model as a constant head boundary.

The lake boundary was screen digitized from a 1995 satellite image of the area. Since that time, the lake boundary has undoubtedly changed, but change has been ignored. The fluctuations in lake level can be modelled by defining the lake as a 'time variant head boundary' as the changes in level, and the time over which these change occurs are known, but this limits the flexibility of the model.

Lake fluctuations are instead modelled by defining the lake as being a large wellfield. The lake area encloses 508 cells, and these 508 cells are defined as wells. This approach offers the advantage of being able to model lake level change based on, for example, discharge into the lake, or other components of the lake water balance.

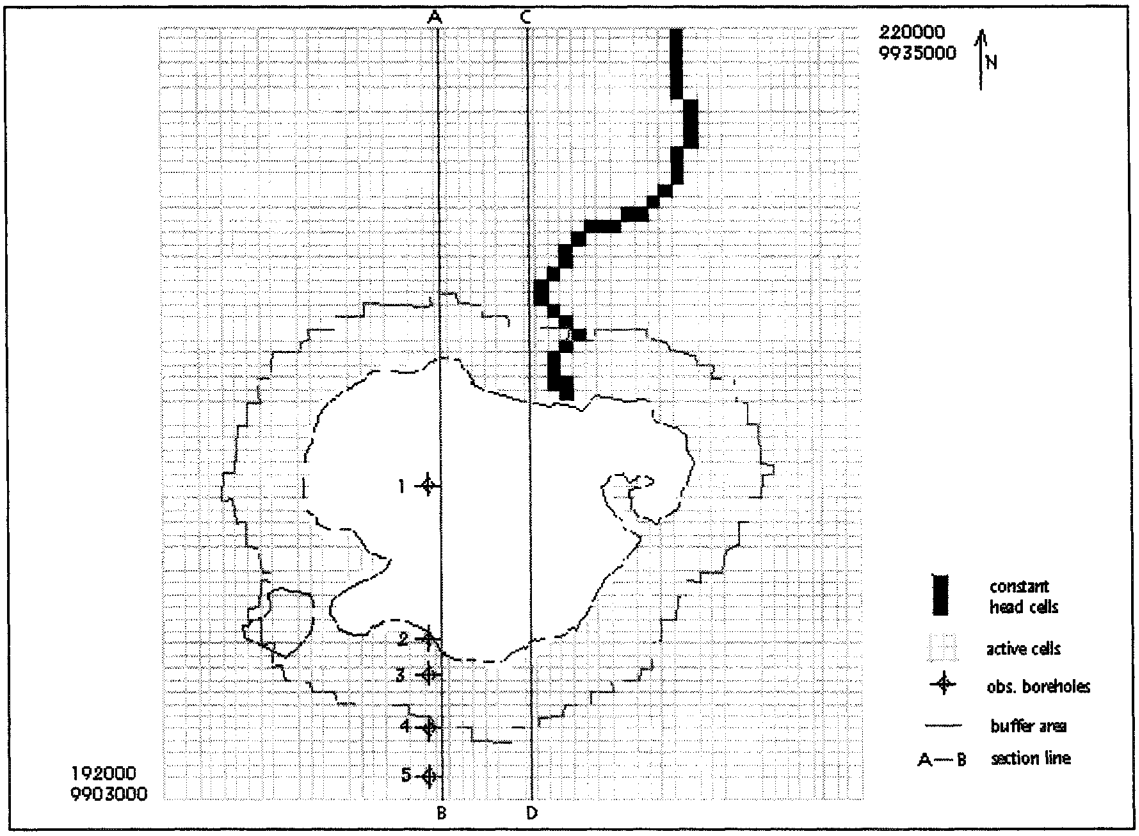
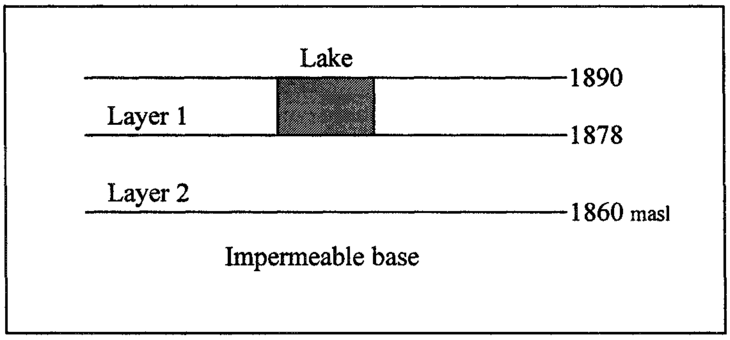


Figure 4.4 - Model layout

At any given time, the stress on each lake cell is the same to ensure that the lake surface remains flat. However, due to errors in rounding off in calculations, the lake surface is never perfectly flat, but varies from cell to cell in the order of 1-3 cm.

The aquifer is 30 m thick. The upper layer lies at 1890 masl and is 12 m thick. The lower layer is 18 m thick.



Pumping test data and geophysical data were both scarce and incomplete, and groundwater level data was not very accurate (boreholes not levelled). Consequentially, values used for the hydraulic parameters in the model were guided by results of a study conducted in the Naivasha area in 1990 by the Kenyan Ministry of Energy (Ministry of Energy, 1990).

AREA	LITHOLOGY	GEOMETRIC MEAN EST. CONDUCTIVITY (m/d)	GEOMETRIC MEAN EST. TRANSMISS. (sq m/d)	TOTAL NO. BOREHOLES
NE Naivasha	sediments & volcanics	12 (33)	307 (1170)	35
SE Naivasha	sediments & volcanics	20 (114)	502 (3082)	22
SW Naivasha	sediments & volcanics	63 (196)	297 (940)	17
NW Naivasha	sediments & volcanics	148 (818)	1601 (5308)	26

Table 4.1 - Average aquifer characteristics and lithologies from borehole data from Min. of Energy Study (Figures in brackets are arithmetic means)

AREA	BOREHOLE ID	LITHOLOGY	TRANSMISS. (sq m/d) (pump. test)
NE Naivasha	C1063	sediments & volcanics	1330
NW Naivasha	C2657	sediments & volcanics	307

Table 4.2 - Aquifer characteristics and lithologies from pumping test data (Untied Nations-Kenya Government Study, McCann, 1974)

AREA	LITHOLOGY	EST. HYDRAULIC CONDUCTIVITY (m/d)
NE Naivasha	sediments & volcanics	10
SE Naivasha	sediments & volcanics	17
SW Naivasha	sediments & volcanics	10
NW Naivasha	sediments & volcanics	53

Table 4.3 - Estimated aquifer characteristics used in the model

Based on the above, the model was assigned 5 hydraulic conductivity zones, one for each quadrant, and a fifth for the lake. The lake however was located in layer one only. In addition to an extremely high conductivity value, lake cells in the model are defined, more importantly, by storage coefficient. Lake cells were given a storage coefficient value of 0.999 to simulate free water flow such as occurs in water bodies (a value of 0.999 was used as opposed to the expected 1.000 as it has been suggested that using 0.999 creates greater numerical stability in the model - groundwater mailing list comment). Layer 1 is hydrogeologically unconfined and has a storage coefficient of 0.12 while the lower layer which is hydrogeologically confined has a storage coefficient value of 0.0001.

Flow between the two model layers is affected by a layer of lower conductivity. This semi-confining layer is not explicitly included as a separate model layer, but rather its effect is modelled by means of a vertical conductivity term. The vertical conductivity value was taken to be 1/10 th of the corresponding K value for the cell (general modelling assumption) as data to suggest otherwise was unavailable.

The various inflows to and outflows from the lake were computed into a single balance figure which was assigned to the wells. In this model situation and the scenarios which follow, the model is always run from a flat water table position.

4.7 Groundwater Storage

Change in groundwater storage is influenced by:

- head change
- surface area over which the change occurs
- aquifer storage coefficient or storativity

In the aquifer surrounding Lake Naivasha, the area over which groundwater head change occurs depends on the magnitude of change in the lake head, as the lake head is the main factor affecting groundwater levels. The larger the change in lake head, the wider the area of the aquifer that will be affected. Vekerdy (1996), in studying the area of influence of a river on groundwater levels in an alluvial basin in Hungary says "the river-influenced zone may be determined as the zone where the abrupt stage changes of the river have direct effect on the groundwater level. For practical reasons the river-influenced zone is determined as the zone where the groundwater head rise generated by the rivers largest possible (in duration and head) flood wave exceeds a minimum which is determined by the random error of the groundwater head measurement". For the purpose of this study, the area of influence of the lake is limited to a 3 km buffer area around the lake, as beyond this, groundwater heads change very little in response to the largest lake level change modeled (4.4 m rise, scenario 7).

Storage coefficient or storativity is the controlling factor governing storage change. Aquifer storativity may be defined as *" the volume of water that an aquifer releases from, or takes into, storage per unit surface area of aquifer per unit change in head"* (Ward & Robinson, 1990). It is thus the ratio of a volume of water to a volume of aquifer and is hence dimensionless. For confined aquifers, it is defined as follows:

$$\mathbf{S = \rho g b(\alpha + n\beta)}$$

S - storativity
ρ - water density (M/L^3)
g - acc. due to grav. (L/T^2)
b - aquifer thickness (L)
β - water compressibility (N/M^2)
α - aquifer compressibility (N/M^2)
n - porosity

In a hydrogeologically unconfined aquifer, the storage coefficient approximates the specific yield since dewatering of the aquifer occurs. In a hydrogeologically confined aquifer however, water is not released through dewatering of the pores, but through compression of the aquifer and expansion of the water. Hence the volume of water released by such an aquifer per unit change in head is significantly less than the volume released by an unconfined aquifer. In otherwords, confined aquifers have much lower storativities than unconfined aquifers.

The storage calculations were done in ILWIS using map and table calculations. The procedure is detailed below.

Procedure for computing groundwater storage in ILWIS using head changes calculated in PMWIN

⇒ maps **lake95** and **buff3** are created

lake95 is a raster map of the lake based on the 1995 satellite image
buff3 is a boolean map of the lake with a 3 km buffer zone around it

⇒ after the model is run, the drawdown at the end of each stress period is saved as an ASCII file and imported into ILWIS as a map called **d1(d2,d3...)**

PMWIN calculates drawdown by subtracting the difference between initial and final heads

⇒ map **suitarea** is created

suitarea = iff ((buff3=1 and lake95=0), d1, 0)

this map selects drawdown values in the 3 km buffer zone around the lake. All other cells, including the lake itself are assigned a value of zero

⇒ the storage maps are created

storage1=suitarea*0.12*sq(500)	storage in layer 1 (unconfined)
storage2=suitarea*0.0001*sq(500)	storage in layer 2 (confined)

⇒ the storage maps are opened as histograms, and columns **no_obs** and **storage** created

no_obs=value*npix	no_obs ensures all lake pixels are counted
storage = cumulative (no_obs)	this adds up the pixels

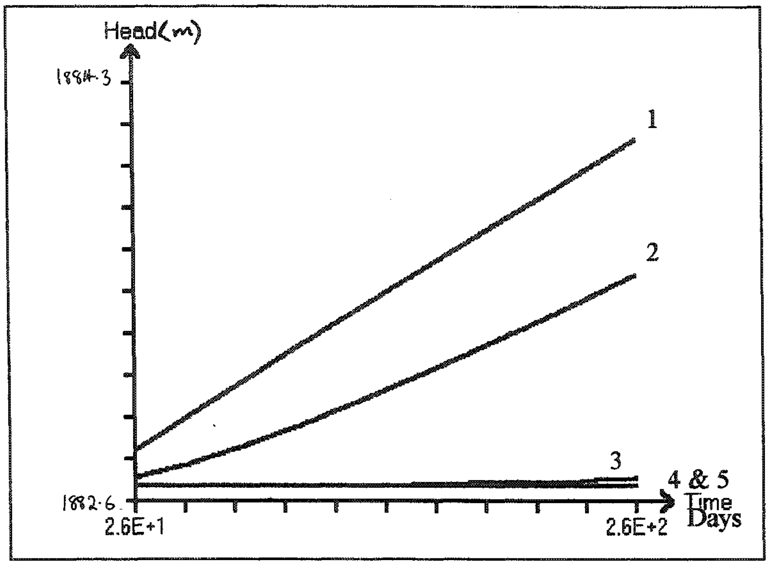
The implications of the differing storage coefficients of the upper and lower layers of the aquifer becomes very obvious when the change in storage of the two layers is compared.

