Growing lake with growing problems: integrated hydrogeological investigation on Lake Beseka, Ethiopia

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To my parents, my father Ayalew Belay and my mother Tsige Gebrehiwot, who have always been my source of inspiration.

ABSTRACT

The complex tectonic and volcanic processes in the Ethiopian rift valley have resulted in the formation of volcano-tectonic structural depressions that became sites for many rift valley lakes. Lake Beseka is one of the rift valley lakes in the northern section of the MER near to the Afar triangle. The lake plays an important role in the ecology of birds and wildlife, as it is located in the northern part of the Awash National Park. Lake Beseka has been expanding at an astounding rate since the late 1960s and early 1970s. This growth has had a detrimental effect on the surrounding physical, hydrological and infra-structural environment. This study is conducted with primary objective of understanding the hydraulic interaction of Lake Beseka with the surrounding groundwater system, and identifying and quantifying the role of groundwater in the hydrology of the lake. Integrated approaches of hydrochemistry, isotope hydrology, recharge estimation (water table fluctuation, EARTH modeling and chloride mass balance), and groundwater modeling (MODFLOW) are applied for hydrogeological characterization of the study area. The flow system in the watershed of Lake Beseka is analyzed by coupling a groundwater level map with the hydrochemical and isotopic composition of the water bodies. Hierarchical multi-element cluster analysis (HCA) is used to classify hydrochemical water samples of the study area into different groups. The lake is characterized by a Na-HCO_3-Cl type of water, and has a hydrochemical signature similar to that of the groundwater system flowing from the western part of the watershed. The groundwater system in the western part of the watershed and hot springs that emerge at the western edge of the lake are characterized by a Na-HCO_3 type of water with an average isotopic composition of -2.8 ‰ in $\delta^{18}O$ and -10.7 ‰ in δ^2H, which is comparable to the isotopic concentration of the input signal. The input signal has an isotopic value of -3 ‰ in $\delta^{18}O$ and -9.3 ‰ in δ^2H, and is derived from the intersection point of the local meteoric water line of Addis Ababa rainfall and the local evaporation line, which plots along evaporated waters of Lake Beseka. Hydrochemical and isotopic evidence indicates that groundwater flows from the western part of the watershed and discharges to the lake in the form of hot springs. This groundwater inflow constitutes the major water inflow to Lake Beseka and forms an integral part of its water budget. This fact is well supported by the groundwater modeling results, which estimate that 51% of the total water inflow to Lake Beseka comes from groundwater seepage to the lake. This seepage is computed by the model to be 33.8Mm^3 annually. It is evident from the recharge estimation that recharge of the groundwater system comes from infiltration of local precipitation in the watershed. However, this recharge is estimated to be only 17.4 Mm^3 annually. Thus, the groundwater recharge that seasonally replenishes the aquifer system within the lake watershed is not significant enough to explain the expansion of Lake Beseka. The model estimated that 30.45Mm^3 of groundwater laterally flows to the aquifer system of the lake watershed annually across the surface boundaries. This lateral inflow might be related to recharge from the mountains. It is most likely that the expansion of Lake Beseka is related to changes in the amount of this groundwater, which first flows to the lake watershed and then to Lake Beseka. Owing to the geologic setting of the study area, tectonically induced modification of hydraulic gradient of the groundwater regime in the region might have resulted in an increase in the discharge of the hot springs that continuously flow to the lake.

KURZFASSUNG

Die komplexen tektonischen und vulkanischen Prozesse im äthiopischen Rift Valley haben zur Bildung von vulkanisch-tektonischen, strukturellen Vertiefungen geführt, in denen sich viele Seen gebildet haben. Der See Beseka ist einer dieser Seen im nördlichen Bereich der ostafrikanischen Riftzone in der Nähe des Afar Dreiecks. Der See spielt eine wichtige Rolle in der Ökologie von wildlebenden Tieren, da er im nördlichen Teil des Awash Nationalparks liegt. Er hat sich seit den späten 1960er und frühen 1970er Jahre mit einer sehr großen Geschwindigkeit ausgedehnt. Diese Ausdehnung wirkt sich nachteilig auf die physikalische, hydrologische und infrastrukturelle Umgebung aus. Diese Studie hat zum wichtigsten Ziel, Erkenntnisse über die hydraulischen Interaktionen vom See mit dem umgebenden Grundwassersystem zu gewinnen sowie die Rolle des Grundwassers in der Hydrologie des Sees zu bestimmen und zu quantifizieren. Integrierte Ansätze der Hydrochemie, Isotopenhydrologie, Bestimmung der Grundwasserneubildung (Grundwasserspiegel-Fluktuationsmethode, EARTH-Modellierung und Chloridmassenbilanz-Methode), sowie Grundwassermodellierung (MODFLOW) werden für die hydrogeologische Charakterisierung des Untersuchungsgebietes eingesetzt. Das Fließsystem im Einzugsgebiet vom See Beseka wird durch die Verbindung der Grundwassersfließrichtung mit der hydrochemischen und isotopischen Zusammensetzung der Wasserkörper analysiert. Mit der hierarchischen Clusteranalyse (HCA) werden die hydrochemischen Wasserproben im Untersuchungsgebiet in verschiedene Gruppen klassifiziert. Der See ist durch einen Na-HCO_3-Cl-Wassertyp charakterisiert und zeigt eine hydrochemische Signatur, die der Signatur des Grundwassersystems, das aus dem westlichen Teil des Einzugsgebiets stammt, ähnlich ist. Das Grundwassersystem im westlichen Teil des Einzugsgebiets sowie heiße Quellen, die am westlichen Rand des Sees entspringen, ist als Na-HCO_3-Wassertyp mit einer durchschnittlichen isotopischen Zusammensetzung von -2.8 ‰ in $\delta^{18}O$ bzw. -10.7 ‰ in $\delta^{2}H$ charakterisiert. Diese Werte sind vergleichbar mit der isotopischen Konzentration des eingehenden Signals. Das Eingangssignal hat einen isotopischen Wert von -3 ‰ in $\delta^{18}O$ bzw. -9.3 ‰ in $\delta^{2}H$ und entsteht aus dem Schnittpunkt der lokalen meteorischen Wasserlinie des Addis Ababa Niederschlags und der örtlichen Evaporationsline, die vom verdunsteten Wassers des Sees Beseka beeinflusst wird. Hydrochemische und isotopische Daten deuten daraufhin, dass das Grundwasser aus dem westlichen Teil des Einzugsgebiets zufließt und in Form von heißen Quellen austritt. Dieser Grundwasserzufluss bildet den Hauptzufluss zum See Beseka und somit einen wichtigen Bestandteil des Wasserhaushalts. Diese Tatsache wird durch die Grundwassermodellierung gestützt, die zeigt, dass 51% des gesamten Wasserzuflusses zum See aus dem Grundwasser stammt. Das Modell berechnete eine Versickerungsmenge von 33.8 Mm^3 pro Jahr. Die Ergebnisse der Modellierung zeigen, dass die Grundwasserneubildung durch die Infil-tration des örtlichen Niederschlags im Wassereinzugsgebiet stattfindet. Diese Neubildung beträgt jedoch nur 17.4 Mm^3 pro Jahr. Daher ist die Grundwasserneubildung, die jahreszeitlich zur Grundwasseranreicherung im Einzugsgebiet des Sees führt, nicht ausreichend genug, um die Ausdehnung des Sees zu erklären. Das Modell berechnete, dass 30.45 Mm^3 des Grundwassers jährlich seitlich in das Grundwassersystem fließen. Dieser laterale Zufluss könnte im Zusammenhang mit der Grundwasserneubildung in den Bergen

stehen. Es ist sehr wahrscheinlich, dass die Ausdehnung des Sees durch eine Erhöhung dieser lateralen Grundwasserströmung verursacht wird, die zunächst in das Einzugsgebiet und dann in den See fließt. Aufgrund der geologischen Lage des Untersuchungsgebietes könnte eine Veränderung des hydraulischen Gradienten des Grundwassersystems in der Region durch tektonische Aktivitäten zu einer Zunahme der Wassermengen, die aus den heißen Quellen ständig in den See hinein fließen, geführt haben.

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LIST OF ABBREVIATIONS

EARs	East African Rift system
FAO	Food and Agriculture Organization of the United Nations
GNIP	Global Network for Isotopes in Precipitation
GMWL	Global Meteoric Water Line
IAEA	International Atomic Energy Agency
LMWL	Local Meteoric Water Line
LEL	Local Evaporation Line
NWP	The North Western Ethiopian Plateau
UNEP	United Nation Environmental Program
UNESCO	United Nation Educational Scientific and Cultural Organization
US $	Dollar of United States of America
VSMOW	Vienna Standard Mean Oceanic Sea Water
WHO	World Health Organization
WWDSE	Water Works Design and Supervision Enterprise
YTVL	Yerer – Tullu Wellel Volcano Tectonic Lineament
ZEF	Center for Development Research

1 INTRODUCTION

1.1 Background information

The Great East African rift system is one of the most extensive rift systems on earth. It is an impressive geological feature extending from the Jordan River in Middle East southward to Mozambique in South East Africa with some 6400 km length and 48 to 64 km width. The Ethiopian rift valley is part of the East African rift system, which extends from the Kenyan border up to the Red Sea. The Ethiopian Rift provides a unique opportunity to study the transition between continental rifting to the south in Kenya and new seafloor spreading to the north in Afar in the onshore extension of the Red Sea and Gulf of Aden (Mackenzie et al., 2005). The seismically and volcanically active northern Main Ethiopian rift (MER) and Afar rifts are virtually the only places worldwide where the transition between continental and oceanic rifting is exposed on land (Keir et al., 2006; Maguire et al, 2006). The Rift Valley is the result of spreading, or rifting, between tectonic plates, which, if it continues, may ultimately transform the Horn of Africa into an island in the Indian Ocean (Kious and Tilling 1996; as cited in UNEP, 2008).

Intra-rift faulting and associated volcanic activities resulted in the formation of volcano-tectonic structural depressions, which became sites for many rift valley lakes. The complex geological processes associated with the rift valley are responsible for the creation of several of East Africa’s largest lakes as well as much of its topography (UNEP, 2008). Ethiopian rift valley lakes are located extending from Lake Rudolf in the south at the Ethio-Kenyan border to the north of Afar at the Djibouti and Eritrea border. The Main Ethiopian rift (MER) contains lakes of different hydrological and morphometric characteristics, which are mainly structurally controlled and fill NNE trending tectonic depressions (Le Turdu et al., 1999).

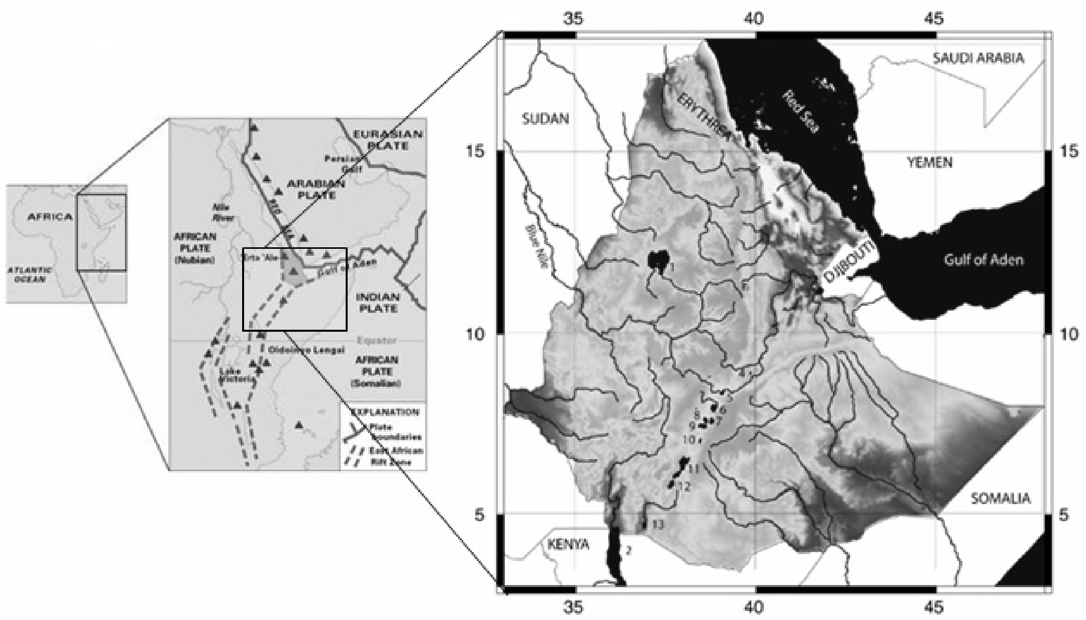


Figure 1.1: Map of East African Rift System (left) and Ethiopia with main lakes, where, l: Tana, 2: Turkana, 3: Abbe, 4: Beseka, 5: Koka, 6: Ziway, 7: Langano, 8: Abiyata, 9: Shala, 10: Awassa, 11: Abaya, 12: Chemo, 13: Chew Bahir (right). (Source: http://en.wikipedia. org/wiki/ Great_ Rift Valley- (left) and Görner et al., (2006) (right)).

The lakes in the rift valley are situated within three basins: Awash basin (lakes Koka, Beseka, Gemari, Abe), which is located in the northern Main Ethiopian Rift, the lakes region (lakes Ziway, Langano, Abiyata and Shalla) occupying a central part of the MER, and the southern basin (lakes Awassa Abaya, Chemo and Chewbahir) in the southern section of the MER. Hydrologically, basins form separate units, but hydrogeologically they may form a unique system within the rift due to the underground interconnection by NE–SW aligned regional faults (Alemayehu et al., 2005).

Lake Beseka is located in the northern part of the MER at the near distance to the Afar triangle, which is a triple junction where three plates (Arabian, Nubian and Somalian) are pulling away from each other along the EARs to form new oceanic crust (Figure 1.1). Lake Beseka is volcanically dammed, endorheic lake, which is situated 190 km from the capital Addis Ababa. It is habitat of a variety of birds, fishes, crocodiles and other aquatic species. The lake plays an important role in the wildlife ecology, as it is located in the northern part of Awash National Park.

Lake Beseka has been expanding at an astounding rate since the late 1960's and early 1970's. The area has increased from 3 km^2 to 40 km^2 within three and half decades (Figure 1.2). The average annual increment of the lake was 0.2 m and the level of the lake has rose by 4 m from 1976-1997 (Zemedeagegneh and Egizabher, 2004). The situation would not be as such looming if the water could be used as drinking or irrigation water, but unfortunately the lake water is saline and alkaline. The drastic expansion of the lake has led to many problems in the surrounding area, and is a severe threat to the wellbeing of the indigenous people and the economic welfare of the nation in general.

1.2 Relevance of the study

The Ethiopian rift valley lakes are characterized by different lake level fluctuation patterns. Some of the lake levels show a declining trend, a few of them have rising trends, while others are still in a stable condition. Lake Abiyata has shown the most drastic level reduction in the MER, while a slight decline is evident in Lake Ziway (Ayenew, 1998). Non-terminal lakes such as Lake Langano and Abaya are characterized by a stable lake level. The lakes Beseka, Awassa and Chamo have been increasing in size over the last few decades; Lake Awassa and Chemo have been growing since the 1980s, while Lake Beseka has been expanding drastically since the late 1960s (Ayenew, 2004).

The expansion of Lake Beseka is alarming and has had detrimental effect on the surrounding biological, physical, hydrological and infra-structural environment. The lake growth has affected the highway and railway structures which run along its northern shore (Figure 1.3), and lake has flooded the highway, the only import-export line to the port of Djibouti, several times. As Ethiopia became a landlocked country in 1993, it entirely depends on this route for importing and exporting commodities. In addition to being Ethiopia's sole access to the port, the highway is the main route that links the central, western, and southern part of the country to the eastern part. The problem has been temporarily overcome by elevating the access through construction of 2-km long and 2-m high embankment (Tessema, 1998). The government had to spend US $1.5 Million to raise the submerged part of the railway and highway (WWDSE, 1999).

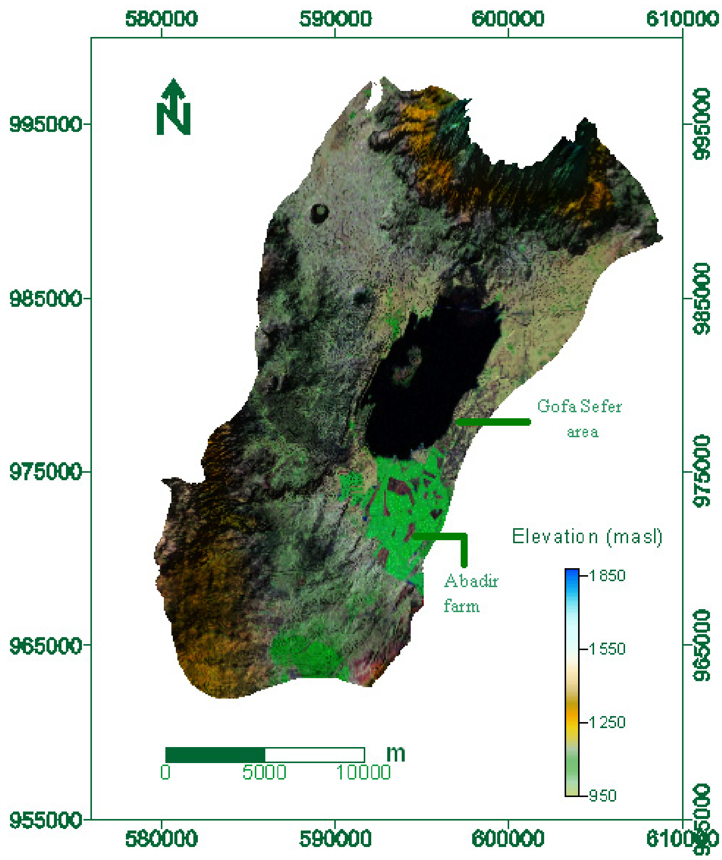


Figure 1.2: Boundaries of Lake Beseka in 2000 and its watershed with Gofa Sefer village and Abadir farm area (DEM of the lake watershed overlaid with LANDSAT imagery).

The lake advancement caused the loss of 57 human lives, and it has already inundated about 35 km^2 of the nearby grazing land and displaced 910 people (WWDSE, 1999). The nearby Awash National Park has been affected indirectly, as displaced nomads encroached into the park, looking for a settlement area, water and grazing land. This is one of the many reasons why the park area has reduced from an original 750 km^2 to the current size of 250 km^2. The flooding of grazing land has led to hardships for local nomadic pastoralists, and caused many camel and cattle herds to move to the national park land, to detriment of the game (Halcrow, 1978).

The expanding lake is in less than 3 km from the River Awash, which is the source of drinking water and irrigation for millions of people downstream. If the lake continues to expand at current rate and other influencing factors remain the same, the lake will cross the natural water divide and invade the town of Addis Ketema and join the River Awash. This would be disastrous, as the quality of the river water will be deteriorated such that agricultural development downstream (such as in Amibara) would be at risk.

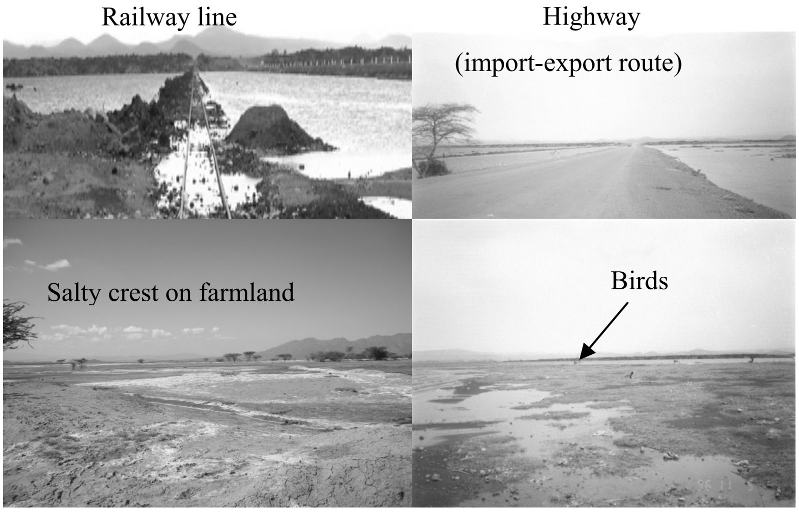


Figure1.3: Effects of Lake Beseka expansion on infrastructures and ecosystems

The growth of Lake Beseka is a natural disaster for the surrounding ecosystems, as the expansion has created an unstable transitional zone between the wetland and nearby terrestrial ecosystem. The Methara sugar plantation has been inundated and lost income from 161.55 ha of land (WWDSE, 1999). The sugar plantation in Abadir suffers from salt-water encroachment and has lost more than 30 ha of farmland (Tessema, 1998) (Figure 1.3).

Many recently conducted studies in the area concentrate on the geological evolution of the northern part of the MER, with special attention to the mechanism of rifting or crustal thinning at the so called magmatic segments. Only a few hydrological studies were carried out on Lake Beseka by Halcrow (1978), Tessema (1998) and WWDSE (1999), with the main objective of identifying the cause of lake expansion and assessing the resulting damage. These studies provide a preliminary overview of the hydrology of Lake Beseka and state that irrigation activity at Abadir farm is the main cause of the lake growth. On the other hand, recent work of Görner et al. (2006) suggested that neo-tectonic activities modified the structure of the basin and triggered the growth of Lake Beseka. However, none of these studies adequately address important groundwater issues such as groundwater recharge, groundwater-surface water interactions and water budget of the lake and its watershed.

Considering the fact that up to now only limited research has been performed by the Ministry of Water Resources and others, a wide knowledge gap exists regarding the importance of groundwater flow to the Lake Beseka and its expansion. This study is conducted with the primary objective of understanding the role of groundwater for the hydrologic system of the lake watershed in general and to the water budget of Lake Beseka in particular. This study applies an integrated approach of hydrochemistry, isotope hydrology, recharge estimation and groundwater modeling for hydrogeological characterization of the lake watershed (Figure1.4). It can bridge the existing gap in the lack of information on quality and quantity of groundwater in the area and illustrate the significance of groundwater for the hydrology of Lake Beseka. The results could also support the development of efficient and alternative lake management plans to monitor Lake Beseka's growth in environmentally acceptable and financially plausible ways.

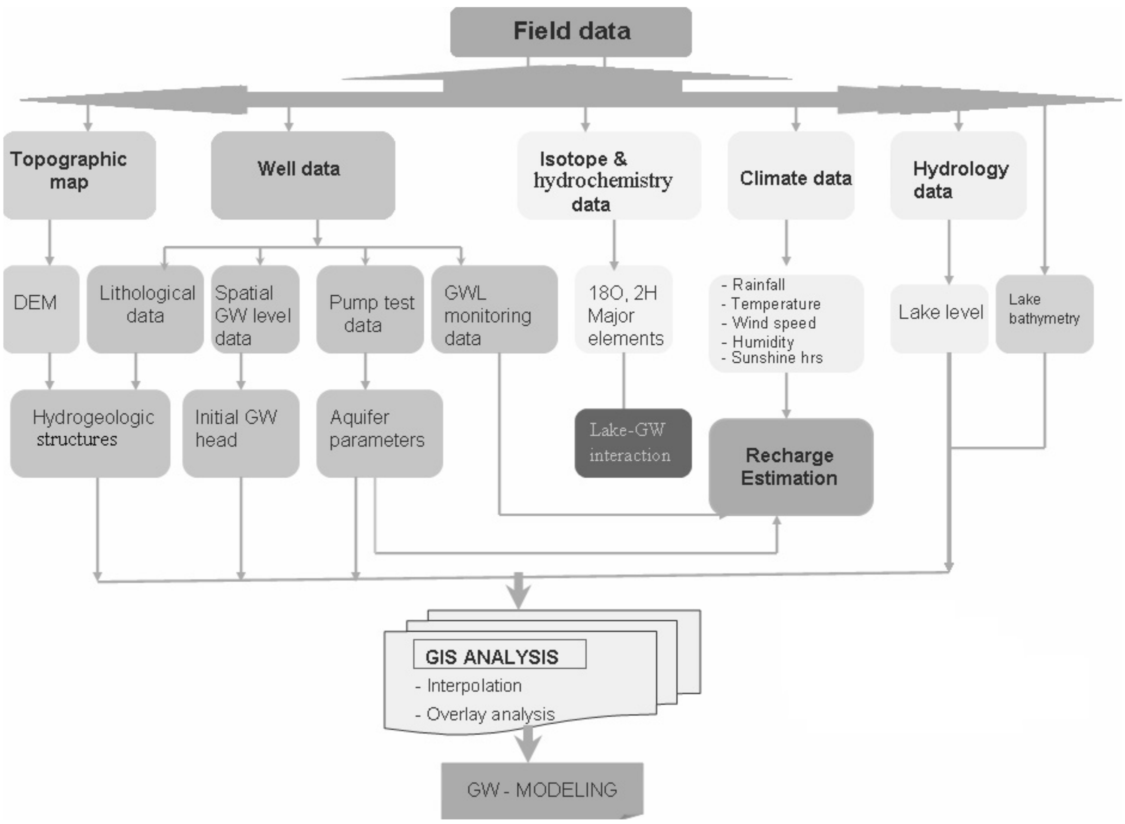


Figure1.4: Flow diagram of the methodology and materials used in this study

1.3 Research objectives

The main research objective is: Detailed hydrogeological characterization of the watershed of Lake Beseka and understanding the hydrology of the lake by identifying and quantifying the role of groundwater in the area.

Specific objectives of the research are:

- To estimate the rate and direction of groundwater flows in the lake watershed,
- To determine the hydraulic connection between the lake and the surrounding groundwater flow system and show the importance of groundwater to the lake hydrology,
- To estimate groundwater recharge of the area,
- To simulate groundwater fluxes from / to the lake and to estimate water budget of the lake.

1.4 Structure of the thesis

The thesis is organized in eight chapters. After the introduction to the general problem and research objectives (Chapter 1), Chapter 2 presents a general description of the study area. Chapter 3 deals with description of the hydrogeological framework of the lake watershed. Hydrochemical and isotopic characteristics of water bodies in the study area are discussed in Chapter 4 and Chapter 5, respectively. Chapter 6 is devoted to estimation of the groundwater recharge using multiple recharge estimation techniques. Numerical groundwater flow modeling of the watershed of Lake Beseka is presented in Chapter 7. The final chapter 8 gives a synthesis of this work. It summarizes the research by converging outcomes from the different approaches used in the study.

2 DESCRIPTION OF THE STUDY AREA

2.1 Location and climate

This study is conducted on Lake Beseka, one of Ethiopian rift valley lakes located in the Horn of Africa. The lake is situated at the center of the Ethiopian rift valley about 190 km east of the capital Addis Ababa. The lake watershed lies between 39°43'-39°59' east longitude and 8°41'- 9°0' north latitude. The surface area of the lake has grown dramatically within the last three and half decades and now covers about 8 % of the total watershed area. The watershed is 505 km^2 in size, and the lake is located at the eastern part of the basin. Two towns, Methara and Addis Ketema are located close to the lake (Figure 2.1).

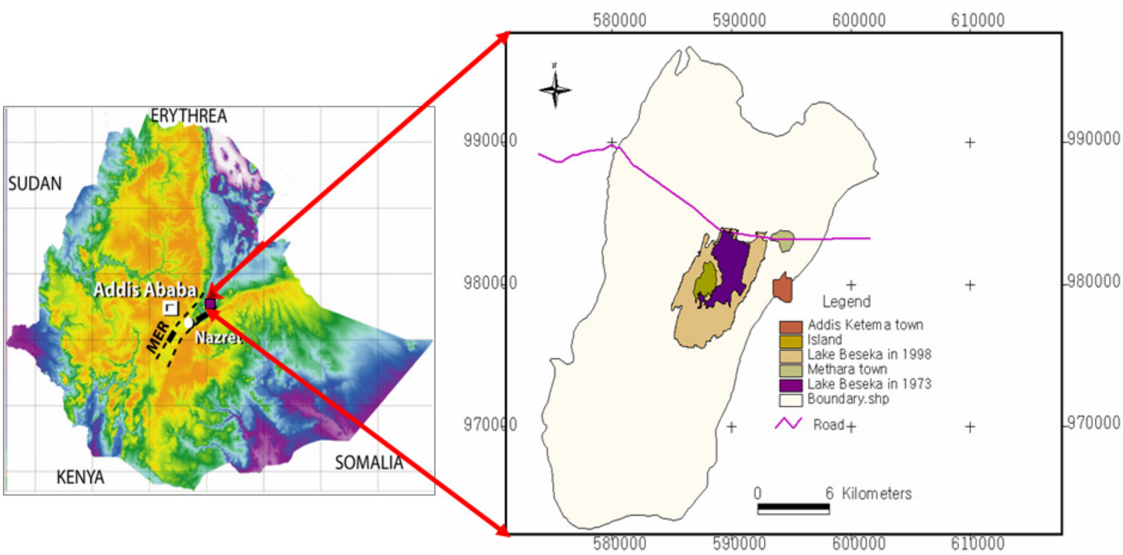


Figure 2.1: Location of the investigated watershed with two nearby towns, Methara and Addis Ketema, and the extension of the lake in 1973 (violet) and 1998 (orange), (Map of Ethiopia is adapted from Görner et al. (2006)).

The Ethiopian climate system is under the influence of the Indian and Atlantic Ocean monsoons (Gemechu, 1977). The seasonal migration of the Inter Tropical Convergence Zone (ITCZ) controls rainfall patterns in most parts of the country. During the summer, the ITCZ is located in northern Ethiopia, and the region is under the influence of moisture from the Atlantic and Indian Ocean (Kebede, 2004). During the months of October to February, the ITCZ migrates to the south of Ethiopia, and most of the country is characterized by dry air. The northward movement of the ITCZ from March to April brings moisture from the Indian Ocean, which results in small spring rain. Rainfall in the western-most sector of Ethiopia is mainly in summer (July - September), while in the eastern part of the country it is in spring (March – May) and in October. The central part of the country is characterized by a bimodal rainfall pattern, with both summer and spring rainfall (Figure 2.2). The study area is located in the center of Ethiopia, is influenced by two moisture sources, and characterized by two rainy seasons in summer (July - September) and spring (March - April). The climate is semi-arid with a mean annual temperature of 25°C and a total mean annual rainfall of 534 mm.

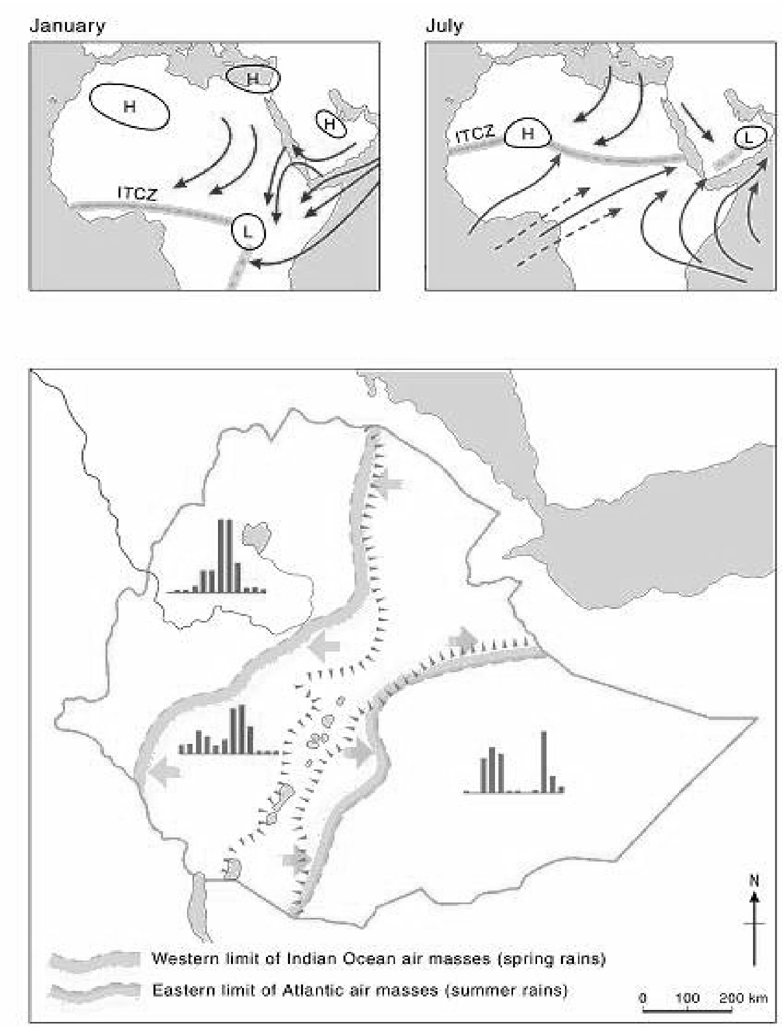


Figure 2.2: Seasonal drifting of the ITCZ in Ethiopia (adapted from Lacaux et al, 1992).