4.8 Field Situation

This example models actual field conditions. Over an observation period of 260 days (13/05/97 - 08/02/98), due to very heavy rainfall at the end of October, the level of the lake rose from 1882.6 masl to 1884.3 masl, a 1.7 m increase. This large increase in level resulted in a significant volumetric increase of the lake. Lake volume increased by 216 x 10^6 m^3. This was calculated as follows:

$$\text{volume} = \text{change in head of lake x lake area } (127 \times 10^6 \text{ m}^2)$$

For this volume increase to occur, approximately 830,580 million m^3 of water has to be added daily to the lake. This volume was input into the lake cells of the model, and the model run. To observe the changes in groundwater level with time at varying distances from the lake, 5 observation boreholes were placed.



BOREHOLE NO.	DISTANCE FROM LAKE (m)
1	in the lake
2	300
3	1800
4	4000
5	6000

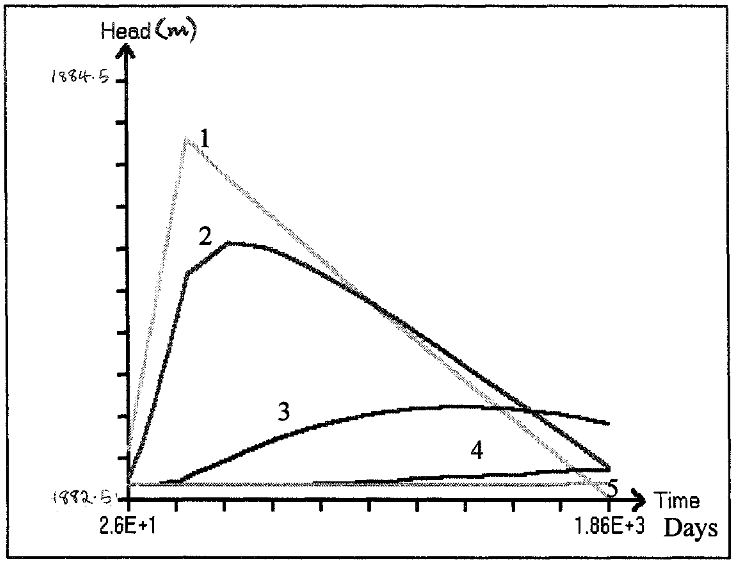
Figure 4.5 - Run 1: Propagation of heads in boreholes spaced away from the lake

From the result of the model run (fig. 4.5), it can be seen that the groundwater level in borehole 2, being closest to the lake, responds fastest to the rising lake level. The groundwater level does not however rise as much as the lake rises. Whereas the lake rises by 1.7 m, the groundwater level at borehole 2 rises by only 1 m (60% of the lake rise). At increasing distance from the lake, the groundwater head propagation becomes less. At borehole 4 (4000 m away), the groundwater table remains stationary, showing no reaction to the rise in lake level.

At the end of the modeled period (260 days), groundwater storage in the upper layer of the aquifer is calculated to have increased by 5.7×10^6 m^3, and in the lower layer by only 4730 m^3. The gain in the lower layer is 0.08 % of the gain in the upper layer, a direct consequence of the difference in storage properties of the two layers.

For the lake to maintain this level to which it has risen, water flowing into it must exceed or equal water flowing out. However, assuming constant dry conditions, with a deficit of 0.135×10^6 m^3/day, after approximately 4 years (1600 days) the lake will resume its former level. This period of four years is derived by dividing the volume increase of the lake (216×10^6 m^3) by the rate of water loss (0.135×10^6 m^3/day).

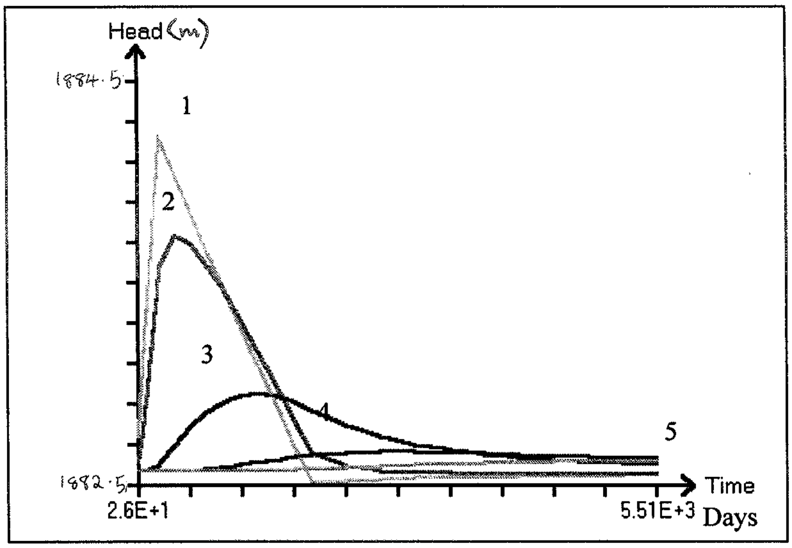
The described situation was modelled, and the result is shown in fig. 3.7. At borehole 4, 4000 m from the lake, it takes some time for the change in lake head to propagate through the aquifer. It is not until approximately 1 ½ years after the lake level has risen that the groundwater table at borehole 4 begins to rise. Near to the lake on the other hand, the aquifer responds quickly to the rise and fall in lake level. After the lake has resumed its original level, storage in the upper layer is reduced by 1.9 x 10^6 m^3, and in the lower layer by 1560 m^3. This represents a 33% groundwater storage loss in the aquifer.



BOREHOLE NO.	DISTANCE FROM LAKE (m)
1	in the lake
2	300
3	1800
4	4000
5	6000

Figure 4.6 - Run 2: Propagation of heads in boreholes spaced away from the lake

A final model run was made to determine the length of time for which groundwater would remain in storage. In this run, the lake level was kept constant (at the original level) for 10 years and the model run (see fig.). (Although the lake level should be constant, it slowly rises due to backflow from the aquifer). Even after this 10 year period, there is still groundwater in storage in the aquifer. The upper layer contains 0.84 x 10^6 m^3 and the lower layer 690 m^3.



BOREHOLE NO.	DISTANCE FROM LAKE (m)
1	in the lake
2	300
3	1800
4	4000
5	6000

Figure 4.7 - Run 3: Propagation of heads in boreholes spaced away from the lake

4.9 Scenarios

Eight climatic scenarios are modelled to see their effects on groundwater storage. These are:

- a wet year (W)
- an average year (A)
- a dry year (D)
- 10 successive average years (10A)
- 5 average years followed by 3 dry years then 2 average years (5A3D2A)
- 3 dry years followed by 7 average years (3D7A)
- 5 average years followed by a wet year and 4 dry years (5A1W4D)
- alternating average and dry years (2A2D2A2D2A)

As already discussed, wet, dry and average conditions relate to the water budget in section 3.0. Wet conditions imply the lake has a positive yearly water balance of 245 x 10^6 m^3. Average conditions imply a positive lake balance of 44 m^3, and dry conditions a yearly deficit of 49 m^3.

4.9.1 Scenario 1 - Wet Year

In a wet year, there is a positive balance of 245 x 10^6 m^3 of water in the lake. In order to gain an appreciation of the storage changes on a monthly, as opposed to yearly basis, the lake water balance was broken down to a monthly time scale. This was done based on discharge and evaporation. Monthly discharge averages (1960-1985) were calculated for station 2GB1 (the Malewa river discharge station nearest the lake - see fig.2.3). The percentage of discharge occurring in each month was then computed and applied to the yearly lake inflow to give a monthly inflow figure. Monthly outflows were deduced in a similar manner. The percentage of evaporation occurring in each month was calculated (based on data in table 3.4) and applied to the yearly lake outflow to obtain a monthly outflow figure. The balance between inflow and outflow for each month was then computed and input into the lake cells of the PMWIN model (see table 4.4).

This method gives an ESTIMATE of the monthly variability of the lake water balance, as the balance depends not only on discharge and evaporation, but several other factors. Additionally, the percentages of the total yearly discharge and evaporation occurring within each month are likely to be different for wet, dry and average conditions, and not constant as has been assumed.

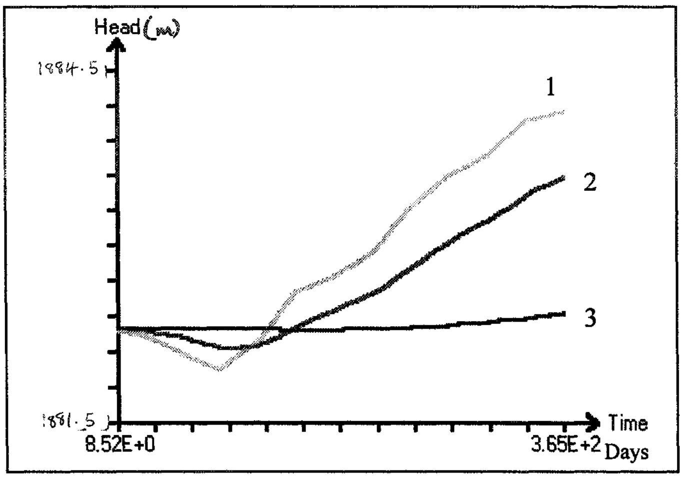
Yearly inflow = 619 x 10^6 m^3
Yearly outflow = 374 x 10^6 m^3
Yearly balance = 245 x 10^6 m^3

MONTH	% of total yearly discharge	INFLOW (cu. m)	% of total evaporation	OUTFLOW (cu. m)	LAKE BALANCE (cu. m)
Jan	3.4	21 046 000	7.8	29 172 000	-8 126 000
Feb	2.7	16 713 000	9.6	35 904 000	-19 191 000
Mar	3.1	19 189 000	10.2	38 148 000	-18 959 000
Apr	9.8	60 662 000	8.2	30 668 000	29 994 000
May	13.1	81 089 000	7.5	28 050 000	53 039 000
Jun	7.0	43 330 000	6.9	25 806 000	17 524 000
Jul	8.6	53 234 000	7.1	26 554 000	26 680 000
Aug	13.1	81 089 000	7.9	29 546 000	51 543 000
Sep	11.1	68 709 000	8.8	32 912 000	35 797 000
Oct	9.9	61 281 000	9.8	36 652 000	24 629 000
Nov	11.3	69 947 000	7.7	28 798 000	41 149 000
Dec	6.9	42 711 000	8.5	31 790 000	10 921 000

Table 4.4 - Monthly lake balance for a wet year

The table shows that from January to March, the lake water balance is negative as there is more water flowing out of, rather than into the lake. The deficit is greatest in February which is normally a dry month. The lake balance is most positive in May, which coincides with the main rainy period (see section 2.3.1).

The model representing the described conditions was run in PMWIN, and the result shown in fig. The level of the lake increased by 1.9 m over the year.