Long-term (1966-2007) mean monthly records of temperature, relative humidity, wind speed and sunshine hours were analyzed (Figure 2.3). The study area is characterized by an average daily maximum and minimum temperature of 33°C and 17°C, respectively. The lowest temperatures are between November and January, while May and June are characterized by higher temperatures. Relative humidity is high during the main rainy season. Wind speed is at a maximum during the months June and July, and at a minimum value during October. Long- term average sunshine hours in the study area are 8.4 hours, and the lowest sunshine hours being during the main rainy season. Potential evapotranspiration exceeds monthly rainfall in nearly all months; thus the climate of the area can be defined as semi-arid according to the Thornthwaite climate classification scheme (FAO, 1984).

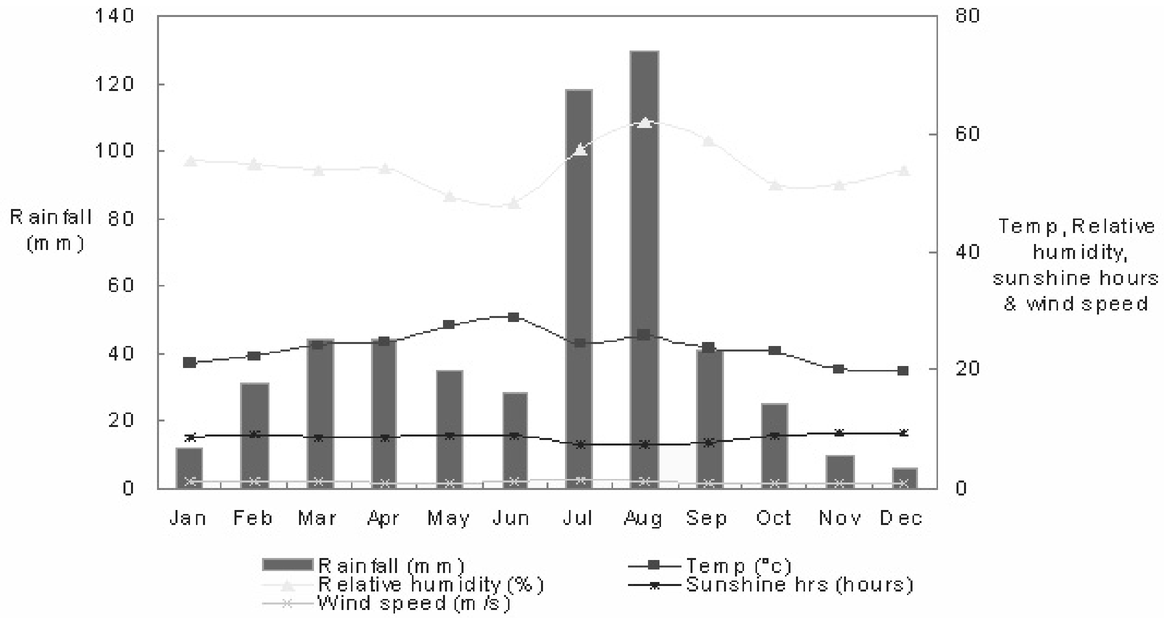


Figure 2.3: Long-term (1966-2007) mean monthly records of meteorological parameters at Methara (data from Research Center of Methara Sugar Estate and National Meteorological Service Agency).

2.2 Topography

The topography of the Lake Beseka area ranges from flat to undulating plains, and from hills to the high Mount Fentale. Most of the watershed is characterized by flat to undulating plains with altitudes ranging from 940 to 1100 m.a.s.l. Plains with small and high gradient are located in the northwestern and western part of the watershed. Volcanic cones are also concentrated in the western part. Hills with medium slopes are located in the south, while a high slope mountain is situated in the northern part of the watershed. The most fascinating and outstanding Quaternary Fentale volcano is located north of Lake Beseka, close to the southwestern corner of the Afar Depression. The Fentale volcano rises 1000 m above the surrounding plain and reaches to 2007 m at the center of the volcano. The last volcano erupted in the 1810s, and today the volcano is characterized by emanating volcanic gasses and is believed to be an active volcano.

2.2.1 Digital Elevation Model (DEM)

A digital elevation model (DEM) is a digital representation of elevation of the terrain over a given area, usually at a fixed grid interval of the earth. The knowledge of surface relief is of a major importance in understanding and evaluating different earth processes. A DEM provides an opportunity to model, analyze and display phenomena related to topography. It is essential in any discipline concerned with process modeling like hydrology, climatology, geomorphology and ecology (Sulebak, 2000). Relief governs the direction and intensity of water movement in watersheds, thus a DEM provides useful topographic parameters for a wide range of hydrological applications.

The DEM of the study area was prepared with a resolution of 10 m through digitizing topographic maps (scale 1:50000). The maps were scanned, and contours and spot heights were digitized and tagged with elevation values in a GIS (geographic information system) environment. The DEM was created in ILWIS first by digitizing contour maps, and then by calculating values through linear interpolation for pixels that were not covered by contours. Contour interpolation in ILWIS, based on the Borgefors distance transform, was used to create the DEM (Figure 2.4).

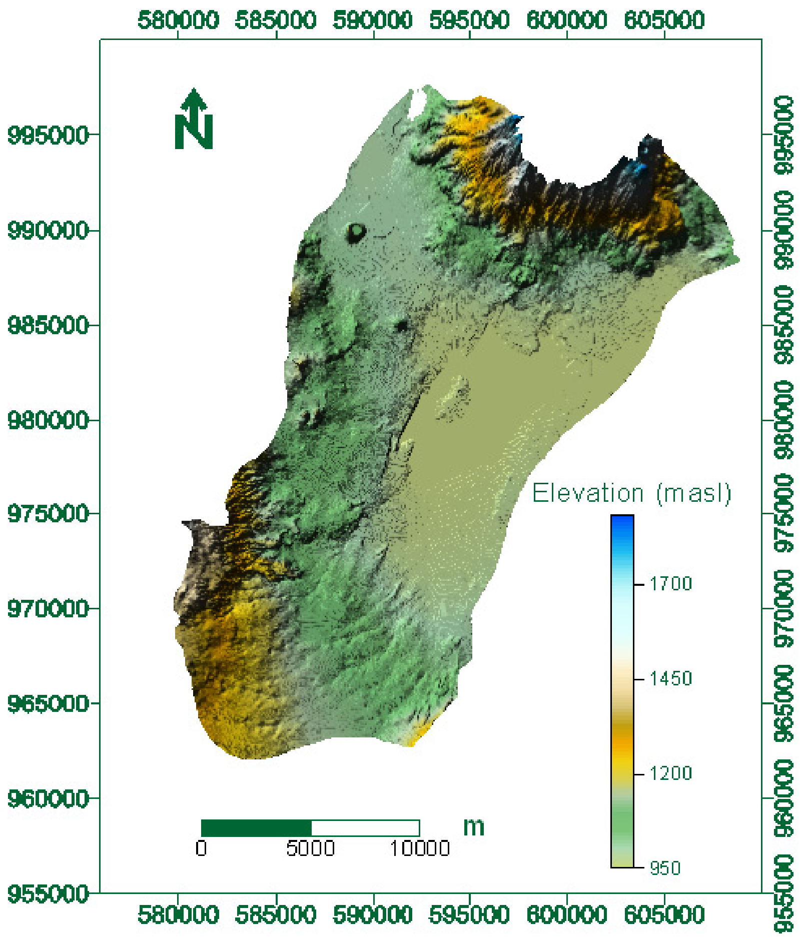


Figure 2.4: Digital elevation model of Lake Beseka watershed (prepared from topographic map of 1:50000 scale).

2.3 Soils

A soil map of the Lake Beseka area exists at the scale of 1:50000 (WWDSE, 1999), and seven major soil units can be identified (Table 2.1). Leptosol (PLLp) is the dominant soil type covering about 33 % of the total watershed area. This soil unit is characterized by shallow soil with weakly developed structures, coarse texture, and is covered mainly by open bushy woodland. Soil developed on mountains (TMLpro) covers about 17 % of the watershed area, commonly found around Mount Fentale and in the western part (Figure 2.5). This soil type is excessively drained, shallow and coarse textured and covered mainly by open and dense bushy woodland.

Table 2.1: Main soil types in Lake Beseka watershed (data from WWDSE, 1999)

Soil class	Soil type	Texture	% by area
PLLp	Lepthsols	Sandy clay loam-sandy loam	33.4
TMLpro		Sandy clay loam-sandy loam	16.9
PLCm	Cambisols	Sandy clay loam	12.6
PLPz	Podzoluvisols	Loam to sandy loam	6.1
CHPLp		Sandy clay loam-sandy loam	4.9
PLLv	Luvisols	Clay loam to clay	4.3
PLFi	Fluvisols	Clay loam to sandy clay	3.6

The cambisol (PLCm) is a well drained, deep, and medium to coarse textured soil type, which is mainly located west of Lake Beseka and in the northeastern part of the watershed. It is mainly covered by open grassland and open bushy land. Podzoluvisol (PLPz) is a well drained soil, medium to course textured and very deep, with moderately developed structures. It is mainly found in the southeastern part and covers about 6 % of the total area. The solonchak soil typically developed along the eastern edge of Lake Beseka. It is poorly drained, deep, weak to moderately developed and characterized by salt crests at the surface, which can be related to capillary rise of groundwater.

The CHPLp soil class is found on the hilly plains of the western watershed and covers about 5% of the total watershed area. This soil is excessively drained, very shallow and coarse textured with coarse fragments and stones. Luvisol (PLLv) is well drained, very deep, moderately developed, fine textured and characterized by none to very few coarse fragments. This soil type contains accumulated clay in the B-horizon and is the dominant soil type in the Abadir sugarcane farm. Fluvisols (PLFi) developed on alluvial deposits, which are found mainly in the northwestern corner of the watershed and cover about 4 % of the study area. They are well drained, very deep, weakly developed and characterized by a wide range of textures with few to frequent coarse fragments.

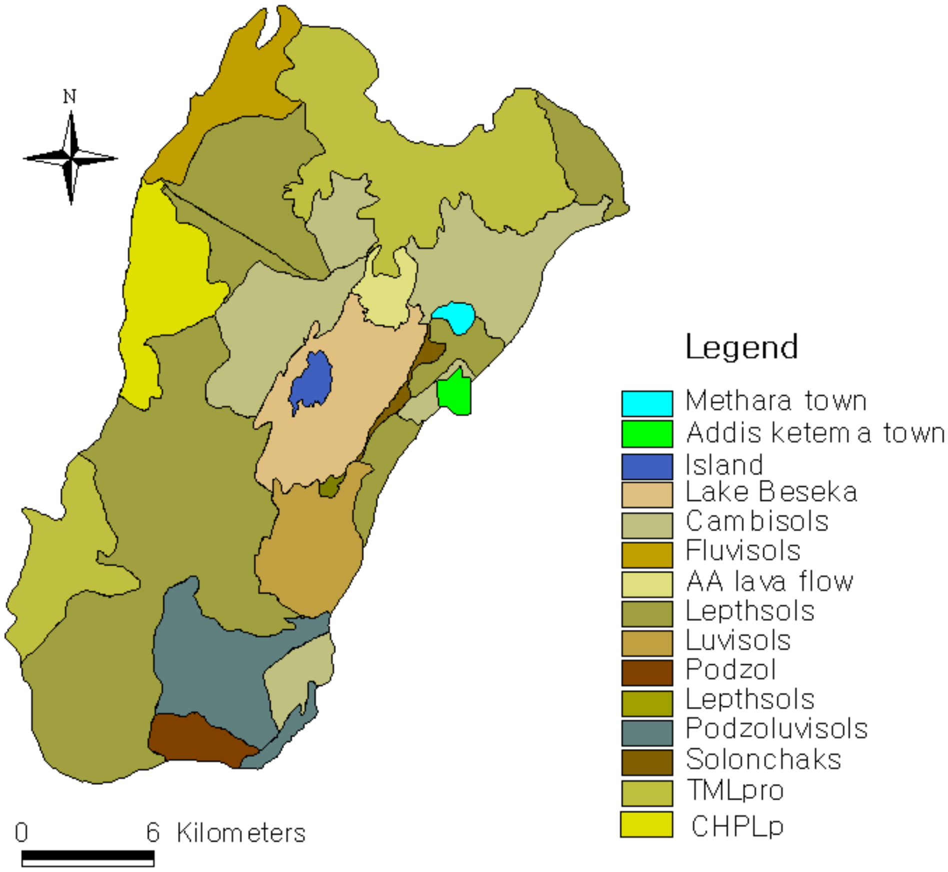


Figure 2.5: Soil types in the lake watershed (data from WWDSE, 1999).

2.4 Land-use and land-cover (LUC) types

The land-use and land-cover (LUC) types of Lake Beseka watershed were mapped in 1999 by WWDSE based on aerial photos and field investigations, and 14 LUC units were identified (Table 2.2). Open bushy woodland is the most dominant LUC type, accounting for 46 % of the total watershed area. This LUC unit is typically found covering the shallow soil of the plain, medium to high gradient hills, volcanic cones and Mount Fentale. This unit is mainly used for grazing, firewood collection, and charcoal production, and for very limited rainfed agriculture (Figure 2.6).

Dense woodland, open grassland and open grassland with bare rock are common LUC types covering 24 % of the total lake watershed area. Dense woodland is commonly found in the northwestern part and covers 8.7 % of the total lake watershed area. Open grassland is mainly found on the eastern side of the lake, and accounts for

8.5 % of the total watershed area. Both LUC units are used for grazing and browsing, and dense woodland is additionally used for firewood production. The northern margin of the lake is bare land consisting of fresh, aa lava which is believed to be the product of the 1820's volcanic eruption in the area (Halcrow, 1978).

Table 2.2: Main LUC types of the lake watershed (data from WWDSE, 1999)

Land-cover	% by area	Land-use
Open bushy woodland	46.0	Grazing, firewood, charcoal production
Dense woodland	8.7	Grazing
Open grassland	8.5	Grazing
Water	7.3	-
Open grassland with bare rock	7.0	Grazing
Mixed	5.1	Grazing, wood harvesting
Dense bushy woodland		Grazing, firewood, charcoal production

The indigenous people in the study area are nomadic pastoralists and depend mainly on cattle and camel rearing. Cultivation is not a common practice in the area, though scattered and insignificant patches of land are used by semi-nomads for rainfed cultivation of sorghum and maize. In the lake watershed, cultivation is mainly confined to the state owned farms in Methara Sugar State (south of the lake) and Nura Era citrus farm (south end of the watershed). The Abadir farm is part of the Methara Sugar Plantation and lies directly south of Lake Beseka. Irrigation farming has been practiced in this area since 1968, however crops changed from cotton and citrus fruits to sugar cane production in 1978 (WWDSE, 1999).

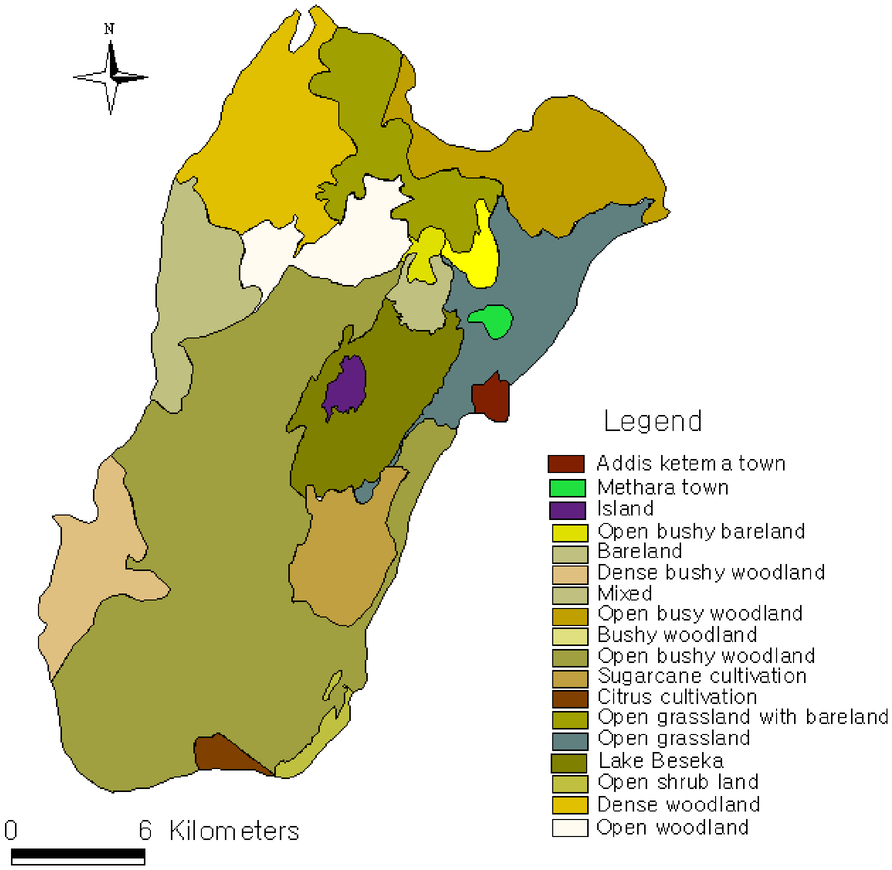


Figure 2.6: Land-use and land-cover (LUC) in the study area (data from WWSDE, 1999).

2.5 Geology

2.5.1 General Geology

Two major tectonic phenomena, i.e., extrusion of basaltic flow following the uplift of Afro-Arabian dome and its subsequent dissection through the development of East African System, (EAS) created major geologic and geomorphic features in Ethiopia. Following the uplift of the Afro-Arabian dome, extensive extrusion of Oligocene basalts associated with ignimbrite, trachyte and rhyolite ascended through fissures related to tensional zones (Mohr, 1962). The Ethiopian plateau is capped by about 2-km thick flood basalts and rhyolites that erupted between 31 and 29 Ma (Hofmann et al., 1997; as cited in Cornwell et al., 2006; Maguirei et al. 2006). The extrusion of Ethiopian flood basalt was earlier, or connected with the onset of rifting in the Red Sea and Gulf of Aden 29Ma (Wolfenden et al., 2004). While rifting is believed to have occurred in Ethiopia later, between 18 and 15 Ma in the southern and central MER (Woldegabriel et al., 1990) and even much later in the northern MER at about 11 Ma (Wolfenden et al., 2004).

Three important rift structures, two oceanic (Red Sea and Gulf of Aden) and one continental (East African Rift), come together to form very complex geological features. The Ethiopian rift system is part of the EARs, a complex tectonic feature with a system of downfaulted, non-continuous but related troughs. The tectonic and volcanic processes of Ethiopian Rift systems are not uniformly continuous, rather they are characterized by distinct periods of increased and decreased activity (Elc electoconsult, 1987). This rift system is of special interest, as it forms the third arm of the Red Sea, Gulf of Aden rift-rift-rift triple junction where the Arabian, Nubian, and Somalian plates join in Afar, exposing the transitional stage of continental rifting worldwide (Bastow et al., 2005; Cornwell et al., 2006). Continental separation has occurred in the Red Sea and Gulf of Aden, while seafloor spreading has yet to begin in the MER (Cornwell et al., 2006).

The MER lies in a unique transitional zone between the continental rifting of East Africa and the seafloor spreading of Northern Afar and the Red Sea (Rooney et al., 2005), and extends from central Ethiopia to Afar in a NNE-SSW direction, cutting the uplifted Ethiopian plateau into the northwestern and southeastern plateau. Throughout the Late Holocene and Quaternary, the MER has been seismically, tectonically and volcanically active, specially confined within a 10-20 km wide subaxial zone known as the Wonji fault belt (Mohr, 1962, 1967). This belt represents continuous volcano-tectonic activity from the Early Pleistocene to recent times, arranged in en echelon[1] fashion and maintaining a NNE-SSW structural orientation along the entire length of the MER (Le Turdua et al., 1999). A series of Quaternary volcanoes, mainly trachytes and pantelleritic rhyolites, occur along the Wonji fault belt, often located at offsets (Williams et al., 2004). Some of these volcanoes collapsed into summit caldera, ejecting widespread pyroclastic deposit covering the rift floor along with fissural basalts and lacustrine sediments.

1 En echelon describes parallel or sub parallel, closely-spaced, overlapping or step-like minor structural features in rock, such as faults and tension fractures, that are oblique to the overall structural trend (source: http://www.glossary.oilfield.slb.com).

The MER is divided into three sectors based on surface geology and geomorphology (Keranen and Klemperer, 2007): the northern (NMER), central (CMER), and southern (SMER) sectors. The NMER covers the area from south of the Afar depression to near Lake Koka, the CMER comprises the region between Lake Koka and Lake Awassa through the lakes region, while the SMER extends from Lake Awassa to the southern rifted zone of Ethiopia.

2.5.2 Geology of Lake Beseka area

Lake Beseka is situated in 20 km graben on the axial position of the rift floor in the northern part of the MER, which in turn is part of the EARs. Being part of the MER, the study area has been under intensive volcanic and tectonic activities during the Quaternary period. The lake watershed is bordered by older volcanoes and a rift margin to the east, and to the north it is bordered by young Quaternary complexes of Fentale, and to the west by Kone. The products of these recent volcano complexes, along with fissural basalt, rhyolite and alluvial and lacustrine sediments, form the common rock types in the study area.

Bofa basalts are the oldest rock formation in the study area, outcropping west and south of Beru and abundantly found in the valley of Awash to the east of the Abadir farm (Halcrow, 1978). They are mostly porphyric with plagioclase phenocrysts, sometimes scoraceous, and are aged between 3.5 and 1.6 million years (Abebe et al., 2005). Dino ignimbrites are the next oldest formation consisting of dark to green ignimbrite, with a thickness of about 30 m (Kebede, 1987). Dino ignimbrites are found mainly in the southeastern part of the watershed and to the north; they are covered by recent sediments in the area of Abadir farm.

Most of the study area and shores of Lake Beseka are underlain by Pleistocene to sub recent basalt, often coarse grained with phenocryst (Halcrow, 1978). Pleistocene basalts are exposed in most fault scarps and form the southeastern and western edges of Lake Beseka. The fault scarp cutting the Pleistocene basalt at the southwestern shore of the lake is characterized by hot springs. Three hot spring locations are indicated in Figure 2.7, of which two are currently submerged; traced when they were located at the western edge of Lake Beseka in 1973. It is assumed that Pleistocene basalts mainly represent lava pouring through fault fissures rather than being the product of volcano eruption (ibid).

A substantial portion of the southwestern part of the lake watershed is covered by undifferentiated lava flow of trachytic to rhyolitic composition with minor ignimbrite intercalations, which are believed to be the product of the Kone volcanic complex (Figure, 2.7). The Fentale volcano complex consists of two caldera volcanoes, the older Tinish Fentale and the younger Fentale volcano.

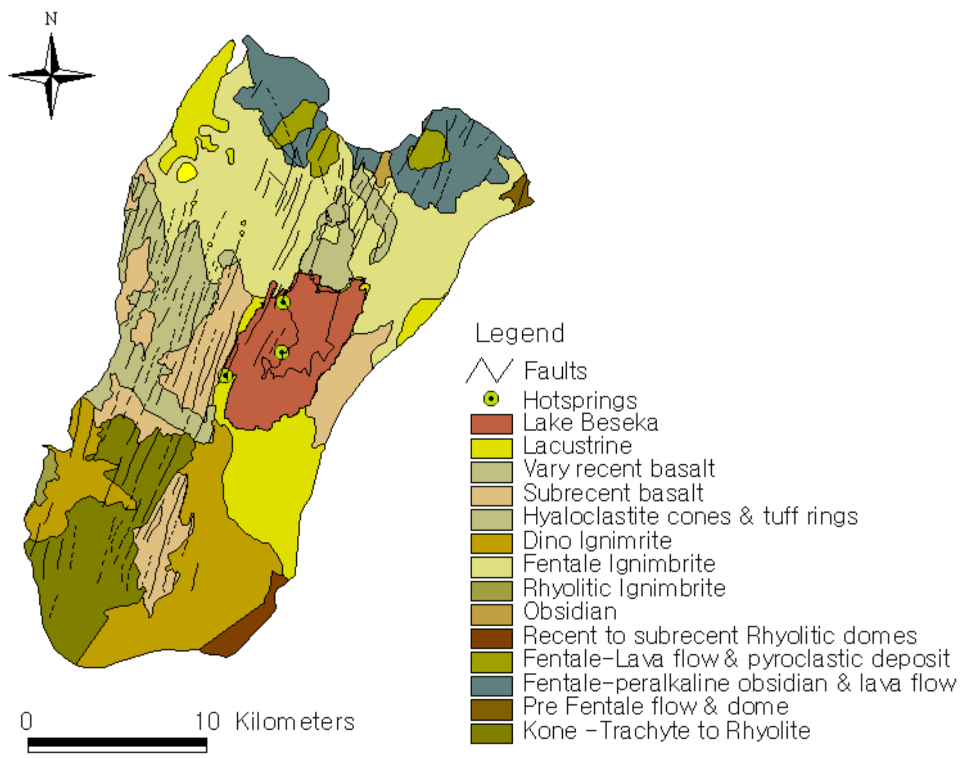


Figure 2.7: Simplified geology map of the study area (adapted from Elc electoconsult, 1989).

The caldera collapse of the Fentale volcano was accompanied by the eruption of voluminous ash flows, which solidified to form welded tuffs (Elc electoconsult, 1987). The pyroclastic deposit of Fentale ignimbrite is a very extensive welded ash-flow system, which caused the summit caldera to collapse (Gibson, 1970). The eruption and collapse of the main Fentale caldera resulted in widespread deposition of Fentale ignimbrite, which is a common rock type in the northern part of the study area. The volcanic sequence of Fentale includes intermediate silicic flow of Tinish Fentale, rhyolitic lava flow, peralkaline obsidian, adventive centers and pyroclastic deposit.

The older welded tuff from Tinish Fentale covers the plains to the east and southeast of the complex, while the rest of the plains are covered by the lavas and tuffs of the Fentale itself (Williams et al., 2004). Fentale ignimbrite is a widely distributed rock formation in the northern part of the watershed, especially occupying the lower slopes of Mount. Fentale and the north shore of Lake Beseka. Fentale ignimbrite is characterized by isolated blisters on its surface that were formed in the presence of steam pockets. Steam was probably formed by the sudden vaporization of water in swampy or shallow lakes trapped beneath the flow at the moment of deposition (Elc electoconsult, 1987).

Holocene aphyric basalts form an island in the lake and are found on the northern shore of Lake Beseka. They are mostly highly vesicular with a cliffy surface (aa lava) and related to fissural basaltic flow. Noticeable occurrence of these rocks is observed at the south of the Fentale crater where a very prominent lava flow produced by the 1820s eruption reached to the north of the lake (Halcrow, 1978).

Lacustrine and alluvial sediments are the youngest rock formation in the lake watershed. Alluvial deposit of clays and silts are found at the Methara and Abadir sugar farms, while lacustrine sediments are found around the perimeter of the lake. Alluvial sediments are deposited by the River Awash during flood times and can reach thickness of about 40 m in the study area (Kebede, 1987). Lacustrine deposits include silts, clays, and tuff, are mostly found associated with volcanoclastic sediments. Lacustrine sediments are mainly fine-grained silty sands often encrusted with salts, which may probably represent an accumulation of volcanic ash, reworked by the lake (Halcrow, 1978). Raised beaches, corresponding to higher lake levels exist around the perimeter of the lake.

2.5.3 Geologic structures

Rifting began in the southern and central MER between 18 and 15 Ma, but in the Northern MER it started to develop after 11 Ma (Woldegabriel et al., 1990; Wolfenden et al., 2004). Structural and stratigraphic patterns indicate a migration of extensional strain after 2.5 Ma from the border faults to a narrow 20-km-wide subaxial zone of narrow, near N-S trending zones of aligned eruptive centers cut by small offset faults and dykes arranged in a right stepping, en echelon pattern (Boccaletti et al., 1998;

Wolfenden et al., 2004). This zone includes rift volcanoes and young fault belts, with a N-NE trend that is oblique to the overall NE trend of the rift (Keranen et al., 2004). The major active fault zones identified within the rift are predominantly orientated N-S to NNE-SSW and NE-SW (Korme et al., 2004). Volcanic activity and structural deformation of the MER has been concentrated in a NNE trending structure formed by young faults and volcanic centers on the rift floor. This volcano-tectonic belt is commonly known as the Wonji fault belt.

Wonji Fault Belt is the principal active rift axis in the MER and was mainly formed during Quaternary period (Mohr, 1967). The Wonji fault belt is characterized mainly by NNE – SSW trending active extension fractures and normal faults, which are associated, in many places, with fissural or central volcanic activity (Mohr, 1987). The Wonji Fault Belt is not a continuous belt but formed from a series of offset segments, each consisting of short and often closely spaced normal faults and fissures (Gibson, 1969, as cited in William et al., 2004). In most cases, normal faults are arranged in a right stepping en echelon configuration (Mohr, 1968; Boccaletti et al., 1998), and the vertical throws are in the order of several tens of meters in the axial zone to several hundreds of meters in the rift margins (Woldegabriel et al., 1990).

The definition of the Wonji fault belt has been recently replaced by the concept of magmatic segments, which are described by Ebinger and Casey (2001) as 20-km-wide, right stepping, en echelon-arranged segments of intense magmatism and faulting. This new concept is based on the fact that the MER is magmatically segmented and seismically active (Kurz et al., 2007), and the extension in the rift is not primarily created by normal faults but by emplacement of magmas and dikes to trigger the normal faults. The MER is commonly considered the typical magma-assisted rift (Keranen and Klemperer, 2007). Kurz et al. (2007) clarified that the development of intra-rift faults in the MER is explained by magma injection where diking induced local stress to produce a new fault generation in the intra-rift. They indicate that magmatic segments are the principle locus of active extension in the Quaternary.

Magmatic segments include aligned silicic volcanic centers such as Gedemsa, Bosetti, Kone and Fentale. Most Quaternary volcanism has occurred in the magmatic segments, beginning ca. 1.6 Ma (Keranen et al., 2004), and more than 80 % of the strain across the rift is accommodated by magmatic segments (Ebinger and Casey, 2001). Keir et al. (2006) indicate that seismicity within the rift is localized within magmatic segments, and that this seismicity is characterized by swarms of low-magnitude earthquakes located in clusters parallel to Quaternary faults, fissures and chain of eruptive centers. They explain that earthquakes in the magmatic segments are predominately less than 14 km deep and may be triggered by dike injection.

Bendick et al. (2006) estimated the relative displacement between Nubia (northwestern) and Somalia (southeastern) plates at a mean annual rate of 4.0 ± 0.9 mm based on global positioning system (GPS) observation across the MER between the year 1992 and 2003. This low rift opening rate is explained by the authors as rather being the predominance of diking as the extension mechanism than the faulting on rift bounding faults. Extension within the rift is accommodated at least partially by aseismic dike injection, and diking is currently active in the area and could concentrate in the magmatic segments (Bendick et al. 2006). After conducting source-controlled seismic experiments in the MER, Keranen et al. (2004) also observed the dominance of magmatic processes in volcanic rift setting before immediate break up. The new three-dimensional seismic velocity model from the MER clearly images mid-crustal intrusions supporting breakup models based on dike intrusion and magma supply (Keranen et al., 2004).

The NNE-SSW fault system of the MER is crosscut by anomalous structure in western boundary of the rift. The E-W running transtensional structure known as Yerer - Tullu Wellel Volcano Tectonic Lineament (YTVL) intersects the western margin of the MER at the latitude of Addis Ababa, between 8° 20’ and 9° latitude (Abebe et al., 1998). The western margin of the MER between these latitudes is not well defined (Woldegabriel et al., 1990) due to this anomalous structure (Abebe, 2000). Currently active N-S structures within the MER combined with the E-W extension along this fault zone results in a regional stress field conducive to volcanic activity (Rooney et al., 2005). This zone described as an older and reactivated volcano-tectonic structure, and is not usually considered as part of the Cenozoic rift system. However, Keranen and Klemperer (2007) recently interpreted the YTVL as an essential part of the MER evolution. They indicate that volcanism in the YTVL and extensions along the Ambo fault (northern border of the YTVL) were coeval with extensions and volcanisms in the NMER. The intersection of these two volcano-tectonic structures has important hydrological implications, and resulted in the development of the River Awash drainage (Abebe, 2000), the only major river originating from the central Ethiopian plateau and flowing toward the rift.

The study area is located in the NMER, at the junction between the Ethiopian rift and the Afar triangle. It is situated within active magmatic segments, which makes it susceptible to intense volcanic and related tectonic activities. The other important structural feature of the study area is its position within latitudes where the E-W trending volcano-tectonic structure of the YTVL intersects with the NNE trending fault system of the MER. The western margin of the MER is not well defined in this zone, which could have significant hydrogeological implications for the area.

NNE tensional faulting dominates the local topography of the study area. The NNE-SSW trending normal faults determined the morphology of Lake Beseka and the general form of the watershed. Normal faults are observed cutting Pleistocene basalts southeast and west of Lake Beseka and displacing Fentale ignimbrite in the northern part of the watershed. Lake Beseka is bounded to the west and east by fault scarps, which run parallel to the regional trend of the axis of the rift valley. On the surface, horizontal spacing of faults is more frequently observed than vertical displacements. To the south of the Fentale caldera, welded tuff is cut by tensional fissures, which are open fractures with little or no vertical displacement oriented in a NNE-SSW direction (Williams et al., 2004). Vertical displacements of faults are more pronounced in the basalts, which are located west of Lake Beseka, and the vertical offsets can reach as much as 30 m (Elc electoconsult, 1987). Faults in Fentale ignimbrite are characterized by low vertical displacement, and are affected by lower degree of weathering compared to the faults in basalts.

The seismically and volcanically active northern MER and Afar rifts are the only places worldwide where the transition between continental and oceanic rifting is exposed on land (Keir et al., 2006). There are frequent earthquakes all over the Ethiopian rift; the epicenters are almost exclusively related to the major rift structures (Ayenew, 2004). The study area has been under seismic and volcanic activities (Figure 2.8). The most recent basaltic volcanism in the Fentale was around 1810 according to historic reports and local oral tradition (Williams et al., 2004). At the time when Lake Beseka is believed to have started to expand, an earthquake with a Mercalli intensity of

4 was recorded by the geophysical observatory of Addis Ababa on March 7, 1967, with the epicenter located about 90 km south of Methara (Halcrow, 1978).

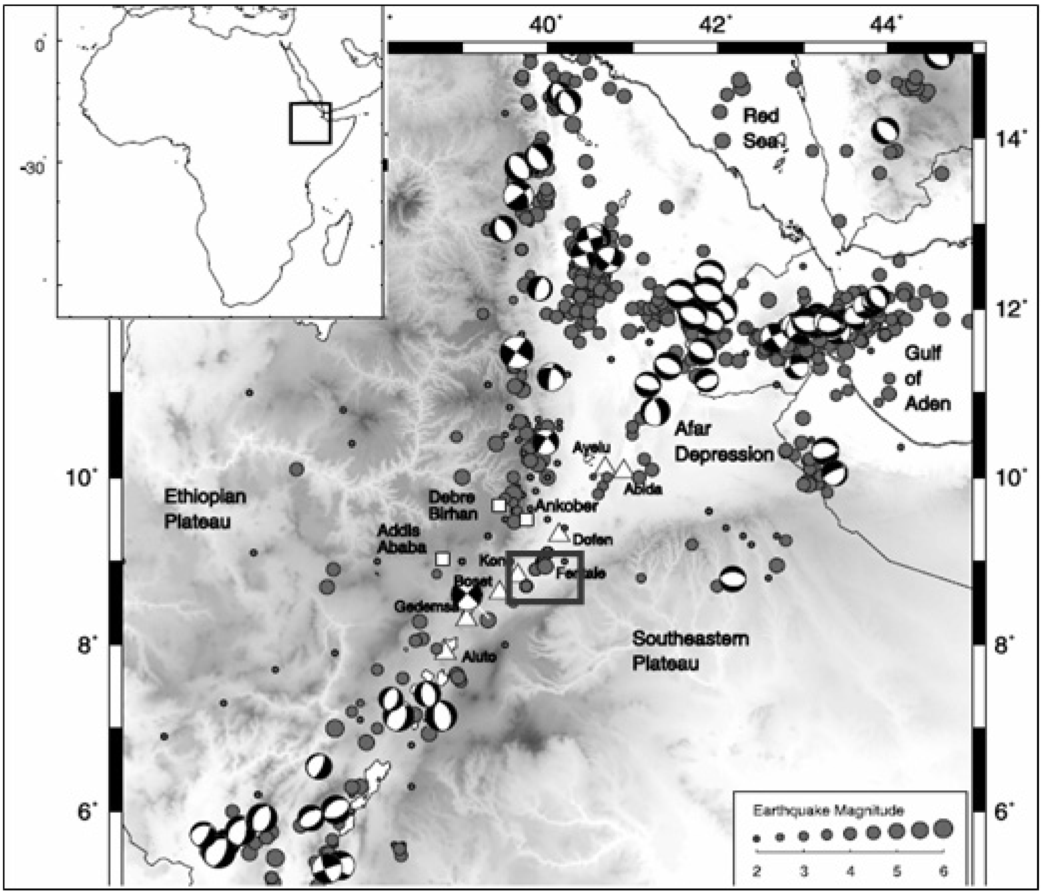
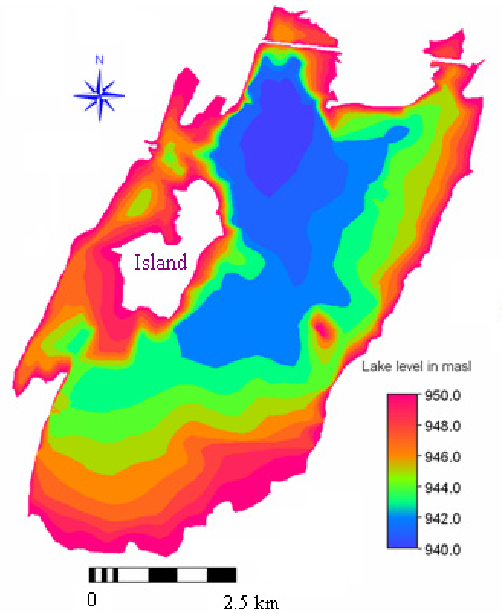


Figure 2.8: Seismic activity of the Horn of Africa since 1960, (adapted from Keir et al., 2006). Quaternary volcanoes are shown by triangles, and study area is represented by a rectangle.

2.6 Bathymetry of Lake Beseka

The earth has undergone a series of climate variations during the Quaternary period, which is characterized by alternating warm, dry and cooler wet periods. Research on Quaternary sediments states that lakes in the MER were larger or even joined together during the pluvial period (Ayenew, 1998). Radiocarbon dating of lacustrine sediments from the northwestern shore of Lake Beseka by Willams et al. (1981) indicates that the lake has undergone water level fluctuation from the late Pleistocene to the Holocene. According to this work, the lake regressed between 18,000 and 12,000 years BP, following the transgression in the late Pleistocene, which resulted in a lake level several meters higher than the present level. Following another transgression, the lake regressed once more after 11,200 years BP, which is associated with the deposition of ash and tuffaceous loam.

For the last few decades, the water level of the Lake Beseka has been rising drastically, and is an immense threat to the surrounding environment. The bathymetry of the lake was conducted at the scale of 1:5000 with a contour interval of 1 m in 1998 by WWDSE. The bathymetry of Lake Beseka as a map was scanned and digitized in a GIS environment and converted to a digital format. Surfer software (Golden software inc., 2002) was used to compute volume and surface area of Lake Beseka at different lake stages, and capacity curve of the lake constructed. With a lake level elevation of 950.7 masl, the volume and surface area of the lake is calculated to be 224 Mm^3 and 40 km^2, respectively (Figure 2.9).



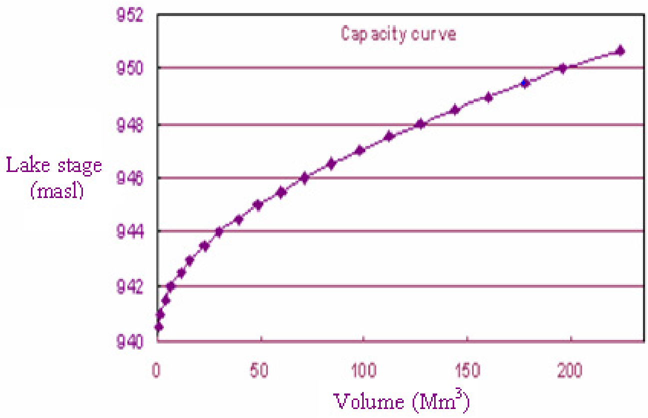


Figure 2.9: Bathymetry and volume capacity curve of Lake Beseka (data from WWSDE, 1999).

The deepest part of the lake is 11 m and located in the northeastern part of the lake. Before its expansion, Lake Beseka was likely to have been located there and then spread toward the higher areas when the equilibrium of the water balance was disturbed by a net influx of water to the lake.

Tectonism and volcanism have affected the morphology of rift lakes (Ayenew, 1998). Like most lakes in the rift, tectonic and volcanic activities have played an important role in the morphology of Lake Beseka. The shoreline of the lake to the west and east is bounded by NNE-SSW trending normal faults. The bottom of the lake is characterized by a series of steps of bedrocks related to different phases of tectonic activity. Recent basaltic lava flows formed the island in the western part of the lake, and dammed the lake in its northern shoreline.

2.7 Data collection and preparation

During the field season, groundwater survey was conducted in May 2006, and three locations were selected for installation of pressure transducer data loggers for groundwater monitoring purpose. Two data loggers were installed in abandoned wells of Gofa sefer (east of Lake Beseka) and Abadir area (south of Lake Beseka), and the third data logger was installed at hotspring site in the southwestern corner of the lake. Data loggers monitored groundwater level and temperature data on daily bases for 15 months (July 2006-September 2007) for Gofa Sefer and Abadir sites while the data from hotspring site was limited only for 6 months (July 2006-December 2006) due to problems faced in the area. Well data including lithological descriptions, static water levels, casing set ups, geophysical and pump test data were collected for wells in and around the lake watershed from Water Well Drilling Enterprise (WWDE), Water Works Design & Supervision Enterprise (WWDSE) and Methara Sugar Estate. Meteorological data (rainfall, temperature, relative humidity, sunshine hours, windspeed, and pan evaporation) were collected from National Meteorology Agency (NMA) for Methara station (1964-2005), and Nura Era station (1977-2005). Additional meteorological data were collected from Methara Research center station (1966-2007). Lake level measurements were collected from Ministry of Water Resource (MWR) (1976-2007) for Lake Beseka.

Water samples were collected from wells, Lake Beseka, River Awash, irrigation drainage lines at Abadir farm, hot spring and cold spring sites, and analyzed for major elements (Na^{+}, K^{+}, Mg^{+2}, Ca^{+2}, Cl^{-}, SO_4^{-2}, HCO_3^{-}, CO_3^{-2} and F^{-}) , stable isotopes ($\delta^{18}O$ and $\delta^{2}H$) and also for limited tritium concentration. In-situ measurements were carried out at field conditions for pH, electrical conductivity (µS/cm) and temperature (°C). Samples were analyzed in the hydrogeological laboratory of Technical University Bergakademie (Freiberg, Germany) for their ion concentration, and in the laboratory of UFZ (Helmholtz-Zentrum für Umweltforschung of Leipzig, Germany) for stable isotopic concentrations. After checking for errors and inconsistencies, data from previous studies (WWDSE, 1999) and (Tessema, 1998) were included to the data set for further analysis. All maps including the soil map, land cover map and geological maps, topographic map of the watershed of the lake and the bathymetry of Lake Beseka were only available as hard copy, and were converted into the soft copy during this study.

3 HYDROGEOLOGICAL PROPERTIES

3.1 General background

The hydrogeology of the volcanic rocks in the Ethiopian rift can be identified as a complex environment owing to the heterogeneity of aquifer composition, irregular distribution of groundwater flow patterns, presence of variable groundwater chemistry, and groundwater depths and occurrence. In the UN hydrogeological map of Africa (UNESCO, 2006), the hydrogeology of the volcanic rocks covering the great East African rift valley is classified as "the most complex and least understood". The disruption of lithologies by cross cutting faults and the variability of volcanic structures make the hydrogeology of the rifted volcanic terrain in Ethiopia very complex (Kebede et al., 2007). The intricate spatial and temporal distribution of volcanic rocks, their wide compositional and textural variability, their different degree of fracturing and weathering and the existence of active tectonic structures in the region create a complex aquifer system in the Ethiopian rift.

Volcanic rocks are characterized by wide range of chemical, structural and hydrogeological properties mainly due to variation in rock types, and the way rocks were ejected and deposited. Volcanic rocks are formed by solidification of magma at or near the ground surface or deposited from ejected fragmentary material. Magma that pours out through the volcanic vents suffers a loss of pressure that allows dissolved gases and water vapor to exsolve (Custodio, 2004). This molten, more or less degassed magma pours out through fissures or volcanic vents to form basic or acidic lava flows. Acidic lava (e.g., rhyolite) is viscous and, therefore, restricted in extent, often forming steep-sided bulbous domes, while basic lava (e.g., basalt) has lower viscosity, and spreads over large areas forming thin, aerially extensive sheets (Cook, 2003). Magma can be fragmented and thrown into the air to form pyroclasts of various sizes, due to huge pressure developed from gases and water vapors within the magma chamber. Pyroclasts may be deposited as cooled fragments that remain loose, or as hot fragments that may be welded to form ash flow (tuff) or ignimbrites.

Hard rock hydrogeology is not applicable to volcanic rocks, as volcanic rocks are mostly characterized by the presence of primary porosities (Larsson, 1984). Primary porosities such as vesicles, flow contacts or interflow spaces, lava tubes or tunnels, shrinkage cracks or columnar joints are common features in volcanic rocks. Total porosity of volcanic formations is generally high, due to voids created by exsolved gases and to the frequent scoraceous and brecciated parts, as well as to their often clastic nature (Custodio, 2004). Primary porosity in volcanic rocks depends on the rate of cooling, extent of degassing during cooling, and on the viscosity of the magma (Cook, 2003). Massive lavas and welded tuffs (ignimbrites) are characterized by low porosities, but they can develop vertical joints that run from top to bottom due to shrinking at the time of cooling, which enhances their porosities. The porosity of the pyroclastic rocks is directly related to fragment size, sorting and degree of cementation. Porosity of pyroclasts varies from impervious cemented volcanic agglomerates to very permeable, young and loose pyroclast deposits, depending mainly on the degree of consolidation and welding.

Although primary porosities are common features in volcanic rocks, they only play a minor role in the permeability of a given formation. This is due to the fact that primary porosities are expected to alter with time when volcanic rocks are exposed to various geological processes, such as weathering and tectonic processes. This is explained by Custodio (2004) as the ‘aging effect’, which describes the importance of secondary properties regarding the hydraulic behavior of volcanic formations through internal adjustment and rock alteration. The presence of secondary porosities could enhance the permeability of volcanic rocks by increasing the interconnectivity of isolated primary porosities that were randomly distributed in the host rock. Secondary porosities that permit flow and storage of groundwater are created mainly by weathering and fracturing processes.

Weathering has considerable influence on the storage capacity of rocks, and includes disintegration and decomposition of primary mineral constituents of host rock as well as the formation of secondary clay mineral products (Larsson, 1984). Weathering of volcanic rocks is influenced by many factors including chemical composition of the rock, relative solubility of original rock, texture of host rock, availability of CO_2 and other gases, etc. Volcanic formations are easily altered on the surface and at depth due to their glassy nature and geochemistry (Custodio, 2004). Fractured rocks are comprised of a network of fractures that cut through a rock matrix. Fractures can be characterized in terms of their dimensions (aperture, length, width), their location (orientation, spacing, etc.) and the nature of the fracture walls (e.g., surface roughness) (Cook, 2003). The capacity of fractures to channel groundwater flow is controlled by interconnectedness of the fractures that are distributed in the host rock, i.e., fracture connectivity, which increases with increasing fracture length and fracture density.

In rocks with distantly spaced fracture patterns, storage capacity of individual fractures plays a more predominant role compared to highly fractured rocks, which can be treated as porous media in hydrological computation (Larsson, 1984). Spacing and distribution of fracture systems in the rock matrix are also important factors for the development of weathered layers. Fractures provide conduits to channel groundwater flow; however, they can also hinder groundwater flows if they are filled by clays or other secondary minerals.

Permeability of volcanic rocks is highly variable, and covers a wide range of values, from almost zero (less than 0.0001 m/day) to more than 1,000 m/day (Custodio, 2004). In the Ethiopian rift valley, hydraulic conductivity of water-bearing formations varies significantly among water wells at only few hundred meters distance from each other (Chernet, 1982; Ayenew, 1998). The study area is affected by several fault systems and covered by volcanic rocks varying from fissural basaltic to viscous acidic lava flow, and pyroclastic deposits to sediments occurring in an inter-fingering manner. Complex aquifer systems are prevalent in this region due to multiple effects of active primary and secondary geological processes. The aquifer system of the study area can be summarized as volcanic rocks with variable composition, texture and stratigraphic relationships affected by different degrees of tectonic and / or weathering processes, which either facilitate or hinder groundwater flow process.

This chapter comprises the hydrogeological characterization of water-bearing volcanic rocks and analyzes the spatial and temporal data to determine the hydrogeological framework of the study area. Hydrogeological physical and stress data are analyzed to develop an understanding of the important aspects of the aquifer system. The hydrogeological framework describes the physical system of the area, such as aquifer thickness, bedrock configuration, and aquifer hydraulic and storage parameters. The hydrogeological stress data provides a description of the source and sink of the aquifer system.

3.2 Determination of hydrostratigraphic layers

Hydrostratigraphic units are comprised of rocks of similar hydrogeological properties, i.e., similar physical and chemical characteristics. In hydrostratigraphic concepts, several lithological units can be combined into one hydrostratigraphic layer, or a single geological formation can be subdivided into aquifer and confining unit. Well data including lithological log descriptions, static water levels, casing set ups, geophysical data and pump test data were collected for wells located in and around the lake watershed from Water Well Drilling Enterprise (WWDE), Water Works Design & Supervision Enterprise (WWDSE) and Methara Sugar Estate. Lithological logs of 44 boreholes were combined with geophysical data to define the aquifer system of the study area. These hydrogeological data are concentrated mainly at the center of the watershed and around Lake Beseka.

An extensive unconfined aquifer is found at a moderately shallow depth consisting of mainly fractured and weathered volcanic rocks. The main water-bearing formation is a mixed aquifer system of fractured and weathered basalts, welded tuffs and pyroclastic deposits. Basaltic rock is the most productive aquifer in the study area. It varies from fine-grained basalt to vesicular and scoraceous basalt. The ability of basaltic formations to transmit water is dependent on the presence of fractures, cracks or tubes, and weathered layers. Fractured basalt, weathered scoraceous basalt and weathered vesicular basalt are common water-bearing formations in the study area. The occurrence of weathered layers and fractures enhance the permeability of the basaltic formations by increasing the interconnectedness of existing primary porosities or other secondary features so as to store and transmit a sufficient amount of groundwater.

Recent and subrecent basalts in the study area are intensely affected by factures, and create highly permeable aquifer zones around Lake Beseka. These basalts are fresh and highly jointed; and have a high degree of permeability and productivity (Chernet, 1990). In addition to being affected by fractures, the basalts are characterized by primary porosities such as vesicular cavities, cooling joints and scoraceous intercalations. These porosities play an important role in creating favorable conditions for the formation of a weathered permeable zone. Weathering helps to intersect isolated vesicular or scoraceous pore spaces so as to enhance permeability of the formation.

Recent aphyric basalts have high permeability, as they are characterized by highly interconnected vesicular pore space (Kebede, 1987).

Pyroclastic deposits are the other water-bearing formation in the area composed of fractured and / or weathered welded tuff, and non-welded tuff with pumice and volcanic breccias. The hydraulic conductivity of pyroclastic deposits depends on the degree of consolidation and welding (Cook, 2003). Sorted pumice is characterized by high inter-granular permeability, which tends to decrease with time due to weathering and compaction. Pyroclastic deposits of the Fentale group show a wide range of hydrogeological features, and are characterized by moderate to low permeability. This is due to the fact that non-welded tuff aquifers have different pyroclast sizes, sorting arrangements, levels of pumice and breccia intercalations and are affected by different extents of weathering. Coarse-grained pyroclastic deposits (pumice) are fairly permeable compared to fine-grained deposits (ashes and cinders) (Elc electroconsult, 1987). Formations such as breccias, lapilli, ash-fall deposits and loosely cemented ash flows easily change into a low-permeability mass in only a hundred thousand to a few million years (Custodio, 2004).

Welded ash flows (ignimbrites) are characterized by very low primary porosities, as they are formed by the fusion of rock fragments at high temperatures. Thus, water circulation in welded tuff depends on the presence of secondary porosity and permeability, which can be developed through fracturing and weathering. Due to their tendency to develop joints to considerable depths, welded tuff can easily permeate water and become productive aquifers. In the study area, fracturing and weathering have affected the welded tuff to different degrees, and as a result, it varies from fresh, massive to jointed, toward slightly to highly fractured and weathered ignimbrite. To the west and south of Fentale, welded tuffs are characterized by moderate permeability due to fractures and the abundance of fresh, columnar joints with blisters and crevasses. However, the permeability of welded tuffs tends to decrease east of Fentale, as joints are filled with clay materials due to weathering (Kebede, 1987).

Methara and Abadir sugar plantation areas are covered by alluvium, with textures varying from sandy clay to sandy gravel. These sediments are characterized by variable permeability, where volcanic and water-deposited sands are highly permeable in contrast to fine-grained sediments. Permeability of lacustrine and alluvial sediments is derived from voids that exist between grains, i.e., intergranular porosity, which in turn is governed by size, shape, sorting and packing of grains. Lacustrine deposits have a limited lateral extent in the study area, and are mostly found interbedded with pyroclastic deposits, especially with tuffs or overlying the fractured basalt.

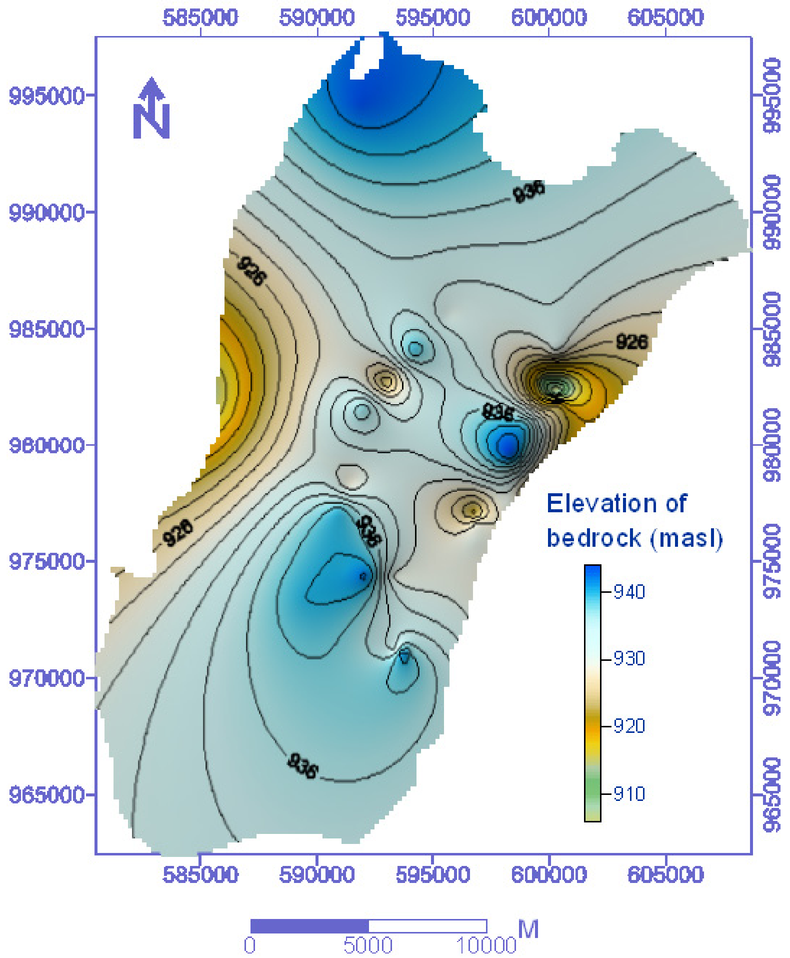


Figure 3.1: Elevation map of the bedrock in the study area (prepared from lithological log of boreholes, DEM and geophysical data)

The elevation map of bedrock was derived from lithological logs, DEM, and geophysical data (Figure 3.1). The bedrock is characterized by significant elevation changes within short distances, and is significantly irregular along the western edge of Lake Beseka, which is characterized by abundant tectonic features and by hot springs (Figure 2.7). The bedrock is a massive volcanic sequence, which is locally fractured by deep faults. Tectonic structures play an important role in controlling the hydrogeological regime of the area, and that could be the reason behind the irregularity of the morphology of the bedrock. Generally, the bedrock elevation is deeper in the western part of the watershed and lowest in the northeastern part.

In the study area, aquifer thickness is spatially variable and mainly governed by aquifer composition, type and magnitude of tectonic structures and extent of weathering, etc. Saturated thickness represents the vertical distance from top elevation of the aquifer to the base of the formation or elevation of the bedrock. In general, the lake watershed is characterized by an average saturated thickness of about 25 m, varying from 9 m of weathered tuff in the Haro adi area (east) to 46 m of aquifer thickness of highly fractured basalt in the Tuttuti area (west).

Two cross-sections were prepared from the lithological logs of boreholes, groundwater levels and geophysical data to determine the hydrogeological setup of the lake watershed. The first section is from the western to northeastern part of the watershed across Lake Beseka with a total length of 16 km (Figure 3.2, a). The second section is from northwest to southeast direction across Lake Beseka with a total length of 18 km (Figure 3.2, b). In the first cross-section, the aquifer is a mixture of basalt and welded tuff. The basaltic aquifer varies from scoraceous and vesicular to jointed basalt, while the welded tuff is mainly characterized by weathered layer mixed with pumice and sand breccia. In the second cross-section basalt is the dominant water-bearing formation. This cross-section illustrates that, although Abadir farm is covered by alluvial and lacustrine sediments, basaltic formation remains to be the main water-bearing formation in the area. Both sections reveal the existence of three hydrostratigraphic layers: overburden, saturated aquifer and bedrock. The composition of the aquifer is mixed in the area, varying from highly fractured basalt to weathered scoraceous and vesicular basalt, and weathered and fractured welded tuff to intercalation of pyroclastic deposits.

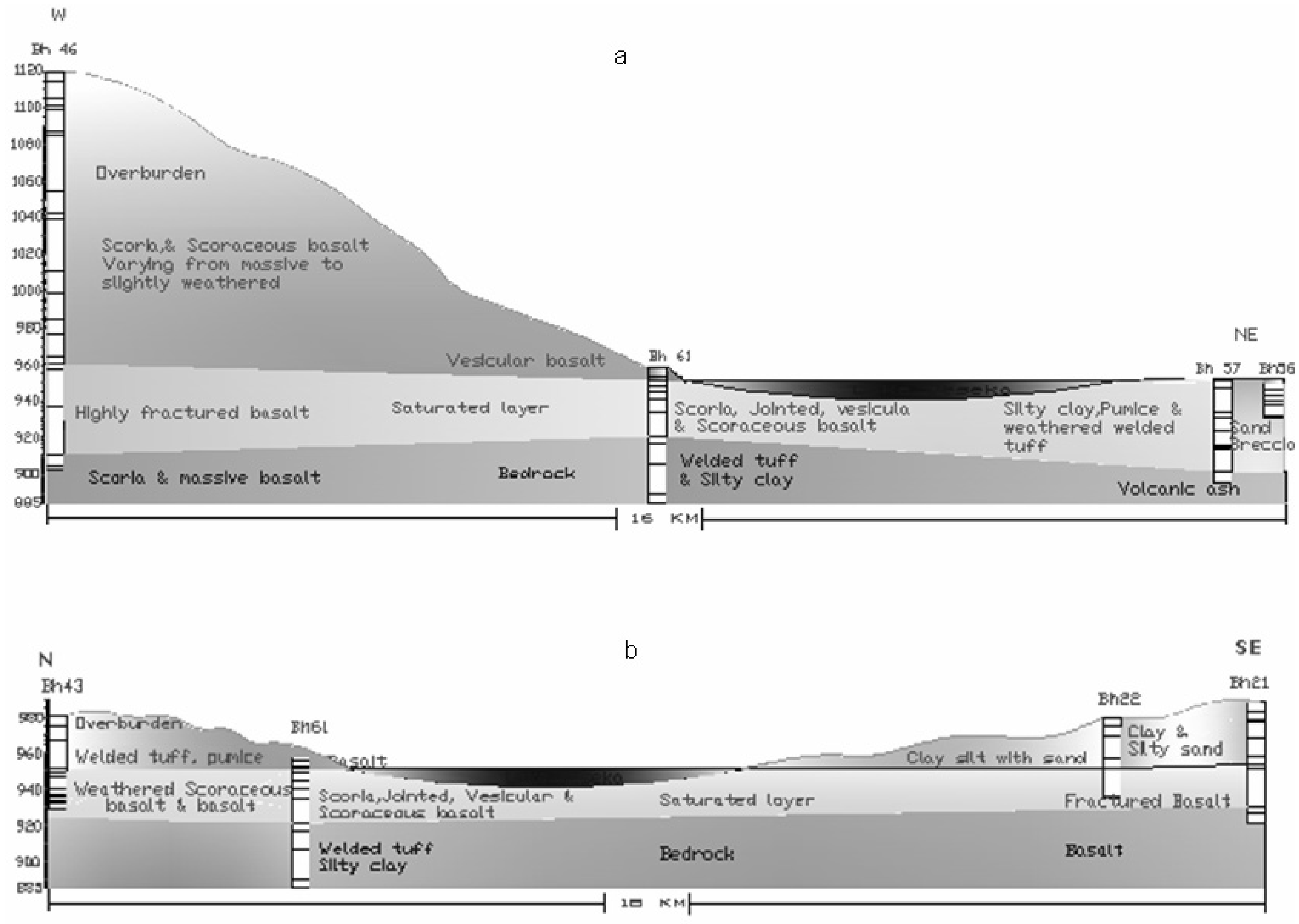


Figure 3.2: Cross-section a: thickness of the overburden is higher west of Lake Beseka than to the east, and the aquifer is a mixture of basalt and welded tuff. Cross-section b: the aquifer is mainly composed of basalt, which varies from weathered scoraceous and vesicular basalt to fractured basalt.

3.3 Spatially dependent flow system analysis

Gravity is the dominant driving force in groundwater movement in an unconfined system. It moves downhill until in the course of its movement, it reaches the surface at a spring or through a seep along the side or bottom of a stream channel or estuary (Heath, 1983). The groundwater flow pattern is controlled by the configuration of the water table, which in turn is controlled by the prevailing climate, topography, and geology (Sophocleous, 2004). Climatic factors, precipitation and evapotranspiration, mainly controls the rate of recharge, while topography and geology determine the spatial distribution of discharge and recharge zones (Dingman, 2002). A groundwater contour map of the study area was prepared from the groundwater level data of 78 boreholes, which are located inside and around the lake watershed (Figure 3.3).

The contour map describes the groundwater flow system where the system is in equilibrium with its fluxes, and the flow is governed by the medium properties of the aquifer. The groundwater level is higher in the topographically elevated area, in the south and northwestern part of the watershed compared to the area around Lake Beseka, and these areas can be considered as recharge zones. These zones are topographically higher areas, and are characterized by a deeper water table compared to the lower areas around the lake (Figure 3.4) In recharge areas, the unsaturated zone between the water table and land surface there is often a deeper than in discharge areas (Fetter, 2001). However, tectonic structures are abundant features in the study area, and can favor direct and localized rainfall recharge to the groundwater system.