BOREHOLE NO.	DISTANCE FROM LAKE (m)
1	in the lake
2	300
3	1800

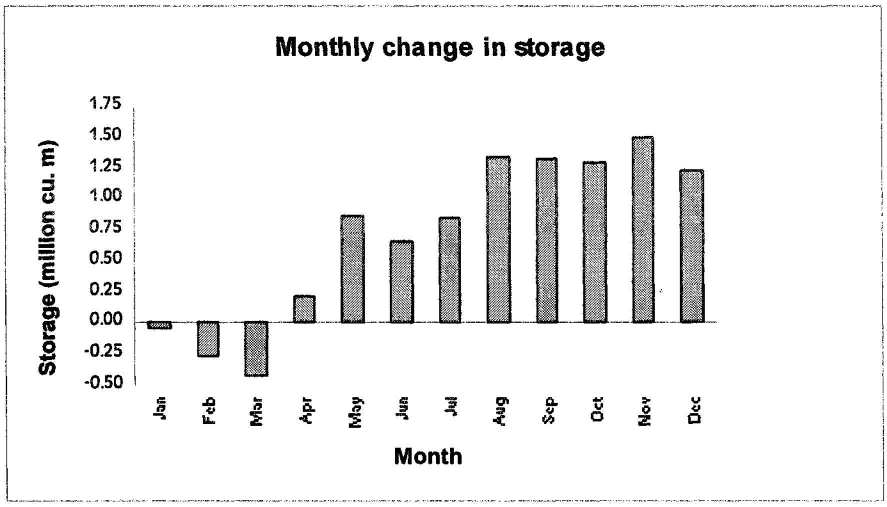
Figure 4.8 - Wet Year: Propagation of heads in boreholes spaced away from the lake

Storage changes were calculated in ILWIS using the procedure described in section 4.7. The results are displayed in the table below. In the table, a negative sign indicates a decrease in storage, and net storage refers to the total additional storage available at the end of the month in question. For example, from the table, at the end of August, the net storage in the aquifer is 3.2 x 10^6 m^3. If the aquifer had an initial storage of 100 x 10^6 m^3 of water, then by August of a wet year, its storage would have increased to 103.2 x 10^6 m^3. Storage is calculated for the unconfined layer only as the storage in the confined layer is very small (approximately 0.08 % of the unconfined storage). An error of 1% is thus larger than the storage changes in the confined layer, and so it has not been included.

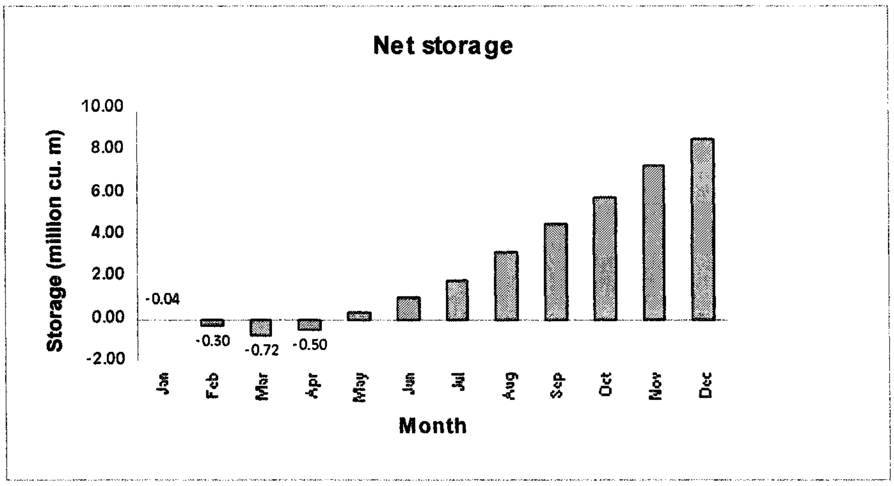
MONTH	CHANGE IN STORAGE (cumecs)	NET STORAGE (cumecs)
Jan	- 35 910	- 35 910
Feb	- 260 580	- 296 490
Mar	- 425 370	- 721 860
Apr	216 090	- 505 770
May	860 820	355 050
Jun	659 700	1 014 750
Jul	845 640	1 860 390
Aug	1 336 650	3 197 040
Sep	1 319 850	4 516 890
Oct	1 289 490	5 806 380
Nov	1 503 450	7 309 830
Dec	1 225 710	8 535 540

Table 4.5 - Monthly groundwater storage change

Graphs of monthly storage changes and net storage are shown in fig. 3.10 below.



a)



b)

Figure 4.9 - Patterns of storage change in a wet year

When figures 3.10 a) and b) are compared, it is noticed that despite the gain of groundwater in April, the overall groundwater storage balance for the month is still in deficit as the groundwater which was lost in the first three months is not yet compensated for. After the rainy period in May, the groundwater resources are replenished.

To summarize, at the end of what is considered a wet year, the additional amount of groundwater available in a 3 km buffer zone around Lake Naivasha is roughly 8.5 million cubic metres. This represents 3.5% of the lake balance for the year.

4.9.2 Scenario 2 - Average Year

The lake water balance in an average year is positive 44 x 10^6 m^3. Daily there is a volume of 0.12 x 10^6 m^3 being added to the lake storage. A monthly water balance for the lake was calculated and is shown in the table below.

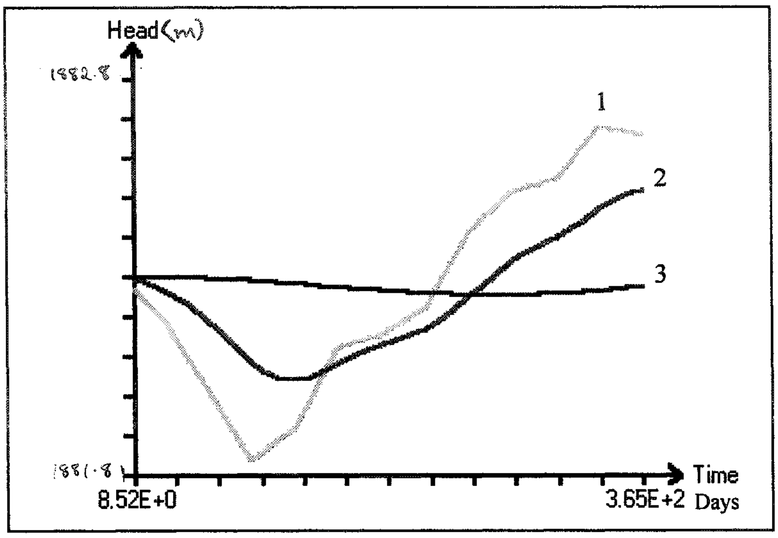
Yearly inflow = 401 x 10^6 m^3
Yearly outflow =356 x 10^6 m^3
Yearly balance = 44 x 10^6 m^3

MONTH	% of total yearly discharge	INFLOW (cu. m)	% of total evaporation	OUTFLOW (cu. m)	LAKE BALANCE (cu. m)
Jan	3.4	13 634 000	7.8	27 768 000	-14 134 000
Feb	2.7	10 827 000	9.6	34 176 000	-23 349 000
Mar	3.1	12 431 000	10.2	36 312 000	-23 881 000
Apr	9.8	39 298 000	8.2	29 192 000	10 106 000
May	13.1	52 531 000	7.5	26 700 000	25 831 000
Jun	7.0	28 070 000	6.9	24 564 000	3 506 000
Jul	8.6	34 486 000	7.1	25 276 000	9 210 000
Aug	13.1	52 531 000	7.9	28 124 000	24 407 000
Sep	11.1	44 511 000	8.8	31 328 000	13 183 000
Oct	9.9	39 699 000	9.8	34 888 000	4 811 000
Nov	11.3	45 313 000	7.7	27 412 000	17 901 000
Dec	6.9	27 669 000	8.5	30 260 000	-2 591 000

Table 4.6 - Monthly lake water balance for an average year

As in the case of the wet year, from January to March the lake loses water, but also in December. The amount of water lost by lake is 40% more than the losses incurred during the wet year.

The lake balance for the different months was input into the lake cells of the model, and the model was run. The result is shown in fig. Although by the end of the year the lake undergoes a net rise in level of 0.4 m, from January to March when it loses water, its level drops by 0.5 m. At borehole 3 (1800 m away), the groundwater table remains relatively stationary during the entire year.



BOREHOLE NO.	DISTANCE FROM LAKE (m)
1	in the lake
2	300
3	1800

Figure 4.10 - Average Year: Propagation of heads in boreholes spaced away from the lake

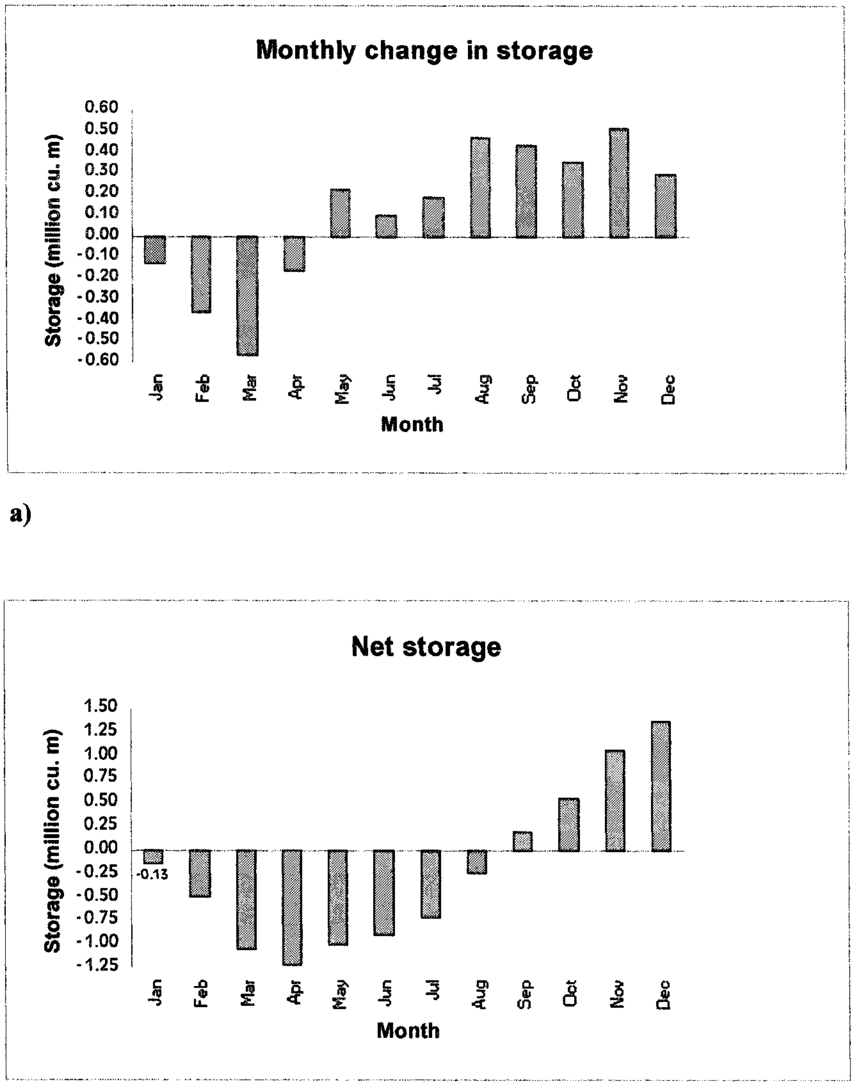
The table below shows monthly changes in groundwater storage during the average year.

MONTH	CHANGE IN STORAGE (cumecs)	NET STORAGE (cumecs)
Jan	- 128 730	- 128 730
Feb	- 358 710	- 487 440
Mar	- 565 380	-1 052 820
Apr	- 163 410	-1 216 230
May	224 850	- 991 380
Jun	96 120	- 895 260
Jul	185 100	- 710 160
Aug	468 900	- 241 260
Sep	433 860	192 600
Oct	354 870	547 470
Nov	513 210	1 060 680
Dec	295 710	1 356 390

Table 4.7 - Monthly groundwater storage change

From the table, the aquifer loses groundwater from January to April. The groundwater lost during this period is not compensated for until September, five months later. By the end of the year, 1.4 million cubic metres of groundwater is added to the existing groundwater resources in the aquifer. This represents 3% of the lake balance for the year.