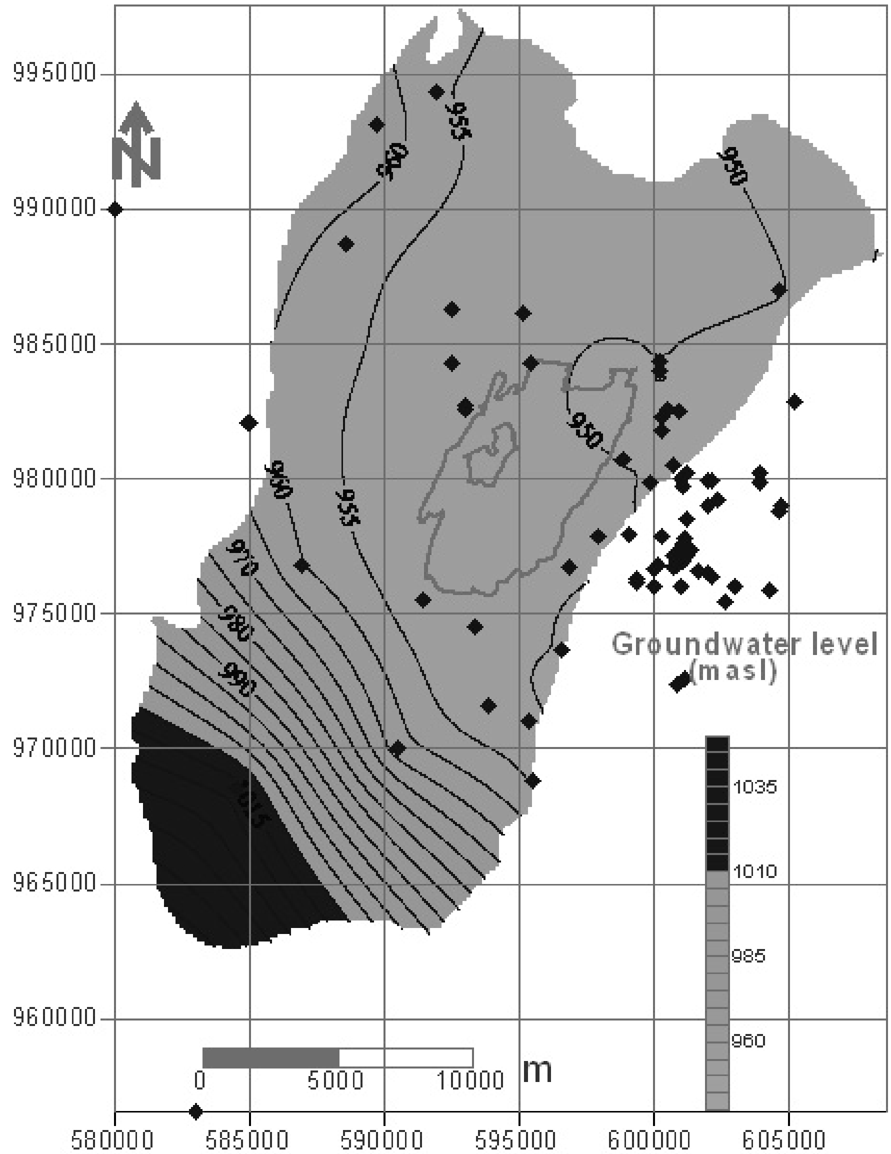


Figure 3.3: Groundwater level and borehole locations in the Lake Beseka watershed (data collected from the offices of WWSDE, Methara Sugar Estate and field data).

In the study area, the groundwater flows from the south and northwestern part of the watershed toward the lake. The groundwater level is high in the south and northwestern part of the watershed and becomes lower toward Lake Beseka and the eastern part of the watershed. The groundwater level is at its lowest east of Lake Beseka, around the Methara town area, and this may induce groundwater outflows from the lake. Part of the basin around Lake Beseka and east of the lake can be considered as a discharge zone. The southwestern and western border of the lake is characterized by widespread hot springs. In discharge areas, the water table is usually at or near the surface, and such areas are usually the sites of streams, lakes, wetlands, or springs (Dingman, 2002).

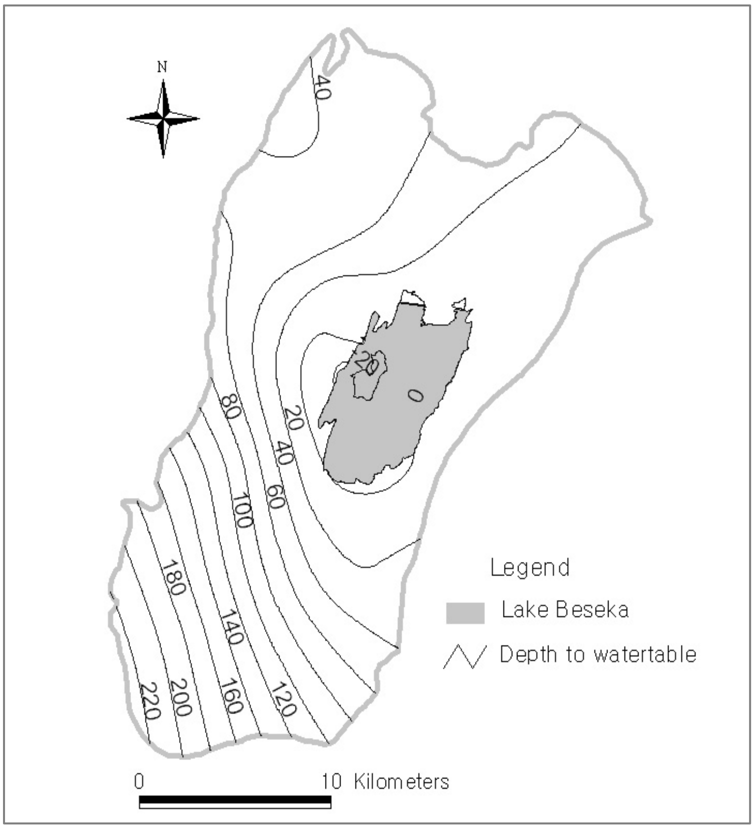


Figure 3.4: Depth to water table (prepared from data from WWSDE, Methara Sugar Estate and field data).

The groundwater contours of the study area are influenced by tectonic structures, as can be seen west of Lake Beseka, where hot springs are aligned along fault lines (Figure 2.7). Hot springs along tectonic lines are evident around the Tone area (southwest) and the northwestern part of the lake, where two of the hot springs are currently submerged. The groundwater flow direction in the study area is comparable to the surface water flow direction, and is controlled mainly by topographic, geologic and tectonic features.

In spatially dependent groundwater flow systems, flows are mostly dependent on the hydraulic conductivity of the aquifers. The rate of movement of groundwater from recharge areas to discharge areas depends on the hydraulic conductivity of the aquifers (Heath, 1983). Thus, assuming similar recharge values, hydraulic conductivity values can be roughly interpreted from groundwater contour maps. The northwestern and eastern parts of the lake watershed are characterized by low hydraulic gradients in contrast to the southwestern part. These gently sloping groundwater contours around Lake Beseka area can be related to the existence of relatively high hydraulic conductive zones compared to the closely spaced contours in the southwest. This can be explained by the difference in the geology of the aquifers, and spacing and distribution of fractures within the lake watershed.

3.4 Temporally dependent flow system analysis

Meteorological and hydrological parameters such as precipitation and evapotranspiration often govern the temporal variability of groundwater in an aquifer system. The depth of the water table at a given location is determined by feedback among precipitation, infiltration, runoff, and evapotranspiration at that location along with the regional flow (Dingman, 2002). In addition to aquifer properties, storativity plays an important role in controlling temporal variation of groundwater in the system. Water-level measurements from wells remain the principal source of information on the effects of hydrologic stresses on groundwater systems (Alley et al., 2002). During this study, three data loggers were installed and groundwater level and temperature data were monitored on a daily basis for 15 months (Figure 6.5).

An analysis of the monitored data indicates that the groundwater system of the area responds to rainfall events during the spring and summer rainy seasons. Temporal variability of the groundwater system is governed mainly by climate factors, specifically by the relative importance of precipitation with respect to evapotranspiration. The groundwater system is recharged when the rain amount is high enough to exceed the rate of evapotranspiration at a given time scale. The amount of water available for groundwater recharge depends on the annual total rainfall, intensity and duration of individual rainstorms, and the evaporation, which is a function of latitude, altitude and temperature (Larsson, 1984).

During the main rainy season, groundwater starts to increase about 3 weeks after the onset of the rainy season, and reaches its maximum value after 90 to 100 days. When recharge occurs following periods of precipitation, the groundwater level rises in response to the arrival of recharging rain water to the saturated zone (Figure 3.5, right).

During the dry season, in contrast, the groundwater level recedes due to the combined effect of groundwater runoff and evapotranspiration where there is little or no rain.

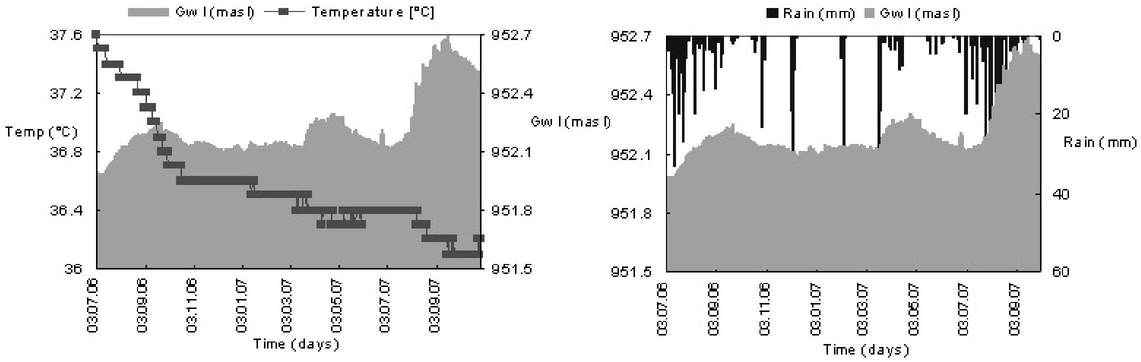


Figure 3.5: Time-series daily groundwater temperature and groundwater level (left) and daily rainfall with groundwater level data (right) from the borehole at Gofa Sefer (right) for the time period 03.07.2006-30.09.2007.

In the study area, the temporal fluctuation of the groundwater levels in a response to prevalent climatic factors is also reflected by the temperature of the groundwater system. The rise of the groundwater level during wet periods is associated with a decline in temperature of the groundwater following the arrival of rainwater to the water table (Figure 3.5, left). The quick response of the groundwater system to rainfall events suggests the importance of preferential flow as a recharge mechanism to the aquifer system. The widespread existence of fractured volcanic rocks can create favorable conditions for preferential flow of rainwater towards the saturated zone.

3.5 Determination of hydraulic parameters

Hydraulic properties of an aquifer system such as hydraulic conductivity, transmissivity, and storativity are useful parameters, as they determine the transmitting and storing capacity of water-bearing formations. Hydraulic conductivity and transmissivity describe the flow property of a medium, indicating how well groundwater can be transmitted to the discharge point (Lubczynski and Roy, 2005). Hydraulic conductivity describes how fast water can move through pore spaces or fractures, while transmissivity measures how much water can be transmitted horizontally; and is the product of hydraulic conductivity times thickness of the aquifer. Storativity is a storage parameter, which describes the volume of water that a permeable unit will absorb or expel from storage per unit surface area per unit change in head.

Hydraulic properties of aquifers can be estimated through laboratory and field techniques.

The aquifer pumping test is a frequently employed in-situ method, which is used to determine the hydraulic properties of water-bearing formations. The test involves putting an artificial stress on the aquifer system by continuously pumping water from the borehole and measuring water level changes in the pumped well. The change in hydraulic head can then be used to estimate hydraulic conductivity and storage parameter of the aquifer. Different analytical methods are available for interpretation of pumptest data. Pumping tests, which are conducted to determine hydraulic properties of volcanic rocks are generally interpreted in the same way as the tests carried out for granular materials (Custodio, 2004).

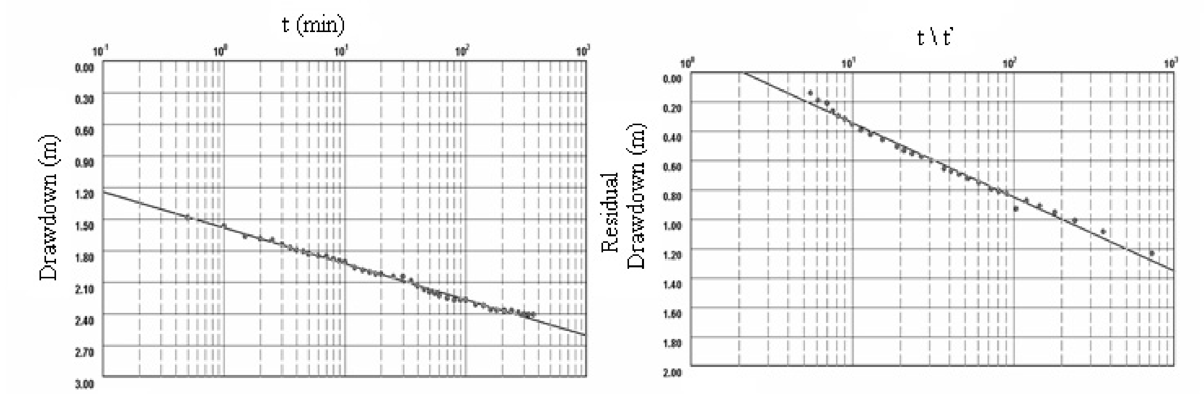


Figure 3.6: Time-drawdown and time-residual drawdown plot of borehole at Haro Adi (east of Lake Beseka) from left to right: the borehole was pumped at a discharging rate of 10 l/s for a total duration of 360 minutes.

Data from pump tests of 30 boreholes that had been conducted by the drilling contractors during the phase of borehole construction were collected. For most boreholes, the pump tests lasted between 6 and 24 hours. Residual drawdown records from recovery tests were available for some wells, and the total duration period varied from 0.5 to 360 min. All pump tests were single well tests, and pumping wells were simultaneously used for monitoring drawdowns. From single well tests, the storage term of the aquifer cannot be determined, thus existing pumptest data are used only for estimation of hydraulic conductivity or transmissivity. In the absence of observation wells, estimating storage properties using analytical methods can be difficult (Kolja et al., 2007).

Aquifer Test Version 2.5 (Waterloo Hydrogeologic Inc) was used in this study for evaluation of pumping and recovery test data. The Cooper and Jacob (1946) straight-line fitting method with Jacob's correction for unconfined aquifers was used for the analysis of drawdown data. The Cooper and Jacob method is based on a linear curve fitting, and is the preferred method for single well tests (Fetter, 2001). The Theis and Jacob recovery test method with Jacob's correction for unconfined aquifer was used for evaluation of residual drawdown data. The interpretation of the pumping and recovery test of an example borehole is presented in Figure 3.6.

The permeability of volcanic rocks is highly variable in space, as it is mainly governed by the presence and extent of secondary porosities like fractures, joints, weathered layers, etc. In the MER, permeability of volcanic rocks shows high spatial variability, both laterally and with depth, mainly due to variation in the degree of fracturing (Ayenew, 1998). The evaluation of pumping test data indicates that recent and subrecent basalts are characterized by the highest hydraulic conductivity. The values vary between 766 m/day for basalt to 0.25 m/day for lacustrine deposit, which characterizes the heterogeneity of the aquifer system. Very high hydraulic conductivities are observed in the aquifers of the highly fractured and scoraceous basalts, while low values are observed for aquifers dominated by silty clay and weathered volcanic ashes. The aquifers in weathered welded tuffs and breccias are characterized by moderate hydraulic conductivity.

Based on pumping test data, the lithological and structural characterization of the study area, and the literature on other parts of the rift valley, the watershed of Lake Beseka can be divided into five hydraulic conductivity zones (Figure 3.7):

Zone 1: Recent and subrecent fissural basalts that are highly fractured and mostly with scoraceous intercalation. This zone is characterized by a high range of hydraulic conductivity mainly due to widespread occurrence of fractures and intercalation of scoraceous basalt. It is located at the center of the watershed, mainly to the west and east of Lake Beseka. Hydraulic conductivities of the recent and subrecent basalts vary from 7.6 to 766 m/day with arithmetic and geometric mean of 163 and 60 m/day, respectively. Fracture characteristics such as fracture intensity, spacing, interconnectedness, opening and etc., vary within the basaltic aquifers, resulting in variations in the hydraulic conductivity.

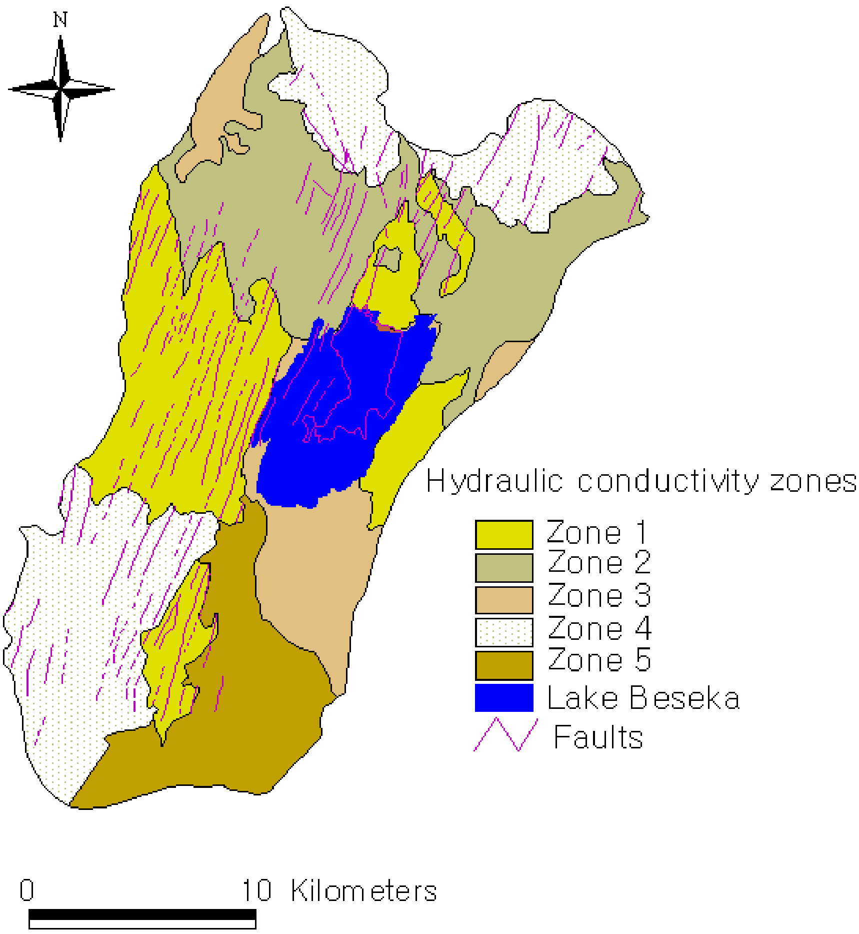


Figure 3.7: Hydraulic conductivity zones in the watershed of Lake Beseka.

Zone 2: Pyroclastic deposit of the Fentale group, which contains fractured and weathered welded tuff, pumice and volcanic breccias, locally mixed with volcanoclastic sediments. This zone occupies the northern part of the watershed, mainly on the lower slopes of Mount Fentale to the north shore of Lake Beseka. The permeability of the pyroclastic deposits ranged from 5.4 to 71.1 m/day with arithmetic and geometric mean of 26 and 17 m/day, respectively. Hydraulic conductivity of the unwelded pyroclast deposit is governed by the extent of weathering, consolidation or welding, while degree of weathering and fracturing play an important role in the permeability of welded tuffs.

Zone 3: Includes lacustrine sediments with subordinate alluvium and volcanoclastic sediments. Lacustrine and alluvial deposits are found south of Lake Beseka around the Abadir farm and in the northwestern corner of the watershed.

However, these sediments become an important part of the aquifer system outside the lake watershed, in the Methara Farm area. Hydraulic conductivity of this zone varies from 0.25 to 20.52 m/day with arithmetic and geometric mean of 5 and 3 m/day, respectively. The hydraulic conductivity of sediments is governed by grain-size distribution, degree of sorting and extent of volcanoclastic intercalations.

Zone 4: Largely composed of slightly fractured Dino ignimbrite, which covers the southeastern part of the watershed. These rocks have some permeability, which is governed by the presence and extent of fractures and weathered permeable layers.

Zone 5: Primarily includes moderately fractured acidic lava flow from deposit of Kone volcanic complex and occupies the southwestern part of the watershed. This zone contains acidic lava flows, which are characterized by low permeability.

4 HYDROCHEMISTRY

4.1 Introduction

The chemical composition of natural water is derived from many different sources of solutes, including gases and aerosols from the atmosphere, weathering and erosion of rocks and soil, solution or precipitation reactions occurring below the surface, and cultural effects resulting from human activities (Hem, 1992). The chemistry of groundwater in the saturated zone is controlled by chemical reaction rate, residence time within the saturated zone, and mineralogy of the rock matrix, where residence time and flow path are determined by factors such as aquifer thickness, permeability and amount of recharge (Griffioen, 2004). These factors combine to different degrees to create diverse water types with compositions that vary with space and time. The objectives of the hydrochemical investigations are to determine the sources, concentration, and fate of dissolved constituents within the physical framework of flow and transport (Griffioen, 2004). The approach divides the water samples into hydrochemical facies (water types), which are groups of samples with similar chemical characteristics that can then be correlated with location (Thyne et al., 2004).

Hydrochemical investigations are conducted to understand the functioning of the hydrogeological system by relating the quality of the groundwater to different processes in the aquifer system. Water chemistry data can be used to infer groundwater flow directions, identify sources and amounts of recharge, estimate groundwater flow rates, and define local, intermediate, and regional flow systems (Anderson and Woessner, 1992). Underlying this hydrochemical approach are a number of assumptions including (1) natural water chemistry is a result of rock-water reaction such as dissolution/precipitation, reactions on aquifer surfaces and biological reactions, (2) distinctive chemical signatures are related to specific sets of reactions, (3) dissolved concentrations generally increase along the surface flow path until a maximum value dictated by mineral equilibrium, and (4) hydrochemical facies are directly related to the dominant processes (Thyne et al., 2004).

In arid or semi-arid areas where hydrological and hydrogeological data are scarce and difficult to access, hydrochemical data can be used as a valuable tool to develop conceptual models of the watershed. In this study, hydrochemical data are used with the main objective of discriminating the origin of different water bodies in the lake watershed, to gain insights on the hydrochemical evolution of the groundwater system in the area, to schematize groundwater flow patterns and to identify hydraulic interactions between Lake Beseka and the surrounding groundwater system. In this context, 38 water samples were collected from wells, Lake Beseka, River Awash, irrigation drainage lines at Abadir farm, hot spring and cold spring sites, and existing hydrochemical data were compiled from previous studies ((WWDSE, 1999) and (Tessema, 1998), Figure 4.1). Charge balance was calculated for chemical data of previous studies, and those with an error of more than 5 % were rejected from the data set. The samples were analyzed for their major ion concentration (Na^{+}, K^{+}, Mg^{+2}, Ca^{+2}, Cl^{-}, SO_4^{-2}, HCO_3^{-}, CO_3^{-2} and F^{-}) in the hydrogeological laboratory of the Technical University Bergakademie, Freiberg, Germany.

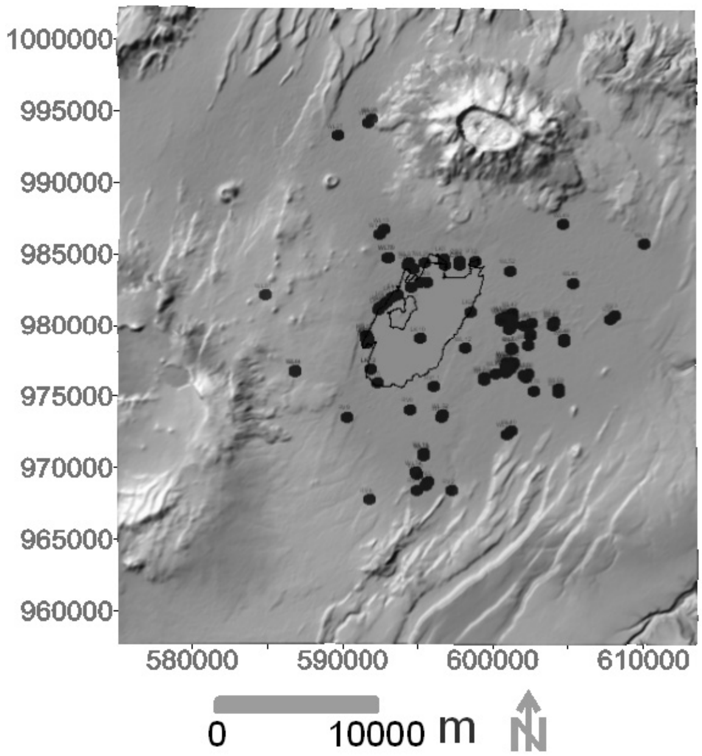


Figure 4.1: Location of hydrochemical sampling points in the study area

4.2 Physiochemical parameters

Physicochemical parameters such as electrical conductivity (EC), pH and temperature were measured in the field. Electrical conductivity is the ability of a substance to conduct electric current (Hem, 1992), and is used as an indirect measure of TDS (mg/l)[2]. Electrical conductivity is a good estimator of salinity due to the fact that TDS is the proportion of the EC in the micromhos (Hounslow, 1995). The EC values of the water bodies in the study area vary significantly. Waters from the lake have higher EC values than those from the surrounding groundwater system. The EC of the lake water varies between 3500 and 7400 μS/cm laterally. The northwestern and southwestern parts of Lake Beseka are characterized by the highest and lowest EC values, respectively. The highest EC value of Lake Beseka was measured at its deepest part, which could be related to the existence of remnant old saline water. On the other hand, the lowest EC values were measured at the part of the lake where numerous less-saline hot springs emerge.

Electrical conductivity of groundwater increases from the western part of the watershed toward the eastern part along the groundwater flow direction, which is also supported by groundwater level map (Figure 3.3). Generally, dissolved solids concentrations in groundwater increase along flow paths, from the surface to the saturated zone and through the aquifer, due to the dissolution of minerals (Freeze and Cherry, 1979). The groundwater system in the western part of the watershed is characterized by relatively low EC values ranging between 1500 and 2000 μS/cm. This can be related to fast flushing of the groundwater, which has a low subsurface residence time. Groundwater in the eastern part of the watershed has relatively higher EC values, with exceptionally high values around Addis Ketema, in the northeastern part. This saline zone is characterized by highly mineralized groundwater with EC values up to 8500μS/cm. This part of the watershed is considered the discharge zone, where groundwater is characterized by high subsurface residence time. Specifically, this mineralized zone has the lowest groundwater level in the lake watershed, and therefore, the groundwater might reside in the system for a considerable period of time. This long residence time could facilitate rock-water interactions causing higher salinity of the water. Previous studies explain the occurrence of the mineralized zone as remnant lake water intrusion to the groundwater system in the past, when the lake was more saline.

[2] Dissolved solid content in water calculated as the total mass of ions including SiO_2

The water from the River Awash is used as irrigation water for the Abadir sugarcane plantation. It is characterized by the lowest EC values (250- 400 µS/cm) in the study area.

The temperature of the lake is mostly about that of the ambient air with exceptional locations where hot springs are either submerged or emerging. In the southwestern corner and western perimeter of the lake, hot springs are abundant, and the lake temperature is higher and can reach up to 35 ^{o}C. The temperature of the hot springs ranges from 38 to 43 ^{o}C. The groundwater system in the western and southeastern part of the watershed is characterized by high temperatures, comparable to those of the hot springs. The wells at Tedecha tera (44^{o}C), Illala (40.8^{o}C), Meldiba (41^{o}C) and Abadir (37^{o}C) are examples of this. These high temperatures in the groundwater system are related to the location of the area in a tectonically and volcanically active region. The temperature of groundwater is a response to the movement of heat from the Earth's interior (Heath, 1983). The Ethiopian rift valley has a positive thermal anomaly (UNDP, 1973), which could be a possible source of heat to a circulating groundwater system at relatively shallow depth. Since rifts are characterized by intense tectonic activity, with crustal thinning, astenosphere uprise and associated volcanism, they are sites for thermal regional anomalies (Elc electoconsult, 1987).

4.3 Cluster Analysis

Hierarchical multi-element cluster analysis (HCA) was used to classify water samples of the study area into different groups. Cluster analysis refers to a set of techniques designed to classify observations so that members of the resulting groups are similar to each other and distinct from the other groups (Swanson et al., 2001). The ability of cluster analysis to classify groundwater chemistry into coherent groups that may be distinguished in terms of aquifer type, subsurface residence time, and degree of human impact on water chemistry provides a good opportunity to understand groundwater geochemical evolution among the different groups or subgroups (Kebede, 2004).

The HCA is designed to group water samples based upon similarities in multidimensional space. Closeness or similarity between samples was measured by using the distance between samples in n dimensional space of variables so that similar items can be agglomerated by using linkage methods. Water samples from previous and present studies were found homogenous, and Statgraphics centurion XV software was used for the cluster analysis. The squared Euclidian distance between Na^{+}, K^{+}, Mg^{+2}, Ca^{+2}, Cl^{-}, $SO4^{-2}$, $HCO3^{-}$, $CO3^{-2}$, pH and EC values were used to calculate similarity among samples with the furthest neighbor method for linkage. The raw chemical data were log transformed and standardized so that variables had equal weight. Water samples were first placed in separate clusters and then different clusters aggregated hierarchically based on the maximum distance measured between any member of one cluster and member of the other. This technique is useful for discriminating between samples (Thyne et al., 2004), which can ultimately assist in the recognition of potentially meaningful patterns (Swanson et al., 2001).

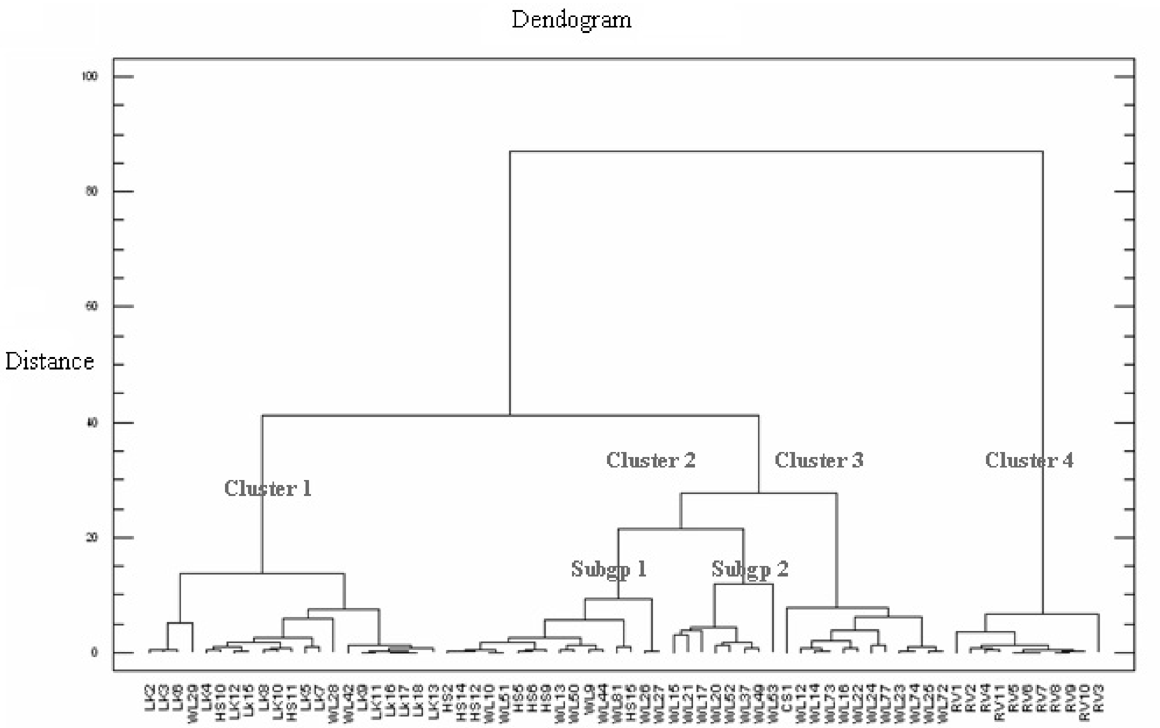


Figure 4.2: Dendrogram of water samples from the study area

The HCA makes it possible to discriminate water bodies of different origins, to understand surface-groundwater interaction, and to identify geochemical evolution and subsurface residence time of groundwater. The cluster analysis identified four major hydrochemical facies in the water samples; these facies are characterized by distinct water chemistries (Figure 4.2). Each cluster group is plotted on a Piper diagram (Figure 4.3), which shows the relative contribution of major cations and anions on a milliequivalent basis to the total ion content of the water. Sodium (Na+) is the dominant cation in all water types, except cluster 4, thus classification is based mainly on anion species and their relative proportion with respect to each other.

Cluster 1

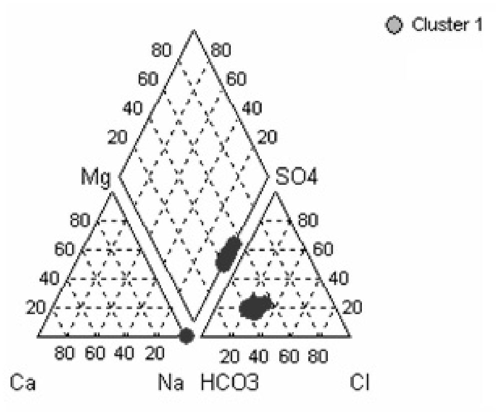
The first hydrochemical cluster is typified by a *Na-*HCO_3*-Cl* and *Na-*HCO_3*-*CO_3*-Cl* type of water. Sodium is the major cation, and bicarbonate is the dominant anion with a concentration range of 61 to 70 % of the total anions. Most of the water samples in this cluster have pH values greater than 9, and at this pH some of the bicarbonate changes to the carbonate forms (Fetter, 2001). In this group, chloride is found in considerable amounts reaching up to 23 % of the total anions. This hydrochemical cluster has typical EC values between 3696 and 7494 µS/cm. Samples from Lake Beseka, a few hot spring samples and well samples from the Addis Ketema area are agglomerated in this cluster.

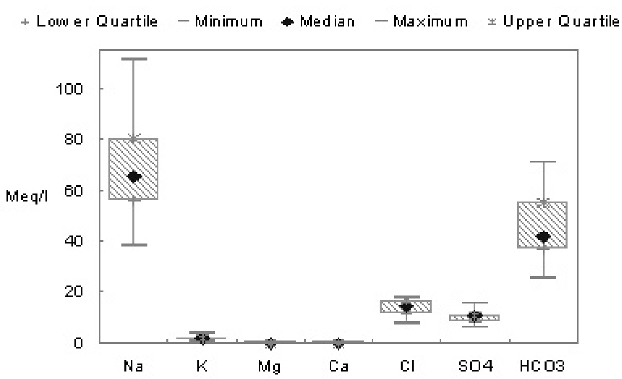
Cluster 2

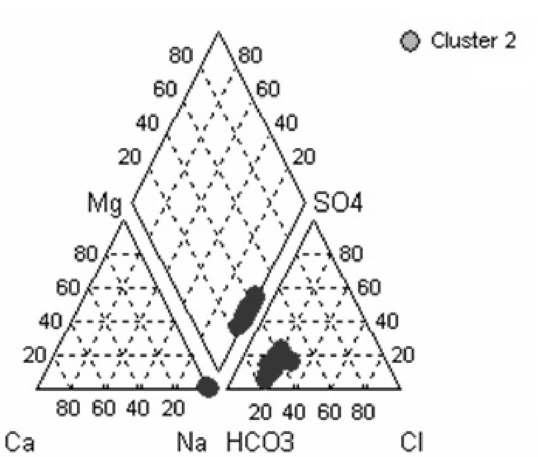
The second cluster is identified by a typical sodium-bicarbonate *Na-*HCO_3 type of water. Sodium and bicarbonate are the dominant cation and anion, and the concentration of bicarbonate varies between 61 to 77 % of the total anions. Groundwater samples from the western part of the watershed and around the lake, and from hot springs that emerge at the southwestern corner of the lake are grouped in this cluster. As can be seen from the dendogram, two subgroups can be observed in cluster 2. The first subgroup (Subgroup 1) is characterized by relatively low saline groundwater with EC values between 1548 and 1990 µS/cm. The second subgroup (Subgroup 2) is more saline groundwater with EC between 1800 and 3814 µS/cm. The Subgroup 1 type of water is observed in wells and hot springs located in the western part of the study area. On the other hand, Subgroup 2 type is found in the wells around the lake, especially in the north and east of Lake Beseka. The salinity of groundwater in Subgroup 2 could be due to the interaction of the lake with the surrounding groundwater system and / or longer subsurface resident time of groundwater system.

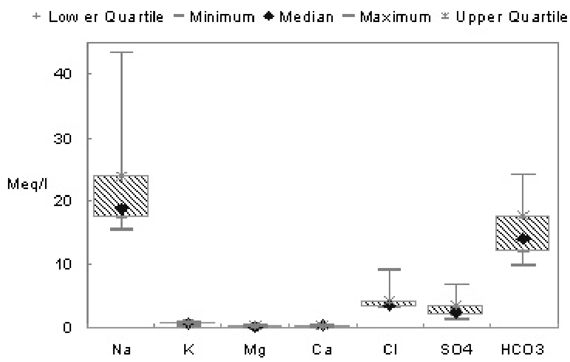
Cluster 3

The third hydrochemical cluster consists of groundwater with *Na-HCO_3-SO_4-Cl* and *Na-HCO_3-Cl* types of water. Bicarbonate is the principal anion with concentrations ranging between 41 and 51 % of the total anions; however sulfate and chloride exist in significant amounts. In this cluster, the level of chloride and sulfate is higher and can reach values as high as 33% and 35% of the total anions, respectively. The groundwater systems in this cluster are located in the eastern part of the study area. There are two subgroups in this cluster. The first subgroup type of groundwater is found in the southeastern part of the basin around Abadir 3rd and 4th camp, and the Abdella tsebele cold spring areas.









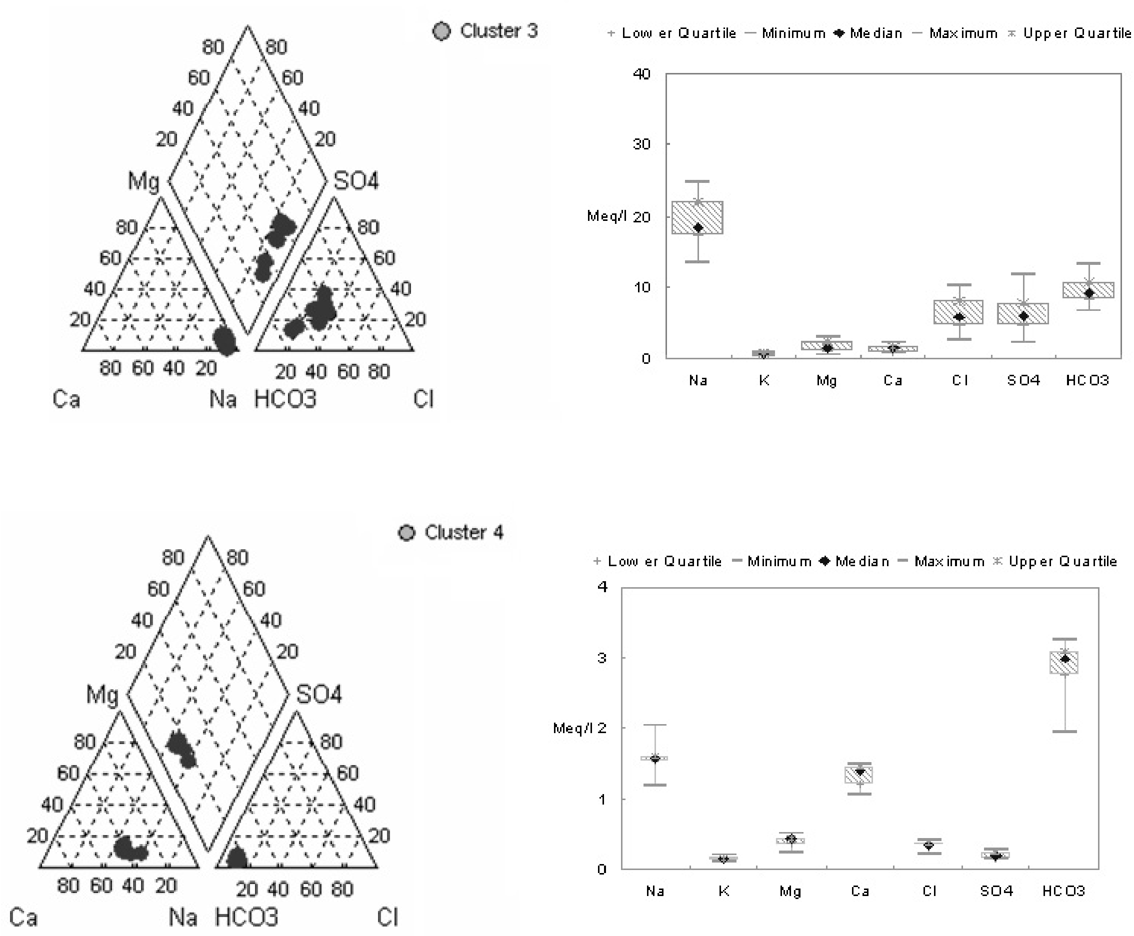


Figure 4.3: Piper plot of hydrochemical clusters and their statistical summary: Cluster 1 mainly contains water samples from Lake Beseka, Cluster 2 is typical of the groundwater system in the western part of the watershed and the hot springs, Cluster 3 contains samples from the groundwater system in the eastern part of the watershed, Cluster 4 contains irrigation water samples from River Awash.

This subgroup is characterized by 29% Cl and 29% SO_4 of the total anions, and by EC values between 2006 and 2980 µS/cm. Sub group 2 comprises the groundwater system located east of Lake Beseka around Gofa Sefer and Haro Adi area, which is characterized by low EC values ranging between 1500 and 1800 µS/cm. The groundwater system in this subgroup has high Cl and SO_4 with 31 % and 18% of the total anions, respectively.

Cluster 4

The fourth hydrochemical cluster is typified as a *Na-Ca-HCO3* type of water, where sodium and calcium are major cations. Bicarbonate is the major anion with a concentration range of 82 to 87 % of the total anions. Samples from River Awash and the irrigation drainage line in Abadir farm are grouped in this cluster. This water type has been introduced to the study area since water from River Awash began to be used for irrigation of the Abadir sugar cane plantation.

4.4 Groundwater chemistry

The majority of groundwater samples falls into the $Na\text{-}HCO_3$ type in the Piper plot, as the dominating cation is sodium (Na^+), while bicarbonate (HCO_3^-) is the principal anion. The $Na\text{-}HCO_3$ type of water is the prominent groundwater type in the Ethiopian rift valley, which could be related to the widespread occurrence of acidic volcanic rocks. The hydrolysis of silicate minerals in volcanic rocks with the presence of CO_2 from atmosphere, soil zone and deep source has resulted in high amounts of Na^+ and HCO_3^- in the groundwater systems:

$$CO_2 + H_2O + \text{Na, K silicates} \Leftrightarrow HCO_3 + \text{Na, K} + \text{H-silicates (UN, 1973)}$$

The above reaction is facilitated by CO_2 out-gassing from the magmatically active section of the Ethiopian rift system. Gizaw (1996) indicated that CO_2 out-gassing from magamatic sources is the principal reason for the high concentration of bicarbonate in the groundwater of the rift. Kebede et al. (2007) also claimed high bicarbonate concentrations in the groundwater system of the rift due to the hydrolysis reaction aided by CO_2 input from deeper sources. Although bicarbonate is the dominant anion in the groundwater system of the study area, its concentration decreases, and it is replaced by chloride and sulfate anions along the groundwater flow direction i.e., from west to east. Two hydrochemical profiles were taken to show how bicarbonate-dominated groundwater has hydrochemically evolved in its flow path toward the eastern part of the watershed (Figure 4.4).

Geochemical reactions along groundwater flow paths can lead to regional variations in water composition that evolve in the direction of flow (Lynn et al., 2005).

Eberts and George (2000) explained how predominant major ions in groundwater can be used as indicators of important chemical and hydrologic processes in an aquifer system. With reference to the work of Chebotarev (1955), they indicated that the general evolutionary sequence of groundwater of the dominant anion is from HCO_3 to SO_4 and *Cl*, with increasing distance along the flow path and an increasing age of water. They also specified that this anion sequence corresponds to an increase in dissolved solids concentration.

Water from the hot springs and wells in the western and northeastern part of the watershed are the *Na-HCO3* type of water. In the western part of the watershed, groundwater is characterized by relatively low salinity and high bicarbonate concentration, ranging from 61 to 76 % of the total anions. Recharge and flushing of groundwater by CO_2-rich meteoric water and the fast circulation of groundwater with low subsurface residence time govern groundwater chemistry in this area. Water containing predominantly bicarbonate is generally present in areas of active groundwater flushing (Eberts and George, 2000). The existences of tensional faults with high vertical permeability facilitate the recharge process in the western part (Elc electoconsult, 1987).

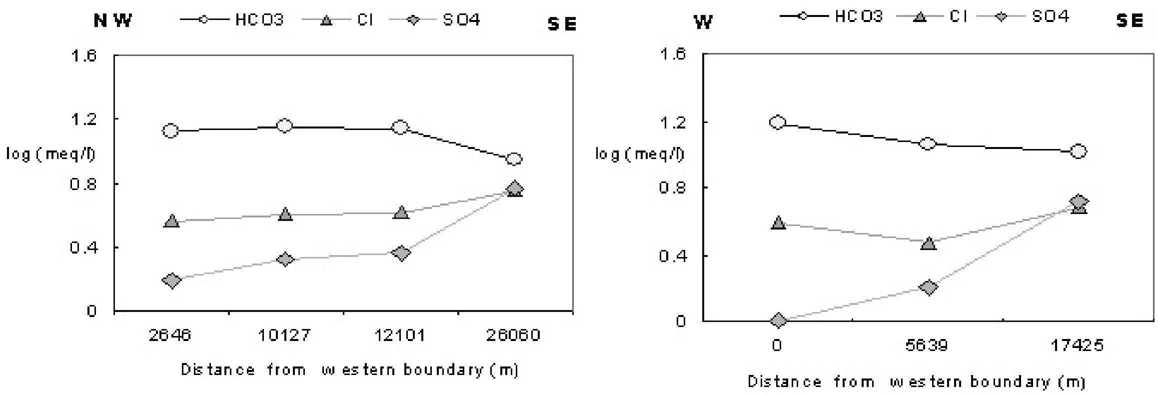


Figure 4.4: Hydrochemical profiles from the northwestern to southeastern (left) and from western to southeastern (right) part of the watershed.

Water that predominantly contains sulfate is present where groundwater circulation is less active, and predominantly chloride where groundwater flow is very sluggish and flushing of the aquifer is minimal (Eberts and George, 2000). Well samples from the eastern part of the watershed have significant sulfate and chloride concentrations, and are characterized as *Na-HCO3-SO4-Cl* or *Na-HCO3-Cl* types of water. Concentrations of bicarbonate range between 41 and 51 % of the total anions. This groundwater system is more saline, and the water chemistry has evolved in such a way that the proportion of chloride and sulfate increased at the expense of bicarbonate to become a significant component of the total anions. This is typical for discharge areas where groundwater has undergone a significant degree of geochemical evolution and is characterized by a longer subsurface residence time with pronounced rock-water interactions.

The wells which are located east of Lake Beseka, around the Gofa Sefer areas have relatively low saline groundwater, while it is still characterized by a *Na-HCO3-Cl* type of water. This could be related to the existence of geological structures in the area, i.e., faults that facilitate local recharge of meteoric water to periodically flush the groundwater system.

Wells around Addis Ketema have high chloride content and are categorized in cluster 1, where lake samples are also grouped. This area is characterized by very high EC and is identified as a zone of highly saline groundwater. As the groundwater is more saline than the current lake chemistry, previous studies suggest that it is old saline lake water intruded into the groundwater system. However, the salinity could be related to the relative position of the mineralized zone in the watershed, i.e., in the discharge zone, where the groundwater had undergone significant hydrochemical evolution, and is characterized by longer subsurface residence times with pronounced interaction of the groundwater with volcanic host rocks (Figure 4.5).

The fluoride content of the groundwater in the study area is very high, with average concentrations of 7.7 and 9.9 mg/l for the western and the eastern part of the watershed, respectively. This is a common feature of most groundwater systems in the Ethiopian rift. Gizaw (1996) discussed the widespread occurrence of acidic volcanic rocks, low calcium (Ca*)* content, high CO_2 pressure and geothermal heating in the rift as the major reasons for the high fluoride (F) content in rift waters. Chernet (2001) explained that thermal waters from volcanic rocks of the MER are characterized by positive alkalinity residual of calcite, a condition that enhances fluoride mobility. With increasing temperature, solubility of calcite decreases and tends to precipitate and *Ca* tends to be fixed in the form of calcite *(*$CaCO_3$*)* or any other Ca-bearing minerals.

Consequently, there would be not enough Ca^{2+} to fix the F^-, which was previously in equilibrium with CaF_2 (fluorite), thus F becomes mobile in the media.

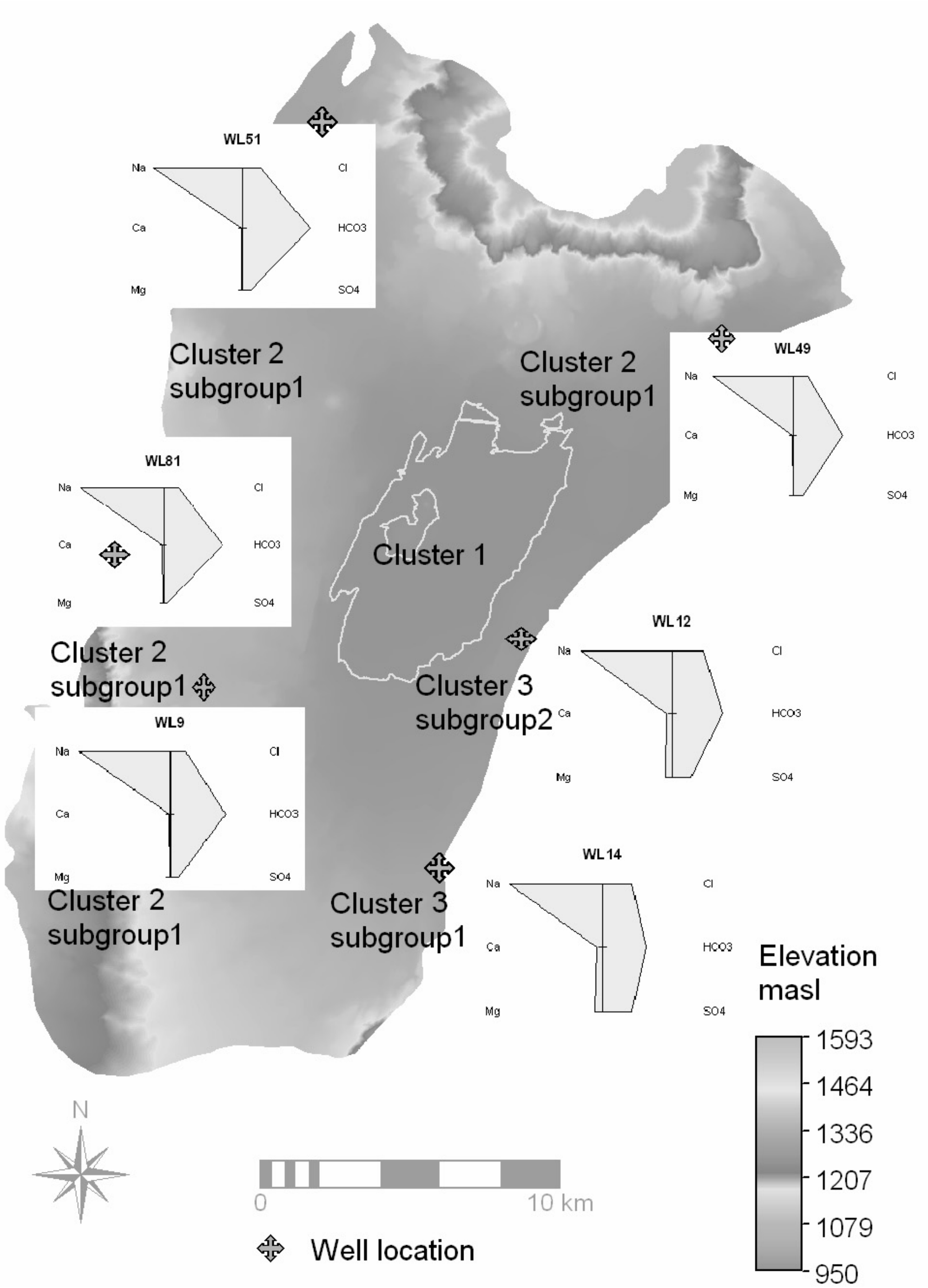


Figure 4.5: Groundwater hydrochemical evolution in the watershed of Lake Beseka: As indicated in the stiff plot, bicarbonate dominated groundwater system changes to bicarbonate-sulfate-chloride groundwater system when one goes from the western part of the watershed to the eastern part.

Prolonged intake of water causes dental and skeletal fluorosis when the fluoride content exceeds the maximum concentration of 1.5 mg/l (WHO, 1984). Dental and skeletal fluorosis is a common health problem in the Ethiopian rift valley (Gizaw, 1996; Chernet, 2001; Ayenew, 2007). The degree of this adverse effect is governed by the fluoride concentration in the water, the quantity of water consumed, the length of consumption time, and temperature, as well as the nutritional and economic status of the community (Gizaw, 1996). In the rift valley towns like Nazareth, Addis Ketema and Methara, boreholes have recently been abandoned due to water quality problems and replaced by surface water supplies.

4.5 Lake chemistry

The Ethiopian rift valley lakes are characterized by typical hydrochemical features. Most lakes are alkaline and have a sodium bicarbonate type of water, are characterized by different degrees of salinity and alkalinity. This hydrochemical variation within the lakes is controlled by the hydrological, hydrogeological and geomorphological setting of the lakes (Alemayehu et al., 2005). The hydrological setting, especially inflow-outflow ratio, plays an essential role in determining the hydrochemical characteristics. Lakes with a substantial amount of outflow in either surface or subsurface form have a lower alkalinity and salinity compared to terminal lakes. Water input-output relationships are the dominant feature of the status in the salinity series of the rift lakes (Wood & Talling, 1988).

For instance the lakes Ziway, Awassa and Abaya are fresh lakes, as these lakes discharge a significant amount of surface and / or groundwater. The degree of freshness within these lakes is governed mainly by the amount of outflow with respect to their total inflow. On the other hand, terminal lakes like Beseka, Abiyata and Shalla do not have substantial outflows; consequently they are characterized by higher salinity and alkalinity. In terminal lakes, evaporation is the only way for water outflows, and as a result, solutes concentration increases. Thus, terminal lakes have higher salinity due to progressive accumulation of solutes in response to evaporation through time. The degree of salinity is controlled by the time since the system became closed, i.e., the existing time of the system.

Two terminal lakes, Lake Beseka and Abiyata have revealed substantial water chemistry changes in the last few decades. They have shown strong temporal variation in their chemistry, while other lakes show little or no change (Alemayehu et al. 2005). The Lake Abiyata ionic concentration and salinity has increased (Kebede et al. 1994). The significant change in the water chemistry of Lake Abiyata has been observed since the late 1980s related to the large-scale withdrawal of water for soda ash (*$NaCO_3$*) production (Ayenew, 2004). Lake Beseka has also exhibited temporal hydrochemical changes, but the trend is completely opposite to that of Lake Abiyata. A comparison of chemical data with previous records showed that a 10-fold dilution of total ionic concentration occurred over 30 years in Lake Beseka and an about 3-fold increase occurred over 65 years in Lake Abiyata (Kebede et al., 1994). The water chemistry of Lake Beseka has been diluted, and its EC has reduced by 13-fold within 45 years.

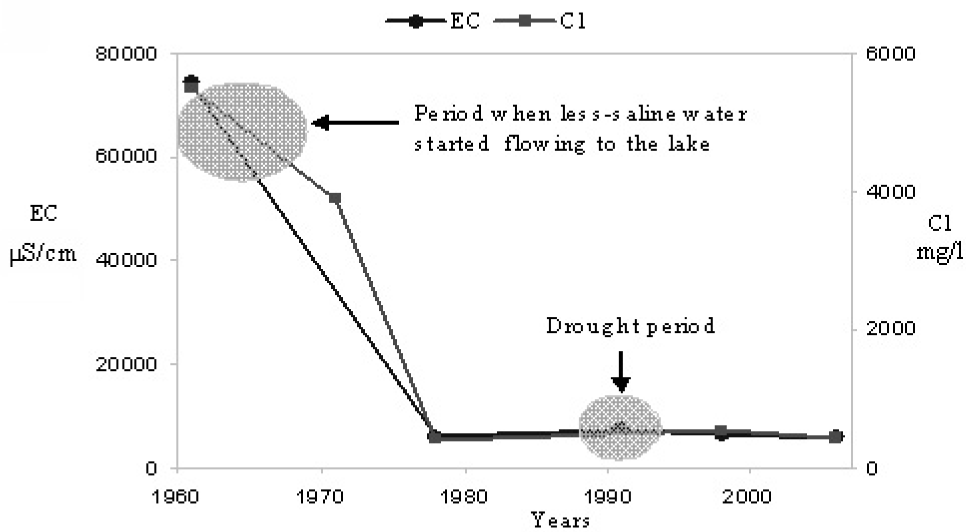


Figure 4.6: Temporal variations of EC and chloride concentration of Lake Beseka, (from Halcrow (1978), Kebede et al. (1994), Ayenew (1998), WWDES (1999) and this study (2006)).

The water chemistry of Lake Beseka has been analyzed by different authors since 1960s. Although sampling and analysis were done in an irregular manner, it is evident from previous studies that Lake Beseka used to be more saline and alkaline than today. As cited in Ayenew (1998), EC and pH of the lake was measured by Talling & Talling in 1961 to be 74170 µS/cm and 10, respectively. Halcrow (1978) compared total dissolved solids (TDS) and noted that TDS in 1971 was 14 times more saline than that measured in 1978. Electrical conductivity of the lake was 5665 µS/cm in 2006, which is about 10 % diluted compared to the EC value measured in 1998.

Electrical conductivity and chloride content of Lake Beseka have dramatically diminished between the 1960s and 1980s (Figure 4.6). This gives a clue to the time when relatively less-saline water started flowing into the lake causing changes its hydrochemical properties. After the 1980s, both EC and chloride content of the lake showed a slight increment, which could be related to either declining water inflow or increased rate of evaporation from the lake. In either way, the situation could be related to the wide-scale drought incidence in the mid 1980s in Ethiopia.

Electrical conductivity and chloride values from the recent years illustrate that Lake Beseka is still being diluted by relatively less-saline water inflows. However, the rate of dilution is now significantly lower, compared to the period when the lake started to expand. This is due to the fact that the volume of the lake has significantly increased since then, and lake water has been concurrently affected by evaporation. Thus, the dilution effect is suppressed, and it is not as noticeable as it was in early times.

4.6 Lake - groundwater relation tracing

Though Lake Beseka is affected by temporal hydrochemical variations, sodium and bicarbonates remain the dominant ions. Lake water is characterized by low bivalent cations, calcium and magnesium, which is also the case for the surrounding groundwater system. The groundwater system in the western part of the watershed plots on the Piper diagram along the same line with the lake but with different salinity values (Figure 4.3, cluster 1 and cluster 2). The lake is plotted on a more saline locus than the groundwater. Lake Beseka and the groundwater system in the western part of the watershed have the same hydrochemical properties, but the lake chemistry has evolved with respect to the groundwater in response to evaporation.

The lake is a stagnant water body, which is continuously losing water to the atmosphere through the process of evaporation. It is believed that evaporation accounts for the dominant water flux in most rift lakes, and plays an important role in their unique hydrogeochemical signature (Alemayehu et al., 2005). As a consequence of evaporation, concentrations of dissolved solutes in Lake Beseka are progressively enhanced in the remaining water, making the lake more saline compared to the concentration of its inflows.

The compositional similarity of Lake Beseka with the hot springs and western wells can also be observed on the Schoeller plot (Figure 4.7, right). Hot springs and groundwater from the western wells plot on the same line on the Schoeller plot, and both show a similar ionic trend to that of Lake Beseka. This indicates that hot springs that emerge at the southwestern corner of the lake and groundwater that flows from the western part of the lake have an almost identical composition, typified as the *Na-HCO_3* type of water. Additionally, they have a similar relative proportion of cations and anions to that of Lake Beseka, which is characterized by the *Na-HCO_3-Cl* type of water. Sodium and bicarbonate are the dominate cation and anion in both groundwater and lake water, though the importance of chloride and the amount of total dissolved solids increase in the lake water. This is due to the fact that Lake Beseka has a similar hydrochemical signature to that of groundwater system, which flows from the western part of the watershed, but lake salinity has increased with respect to the groundwater chemistry due to evaporation. This evaporation effect is demonstrated on the Schoeller plot as the vertical concentration gap between the groundwater and the lake water.

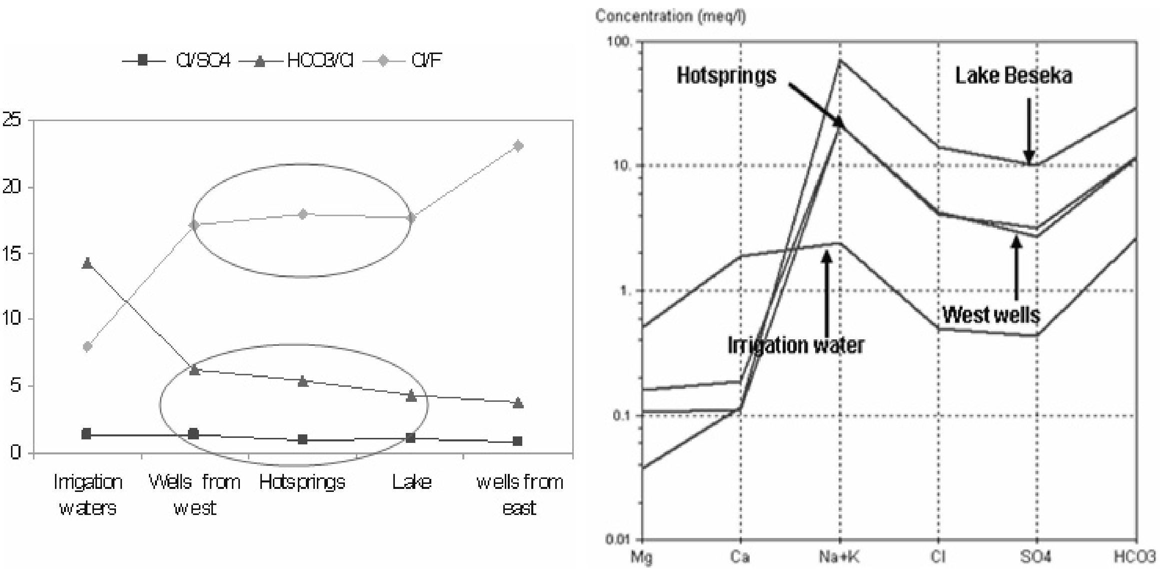


Figure 4.7: Plots of ionic ratios (left) and Schoeller plot of different water bodies in the study area (right).

On the other hand, irrigation water exhibits different hydrochemical signals, being dominated by two cations species (sodium and chloride) and bicarbonate as the dominant anion. The lake is characterized by the *Na-*HCO_3*-Cl* type of water, while irrigation water is typified as a *Na-Ca-*HCO_3 type of water. This hydrochemical difference is clearly evident on the Schoeller plot, where irrigation water plots differ from the lake in terms of ionic trend and concentration. This fact is also clear on the Piper plots (Figure 4.3, cluster 1 and cluster 4), and also on the ionic ratio plot, (Figure 4.7, left). This indicates that Lake Beseka has a similar hydrochemical signature to that of the groundwater that flows from the western part of the watershed rather than to that of irrigation water.

Ionic ratios of chloride-sulfate, bicarbonate-chloride and chloride-fluoride were used to compare the water chemistry of Lake Beseka with that of the surrounding groundwater system and irrigation water used at Abadir farm. The ionic ratios were used, based on the assumption that evaporation increases the concentration of individual ions and total dissolved salts in the lake water, but that the relative proportion of ions remains constant, unless the lake attains a saturation level with respect to the given salts. Analysis of ionic ratios further strengthen the fact that Lake Beseka has comparable hydrochemical properties to that of the groundwater in the western part of the watershed and the hot springs than the irrigation water at Abadir farm (Figure 4.7, left).

The hydrochemical evolution of groundwater is also evident from the trends of the ionic ratios from the western wells through Lake Beseka to eastern wells (Figure 4.7, left). The bicarbonate-chloride ratio shows a declining trend along this line, while an increasing trend is observed for the chloride-fluoride ratio. This is because chloride increases at the expense of bicarbonate along the groundwater flow direction, and both chloride and fluoride are expected to increase for waters with longer residence time (eastern groundwater). The hydrochemical analysis shows two important features of the groundwater dynamics of the lake watershed. The first is that groundwater flows from the northwestern and southwestern part of the watershed toward the east, and discharges into the lake in the form of submerged and emerging hot springs. This groundwater governs not only the hydrochemical properties of Lake Beseka, but also the hydrological characteristics of the lake, as most of these dissolved solutes in Lake Beseka have been conveyed to the lake via the groundwater inflows.