The pattern of groundwater storage change is shown in the graphs below.



a)

b)

Figure 4.11 - Patterns of storage change in an average year

4.9.3 Scenario 3 - Dry Year

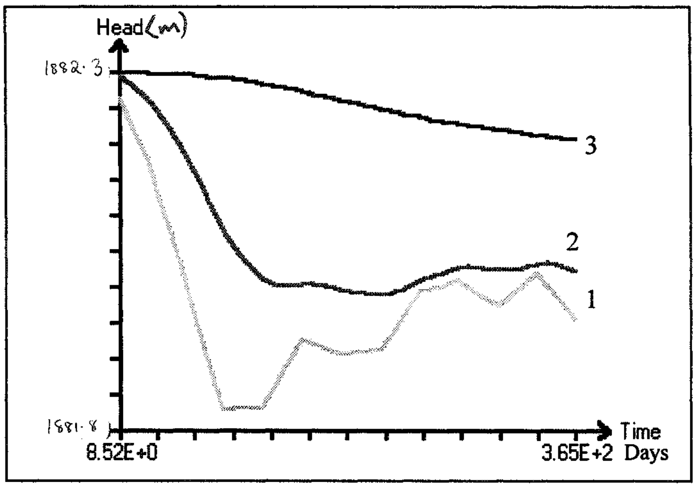
In a dry year, the lake loses on average 0.14 x $10^6 m^3$ of water per day, resulting in a yearly deficit of 49 x 10^6 m^3. A monthly lake water balance was calculated (see table 4.8).

Yearly inflow = 267 x 10^6 m^3
Yearly outflow = 316 x 10^6 m^3
Yearly balance = 49 x 10^6 m^3

MONTH	% of total yearly discharge	INFLOW (cu. m)	% of total evaporation	OUTFLOW (cu. m)	LAKE BALANCE (cu. m)
Jan	3.4	9 078 000	7.8	24 648 000	-15 570 000
Feb	2.7	7 209 000	9.6	30 336 000	-23 127 000
Mar	3.1	8 277 000	10.2	32 232 000	-23 955 000
Apr	9.8	26 166 000	8.2	25 912 000	254 000
May	13.1	34 977 000	7.5	23 700 000	11 277 000
Jun	7.0	18 690 000	6.9	21 804 000	-3 114 000
Jul	8.6	22 962 000	7.1	22 436 000	526 000
Aug	13.1	34 977 000	7.9	24 964 000	10 013 000
Sep	11.1	29 637 000	8.8	27 808 000	1 829 000
Oct	9.9	26 433 000	9.8	30 968 000	-4 535 000
Nov	11.3	30 171 000	7.7	24 332 000	5 839 000
Dec	6.9	18 423 000	8.5	26 860 000	-8 437 000

Table 4.8 - Monthly lake water balance for a dry year

From the table above, in a dry year the lake loses water in six months of the year, from January to March (as in the other examples), and in June, October and December. Most of the gain in lake storage occurs in May and August. The monthly lake balance amounts were input into the lake cells of the model, and the model was run. The result of the model run is shown below.



BOREHOLE NO.	DISTANCE FROM LAKE (m)
1	in the lake
2	300
3	1800

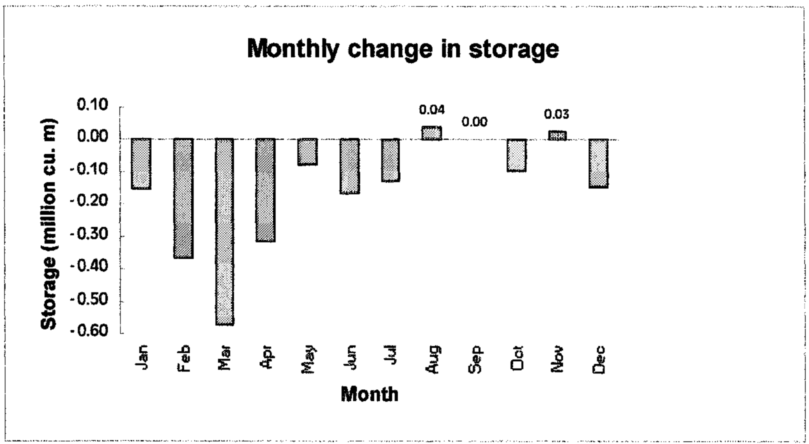
Figure 4.12 - Dry Year: Propagation of heads in boreholes spaced away from the lake

Monthly groundwater storage changes were calculated in ILWIS and are tabulated below.

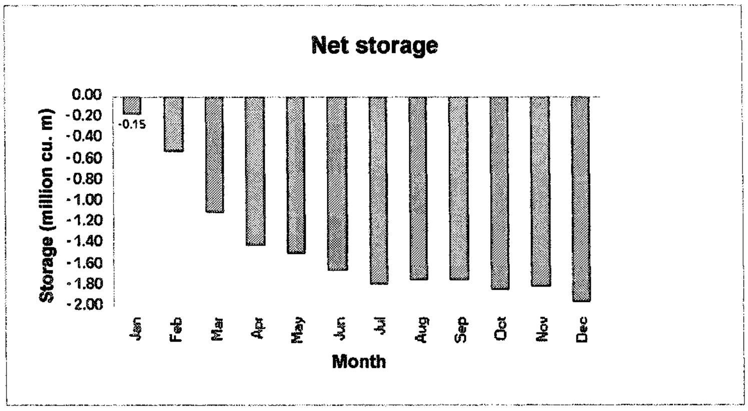
MONTH	CHANGE IN STORAGE (cu. m)	NET STORAGE (cu. m)
Jan	- 152 130	- 152 130
Feb	- 364 500	- 516 630
Mar	- 573 090	-1 089 720
Apr	- 316 380	-1 406 100
May	- 77 910	-1 484 010
Jun	- 165 240	-1 649 250
Jul	- 127 590	-1 776 840
Aug	40 830	-1 736 010
Sep	90	-1 735 920
Oct	- 93 990	-1 829 910
Nov	25 950	-1 803 960
Dec	- 145 350	-1 949 310

Table 4.9 - Monthly groundwater storage change

Despite the fact that the lake gains water for half of the year, the amount gained is not enough to compensate for the months when it loses water, and as a result, the groundwater storage is in deficit for the entire year. By the end of the year, existing groundwater resources are depleted by 2×10^6 m^3 . The pattern of storage loss is shown in fig. below.



a)



b)

Figure 4.13 - Patterns of storage change in a wet year

4.9.4 Discussion

The changes in groundwater storage in a wet year, an average year, and a dry year are summarized in figure. As can be clearly seen, there is a large volume of water available at the end of a wet year, as compared to average and dry years. Based on the calculations, at the end of a wet year, groundwater storage increases by 8.5 million m^3. At the end of an average year the increase is 1.4 million m^3 (16 % less), and after a dry year, the groundwater resources are actually depleted, by 2.0 million m^3. The implications of this are obvious. In the average year, not more than 1.4 million m^3 of groundwater can be safely abstracted, and in a dry year, as far as is possible, use of groundwater resources should be limited.

Acknowledging the monthly pattern of groundwater storage change is important, especially during the average year. Although excess groundwater is available at the end of the average year, and the lake gains water from April to November, there is a groundwater deficit in the first eight months of the year. During these first eight months, it is thus not advisable to put pressure on groundwater resources.

These conclusions are all based on groundwater storage available at the end of one year. This is not very realistic as the effect of storage change is cumulative. The following scenarios therefore model changes in groundwater storage over ten year periods under varying climatic conditions.

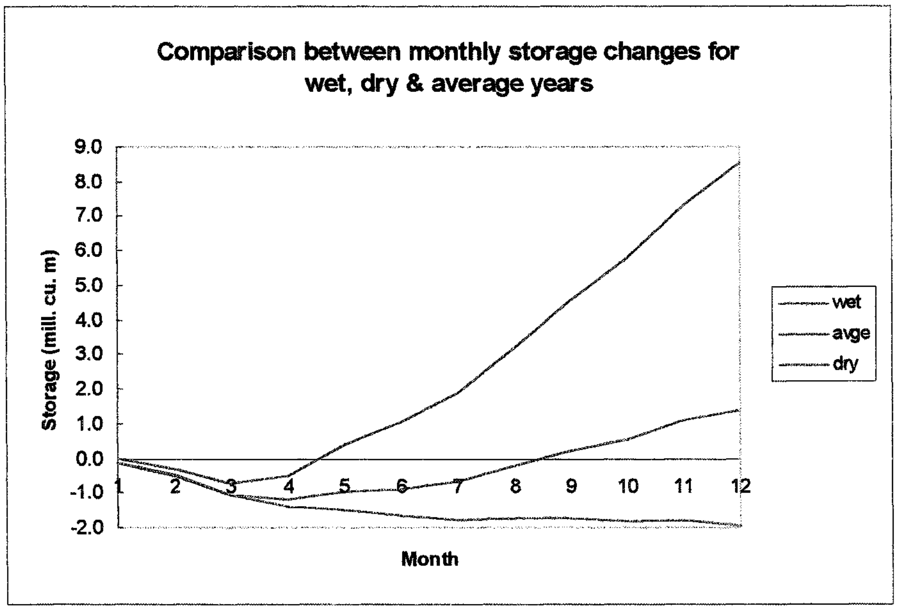


Figure 4.14 - Storage change: wet,average, dry years

4.9.5 Scenario 4 - 10A

In this scenario, 10 successive average year conditions are modeled. At the end of such a period, the lake level rises by 3.1 m, and groundwater resources in the aquifer surrounding the lake are augmented by 34 million m^3 (see table). This is a considerable volume of water, and represents 8% of the lake water balance over the ten year period (440 x 10^6 m^3). The pattern of groundwater storage change is regular as figure shows.

YEAR	YEAR TYPE	CHANGE IN STORAGE (cu. m)	NET STORAGE (cu. m)
1	Avge	2 188 710	2 188 710
2	Avge	2 894 520	5 083 230
3	Avge	3 239 490	8 322 720
4	Avge	3 433 410	11 756 130
5	Avge	3 551 340	15 307 470
6	Avge	3 626 910	18 934 380
7	Avge	3 676 590	22 610 970
8	Avge	3 708 090	26 319 060
9	Avge	3 729 060	30 048 120
10	Avge	3 741 630	33 789 750

Table 4.10 - Yearly groundwater storage change (10A)

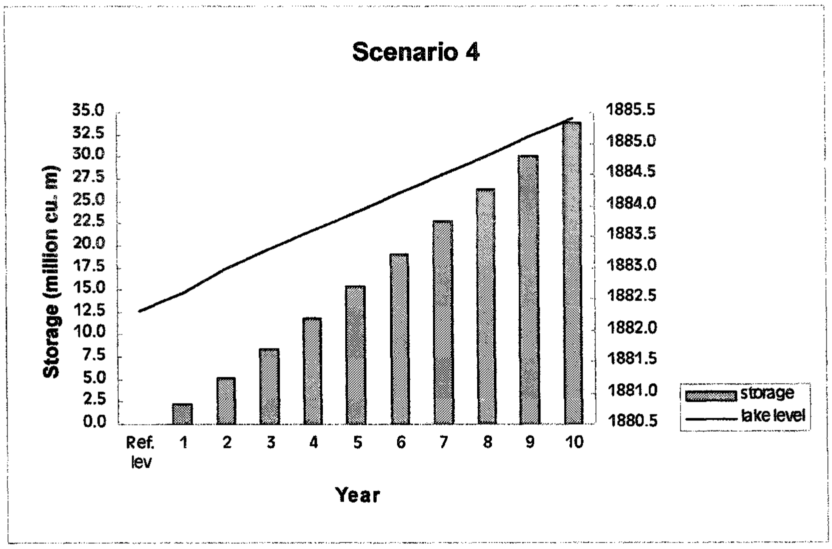


Figure 4.15 - Pattern of yearly groundwater storage change

4.9.6 Scenario 5 - 5A3D2A

In this scenario, a climatic pattern of 5 successive average years followed by 3 dry years and then 2 average years is modeled. This climatic pattern is more realistic than the previous scenario. At the end of the period, the lake level has undergone a net rise of 1.1 m and an additional 13 million m^3 of water is added to the groundwater storage (see table 4.11). The lake water balance over the ten year period is $+161 \times 10^6$ m^3. The volume of groundwater taken into storage in the aquifer thus represents 8% of the lake water balance. The pattern of groundwater storage change is shown in figure. As the figure indicates, it takes approximately one year for the groundwater storage to respond to the fall in lake level. In year 6 when the lake level falls by approximately a half of a metre, the groundwater storage does not change. It is not until year 7, one year later that the storage decreases.