5 ISOTOPE HYDROLOGY

5.1 Introduction

Isotope techniques are effective tools for obtaining a variety of hydrological information such as origin of groundwater, determination of its age, velocity and direction of flow, interrelations between surface water and groundwater, possible interconnections between different aquifers, aquifer characteristics, etc. (Aggarwal et al., 2004). The isotopes of oxygen and hydrogen provide useful information for tracing the source of waters, as they are constitute of the water, and conservative in aquifers. Stable isotopes in water (^{18}O and ^{2}H), which are affected by meteorological processes, provide a characteristic fingerprint of their origin (Clarke and Fritz, 1997). Tracing groundwater by means of environmental isotopes offers unique information on the origin and movement of groundwater and on physical processes such as evaporation and isotopic exchange processes (Mook, 2000). Water is tagged with isotopic signals, which vary according to history and pathway through the hydrologic cycle. This is the key feature of stable isotopes, and explains why they are important to identify the origin of groundwater.

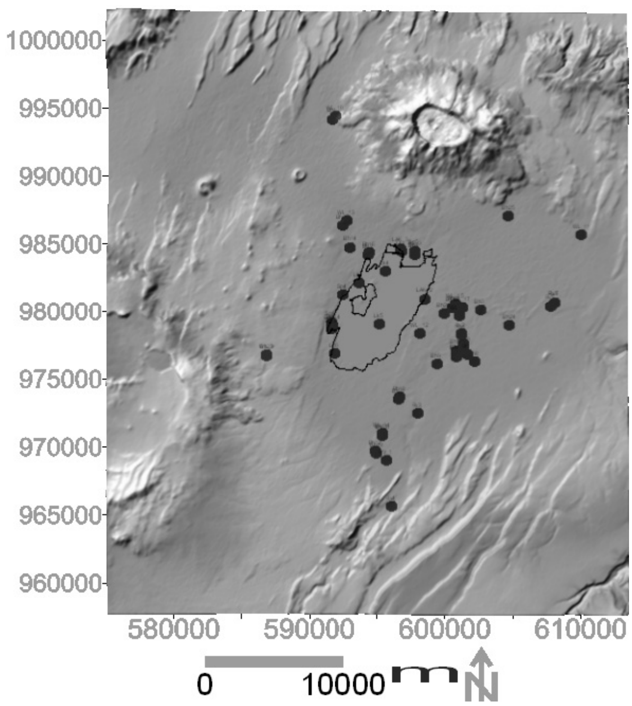


Figure: 5.1 Location of isotope sampling points in the study area

In this study, stable isotopes of water (^{18}O and ^{2}H) and limited tritium data were used to investigate groundwater origins and to identify groundwater relation with the water bodies in the region. Collection of 39 water samples were conducted from wells, Lake Beseka, River Awash, irrigation drainage lines at Abadir farm, hot spring and cold spring sites and analyzed for stable isotopes and limited tritium concentration. Samples were analyzed in the laboratory of UFZ (Helmholtz-Zentrum für Umweltforschung of Leipzig, Germany) for isotopic concentrations with the help of hydrogeological laboratory of the Technical University Bergakademie, Freiberg, Germany. Data from previous works (WWDSE, 1999 and Tessema, 1998) were compiled and included in the dataset for further analysis (Figure 5.1). Isotopic data for the Addis Ababa rainfall station was downloaded from the IAEA database (http://isohis.iaea.org).

5.2 Theoretical background

Variations in stable isotope ratios of natural compounds are governed by chemical reactions and phase changes due to the energy difference between chemical bonds involving different isotopes of an element (Aggarwal et al., 2004). Such energy differences are the result of relative mass differences between isotopes. A variation in the number of neutrons within an element results in different masses or atomic weights of the elements, and this mass difference controls the mobility and binding energy of isotopic molecules. Heavier isotopic molecules are characterized by lower mobility and higher binding energy than lighter ones of the same composition (Mook, 2000). This leads to the process of mass-dependent fractionation[3] or partitioning of isotopes during certain physical or chemical processes. Isotopes of H and O combine to form isotopically different water types, which react in a different way within the hydrologic system. Upon evaporation, for instance, lighter species of water tend to move faster and concentrate toward the vapor state and result in more heavier isotopic species ($H_2{}^{18}O$) in the remaining water.

[3] Fractionation is any process that causes the isotopic ratios to differ from one phase to the other

Isotopic composition is often expressed as the ratio of heavy to light isotopes. The calculation of the ratio is a commonly used for expressing an isotopic composition in which the isotopic ratio of water molecules is compared to the isotopic ratio of standard water. For convenience, the measurements are reported as ratios in δ notation as per mil (per thousands, ‰) relative to the standard. Hydrogen and oxygen isotopes are compared to the reference of the Vienna Standard Mean Ocean Water (VSMOW):

$$\delta^{18}O(permil) = \left[\frac{(18O/16O)_{Sample}}{(18O/16O)_{SMOW}} - 1 \right] * 1000 \text{ and } \delta D(permil) = \left[\frac{(D/H)_{Sample}}{(D/H)_{SMOW}} - 1 \right] * 1000$$

A positive δ-value signifies that the sample is 'enriched' in heavy isotopes compared to the standard, VSMOW, while negative δ-value indicates depletion in the heavy isotopes.

Tritium (^{3}H) is a radioactive isotope of hydrogen with a half-life of 12.43 years. It is produced naturally in the upper atmosphere from nitrogen by neutron impacts of cosmic radiation: $^{14}N + n \Rightarrow {}^{3}H + {}^{12}C$. Following atmospheric detonation of thermonuclear bombs between 1952 and 1962, a large amount of tritium was introduced to the atmosphere and entered the hydrological cycle through precipitation. This bomb-produced tritium is used as an indication of recent groundwater replenishment and for determining the age of recent groundwater recharges. Presence of tritium in groundwater provides evidence of active groundwater recharge (Clarke and Fritz, 1997). However, its use has been limited by its uneven global distribution, as tritium fallout is less pronounced in the southern hemisphere. Tritium concentrations are expressed as tritium units (TU), where 1TU = 1 atom of ^{3}H per 10^{18} atoms of ^{1}H.

5.3 Isotopic composition of precipitation

Before using isotopes for tracing groundwater or groundwater-lake interactions, it is important to understand how isotopic composition of meteoric water changes with time and space. Kebede (2004) indicated that the $\delta^{18}O$ and δD composition of Ethiopian meteoric waters, and their spatial and temporal variations are influenced by continental or global factors as well as by regional or local factors. Global factors include seasonal drifting of the ITCZ and associated shifts in the source of moisture, while local factors include altitude effect[4], evaporation effect, land-surface moisture feedback, etc.

Rain forming vapor that evaporates from an ocean is isotopically depleted with respect to the ocean water by about 12-15‰ in $\delta^{18}O$ and 80-120‰ in deuterium (Aggarwal et al., 2004). Once the vapor has evaporated from oceanic surfaces, progressive raining out proceeds through successive cooling and condensation with production of precipitation in response to decreasing temperature when the vapor mass moves toward continental regions. Precipitation that is derived from the vapor is always enriched in its isotopic composition relative to the vapor. Consequently, the progressive rainout process from the coast causes increasingly lighter rains along its way to continental areas. As a general rule, $\delta^{18}O$ becomes more negative the further the rain is moved from the main source of the vapor in the equatorial regions (Mook, 2000). Because the degree of condensation of a vapor mass depends on temperature, a relation between isotope composition of precipitation and its temperature of formation should be expected: as the formation temperature decreases, the δ-values of precipitation decrease (Dansgaard, 1954; as cited in Aggarwal et al., 2004).

Craig (1961) observed that the isotopic compositions of meteoric waters in the global hydrological cycle behave in predictable ways. He found that δ^2H and $\delta^{18}O$ correlate on a global scale, and plot on a line with a slope of 8 and an intercept of 10. This line is called the Craig line or global meteoric water line (GMWL). It is derived from averages of many local or regional meteoric water lines, which differ from the global line due to varying climatic and geographic parameters (Clarke and Fritz, 1997). The slope of the GMWL expresses the equilibrium fractionation factor, which is eight times greater for oxygen than for hydrogen. The intercept is commonly known as deuterium excess (d-excess), which is calculated as: $d = \delta^2H - 8\delta^{18}O$ and averaged at the value of 10 on a global scale. At regional or local scale, the d-excess varies due to variations in humidity, wind speed, sea surface temperature (SST) during primary evaporation (ibid). Thus, the d-excess can be used to identify the origin of moisture.

[4] Altitude effect: temperature related effect when isotopic composition of precipitation changes with altitude of the terrain to become more and more depleted at higher elevations (Mook, 2000).

Meteoric waters do not always plot on the GMWL, as they are affected by various local and regional factors such as source of moisture, extent of evaporation during rainfall, altitude and amount effect[5]. The line that is formed by plotting δ^2H and $\delta^{18}O$ for specific regions is called the local meteoric water line (LMWL). As it is not always possible to monitor precipitation over representative periods of time, meteoric water lines must be taken from the closest available station (Clarke and Fritz, 1997). The isotopic composition of Addis Ababa precipitation has been monitored for 37 years, and it is the nearest monitored station for the Lake Beseka area. The station is located under the same moisture sources as the study area, thus the isotopic concentrations of the rainfall are used to set the local meteoric water line. Short-term $\delta^{18}O$ and δ^2H monitoring of the MER rainfall shows comparable isotopic composition and seasonal variation with Addis Ababa rainfall (Kebede, 2004).

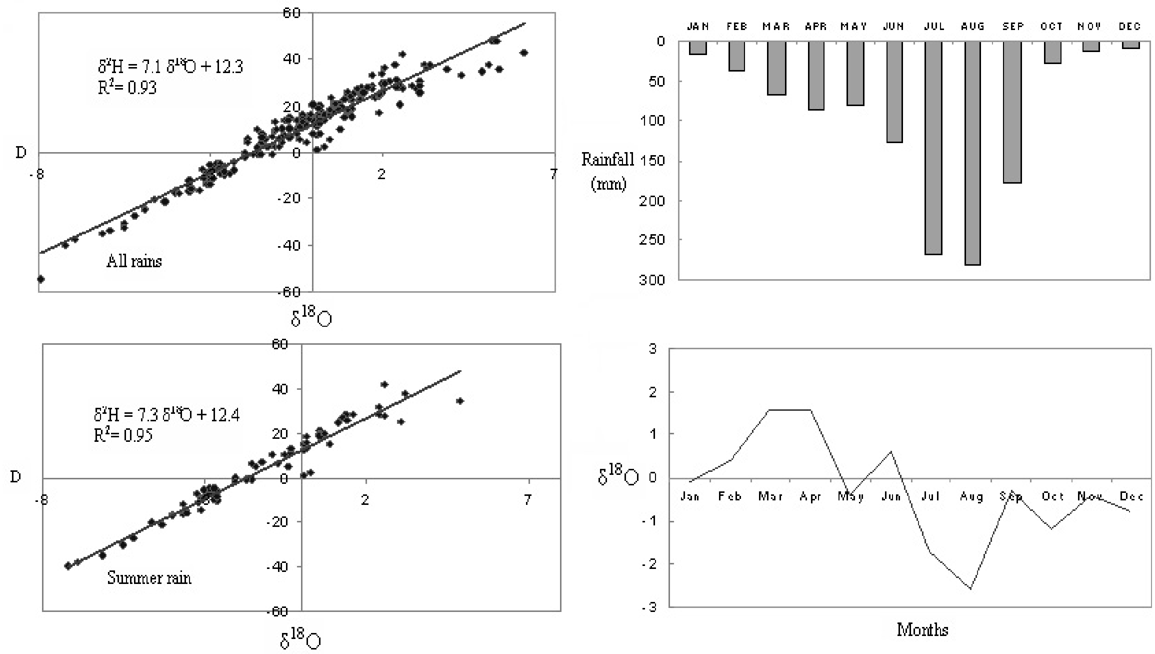


Figure 5.2: δ^2H - $\delta^{18}O$ plot of Addis Ababa rainfalls (January 1965-December 2002), all rainfalls (left top), summer rainfalls (left bottom), long-term average of monthly rainfall amount of Addis Ababa station (right top) and its seasonal isotopic variability (right bottom).

[5] Amount effect is an isotopic effect that takes place due to evaporation that mainly enriches small rains compared to heavy ones (Mook, 2000).

Isotopic composition ($\delta^{18}O$, $\delta^{2}H$ and ^{3}H) of Addis Ababa rainfall (at 2360 masl, latitude 9° and longitude of 38.73°) has been monitored monthly since 1965 by IAEA (International Atomic Energy Agency) under the framework of the GNIP (Global Network of Isotopes in Precipitation) program. The record from January 1965 to December 2002 was used to establish the local meteoric water line (LMWL) and to evaluate isotopic variations within the water bodies of the study area (Figure 5.2, left).

In a $\delta^{2}H$ - $\delta^{18}O$ plot, the monthly rainfalls in Addis Ababa plots on the line defined by the relation $\delta^{2}H = 7.1\ \delta^{18}O + 12.3$ with a regression coefficient (r^2) of 0.93 (Figure 5.2, left). The isotopic composition of Ethiopian rainfall is the most enriched compared to other Eastern Africa regions, which is a known phenomenon (Rozanki et al., 1996). The enrichment is attributed to the contribution of already enriched moisture from the transpired moisture of the vegetated Congo basin (ibid). However, a satisfactory explanation behind the enrichment of the Addis Ababa rainfall is yet to be found (Kebede, 2004). In this region, source of moisture and local meteorological processes govern seasonal variation in isotopic composition of rainfall. The seasonal isotopic plot of Addis Ababa rainfall shows that summer rains have a relatively depleted isotopic composition compared to the spring rainfall (Figure 5.2, right). The depletion in the summer compared to the spring rainfall is related to the difference in the source of moisture and to local meteorological processes. Kebede (2004) stated that summer rainfall is derived from an admixture of Atlantic Ocean and south / equatorial Indian Ocean air masses, while enriched spring rainfall originates from the northern part of the Indian Ocean. He explains the isotopic enrichment of spring rainfall as the multiple effect of high temperature and low atmospheric humidity of the local climate during that time, rainfall amount effect, and closeness of the region to the source of moisture as well as the high sea surface temperature (SST) condition of the northern Indian Ocean.

In 2003, the isotopic composition of Addis Ababa rainfall was monitored by Kebede (2004) on a biweekly basis for the period between June and September. The analysis indicates that the period that is characterized by intense rainfall amount (July and August) has the most depleted isotopic composition compared to the other months (Figure 5.3, left).

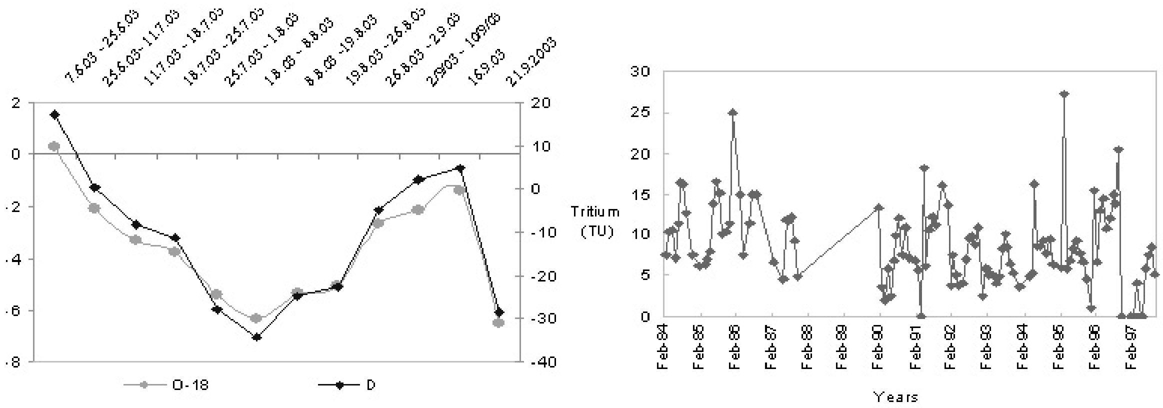


Figure 5.3: δ^2H - $\delta^{18}O$ plot of isotopic composition of Addis Ababa rainfall, between June and September, 2003, (data from Kebede, 2004) (left), and tritium content of monthly Addis Ababa rainfall (1984-1997) (right).

Even if the period between October and December is characterized by small amount of rainfall, the isotopic composition of rainfall is not as enriched as that of the spring rainfall (Figure 5.2, right bottom). This rainfall can be of the same origin as depleted summer rainfall but modified by local meteorological processes such as re-evaporation. Since the summer monsoon is retreating at this time, the isotopic composition of the rainfall reflects local convection and moisture source from locally recycled moisture (Kebede, 2004).

5.4 Isotopic composition of water bodies

The isotopic compositions of Lake Beseka, River Awash and the groundwater in the lake watershed, upstream of the study area and in the Merti areas were investigated. Isotopic composition of the water bodies in the region vary from isotopically depleted groundwater to isotopically enriched lake water (Figure 5.4). Based on their isotopic composition, the water bodies are categorized into five main groups. The first group comprises groundwater samples collected upstream from the study area (outside the lake watershed, to the south). Groundwater samples that were collected mainly from the western and the northeastern part of the lake watershed constitute the second group, and groundwater samples from the eastern part of the watershed, specifically from the Abadir and Addis Ketema areas, comprise the third group. Samples from the River Awash and groundwater system around the Merti area (outside the lake watershed, to the east) are categorized as the fourth group. The fifth group contains water samples from Lake Beseka.

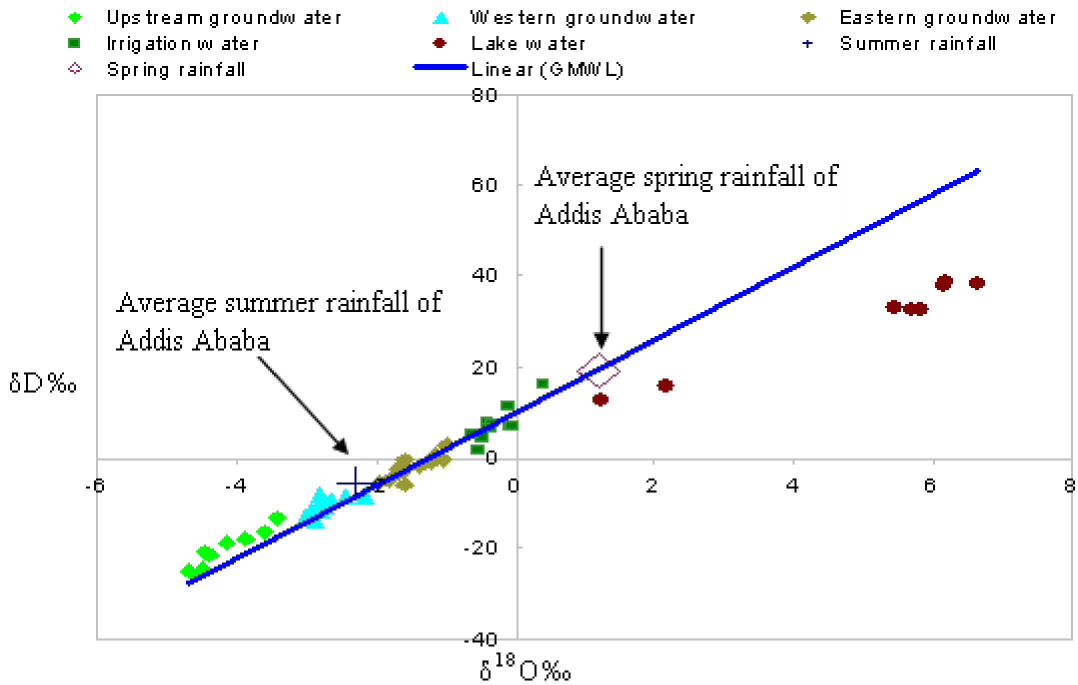


Figure5.4: Isotopic plot of all water bodies in the study area, together with weighted average isotopic composition of summer and spring rainfall in Addis Ababa.

5.4.1 Isotopic pattern of groundwater

The average isotopic composition of the groundwater in the study area is -2.4 ‰ in $\delta^{18}O$ and -8.5 ‰ in $\delta^{2}H$, which is comparable to the weighted average isotopic values of summer rainfall of Addis Ababa, (Figure 5.3). The weighted average isotopic value of $\delta^{18}O$ and $\delta^{2}H$ of Addis Ababa's summer rainfalls (July and August) is -2.3 ‰ and -5.6 ‰ respectively, however, the spring rainfall (March and April) is characterized by the values of + 1.2 ‰ in $\delta^{18}O$ and 19 in $\delta^{2}H$. The difference in isotopic composition of rainfall between the two rainy seasons provides an opportunity to trace seasonal groundwater recharges. The isotopic similarity of summer rainfall with groundwater indicates the significance of summer rains in recharging the groundwater system of the area. The isotopic signature of spring rainfall is not observed in the groundwater, which means that it plays only a minor role in recharging groundwater system of the study area. The importance of summer rainfall as a principal source of recharge is also evident in another parts of the country, including the northwestern Ethiopian plateau, the MER (Kebede, 2004), and the Akaki area, a small town close to Addis Ababa (Demlie et al., 2006).

Groundwater samples collected outside the lake watershed to the south, between the towns of Nazareth and Wolonchiti, are characterized by highly depleted isotopic concentrations (Figure 5.5). They are characterized by average isotopic $\delta^{18}O$ and $\delta^2 H$ value of -4.3‰ and -20.8 ‰, respectively. This groundwater is described as being of an older age (Kebede et al., 2007), and its origin is suggested as either from high altitude source or a recharge that took place under colder climatic conditions.

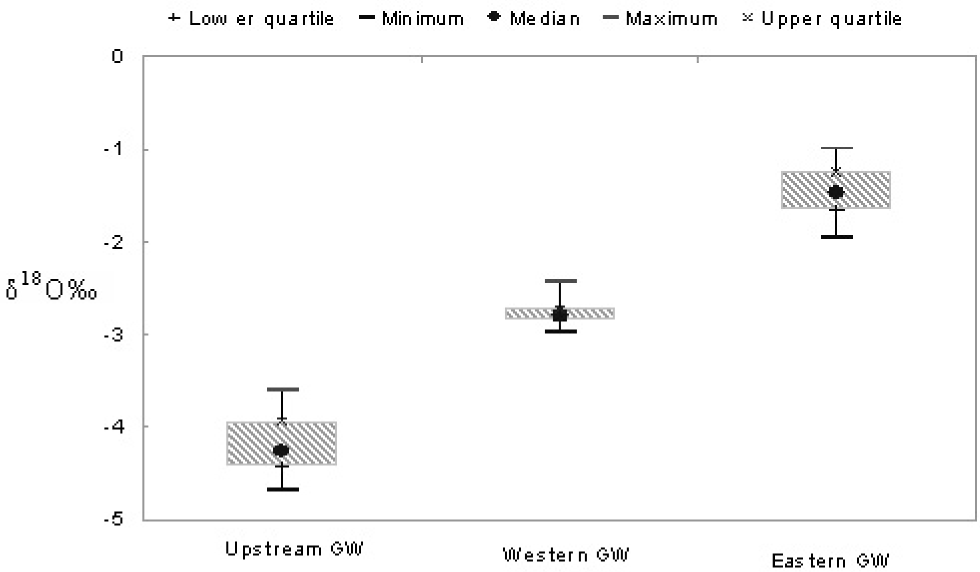


Figure 5.5: Statistical summary of isotopic composition ($\delta^{18}O$) of groundwater from upstream, and western as well as eastern part of Lake Beseka watershed.

The groundwater system in the western and the northeastern part of the watershed is characterized by an average isotopic composition of -2.8 ‰ in $\delta^{18}O$ and -10.7 ‰ in δ^2H (Figure 5.5). Hot springs at the southwestern corner of the lake also belong to this group. The relative isotopic depletion of groundwater compared to the weighted average isotopic composition of the summer rains can be related to the selective recharge of heavy rainfall during July and August (Figure 5.3, left). The depletion of the groundwater in this zone indicates that there is no evaporative enrichment of rainfall prior to recharge. There is only little tritium in this groundwater system, indicating recharge possibly prior to 1953. This might suggest that the groundwater may have a high altitude source and followed deeper circulation pathways for some time. Kebede et al. (2007) described the E-W transverse faults and the NNE-oriented rift faults, which are truncated against the E-W ridges in the area, as possible conduits to channel isotopically depleted waters from the plateau to these aquifers.

The groundwater system in the eastern part of the watershed, around the Abadir farm and the Addis Ketema areas, is characterized by an average isotopic composition of $\delta^{18}O$ and $\delta^{2}H$ of -1.5 ‰ and -2.5 ‰, respectively. The isotopic signature of these waters is relatively enriched with respect to summer rainfall but still depleted compared to the spring rainfall. The tritium content of this groundwater is slightly significant (1.6-5.4 TU), indicating a mixture of recharge from sub-modern (prior to 1953) and modern (after 1953) meteoric waters. This group plots at the mixing line between depleted groundwater in the western part of the watershed and enriched river water (Figure 5.4). The use of water from River Awash for irrigation on the sediments of the Abadir farm may induce infiltration of river water to the underneath groundwater system, which might create a conducive environment for mixing modern meteoric water with the existing sub-modern groundwater system.

The groundwater from the Merti area (outside the lake watershed, to the east) and the river water from River Awash are characterized by a comparable isotopic composition. Groundwater from the Merti area is typified by an average isotopic composition $\delta^{18}O$ and $\delta^{2}H$ of -0.5 ‰ and + 6.3‰, respectively, which is similar to the average isotopic values of the river water (-0.7 ‰ and + 3.2 ‰, respectively) (Figure 5.6). This indicates the importance of River Awash in recharging groundwater system in the Merti area. In addition, the tritium concentration is significant for this groundwater (6.7-10 TU), verifying the importance of modern meteoric water for the aquifer system in that area. The river water is introduced to the lake watershed when it is used for irrigation of the Abadir farm.

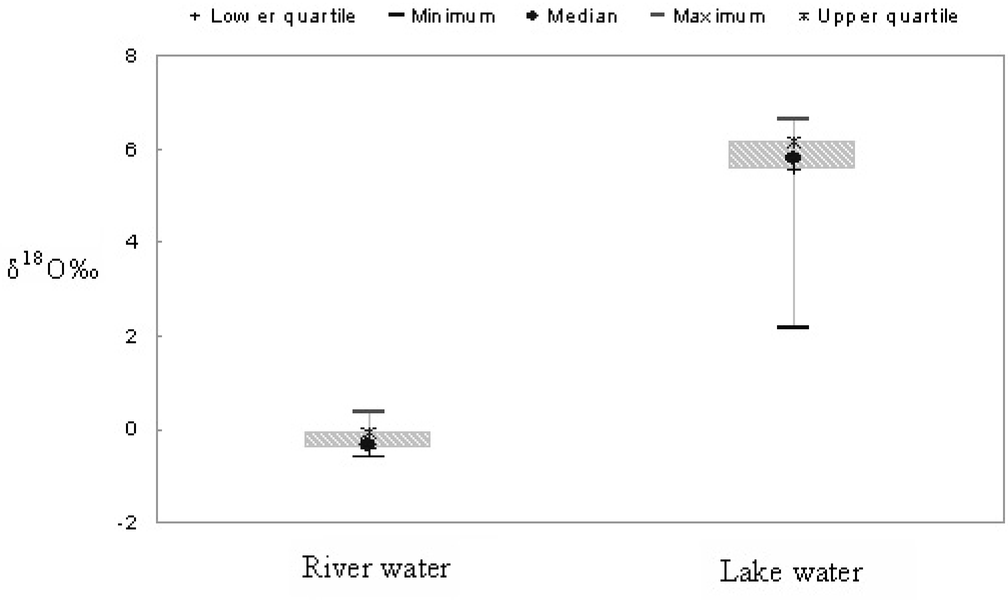


Figure 5.6: Statistical summary of isotopic composition ($\delta^{18}O$) of River Awash water, which is introduced to the lake watershed by irrigation, and Lake Beseka.

Evaporated water from Lake Beseka is isotopically enriched compared to the surrounding groundwater system. Lake water is characterized by an average isotopic composition of + 5.2 ‰ in $\delta^{18}O$ and + 32.0 ‰ in $\delta^{2}H$. Isotopically enriched groundwater is not observed in the study area in any part of the lake watershed. This reflects that Lake Beseka has little or no outflow to the surrounding groundwater system. This is an independent evidence that Lake Beseka is a terminal lake, which loses water only through evaporation.

5.4.2 Isotopic pattern of Lake Beseka

Lakes are characterized by an enriched isotopic composition compared to their inflows due to the dominance of non-equilibrium isotopic fractionation during the evaporation process. The degree of evaporative enrichment is controlled by meteorological variables such as atmospheric relative humidity, surface water temperature and water balance of the lakes (Mook, 2000). Consequently, the isotopic composition of evaporating lakes plots below the GMWL with a slope less than 8, along a line called the local evaporation line (LEL). The slope of LEL is found mainly between 3.5 and 6, and depends on local climate factors.

Lake Beseka is isotopically enriched compared to modern meteoric water (precipitation and surrounding groundwater), due to the effect of evaporative enrichment. Evaporated water from Lake Beseka plot along the LEL defined by the relation $\delta^2H = 5.1\ ^{18}O + 5.5$ ($r^2 = 0.98$). Lake samples from the Ziway-Shalla basin (south of the study area) plot along the LEL represented by the equation of $\delta^2H = 5.5^{18}O + 8.5$ (Ayenew, 1998), while lake samples from Afar region (north of the study area) plot on the LEL defined by $\delta^2H = 3.5^{18}O + 7.5$ (Kebede, 2004). The difference in slope and intercept of LEL within the Ethiopian rift valley lakes can be related to variations in hydrologic setting of the lakes (input-output ratios), which determines their residence time, and also by variations in local climate parameters.

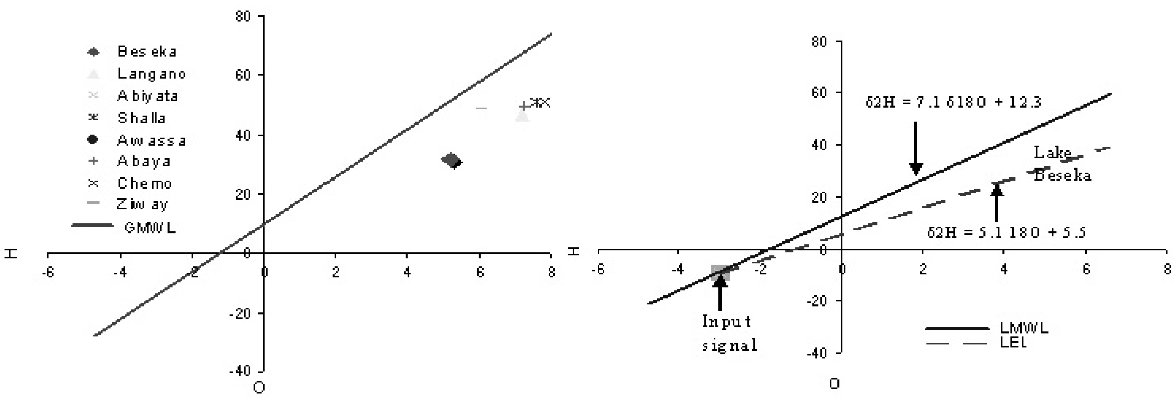


Figure 5.7: Isotopic plot of lakes in the MER (left) and local evaporation and local meteoric water lines (right), with input signal plotting at isotopic value of -3 ‰ in $\delta^{18}O$ and -9.3 ‰ in δ^2H.

The LEL that plots along the evaporated waters of Lake Beseka intersects the LMWL of Addis Ababa rainfall at the isotopic value of -3 ‰ in $\delta^{18}O$ and -9.3 ‰ in δ^2H (Figure 5.7). This value is referred as the input signal. The isotopic concentration of the input signal is comparable to the groundwater system in the western part of the lake watershed, which is characterized by an average isotopic composition of -2.8 ‰ in $\delta^{18}O$ and -10.7 ‰ in δ^2H. This suggests that the major inflow to Lake Beseka comes from the groundwater system, which flows mainly from the western part of the watershed. The possible importance of groundwater flow to the watershed of Lake Beseka from Lake Koka, man made lake situated at some 90 Km upstream of the study area is not evident. Lake Koka was created by construction of Koka dam across River Awash for regulation of the flow of the river at the beginning of 1960's. Modern meteoric water plays an important role to the total water and chemical budget of Lake Koka, and consequently it has isotopic composition (-1.62 ‰ in $\delta^{18}O$ and - 2.2 in $\delta^{2}H$) which is different from the isotopic composition of the input signal of Lake Beseka.

Isotopic depletion of the input signal compared to summer rainfall can be explained either by selective recharge with heavy and depleted summer rainfalls along fractures and fault zones, or related to high altitude source of recharge to the groundwater system. Input signals for Lake Beseka significantly vary from the isotopic composition of River Awash (-0.7 ‰ in $\delta^{18}O$ & + 3.2 ‰ in $\delta^{2}H$), which is used in the Abadir farms for irrigation. This rules out the importance of irrigation water as the main input to the hydrology of Lake Beseka, although irrigation water claimed to be the major source of water for the lake expansion in previous studies.