Two groundwater head profiles were drawn along section lines A-B and C-D (see fig. 4.4) so that the propagation of groundwater heads can be more clearly seen. A profile was drawn for each year of the ten year time series (see fig 4.17). Figure shows the influence of the Malewa river on the groundwater levels. When the head in the river is higher than the head in the aquifer, water flows from the river into the aquifer. When the head in the aquifer is higher than the head in the river, the opposite occurs.

YEAR	YEAR TYPE	CHANGE IN STORAGE (cu. m)	NET STORAGE (cu. m)
1	Avge	2 188 710	2 188 710
2	Avge	2 894 520	5 083 230
3	Avge	3 239 490	8 322 720
4	Avge	3 433 410	11 756 130
5	Avge	3 551 340	15 307 470
6	Dry	- 100 260	15 207 210
7	Dry	-2 051 730	13 155 480
8	Dry	-2 845 770	10 309 710
9	Avge	457 860	10 767 570
10	Avge	2 205 330	12 972 900

Table 4.11 - Yearly groundwater storage change (5A3D2A)

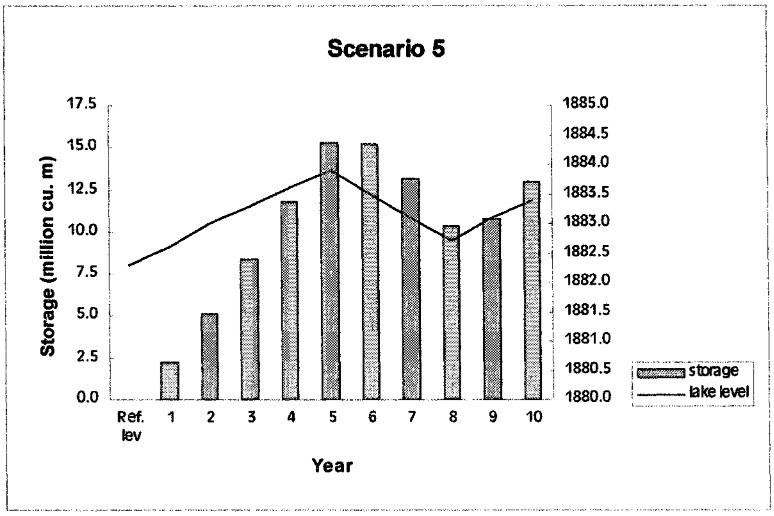
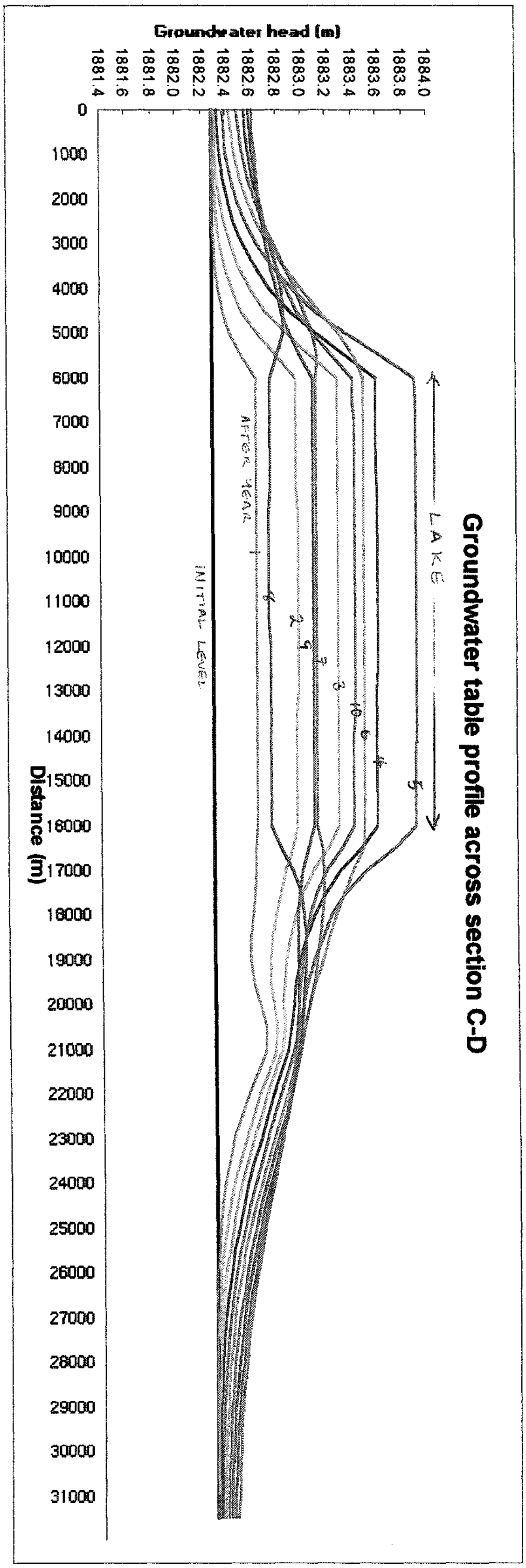


Figure 4.16 - Pattern of yearly groundwater storage change

b)



a)

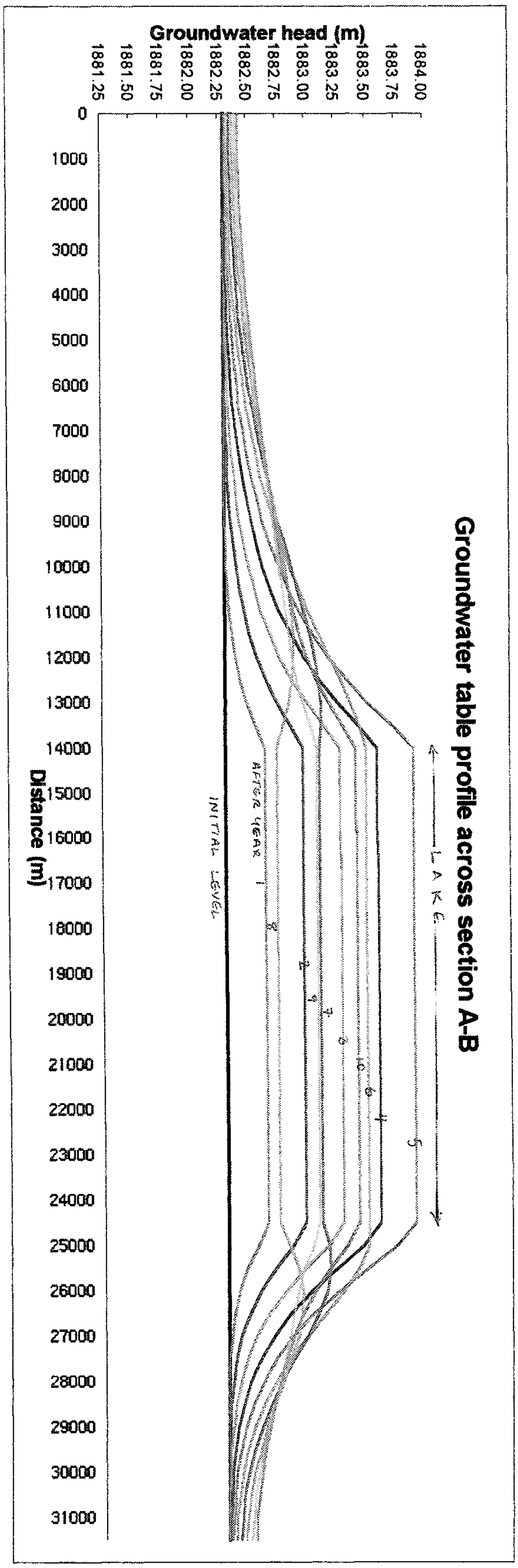


Figure 4.17 - Groundwater head profiles (Scenario 5)

4.9.7 Scenario 6 - 3D7A

In this scenario, a climatic condition of 3 dry years followed by 7 average years is modeled. At the end of the period, the lake level has risen by 1.3 m. Groundwater resources in the aquifer increased by 11 million m^3 (see table 4.12). This represents 7% of the lake water balance over the ten year period (161 x 10^6 m^3). In the previous example, the number of wet and dry years was the same, but their pattern of occurrence was different. This makes a difference in the groundwater storage. In the previous scenario, 2 million m^3 more groundwater was taken into storage. The pattern of groundwater storage change is shown in fig. 4.18. Although only the first 3 years are dry, it takes approximately 3 years for the groundwater lost during this period to be replenished. A profile of the groundwater heads at the end of each year is shown in fig. 4.19.

YEAR	YEAR TYPE	CHANGE IN STORAGE (cu. m)	NET STORAGE (cu. m)
1	Dry	-1 520 880	-1 520 880
2	Dry	-2 832 690	-4 353 570
3	Dry	-3 315 390	-7 668 960
4	Avge	162 210	-7 506 750
5	Avge	2 015 070	-5 491 680
6	Avge	2 749 080	-2 742 600
7	Avge	3 127 050	384 450
8	Avge	3 354 270	3 738 720
9	Avge	3 496 980	7 235 700
10	Avge	3 593 010	10 828 710

Table 4.12 - Yearly groundwater storage change (7D3A)