5.5 Summary

Isotopic and chemical composition of groundwater can be used to localize recharge areas and to determine the origin of groundwater and of individual chemical components (Mook, 2000). Hydrochemical and environmental isotope data are used in this study to gain a general picture of the groundwater circulation in the study area, and to assess the importance of groundwater to the hydrology of Lake Beseka. Silicate hydrolysis of volcanic rocks is the key process that controls the hydrochemical characteristics of the groundwater, and resulted in the dominance of the $Na\text{-}HCO_3$ type of water in the study area. Recharge of the groundwater system by isotopically depleted summer rainfall, mixing of modern and sub-modern meteoric water, and the evaporation of the lake water with respect to its inflows dominate the isotopic processes and govern the isotopic composition of the water bodies in the study area.

The groundwater system in the western part of the watershed is characterized by a $Na\text{-}HCO_3$ type of water and by an average isotopic composition of -2.8 ‰ in $\delta^{18}O$ and -10.7 ‰ in $\delta^{2}H$. Bicarbonate is the predominant anion in this groundwater, which is a typical characteristic of recharge areas where groundwater is flushed by CO_2-rich meteoric water. The isotopic composition of this groundwater is comparable to the weighted average isotopic composition of Addis Ababa summer rainfall, signifying the latter's importance in recharging the groundwater system. The groundwater lacks a significant amount of tritium, which suggests a sub-modern nature of the system, which has followed deeper circulation pathways from a high altitude source. The fact that the study area is located at the structural intersection zone between the NNE-SSW trending rift faults and the E-W trending YTVL could create favorable condition to conduit isotopically depleted groundwater from the central part of the western plateau. Kebede et al. (2007) described the intersection between the NNE rift structures and E-W ridges as a hydrogeologic window that channels depleted groundwater from the mountain to the rift.

The groundwater system in the eastern part of the watershed is characterized by a *Na-HCO3-SO4-Cl* and *Na-HCO3-Cl* type of water. In this groundwater system, sulfate or chloride or both anions form significant part of the total anions. This is a distinctive feature of discharge areas where groundwater circulation is less active or slow. Hydrochemical and isotopic composition of the groundwater in the eastern part identifies two subclasses.

The first subclass is characterized by a *Na-HCO3-Cl* type of water, insignificant amount of tritium, and isotopic composition comparable to the groundwater system in the western part of the watershed (-2.8 ‰ in $\delta^{18}O$ and -10.7 ‰ in $\delta^{2}H$). The groundwater in this subclass has similar origin to that in the western part of the watershed, except that it is hydrochemically evolved in its pathway to become a chloride-significant groundwater.

The second subclass in eastern part of the watershed is the groundwater system with a Na-*HCO3-SO4-Cl* type of water, an average isotopic composition of -1.5 ‰ in $\delta^{18}O$ and -2.5 ‰ in $\delta^{2}H$, and some tritium. The importance of the chloride and sulfate in this subclass indicates that this groundwater system is characterized by longer subsurface residence time with sluggish groundwater movement. Both its isotopic composition and tritium content suggest that the groundwater system has undergone mixing of sub-modern (prior to 1953) and modern (after 1953) meteoric water. This groundwater is mostly located under the Abadir farm, where modern meteoric water from River Awash is used for irrigation purposes. This might create suitable conditions for infiltration of irrigated water to an already existing sub-modern groundwater system with possibility of mixing processes.

Evaporated water from Lake Beseka is characterized by a *Na-HCO3-Cl* to *Na-HCO3-CO3-Cl* type of water, and by an average isotopic composition of + 5.2 ‰ in $\delta^{18}O$ and + 32.0 ‰ in $\delta^{2}H$. Lake Beseka is isotopically enriched compared to modern meteoric water due to evaporative enrichment. Ionic ratio and Schoeller plot indicate that the lake has a hydrochemical signature comparable to that of groundwater system flowing from the western part of the watershed. This concept is independently supported by the isotopic plot where the LEL intersects the LMWL at the isotopic value referred as the input signal (Figure 5.6). The input signal is characterized by the isotopic values of -3 ‰ in $\delta^{18}O$ and -9.3 ‰ in $\delta^{2}H$, which is comparable to the average isotopic values of the groundwater system flowing from the western part of the watershed, i.e., -2.8 ‰ in $\delta^{18}O$ and -10.7 ‰ in $\delta^{2}H$. Both hydrochemical and isotopic signature suggest that the groundwater system that flows from the western part of the watershed forms the main source of water input to the water budget of Lake Beseka. Once this groundwater enters the lake, the water evolves in response to evaporation, becoming a saline and isotopically enriched water type with respect to its inflows.

Waters from the River Awash is classified as a *Na-Ca-HCO3* type of water, and shows an average isotopic composition of - 0.7 ‰ in $\delta^{18}O$ and + 3.2 ‰ in $\delta^{2}H$. River Awash is located outside the lake watershed, but the water is introduced to the study area for irrigation use. The groundwater system around this river (outside the study area, to the east) has a similar isotopic and hydrochemical composition with the river, which implies that the aquifer system of the Merti area is hydraulically linked with River Awash. On the other hand, ionic ratio, Schoeller and isotopic plots indicate that this river water has different isotopic and hydrochemical signatures compared to the lake, indicating that it plays no important role regarding being the main input to the hydrology of Lake Beseka.

5.6 Conceptual model

A conceptual model is pictorial representation of the groundwater flow system and is built with the purpose of simplifying field problems and organizing associated field data (Anderson and Woessner, 1992). While the conceptual model is an idealized summary of the current understanding of the watershed conditions, and the key aspects of how the flow system works, it is subject to some simplifying assumptions (Middlemis, 2001).

Simplifying assumptions are necessary, as complete reconstruction of the field system is not feasible, and there are rarely sufficient data for a comprehensive model (PDP, 2002). Existing hydrogeologic data were combined with hydrochemical and isotopic information to set up a conceptual model of the study area (Figure 5.8). The flow system was analyzed by combining groundwater level measurements with the hydrochemical and isotopic composition of the water bodies in the study area.

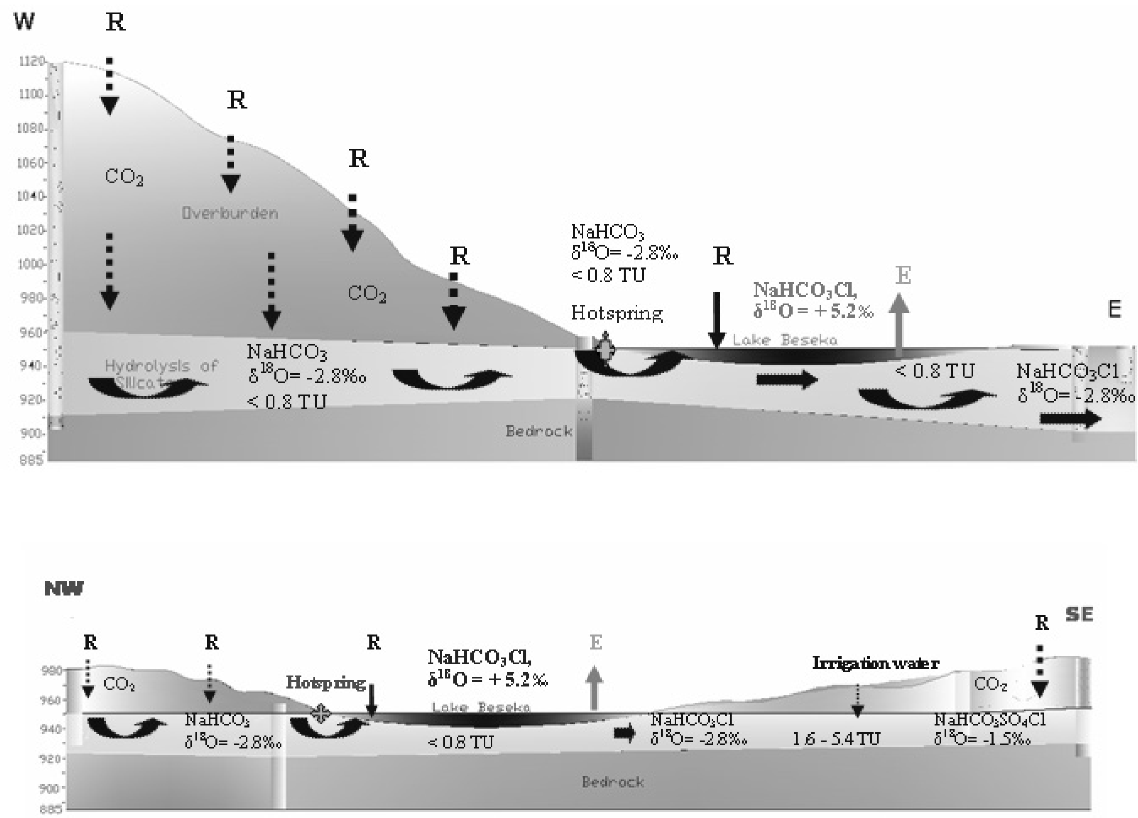


Figure 5.8: Conceptual model of the study area, from west to east (top); from northwest to southeast (Abadir farm areas) (bottom) part of the watershed. **R**: Recharge from summer rains, **E**: Evaporation, TU: Tritium unit, : Groundwater flows, : Hot springs

An extensive unconfined aquifer is found at a moderately shallow depth from the surface. The main water-bearing formation is a mixed aquifer system of fractured and weathered basalts, welded tuffs, and pyroclastic deposits. The groundwater, which is recharged mainly by summer rainfall, flows from the northwestern and southwestern part of the watershed toward the east. This groundwater is characterized by a *Na-HCO*$_3$

type of water, by an average isotopic composition of -2.8 ‰ in $\delta^{18}O$, and by a low tritium concentration (< 0.8 TU). Part of this groundwater flows to the lake and emerges along the western edge in the form of hot springs. Once this groundwater enters the lake, it is affected by evaporation, becoming a *$NaHCO_3Cl$* type and isotopically enriched water with an average isotopic composition of + 5.2 ‰ in $\delta^{18}O$.

Part of the groundwater that flows further to the east under the lake has the same isotopic signature and tritium content as the groundwater from the west, but hydrochemically evolved to become a NaHCO3Cl type of water. However, some part of the groundwater system in the eastern part of the watershed is characterized by an average isotopic composition of -1.5 ‰ in δ18O with significant tritium content (1.6-5.4 TU) and a Na-HCO3-SO4-Cl type of water. This can be explained by the mixing of hydrochemically evolved sub-modern groundwater that flows from the west with recent meteoric to become isotopically distinct water. Infiltration of irrigating water at Abadir farm could be one source of meteoric water for possible mixing. However, it does not play an important role in the water budget of the lake.

6 GROUNDWATER RECHARGE

6.1 Precipitation

Precipitation is the ultimate source of all water on the earth's landmasses (Healy et al., 2007), and the principal source of fresh water for replenishing soil moisture and groundwater. All water enters the land phase of the hydrologic cycle as precipitation (Dingman, 2002). Thus, to assess the response of different hydrological processes, it is imperative to understand the temporal and spatial distribution of precipitation. Like most parts of Ethiopia, seasonal distribution of rainfall is governed by the position of the Inter Tropical Convergence Zone (ITCZ) in the study area. Meteorological data (rainfall, temperature, relative humidity, sunshine hours, wind speed, and pan evaporation) were collected from National Meteorology Agency (NMA) of Methara (1964-2005), and Nura Era (1977-2005) stations, and additional meteorological data were collected from Methara Research center station (1966-2007).

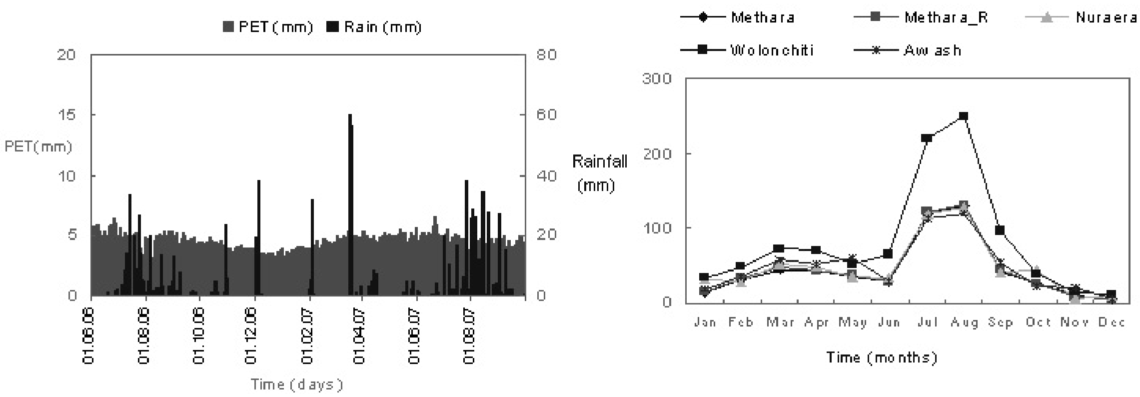


Figure 6.1: Daily rainfall and potential evapotranspiration at Methara research center station (left) and temporal variability of long-term mean monthly rainfall (1966-2006) at nearby meteorological stations (right).

Rainfall distribution in the region is bimodal (spring and summer rains) following the convergence of air flows from the Atlantic and the Indian Ocean. The main rainy season is in summer, from July to September, when the ITCZ lies in northern Ethiopia, and the area is under the influence of moisture from the Atlantic and the Indian Ocean. For March and April, a southeasterly wind from the Indian Ocean brings a small amount of spring rain (Figure 6.1, right).

Analysis of long-term mean annual precipitation in the region indicates spatial variation of precipitation according to topography. The amount of precipitation tends to increase toward the south with increasing elevation. The areal depth of precipitation for the watershed is estimated using the Thiessen polygon method, which was constructed based on the three nearby meteorological stations (Methara, Methara research center and Nura era stations; Figure 6.2). Long-term mean annual precipitation in the lake watershed from 1966 to 2007 was estimated 534 mm. Within this period, 1982 and 2002 were the wettest and driest years, with 864 mm and 309 mm of precipitation, respectively.

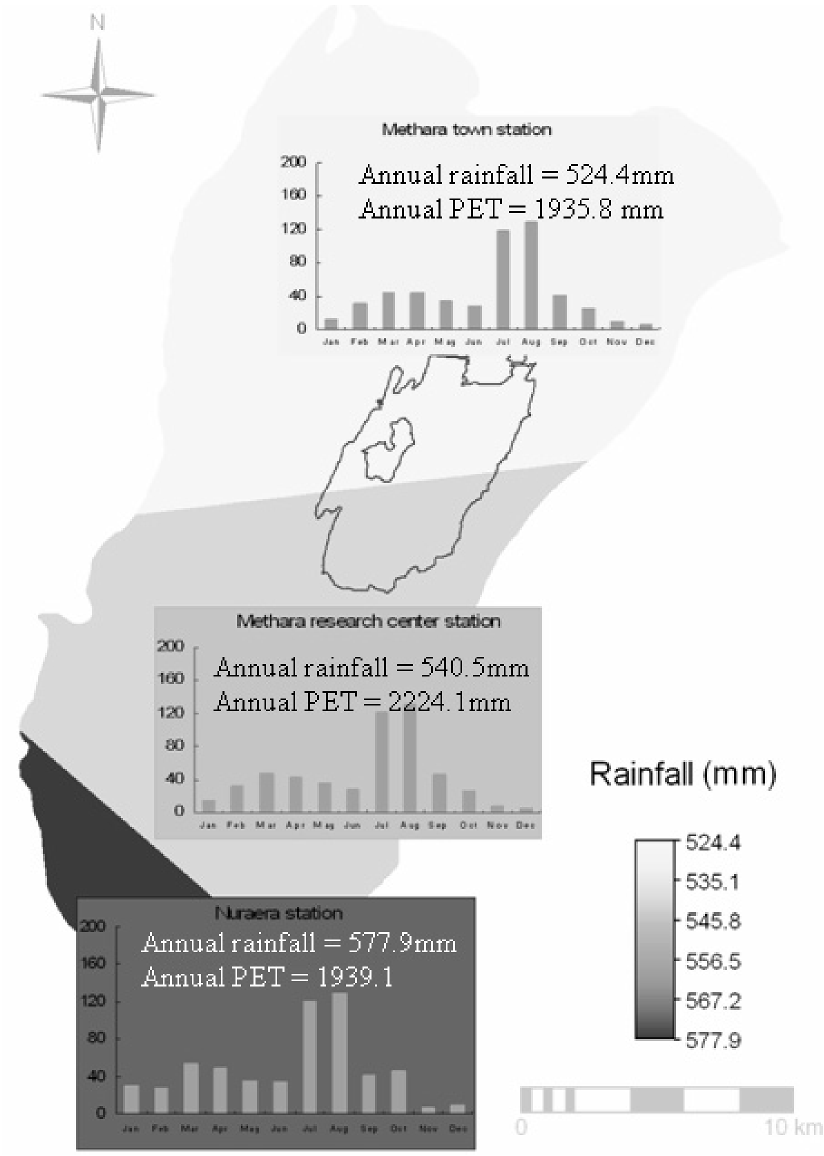


Figure6.2: Areal depth of precipitation for the lake watershed (Thiessen polygon).

Comparison of the daily rainfall with potential evapotranspiration during the period of June 2006 – September 2007 at Methara station shows that the rainfall in the area has extreme and intensive characteristics. Intensive rainfall events are observed during both summer and spring rainy seasons (Figure 6.1, left). For instance during the spring rainy season on 18/3/2007, maximum daily rainfall was 60.10 mm while the potential evapotranspiration was only 5.04 mm. Similarly, rainfall during the summer rainy season on 12/7/2006 was 33.40 mm while potential evapotranspiration was only 4.98 mm. The significant gap between daily rainfall and potential evapotranspiration during the rainy seasons points out the potential of these short and extreme rainfall events to replenish the groundwater system of the area.

6.2 Evapotranspiration

Potential evapotranspiration (ET_o)[6] is computed from meteorological data using the FAO Penman – Montheith method. This method is recommended as the standard method for the definition and computation of the potential evapotranspiration (Allan et al., 1998). It requires standard climatological records of radiation, air temperature, air humidity and wind speed. The result shows that the long-term (1966-2007) mean annual potential evapotranspiration in the study area is 2052 mm. The analysis of potential evapotranspiration of the area indicates a seasonal trend, in which the highest value of 7.9 mm/day occurred at the end of the dry season in June, while the lowest value of 5.2 mm/ day occurred in December (Figure 6.3, right).

6.3 Recharge estimation

6.3.1 Background

Groundwater recharge is a key component in any model of groundwater flow or contaminant transport, and its accurate quantification is crucial to proper management and protection of valuable groundwater resources (Healy and Cook, 2002). Balek (1988) defined recharge as the process of downward movement of water through the saturated zone under the force of gravity or in the direction determined by hydraulic

[6] Potential evapotranspiration is the evapotranspiration that would occur if water were plentiful (Healy et al., 2007)

conditions. Precipitation, surface waters, urban water sources, and irrigation losses can be source of groundwater recharge. Recharge mechanisms from theses sources can be direct or indirect.

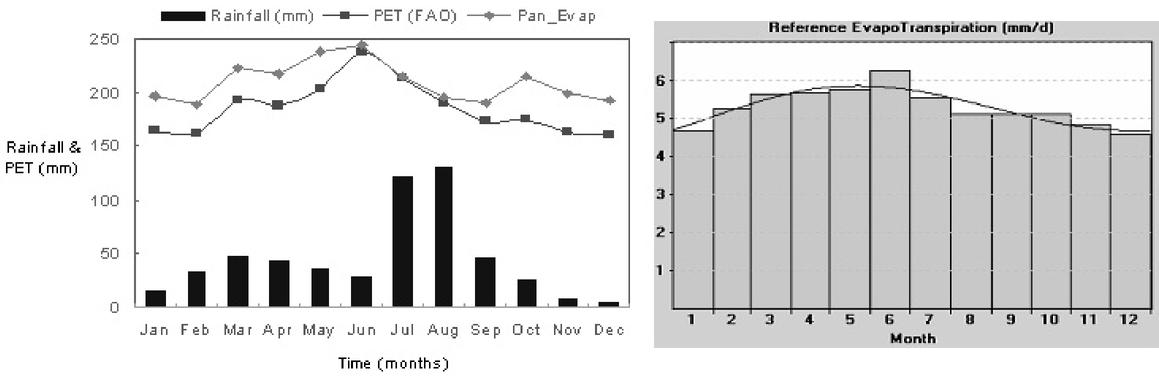


Figure 6.3: Long-term (1966-2007) mean monthly potential evapotranspiration compared with monthly precipitation (left) and daily potential evapotranspiration calculated by CROPWAT for each month (right).

The direct or diffuse recharge mechanism refers to the water added to the groundwater reservoir in excess of soil-moisture deficit and evapotranspiration. The indirect or localized recharge mechanism on the other hand refers to the concentrated percolation of water to the water table following runoff and localization in joints, low lying areas, on lakes or through the beds of surface water courses (Lerner et al., 1990). Preferential recharge takes place via pathways such as macropores opposed to diffuse recharge, which takes place through the entire vadose porous medium (Beekman and Xu, 2003; Sophocleous, 2004).

Recharge is governed by the intricate balance between several components of the hydrological cycle, each of which is a function of several factors (Bredenkamp and Xu, 2003). Environmental factors like climate, geomorphology (including topography, soil and vegetation), depth of water table, and geology play an important role in controlling the location and timing of recharge (Alley et al., 2002; Scanlon et al., 2002). These parameters are crucial in determining flow mechanisms and recharge rates, and thus, understanding these environmental factors helps to adapt appropriate recharge estimation methods. Recharge rates and mechanisms that are determined by the aforementioned factors are significantly different in arid and humid regions. In humid areas, diffuse recharge is the dominant flow mechanism where aquifers are often full, and the rate of recharge is often limited by subsurface geology. In contrast, in arid regions, focused recharge dominants and recharge rate is mainly controlled by climate factors (Scanlon et al., 2002).

Recharge estimation in arid and semi-arid areas is challenging due to time variability of precipitation, small amount of the recharge, and spatial variability of topography, soil characteristics, vegetation, and land use (Beekman and Xu, 2003). These areas are characterized by high evapotranspiration and sporadic rainfall with high temporal and spatial variability. According to Lerner et al. (1990), when aridity increases, direct recharge is likely to become less important, and indirect recharge is more important in terms of total recharge of the aquifer. Thus, indirect or localized recharge is often the most important source of natural recharge in these regions (Sophocleous, 2004). Indirect recharge is assumed to be more important with increasing aridity, because in more arid environments only localized concentration of runoff and subsequent infiltration can exceed the high thresholds for deep percolation (Külls, 2000). Recharge estimation in arid and semi-arid regions, therefore, needs a different approach to those used in humid areas (Scanlon et al., 2002).

For quantification of natural groundwater recharge, several methodologies are available (Lerner et al., 1990; Simmer, 1988; and Scanlon et al., 2002). Each technique has its own limitations with respect to accuracy and reliability (Beekman and Xu, 2003). Recharge estimation techniques can be classified according to hydrological zones, hydrogeological provinces and approaches. Based on hydrological sources or zones from which section that the data are obtained, recharge estimation techniques can be classified into surface water, unsaturated zone, and saturated zone (Beekman and Xu, 2003). Surface-water and unsaturated-zone approaches usually provide estimates of potential recharge, whereas groundwater techniques generally provide information on actual recharge, as this is based on the water that has already reached the water table (Scanlon et al., 2002). Saturated-zone-based techniques provide spatially integrated recharge estimates compared to the point estimates of unsaturated-zone methods. For each hydrological zone, recharge estimation techniques can further be classified into physical and tracer approaches (Scanlon et al., 2002; Beekman and Xu, 2003; Sophocleous, 2004).

Physical methods for recharge estimation range from direct measurements of hydrological parameters (lysimeter) to indirect physical techniques of water-balance methods and numerical modeling. Direct measurement of recharge requires elaborate instrumentation and is feasible only in a research setting (Dingman, 2000). Water-balance methods estimate recharge as the residual of all other fluxes based on the principle that other fluxes can be measured or estimated more easily than recharge (Lerner et al., 1990). The advantages of water-balance methods are that they use readily available data (rainfall, runoff, water levels), are easy to apply, and account for all water entering a system (Lerner et al., 1990; Sophocleous, 2004).

The major limitation of the residual approach is that the accuracy of the recharge estimate depends on the accuracy with which the other components in the water-balance equation are measured (Scanlon et al., 2002). The fact that the calculated recharge is the result of the difference in measurements that are themselves subject to significant errors leads to a comparatively high recharge estimation error in many cases (Sophocleous, 1991). This limitation is critical when the magnitude of recharge is small with respect to the other variables, particularly evapotranspiration (Lerner et al., 1990; Scanlon et al., 2002). Water-balance methods include soil moisture budget, river-channel water balance (base flow - separation techniques) and water table fluctuations, etc.

Calibrated groundwater models or inversion models are used to estimate recharge rate based on hydraulic heads and hydraulic conductivities. Numerical modeling methods take transient flows and storage changes into account, and they can include spatial variability of physical properties, of which determination of hydraulic conductivity is one of the most important issues (Lerner et al., 1990). Thus, the reliability of recharge estimates via model inversion depends on the accuracy of hydraulic conductivity (Scanlon et al., 2002). However, it is not easy to obtain reliable values from field methods.

Environmental and applied tracers have been widely used for recharge estimation in arid and semi-arid areas. Environmental tracers most commonly used in recharge studies include ^{3}H, ^{14}C, ^{36}Cl, ^{15}N, ^{18}O, ^{2}H, ^{13}C and Cl. Tracers do not measure water flow directly (Lerner et al., 1990). As a result, recharge can be over- or underestimated due to problems related to secondary tracer inputs, mixing and flow mechanisms. Environmental tracers sum recharge over long periods in arid and semi-arid areas, tens of years in unsaturated zones, and upto thousands of years in aquifers (ibid). Chloride is a widely used environmental tracer in recharge estimation in arid and semi-arid regions using the chloride mass balance (CMB) approach for both unsaturated and saturated zones.

Direct recharge measurements are not easy and not always an applicable procedure due to time and money constraints. On the other hand, indirect recharge estimation approaches are subjected to limitations. All techniques have their own limitations in terms of applicability and reliability (Beekman and Xu, 2003). Lerner et al. (1990) stated that developing incorrect conceptual models, neglecting spatial and temporal variability, and measurement and calculation errors are the main causes of error during recharge estimation. Uncertainties associated with each recharge estimation approach underscore the need for application of multiple techniques so as to increase reliability and confidence of recharge estimates (Scanlon et al., 2002). Due to the difficulty in assessing the accuracy of any recharge estimation method, it is essential to use multiple techniques, although consistency among these techniques should not be taken as an indication of accuracy (Healy and Cook, 2002).

Recently, in addition to combining different recharge estimation approaches, surface and groundwater models have been integrated to get more reliable recharge estimates. The most widely known coupled surface-groundwater numerical models vary from the complex and data demanding MIKE-SHE code to the relatively simpler IGSM[7] code (Lubczynski and Gurwin, 2005). Integrated watershed and groundwater models allow a complete analysis of the land-based hydrological cycle, thus providing the means for evaluating the impacts of land use, irrigation development and climate change on both surface water and groundwater resources (Sophocleous, 2004). However, extensive data and expertise are required to effectively utilize coupled surface-groundwater numerical models for sound watershed management.

The objective of a recharge study should be known prior to the selection of the appropriate recharge estimation methods, as the objective determines the required space and time scale (Scanlon et al., 2002; Beekman and Xu, 2003). Recharge is rarely estimated in isolation, but is usually just one aspect of a wider study such as on groundwater resources, pollution transport, subsidence or well field design (Lerner et

[7] ISGM is three-dimensional, finite-element-based integrated groundwater and surface water model

al., 1990). Recharge estimation has rarely been a final objective of research studies but always an intermediary stage to a final step (Sandwidi, 2007). The primary objective of recharge estimation in this study is to determine the amount of water that replenishes the aquifer system in the study area, and to assess whether this recharge amount is substantial enough to explain the continuous groundwater input to the lake. The other objective is to use the recharge estimate as an input to a groundwater model that has been developed to understand the lake - groundwater dynamics in the study area. In addition, the recharge estimate can also be used for future groundwater studies related to evaluation and proper management of the resources. In this section, the recharge was estimated by combining the water table fluctuation method, chloride mass balance and EARTH modeling.

6.3.2 Water table fluctuation method

Introduction

Recharge and lateral inflow of groundwater increase groundwater storage, whereas groundwater evapotranspiration, exploitation, base flow to streams, and lateral outflow reduce groundwater storage. The water table fluctuation method provides an estimate of groundwater recharge from changes in groundwater storage by analyzing groundwater level fluctuations in a given time without considering the lateral flow. Thus, this method requires knowledge on how water levels change in response to rainfall and storage property of the aquifer. It is the most widely used technique for groundwater recharge estimation due to its simplicity and ease of use (Healy and Cook, 2002).

Theoretical background

The water table fluctuation method is based on the assumption that rises in groundwater levels of unconfined aquifers are caused by the recharging rainwater arriving at the water table. Recharge is calculated as the change in groundwater levels over time multiplied by specific yield (Figure 6.4):

$$R = Sy\frac{\Delta h}{\Delta t} \tag{6.1}$$

where R is recharge, S_y is specific yield, h is water table height, and t is time

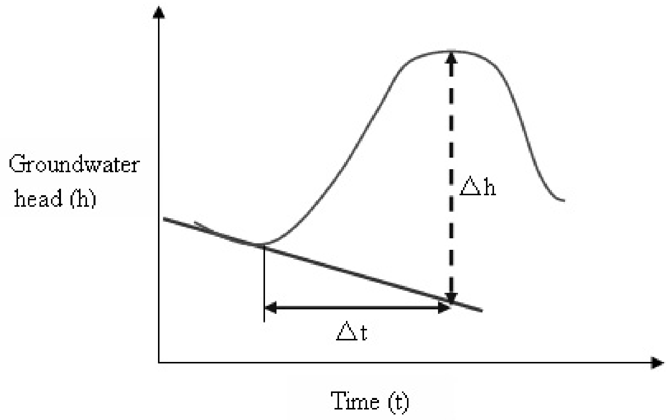


Figure 6.4: Hypothetical water level rise in response to rainfall in water table fluctuation method

In equation 6.1, it is assumed that water arriving at the water table goes to storage, and all other components (evapotranspiration from groundwater, base flow to streams and springs, lateral flows, etc.) are negligible during the period of recharge (Healy and Cook, 2002). Thus, the method is best applied over a short period of time and in shallow water tables that display sharp rises and declines in water levels (Scanlon et al., 2002). Lerner et al. (1990) also emphasized that the water table fluctuation method should only be applied to aquifers with well defined recharge seasons. Total recharge is quantified when the water table fluctuation method is applied for each individual water level rise over a given time, while net recharge is estimated if the method is applied to seasonal water level fluctuations. Net recharge is the difference between total recharge and groundwater abstraction if the lateral inflow equals the outflow.

Beside its simplicity, the advantage of the method is its insensitivity to the mechanism by which water travels through unsaturated zones (whether the recharge mechanism is diffuse or preferential), and its ability to be viewed as an integrated approach rather than as a point recharge estimation approach (Healy and Cook, 2002). A limitation is the difficulty in determining representative values for specific yield and in ensuring that fluctuations in water levels are due to recharge and are not the result of changes in atmospheric pressure, the presence of entrapped air or other phenomena, such as pumping (Scanlon et al., 2002).

In groundwater hydrology, the concept of specific yield or aquifer storage is used to transform a change in groundwater level to equivalent water storage, and hence recharge (Sophocleous, 1991). Specific yield is defined in Dingman (2002) as the volume of stored groundwater released (taken up) per unit surface area of aquifer per unit decline (increase) of water table. Determination of proper specific yield values is a difficult endeavor and significant uncertainty remains as to what value of specific yield to use in a particular study (Healy and Cook, 2002). A characteristic value of specific yield for the aquifer material is often assumed, but this can range widely (Dingman, 2002). Significant variation of specific yield values is observed even within the same textural class when determined through laboratory and field methods. This is due to the fact that drainable porosity measurements in the laboratory usually run for longer periods than pumping test methods. Consequently, Lerner et al. (1990) suggest using standard, literature-specific yield values rather than pump test values if laboratory measurements are not available.