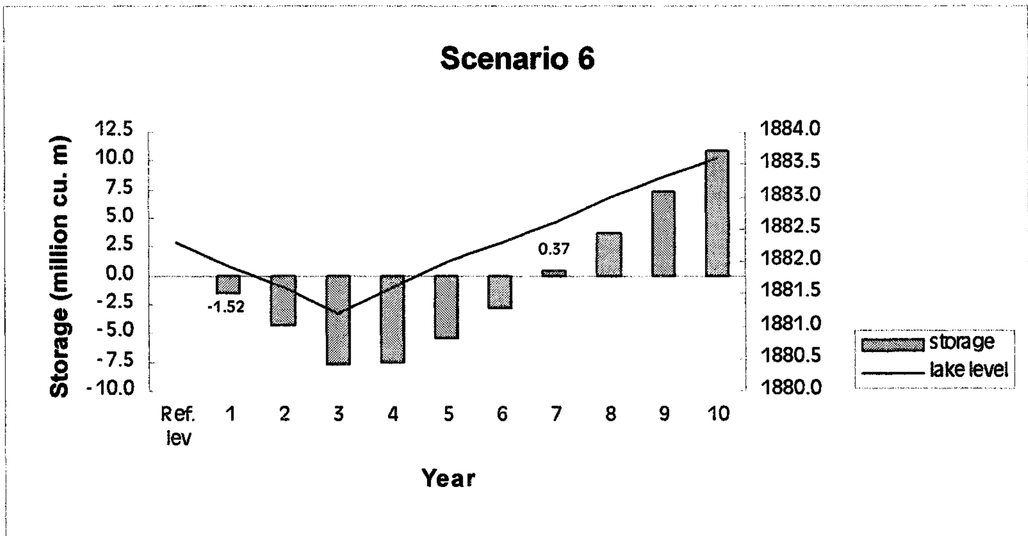


Figure 4.18 - Pattern of yearly groundwater storage change

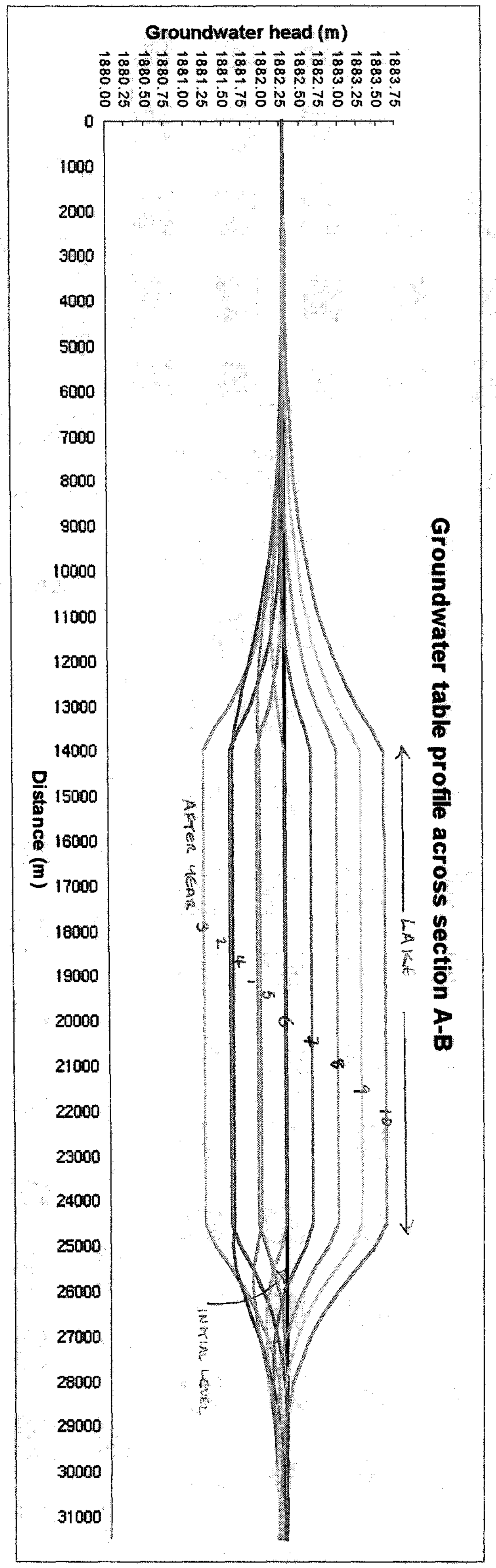


Figure 4.19 - Groundwater head profile across section A-B (Scenario 6)

4.9.8 Scenario 7 - 5A1W4A

In this scenario, 9 average years and a wet year are modeled. At the end the ten years, the lake level rises by 4.4 m, and groundwater resources in the aquifer surrounding the lake are increased by 49 million m^3 (see table 4.13). The occurrence of the wet year makes a significant difference to the groundwater resources, as when 10 average years were modelled (scenario 4), 34 million m^3 of groundwater (15 million m^3 less) was added to the existing storage. The periodic occurrence of a wet year would be good as it would significantly augment existing groundwater storage volumes. In this scenario, the volume of groundwater taken into storage over the ten year period again represents 8% of the lake water balance over the corresponding period. The pattern of groundwater storage change and the groundwater head profile across section line A-B are shown in figs. 4.20 & 4.21.

YEAR	YEAR TYPE	CHANGE IN STORAGE (cu. m)	NET STORAGE (cu. m)
1	Avge	2 188 170	2 188 170
2	Avge	2 895 060	5 083 230
3	Avge	3 239 490	8 322 720
4	Avge	3 433 410	11 756 130
5	Avge	3 551 340	15 307 470
6	Wet	11 553 030	26 860 500
7	Avge	7 927 800	34 788 300
8	Avge	5 467 590	40 255 890
9	Avge	4 672 740	44 928 630
10	Avge	4 306 020	49 234 650

Table 4.13 - Yearly groundwater storage change (5A1W4A)

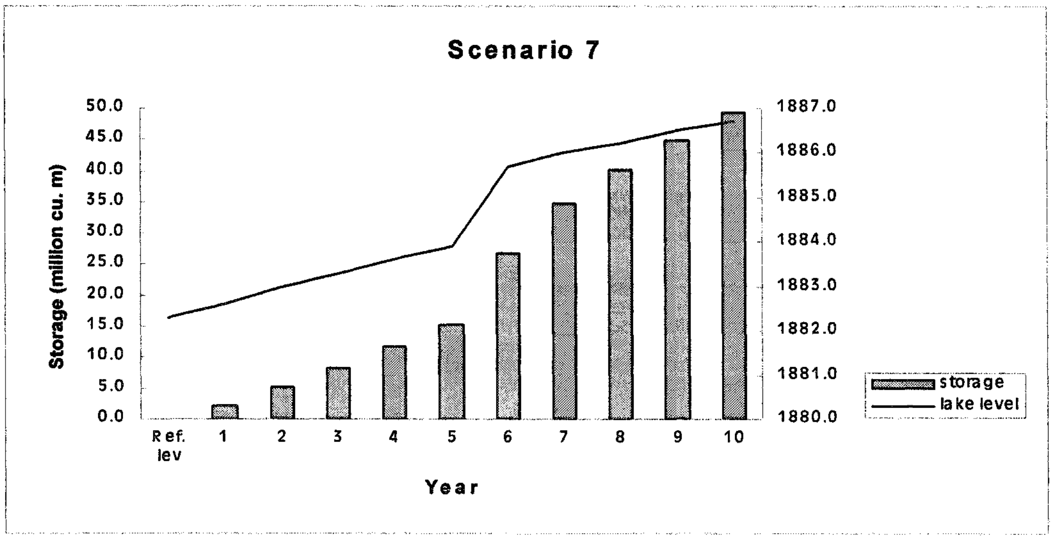


Figure 4.20 - Pattern of yearly groundwater storage change

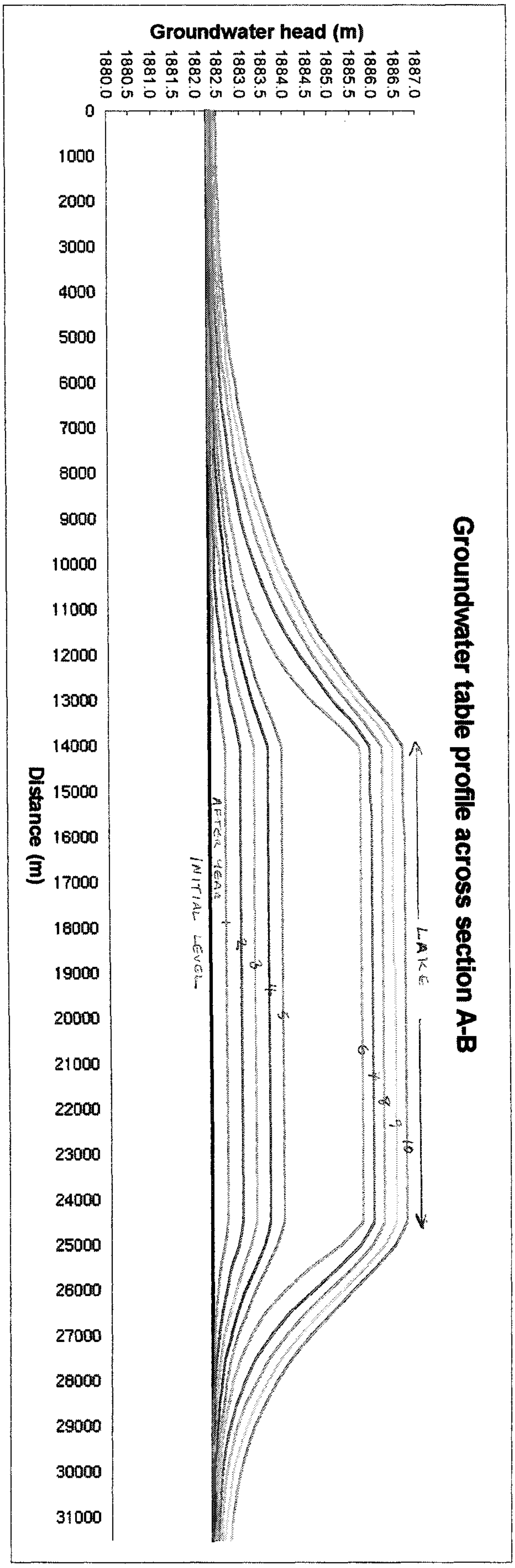


Figure 4.21 - Groundwater head profile across section A-B (Scenario 7)

4.9.9 Scenario 8 - 2A2D2A2D2A

In this final scenario, an alternating sequence of average and dry years is modelled. As would be expected, the lake level fluctuation is cyclic as is also the groundwater storage change (see fig. 4.22). At the end of the period, the lake level has risen by 0.5 m, and 5 x 10^6 m^3 of groundwater has been taken into storage (table 4.14). The lake water balance at the end of the ten year time series is 68 million m3. The volume of groundwater taken into storage represents 7% of the lake water balance. The groundwater head profile in fig. 4.23 shows how the groundwater heads fluctuate under the described climatic conditions.

YEAR	YEAR TYPE	CHANGE IN STORAGE (cu. m)	NET STORAGE (cu. m)
1	Avge	2 188 170	2 188 170
2	Avge	2 895 060	5 083 230
3	Dry	- 487 740	4 595 490
4	Dry	-2 292 300	2 303 190
5	Avge	722 760	3 025 950
6	Avge	2 355 480	5 381 430
7	Dry	- 760 650	4 620 780
8	Dry	-2 449 080	2 171 700
9	Avge	621 660	2 793 360
10	Avge	2 288 580	5 081 940

Table 4.14 - Yearly groundwater storage change (2A2D2A2D2A)

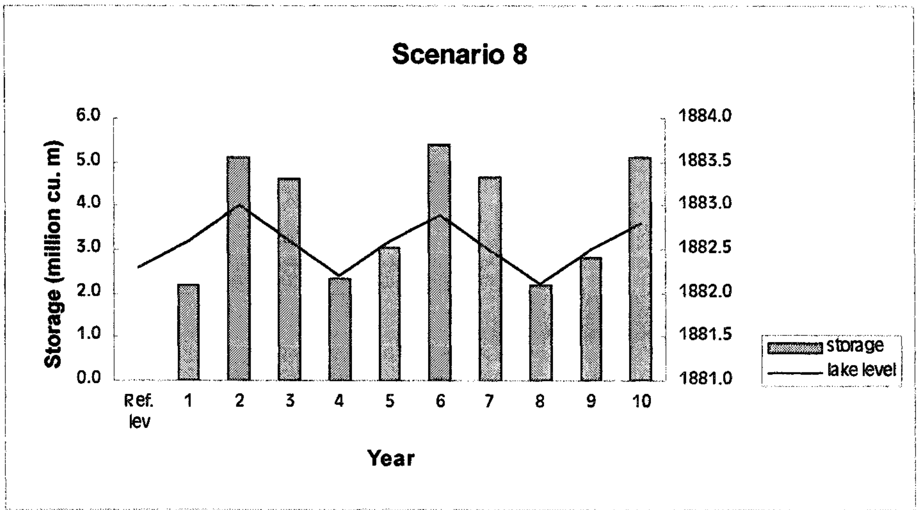


Figure 4.22 - Pattern of yearly groundwater storage change

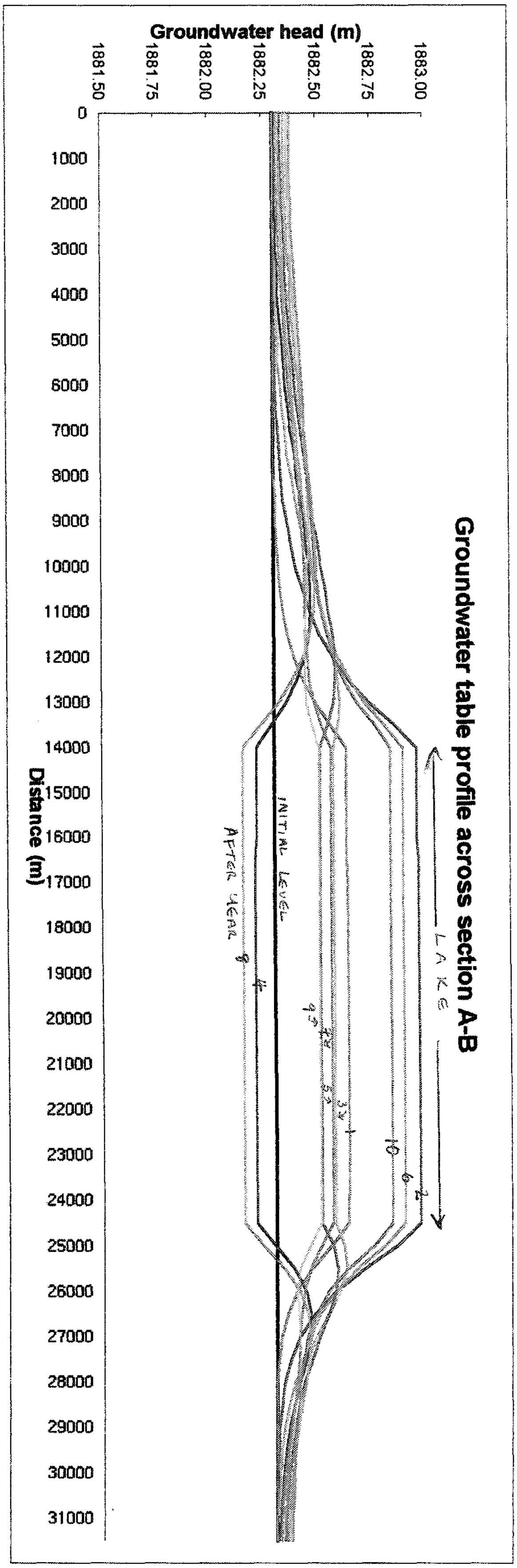


Figure 4.23 - Groundwater head profile across section A-B (Scenario 8)

4.9.10 Discussion

Scenarios 4-8 modelled varying patterns of lake level change over ten year time series to determine the resultant effect on groundwater storage volumes. From the exercise, it is clear that the pattern of change in the lake level from year to year is important in influencing groundwater storage. The lake level rising then falling, as opposed to falling then rising results in different changes in aquifer storage. This is because in general the groundwater table does not respond at once to the change in lake level (except immediately surrounding the lake). It takes time for the heads to propagate through the aquifer.

For the modelled scenarios, the volume of groundwater taken into storage at the end of the ten year series was found to range between 7-8% of the lake water balance over the period. This is a significant volume of water and it can be exploited and used to help meet water demands.

5.0 CONCLUSION AND RECOMMENDATIONS

This thesis is the product of the first of a four year project being carried out around Lake Naivasha. Under this four year project, the various components of the catchment water budget will be studied in detail, and a detailed groundwater model of the area created. The groundwater model presented here and the results derived from it are therefore preliminary.

In calculating the model inputs, many assumptions and generalizations had to be made. For the lake water balance, the factors contributing to lake inflow were precipitation, river discharge and groundwater inflow. Average catchment runoff coefficients had to be used to calculate river discharge into the lake. The catchment characteristics, especially for the Malewa catchment (the most important catchment) are highly variable and so using an average catchment runoff coefficient would have produced some errors. In future studies, aerial photo interpretation could be used to aid in estimating runoff coefficents for the different sections of the catchment. Groundwater inflow to the lake was also estimated, based on studies conducted by previous researchers. The estimate was low and in future study it could be verified or updated by using the model or the piezometric map to calculate the inflow. In order to do this though, accurate groundwater head measurements need to be available.

The factors contributing to lake outflow were evaporation, evapotranspiration, groundwater outflow, and abstractions. Evaporation and evapotranspiration play a large role in affecting lake levels and hence groundwater storage. These parameters were estimated based on averaged data. Again, this is another source of error. To generate primary data, an evaporation pan has been installed around the lake. More pans could be placed around or in the lake and a meterological station set up to record climatic variables. Optionally, remote sensing techniques could be used to help better estimate actual evapotranspiration using the surface energy balance approach.

Estimates of groundwater outflow rates were used. Seepage meters could be reinstalled at key positions in the lake (especially along the south shore where most outflow occurs) to directly measure the groundwater outflow.

In the setup of the model, a number of simplifying assumptions were made. The aquifer was assumed to be uniformly thick and this may not be the case. As proper lithological logs do not exist for most of the boreholes in the study area, it is recommended that geophysical surveys be carried out to accurately determine the aquifer geometry.

Groundwater table fluctuations were monitored in the upper layer of the aquifer. Piezometers should be installed in the semi-confining layer so that the groundwater table fluctuations in this layer can be monitored as well. This will allow the vertical flow between the two layers to be better modelled.

The hydraulic parameters of the aquifer had to be estimated both quantitatively and spatially. Attempts were made to use the model to determine aquifer transmissivity and storativity, but the solution was non-unique. Pumping tests should be carried out on spatially well distributed boreholes and wells so that the these parameters, and their spatial variation can be more accurately determined.

The boundaries of the model were defined as no-flow boundaries. Although in this study, the boundaries are far away enough to not affect the area of interest which is immediately surrounding the lake, in the detailed studies which will follow, it is recommended that the entire catchment be modelled in steady state and the fluxes across the boundaries of the area of interest be applied to those boundary cells which could be optionally set as general head boundary cells or flux boundaries depending on the steady state model solution.

Notwithstanding all the assumptions and simplifications made, the following conclusions were drawn.

- Groundwater storage in the lower layer of the aquifer is low compared to groundwater storage in the upper layer. Groundwater storage in the lower layer is only approximately 0.08% of the groundwater storage in the upper layer.

- At increasing distance from the lake, the groundwater table takes longer to react (if at all) to changes in the lake level. As a result, changes in groundwater storage do not immediately follow changes in the lake level. There is a delay period which depends on the magnitude of the change in lake level.

- The climatic pattern affects groundwater storage volumes. For example, wet years followed by dry years, as opposed to dry years followed by wet years results in different volumes of groundwater being available.

- In a wet year, the volume of groundwater stored in that part of the aquifer lying within a 3 km buffer zone around the lake is increased by 8.5 million m^3. This represents 3.5% of the lake balance for the year. In an average year, the groundwater stored in this same zone is increased by 1.4 million m^3. This represents 3% of the lake balance for the year. In a dry year, groundwater resources are depleted by approximately 2 million m^3. The volume of groundwater stored in the aquifer surrounding the lake is thus significant, and so groundwater can provide a buffering effect for dry year conditions. The occasional wet year significantly augments groundwater resources and provides a sort of safe yield reserve for dry years.

The above conclusions are all based on a specific set of model parameters. A model will always produce an output. The reliability of that output however depends on the quality of the input. As the saying goes, "garbage in, garbage out". With this in mind, it is strongly recommended that emphasis be placed on generating and collecting good quality, and trustworthy data. Otherwise, groundwater model outputs will just be meaningless figures. If model input data is reliable, then a groundwater model can be a useful tool to investigate the reaction of groundwater storage volumes to changes in lake levels. Knowledge of how groundwater resources behave in response to lake level changes, and being able to quantify the volume of groundwater stored in the aquifer is necessary information for the creation of any water resource management plan for Lake Naivasha.

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PLATES

PLATE 1

Vertical profile through lake sediments at the Yacht Club. Sediments in the photo consist of very permeable, medium-grained sands with a pebble layer.

PLATE 2

Yacht Club at the time of field visit (Oct. '97). In the 3 months following the field visit, this area was submerged as the lake level rose.

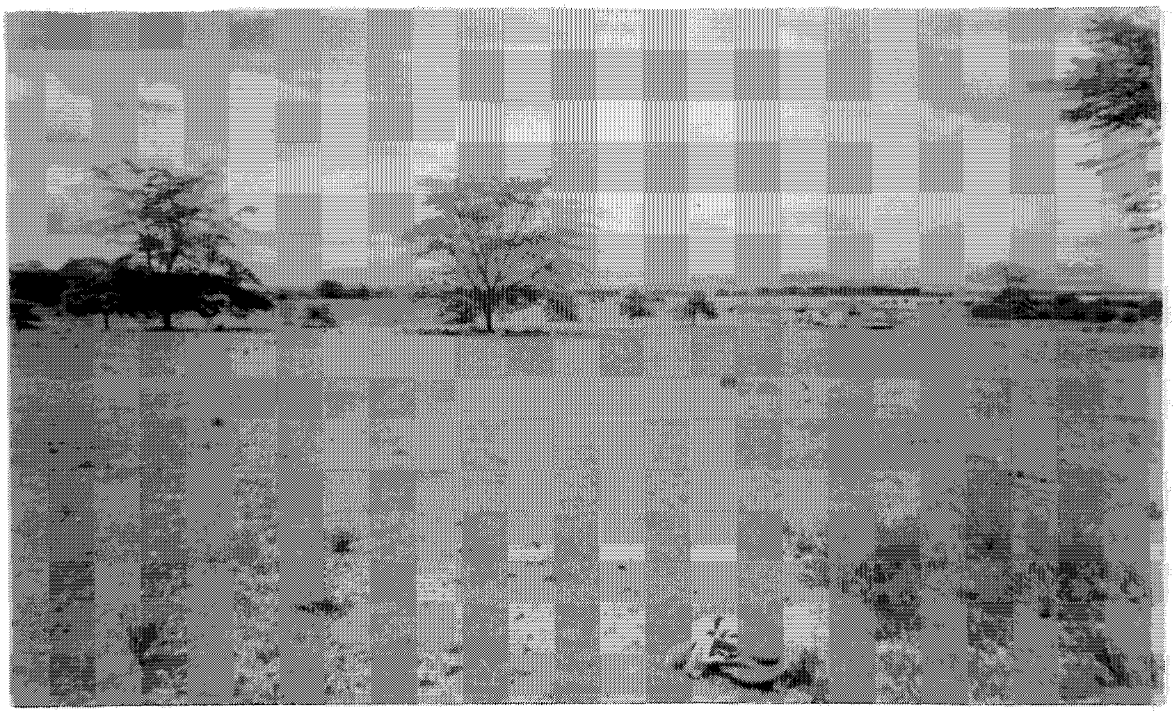


PLATE 3

The tree shown here grew in the 1940's when the lake level was low. When the level rose in the 1060's it was covered under water and died.Debris/sediments can still be seen in the tress, indicating the former lake level.

PLATE 4

Photo of a seepage meter which was installed in the lake during thefield visit, to measure groundwater in/outflows.

APPENDICES

APPENDIX 1 - About the Computer Disk

A computer disk is provided with this thesis. It contains logger data for the Yacht Club (**Yacht**), Manera (**Manera**) and Loldia (**Loldia**) loggers (section 2.6). Barometric pressure recorded by the reference logger is in the file **Reflogg**. Lake level fluctuation data for the period May, 1997 to February, 1998 is included in the file **Lak_lev**. Rainfall data (1985-93) for the stations used in creating the Theissen polygon map (section 3.1) is in file **Rainfall**. The data files are saved as text files and should be readable on most DOS or Windows-based personal computers.

APPENDIX 2 - Malewa River Discharge Data

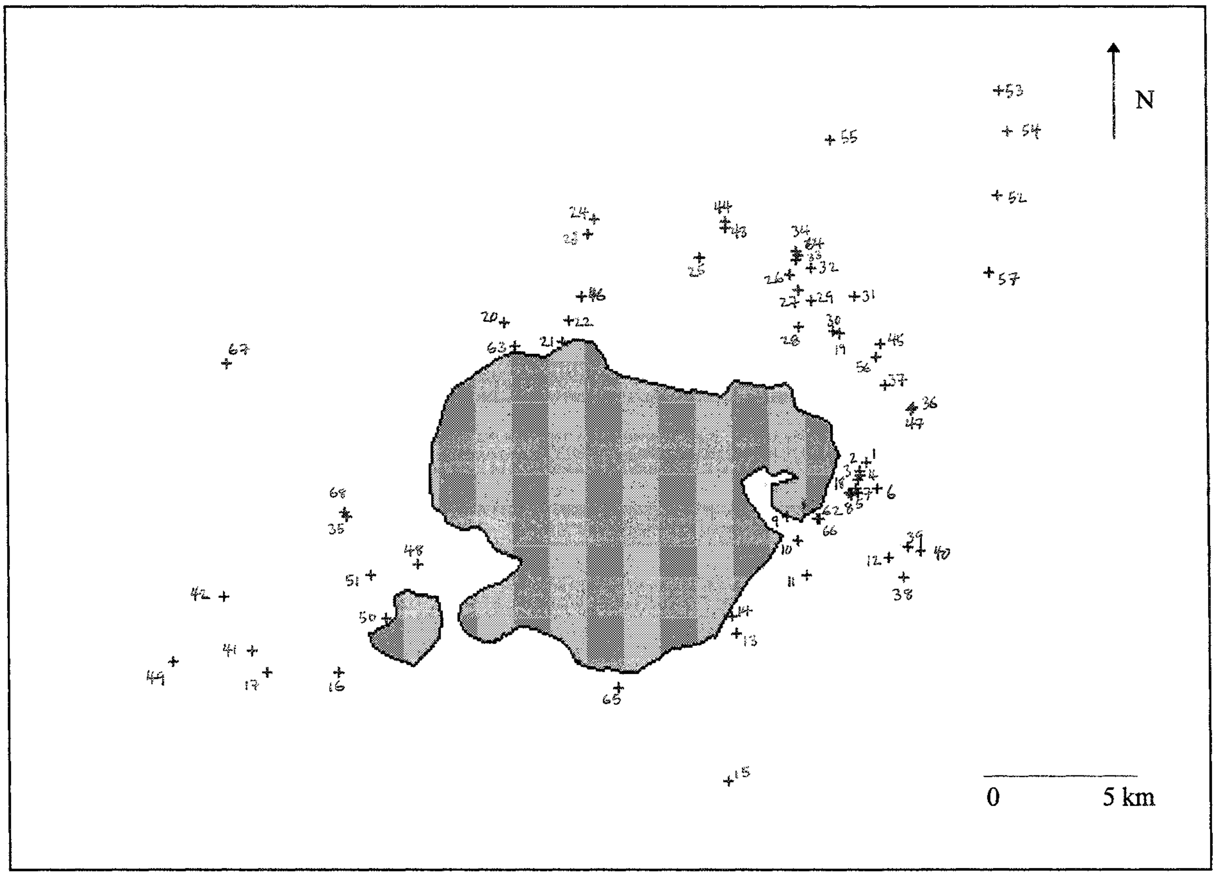
1960-1984 Discharge Measurements - Malewa River
Gauging Station - 2GB1
Catchment Area - 1588 km^2

Year	Total yearly discharge (cumecs)	Average daily discharge (cumecs)
1960	1094.1	3.0
1961	4423.6	12.1
1962	3878.1	10.6
1963	3753.6	10.3
1964	3766.8	10.3
1965	1015.0	2.8
1966	1976.4	5.4
1967	2421.3	6.6
1968	3844.1	10.5
1969	776.5	2.1
1970	2599.3	7.1
1971	2735.4	7.5
1972	1341.4	3.7
1973	1031.1	2.8
1974	2134.2	5.8
1975	2424.5	6.6
1976	1101.8	3.0
1977	2638.8	7.2
1978	3269.2	9.0
1979	2333.7	6.4
1980	1340.8	3.7
1981	2496.3	6.8
1982	1681.0	4.6
1983	2981.2	8.2
1984	759.3	2.1

Average yearly discharge = 2312.7 cumecs
Average daily discharge = 6.3 cumecs

APPENDIX 3 - Borehole Database

Xcoord	Ycoord	Well ID	Location name	Yield (cu. m/hr)	Well Depth (m)	EC (micro siemens/cm)	Rest water level (m)	Elevation (masl)	Groundwater head (m)
214349	9916714	1	-	-	10	-	1896	9.1	1886.9
214115	9916392	2	-	-	15.09	-	1892	13.5	1878.5
214083	9916180	3	-	-	25	-	1895	18.0	1877.0
214041	9916056	4	-	-	21	-	1900	12.0	1888.0
214042	9915576	5	-	2.67	45.9	2260	1907	43.5	1863.5
214789	9915670	6	-	2	65	1965	1921	37.7	1883.3
213804	9915486	7	-	1.8	12.3	1594	1902	11.6	1890.4
213684	9915525	8	-	1.3	13	1665	1897	12.0	1885.0
211300	9914600	9	-	3.4	18	494	1883	15.0	1868.0
211736	9913755	10	-	14	6	818	1884	4.8	1879.2
212053	9912421	11	-	2.3	18	1223	1899	12.7	1886.3
215200	9913100	12	-	15	100	-	1964	75.0	1889.0
209355	9910245	13	-	2.5	60	-	1902	9.0	1893.0
209208	9910831	14	-	35	32.4	-	1887	12.0	1875.0
209054	9904649	15	-	13.09	19.5	-	1994	12.9	1981.1
194187	9908739	16	-	6.52	74	1050	1931	32.0	1899.0
191487	9908739	17	-	5.93	79	1046	2050	32.0	2018.0
213971	9915708	18	-	2.4	15.46	2190	1904	14.4	1889.6
213394	9921504	19	-	72	-	1128	1892	14.5	1877.5
200538	9921954	20	-	36	61.5	664	1911	38.1	1872.9
202789	9921176	21	-	4.8	7.1	563	1883	6.1	1876.9
203036	9922058	22	-	72	11.33	681	1893	10.6	1882.4
203786	9925304	23	-	69.54	12	657	1894	11.1	1882.9
204040	9925879	24	-	62.6	10.4	720	1898	9.8	1888.2
208038	9924408	25	-	1.44	16.3	443	1894	14.3	1879.7
211455	9923721	26	-	80.1	70	582	1895	15.5	1879.5
211822	9923166	27	-	189	75.6	1128	1893	15.6	1877.4
211838	9921745	28	-	78	60	1461	1888	4.8	1883.2
212384	9922728	29	-	171	33	463	1893	10.8	1882.2
213126	9921647	30	-	72	42	645	1892	4.8	1887.2
213929	9922953	31	-	350	63	536	1907	24.0	1883.0
212346	9923999	32	-	148.5	33	810	1899	13.5	1885.5
211769	9924324	33	-	14.4	63	768	1900	20.4	1879.6
211769	9924682	34	-	121	42.4	1232	1902	22.6	1879.4
194535	9914610	35	-	4.2	27	1102	1905	22.6	1882.4
216194	9918712	36	-	8.5	180	765	1996	165.0	1831.0
215119	9919522	37	-	8.32	150	79	1951	76.5	1874.5
215784	9912357	38	-	36	-	1110	1986	96.0	1890.0
215884	9913478	39	-	10.29	130	1083	1971	81.0	1890.0
216441	9913361	40	-	8	138	605	1986	81.0	1905.0
190925	9909573	41	-	10	120	555	2007	98.0	1909.0
189794	9911609	42	-	11.54	120	388	1984	101.0	1883.0
209044	9925558	43	-	1.4	36	398	1899	19.4	1879.6
208994	9925762	44	-	4.5	42	408	1900	20.3	1879.7
214903	9921142	45	-	7.1	100	875	1931	40.0	1891.0
203526	9922907	46	-	50	12	854	1885	10.0	1875.0
216056	9918670	47	-	14.4	180	592	1993	168.0	1825.0
197205	9912864	48	-	-	-	829	1912	7.1	1904.9
187845	9909103	49	-	1.62	240	668	2100	210.0	1890.0
195957	9910799	50	-	5	50	839	1902	42.0	1860.0
195403	9912384	51	-	5	50	792	1914	42.0	1872.0
219411	9926765	52	-	6.5	190	489	2190	131.0	2059.0
219525	9930835	53	-	-	200	423	2438	137.0	2301.0
219848	9929261	54	-	18.2	180	393	2284	120.0	2164.0
213052	9928937	55	-	8.1	108	649	1948	60.0	1888.0
214733	9920622	56	-	7.3	90	714	1932	60.0	1872.0
219112	9923847	57	-	4.5	160	657	2122	130.0	1992.0
211115	9910901	58	-	-	-	-	1911	42.0	1869.0
212730	9914282	59	-	-	-	-	1889	7.5	1881.5
214780	9915467	60	-	-	-	-	1924	37.7	1886.3
211300	9924682	61	-	-	-	-	1900	-	-
212455	9914880	62	-	-	-	-	1883	-	-
200981	9921057	63	-	-	-	-	1890	-	-
211858	9924486	64	Manera	-	-	-	1901	-	-
204859	9908158	65	-	-	-	-	1896	-	-
212553	9914567	66	Yacht near gate	-	-	-	1886	5.3	1880.7
189990	9920342	67	Ndabibi	-	-	-	1993	104.9	1888.1
194448	9914779	68	Crater lake	-	-	-	1907	22.8	1884.2



Borehole Location map (corresponding to table)

APPENDIX 4 - Evapotranspiration Data

Method 1

------------ MONTHLY REFERENCE EVAPOTRANSPIRATION PENMAN MONTEITH ------------

Country: Kenya Meteostation: NAIVASHA
Altitude: 1900 m. Coordinates: 0.43 South 36.26 East

Month	MinTemp °C	MaxTemp °C	Humid %	Wind km/day	Sunshine Hours	Radiation MJ/m²/day	Eto mm/day
January	8.0	27.6	62	104	5.3	17.1	3.8
February	8.1	28.2	61	104	5.9	18.5	4.1
March	9.7	27.2	65	104	5.3	17.8	3.9
April	11.5	25.0	75	104	4.7	16.3	3.4
May	11.2	23.7	80	121	4.9	15.8	3.1
June	9.8	23.0	79	121	4.8	15.0	3.0
July	9.2	22.5	77	121	4.2	14.4	2.9
August	9.3	22.8	76	130	4.7	15.9	3.2
September	8.7	24.5	74	130	5.4	17.7	3.6
October	9.0	25.5	72	130	5.5	17.9	3.8
November	9.2	24.6	77	104	4.4	15.8	3.3
December	8.6	25.7	72	104	4.2	15.1	3.3

Source - FAO's CROPWAT

Method 2
Thornthwaite Method

If tn = average mothly temperature of the consecutive months of the year in °C (where n = 1,2,3...12) and j = mpnthly 'heat index', then:

$$\mathbf{j = (t_n/5)^{1.514}}$$

and the yearly 'heat index' **J** is given by:

$$J = \sum_{1}^{12} j$$

The potential evapotranspiration (PEx) for any month with average temperature t (°C) is then given by:

$$\mathbf{Pex = 16\ (10t/J)^a}$$

$a = (675 \times 10^{-9})J^3 - (771 \times 10^{-7})J^2 + (179 \times 10^{-4})J + 0.492$
A correction factor is applied monthly to values of PEx to correct for latitude.

	Jan °C	Feb °C	Mar °C	Apr °C	May °C	Jun °C	Jul °C	Aug °C	Sep °C	Oct °C	Nov °C	Dec °C
Avge yr.	17.6	17.7	18.3	18.0	17.1	16.0	15.5	15.8	16.2	17.2	16.8	17.0
Dry yr.	27.3	27.3	27.2	25.0	23.6	22.8	22.4	23.0	24.5	25.5	24.5	25.7
Wet yr.	7.9	8.1	9.4	11.0	10.6	9.2	8.6	8.6	7.9	8.9	9.1	8.3

Monthly potential evapotranspiration (mm).

	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec	Total
Avge yr.	71	64	74	69	65	56	55	57	58	67	64	67	768
Dry yr.	158	141	153	112	95	82	80	88	105	125	108	130	1377
Wet yr.	44	41	51	58	57	48	46	47	42	49	49	47	579

Method 3
Langbein-Turc Method

Potential evapotranspiration

$$\mathbf{PET = 0.9t^2 + 21t + 325}$$

t = average yearly temperature (°C)

	PET (mm)
Avge yr.	625
Dry yr.	925
Wet yr.	439