Results

During the field season, groundwater survey was conducted in May 2006, and three locations were selected for installation of data loggers for monitoring groundwater levels. Two data loggers were installed in abandoned wells of Gofa sefer (east of Lake Beseka) and Abadir area (south of Lake Beseka), and the third data logger was installed at the southwestern edge of the lake in the Tone area, where numerous hot springs are emerging from the bed and wall of the lake periphery (Figure 6.5). Data loggers monitored groundwater level and temperature data on daily bases for 15 months (July 2006-September 2007) for Gofa Sefer & Abadir sites while the data from hotspring site was limited for only 6 months (July 2006-December 2006) due to various problems faced in the area.

The wells were monitored by pressure transducers that recorded groundwater level and temperature data on a daily basis. The two well sites (Gofa Sefer and Abadir) had not been under any pumping activity during the monitoring period, and fluctuation of groundwater levels was due to natural factors. The groundwater level data from the two sites were used for estimating recharge through the water table fluctuation method. Seasonal water level fluctuations of the monitored wells indicate that groundwater heads rose and fell following the rainfall pattern (Figure 6.6, left).

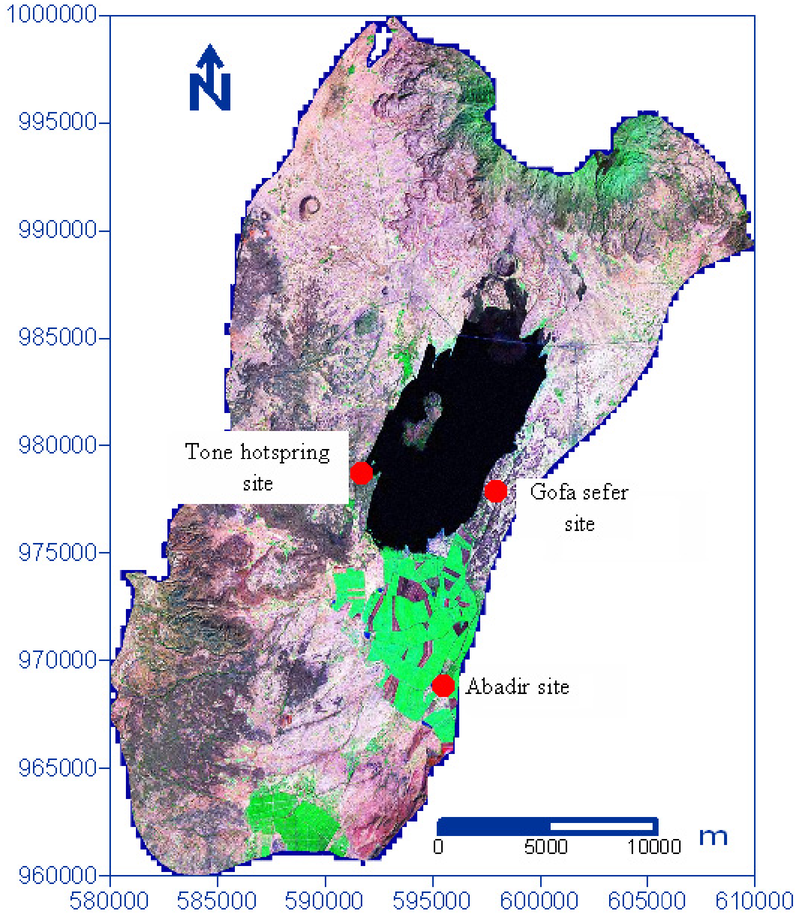


Figure 6.5: Location of groundwater monitoring sites (LANDSAT 7, 2000).

The water level rise (△h) is estimated as the difference between the peak of the water level rise and the value of the extrapolated antecedent recession curve at the time of the peak. The main uncertainty of water table fluctuation method arises from the difficulty in obtaining reliable specific yield values. It was not possible to derive such values for the study area from the existing pumping test data given that all pumping tests were single-well tests. Fractured basalt is the main water-bearing formation in the monitored sites, and the standard specific yield of 0.02 – 0.1 was taken from literature (Johnson, 1967; Freeze and Cherry, 1979; Driscoll, 1986; Domenico and Schwartz, 1998) for estimation of the recharge.

Groundwater recharge was estimated using the water table fluctuation method from groundwater level fluctuation data of one hydrological year (July 2006-July 2007) for the study area. An average annual recharge value of 42 mm was estimated for the Gofa Sefer area, while 42.6 mm was estimated for the Abadir camp vicinity. This recharge is 6.6 % and 6.7 % of the total annual precipitation for the GofaSefer and Abadir areas, respectively.

6.3.3 EARTH Modeling

Introduction

EARTH is a lumped[8] hydrological model for simulating recharge and deep groundwater level fluctuation. It was developed by Van der Lee et al. in 1989 for use in the GRES (groundwater recharge evaluation study) in a project in Botswana. EARTH is the acronym for **E**xtended model for **A**quifer **R**echarge and soil moisture **T**ransport through the unsaturated **H**ardrock. It is a one-dimensional model designed for semi-arid areas and tested under different climate conditions for deep groundwater levels of up to 40 m (Van der Lee and Gehrels, 1990). EARTH combines direct and indirect methods. The direct method describes the recharge process from the atmosphere toward the soil zone, while the indirect method uses groundwater level fluctuations as an indication of the recharge process. The direct method determines recharge by using physical processes above the groundwater table, whereas the indirect method calculates groundwater level from the recharge estimated in the direct part (van der Lee and Gehrels, 1990).

The EARTH model has five modules, MAXIL, SOMOS, LINRES, SUST and SATFLOW (Figure 6.7). The first three modules, i.e., MAXIL, SOMOS and LINRES, are combined in the direct part of the model and will be explained below. In this part, recharge is simulated for the unsaturated zone and calibrated with measured time series soil moisture data. SATFLOW is the indirect part of the model that calculates the groundwater level from the recharge estimate of the direct part. EARTH assumes that the rise of the water level in the groundwater represents the system response to recharge

[8] A lumped model is a model where the watershed is regarded as a single unit, and parameters are estimated from calibration by applying concurrent input and output time series data (Abbott and Refsgaard, 1996).

from precipitation but not lateral ground water fluxes (Lubczynski and Gurwin, 2005). SUST calculates surface ponding and runoff.

The model takes into the account the physical processes in both saturated and unsaturated zones, and simulates water level fluctuation by coupling climatic, soil moisture and groundwater level data. Since the model simulates recharge by linking information from the atmosphere, unsaturated and saturated zones, it has a good potential to forecast future groundwater recharge (Beekman and Xu, 2003). The model is proved to be an efficient and quite accurate recharge estimation tool (Van der Lee and Gehrels, 1997).

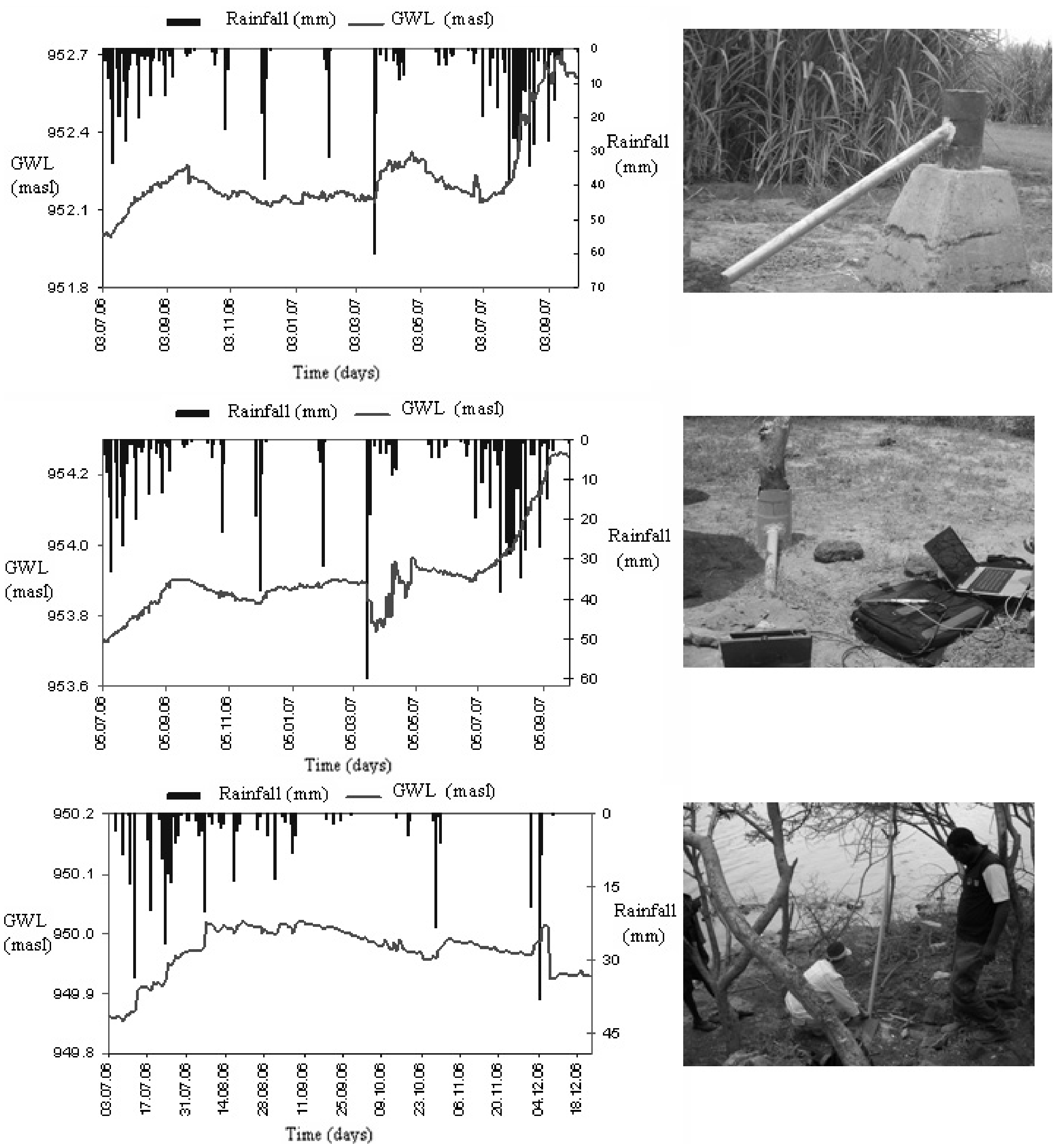


Figure 6.6:Groundwater hydrographs and data logger sites (Gofa Sefer, Abadir camp and hot spring site at Tone, from the top to the bottom).

Beekman and Xu (2003) listed EARTH as a promising recharge estimation method that can be applied in arid and semi-arid regions with greater certainty. The model has been used to estimate recharge in various hydrogeological conditions, starting from the sandstone aquifer of the Karoo formation in South Africa (Usher et al., 2006) to the granitic aquifer of the Sardon watershed in Spain (Lubczynski and Gurwin, 2005). It requires rainfall and potential evapotranspiration (PET) data as an input, and the model should be calibrated by groundwater level or / and soil moisture data.

Theoretical background

EARTH combines four main modules to simulate recharge in saturated and unsaturated zones by coupling climatic, soil moisture and groundwater level data. The first two modules, MAXIL and SOMOS, represent the agro-hydro-meteorological zone of the modeled area, while the other two modules, LINRES and SATFLOW, stand for hydro-geological zone (Van der Lee and Gehrels, 1990).

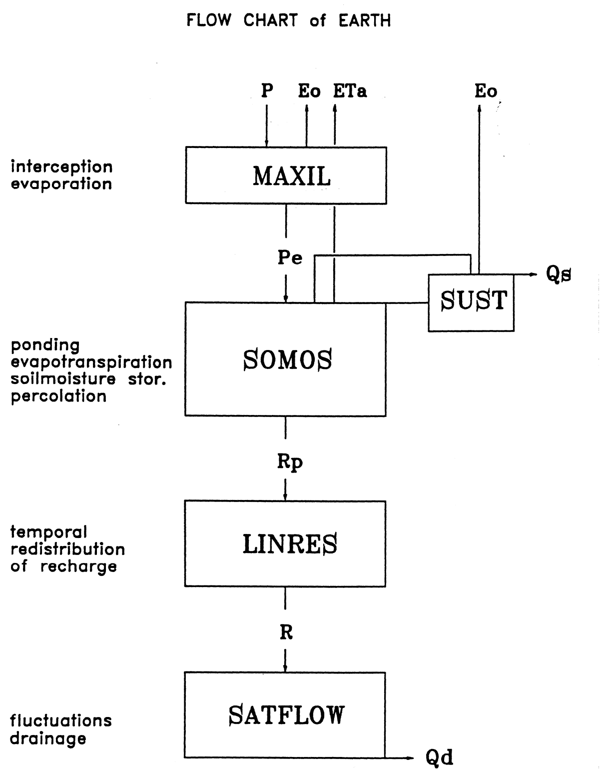


Figure 6.7: Flow chart of EARTH model (adapted from Van der Lee and Gehrels, 1990), where Pe is precipitation excess, AET is actual evapotranspiration rate, R_p is percolation flux below the root zone, Eo is evaporation, Qs is surface runoff and R is recharge flux.

I. MAXIL

In this module, precipitation excess, P_e (i.e., part of the precipitation that reaches the surface and infiltrates), is estimated using a parameter called MAXIL (maximum interception loss). MAXIL estimates the retained water by surfacial features like leaves, steams, depression storage, and etc. The module computes P_e according to:

$$Pe = P - MAXIL - Eo \tag{6.2}$$

where Pe is precipitation excess, P is total precipitation, MAXIL is total surface retention, and Eo is total evaporation. MAXIL is usually estimated by observation or can be used as an optimization parameter.

II. SOMOS

SOMOS stands for Soil Moisture Storage, which describes a reservoir with the root zone depth and capacity of $S_m - S_r$, where S_m and S_r are the saturated and the residual soil water content, respectively. In this module, changes in soil moisture storage are calculated from the infiltrated precipitation excess (P_e) after it is divided into actual evapotranspiration, percolation, ponding and surface runoff processes. Change in soil water storage is calculated from the water balance as:

$$\frac{dS}{dt} = Pe - AET - Rp - Eo(SUST) - Qs \tag{6.3}$$

where dS/dt is change in soil water storage over time, Pe is precipitation excess, AET is actual evapotranspiration rate, Rp is percolation flux below the root zone, Eo (SUST) is evaporated fraction of ponding water, and Qs is surface runoff.

In EARTH it is assumed that actual evapotranspiration can be determined from soil moisture and potential evapotranspiration. The relationship between actual evapotranspiration (AET) and soil moisture content (S) is often taken as being linear (Van der Lee and Gehrels, 1990), and AET is expressed as a function of potential evapotranspiration and soil moisture status as follows:

$$AET = ETp\left|\frac{S - Sr}{Sm - Sr}\right| \tag{6.4}$$

where ETp is potential evapotranspiration, S is actual soil water content, Sr is residual soil water content, and Sm is saturated soil water content.

Percolation (R_{p}) is the downward flow of water to the saturated zone and is expressed in EARTH by the Darcy equation as:

$$Rp = K\left|\frac{dhp}{dz} + 1\right| \tag{6.5}$$

where R_p is percolation, K is hydraulic conductivity which is a function of soil moisture, hp: pressure head, and dh_p/dz: gradient of the hydraulic potential, taken positive downward.

When water flow is purely controlled by gravity (in the case of soils at field capacity), equation 6.5 can be rewritten as:

$$Rp = K \tag{6.6}$$

$$K = Ks\left|\frac{S - Sfc}{Sm - Sfc}\right|^{n} \tag{6.7}$$

where Ks is saturated hydraulic conductivity, n is soil constants, S is actual soil moisture content, Sfc is soil moisture at field capacity, and Sm is maximum soil moisture content.

In EARTH, surface ponding is considered by the SUST module. When the water storage in SOMOS reaches saturation level, and rate of infiltration exceeds rate of percolation, surface ponding or runoff may occur. This module uses a parameter called $SUST_{max}$, which is the maximum amount of ponding water that can be stored at the surface. If $SUST_{max}$ equals zero, all ponding water will become surface runoff, while runoff will be minimum if $SUST_{max}$ is large.

$$SUST_t = S - Sm + SUST_{t-1} - Eo \qquad (6.8)$$

where S is actual soil moisture content, Sm is maximum soil moisture content, and Eo is open water evaporation from the pond, which is limited to the actual presence of water at the surface.

III. LINRES

LINRES (linear reservoir routing) redistributes the output of SOMOS (i.e., the percolation in time) using the parametric transfer function. For temporal redistribution of percolation, LINRES uses two parameters: number of reservoirs and unsaturated recession constant. This module spreads the output over time by routing the water through a series of conceptual linear reservoirs, and is useful for accurate optimization of the model to fit the calculated groundwater level with the measured one (Van der Lee and Gehrels, 1990).

IV. SATFLOW

SATFLOW (saturated flow model) is a one-dimensional parametric groundwater model, where the parameters have a semi-physical meaning (Van der Lee and Gehrels, 1990). It uses the output from previous modules, and recession and storage coefficient to transform the resulting flux into change in groundwater level:

$$\frac{dh}{dt} = \frac{R}{STO} - \frac{h}{RC} \qquad (6.9)$$

where h is groundwater level, RC is recession coefficient, R is recharge, STO is storage coefficient, and dh/dt is change in groundwater level with time.

In the EARTH model, the rate of lateral inflows in the aquifer is assumed to be equal to the rate of lateral outflows. The 1-D model allows for better understanding of the groundwater regime and for reasonable simulation of temporal flux variability; however, the disadvantage of all 1-D models is that they do not account for lateral fluxes (Lubczynski and Obakeng, 2004). EARTH generates total recharge instead of net recharge due to the fact that groundwater abstraction is not considered by the model.

Model input and calibration

Daily precipitation and potential evapotranspiration data are required to run EARTH, and optionally groundwater table and / or soil moisture data are needed for calibration purposes. The model needs 11 input parameters, which can be determined either from field measurements or derived from optimization of the model. Soil moisture data were not available in this study, and therefore daily groundwater level data are used for calibration. Soil moisture parameters that were estimated in previous studies and literature values were used as initial estimates in the model, and adjusted later during the calibration period. Daily precipitation and daily potential evapotranspiration were used as input to the model. Potential evapotranspiration was estimated by the FAO Penman Montheith method from climate data of nearby meteorological stations. EARTH was applied to two locations near the lake where groundwater levels were monitored during the field season July 2006 to September 2007.

Model calibration is a process of finding a set of parameters that produce simulated groundwater levels that match field-measured values (Anderson & Woessner, 1992). Calibration of the model can be performed manually or automatically. The EARTH model is calibrated manually in such a way that parameter values are initially assigned and adjusted in sequential model runs to fit the simulated and measured groundwater levels. Calibration of the model is conducted within the calibration target of $\pm$ 0.15m difference between simulated and measured groundwater heads. The well at Abadir was successfully calibrated well below the calibration target in which 97 % of the population was found within $\pm$ 0.10m difference between simulated and measured heads. Calibration of the well at Gofa Sefer area was conducted under the same calibration target in which 87 % of the population was found within $\pm$ 0.15m difference between the simulated and measured groundwater heads.

Results

The EARTH model provided daily recharge and actual evapotranspiration values as output. The calibration results from both locations show temporal variability of recharge, actual and potential evapotranspiration, and simulated and measured groundwater heads with daily rainfall data (Figures 6.8 - 6.11). The results show variability of daily actual evapotranspiration ranging between of 0.31 and 2.7 mm/day. For one hydrologic year (July 2006 - July 2007), mean actual evapotranspiration was estimated to be 1.1 and 1.0 mm/ day for the Gofa Sefer and Abadir sites, respectively. Computed actual evapotranspiration was relatively higher during the rainy seasons compared to the dry season, especially for Gofa Sefer, which can be explained by the dependency of evapotranspiration on soil moisture.

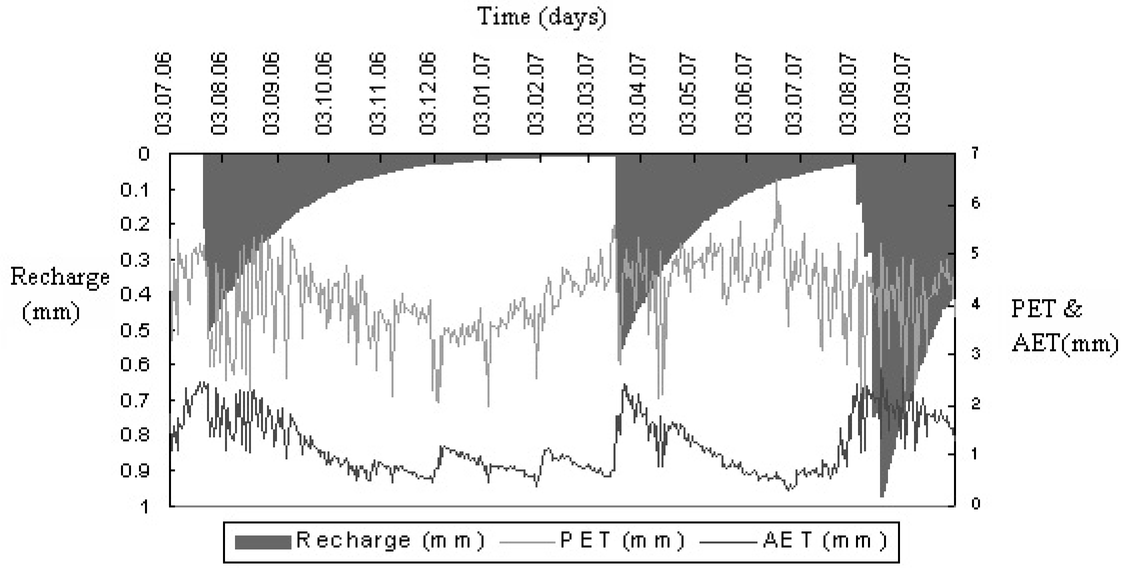


Figure 6.8: EARTH model output of daily recharge, AET, and PET, for Gofa Sefer site (east of Lake Beseka).

Temporal fluctuation of actual evaporation was not very significant for Abadir compared to Gofa Sefer. Differences in soil and land-use characteristics between the two locations are two of the several factors that influenced the observed variations. Luvisol is the type of soil around the Abadir area, characterized by well drained, very deep (up to 20 m) and fine texture, which is used for irrigated sugar cane plantation. Leptosol is the common soil type at the Gofa Sefer site, which is typified by shallow depth (< 2 m), coarse texture, and is covered mainly by open bushy woodland.

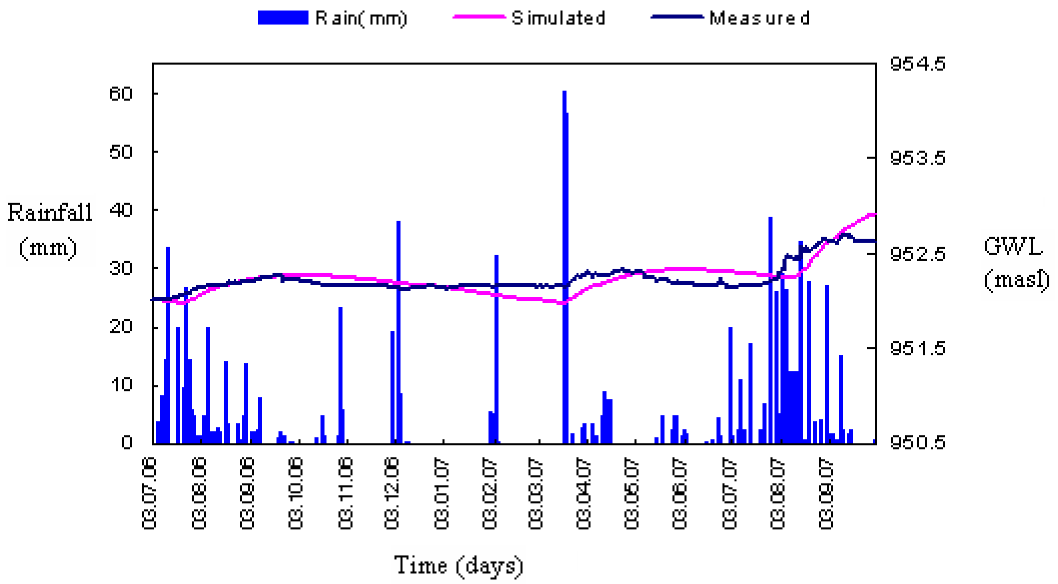


Figure 6.9: Simulated and measured groundwater heads and daily rainfall for a site at the Gofa Sefer (east of Lake Beseka).

According to the calibrated EARTH model, groundwater recharge takes place five weeks after the first rainfall event in the area following the dry season. This can be due to the fact that during the first five weeks, rainfall is mainly used by evapotranspiration and replenishment of soil water storage. However, it is important to note that actually groundwater levels start to respond or to rise in less than five weeks, in about three weeks, after the first rainfall event.

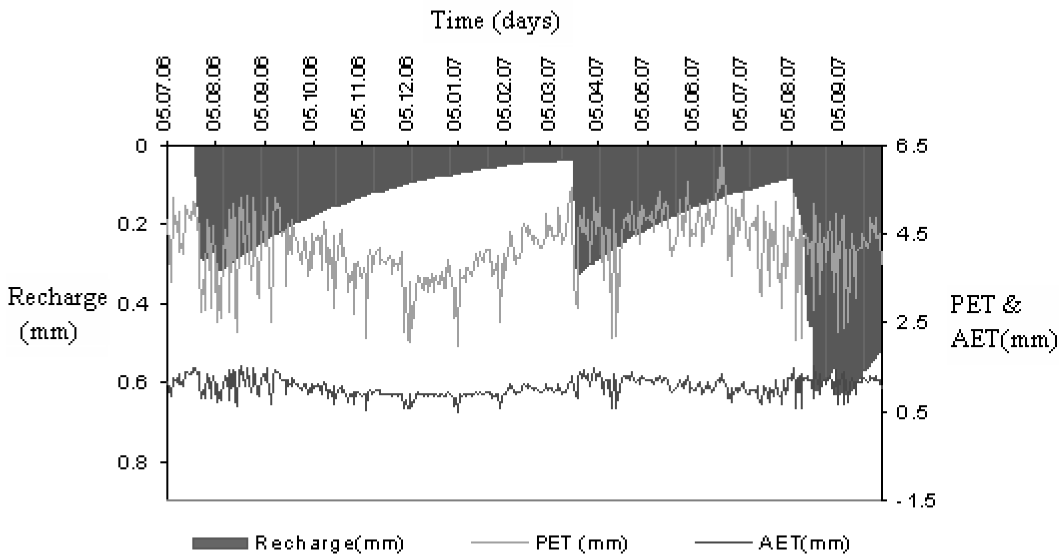


Figure 6.10: EARTH model output of daily recharge, AET, and PET for Abadir site (south of Lake Beseka).

This can be explained by the quick arrival of rainwater to the saturated zone through preferential pathways before the soil reaches its field capacity. Fractures and cracks can provide preferential flow paths for percolating water and create efficient shortcuts from the soil to the groundwater table. The velocity of preferential recharge, which preferentially takes place through macropores (cracks, fractures, fissures, solution channels), is often several orders of magnitude higher than diffuse recharge through the soil matrix (Sophocleous, 2004). Numerous studies have shown that in semi arid areas, very little flow percolates through the soil matrix to any significant depth (Murray, 1996).

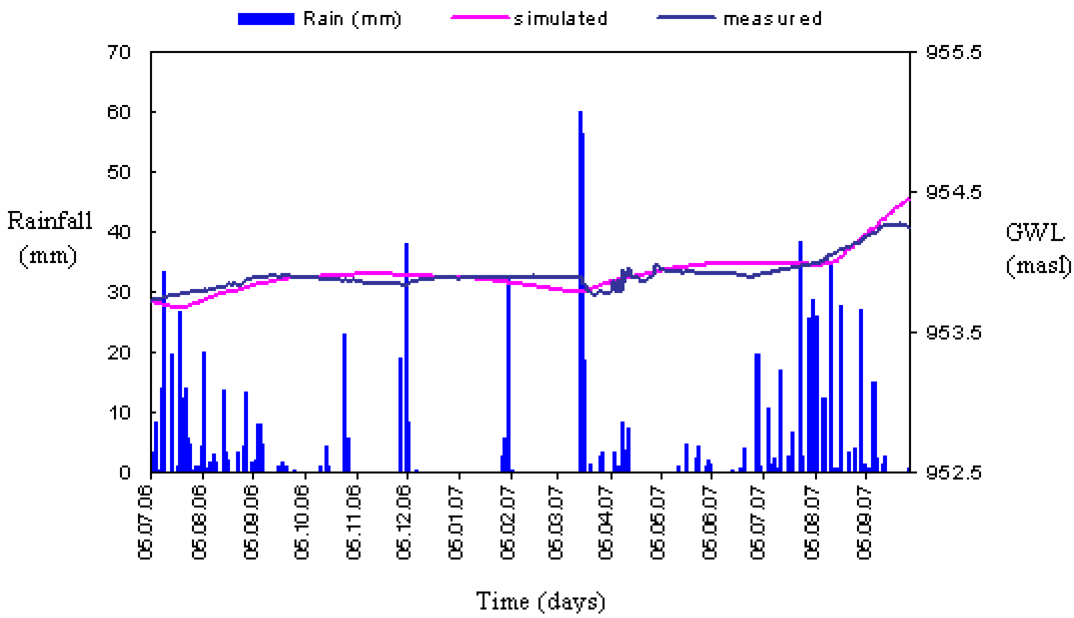


Figure 6.11: Simulated and measured groundwater heads and daily rainfall for Abadir site (south of Lake Beseka).

Groundwater recharge in the study area took place during both summer and spring rainy seasons (Figure 6.8-6.11). During these rainy seasons, high intensity rainfall events are important for creating favorable conditions for initiation of the recharge process. This is because large storm thresholds are required to overcome the substantial soil moisture deficit and to start direct recharge through the soil matrix (Lloyd, 1986). For the period of one hydrological year (July 2006 - July 2007), EARTH estimated an annual groundwater recharge of 47 mm for Gofa Sefer and 54 mm for Abadir site. Daily recharge varied from no recharge to 0.63 mm for Abadir, and from no recharge to 0.98 mm for Gofa Sefer. The results indicate spatial variation of recharge within these two locations despite the fact that they are not far from each other and are both located in similar climatic conditions. This difference can be explained by the heterogeneity in geomorphological features and subsurface hydrogeological conditions of the two locations.

The site in the Gofa Sefer area is characterized by open bushy woodland, a thin soil cover and a relatively shallow water table, while the Abadir area is covered by thicker soil, has a deeper water table, and is used for irrigation purposes. Assuming all other factors are similar, groundwater recharge is expected to be reduced in the presence of shrubs and bushes due to their deeper roots and longer vegetation period.

Table 6.1: water balance of EARTH model for simulation periods, July 3, 2006 – September 30, 2007 for Gofa Sefer site, and July 5, 2006 – September 30, 2007 for Abadir site.

Gofa sefer well	
Fluxes	Amount (mm)
Total precipitation	948.8
Precipitation excess	638.3
Actual evapotranspiration	542.3
Change in soil moisture	14.9
Recharge	81.1

Abadir well	
Fluxes	Amount (mm)
Total precipitation	948.8
Precipitation excess	638.3
Actual evapotranspiration	472.6
Change in soil moisture	81.3
Recharge	84.4

The estimated recharge amount in the EARTH model is lower for Gofa Sefer than for Abadir. On the other hand, analysis of actual transpiration values indicates that it is higher for Gofa Sefer than for Abadir area, which was as expected. However, it is important to note that the amount of recharge that seasonally replenishes the groundwater system of the area is governed not only by a single variable (vegetation difference), but also by the interplay of a variety of hydrogeological and geomorphological factors. The EARTH model has been used in the estimation of recharge in different arid and semi-arid parts of the world, and some of these results are listed below for comparison (Table 6.2).

Table 6.2: Recharge rate determined by EARTH for different aquifer types

Country	Aquifer type	Rainfall (mm/year)	Recharge (%)	Reference
South Africa (Western capetown)	Unconsolidated and fractered rock	380	0.5 - 3.4	(Conrad et al. 2004)
Palastine (Gaza)	Calcareous sandstone and sand	321	36.9	(Baalousha, 2003)
Spain (Sardon area)	Fractured granitic rock	500	4.8 - 44	(Lubcznski & Gurwin, 2005)
South Africa (Kalkveld area)	Fractured sandstone (Karoo formation)	559	7.4	(Usher et al. 2005)

Sensitivity analysis

The efficiency of parameter calibration would clearly be enhanced if it were possible to concentrate the effort on those parameters to which the model simulation results are most sensitive (Beven, 2000). Sensitivity analysis is conducted on the parameters of the calibrated model with the aim of determining those parameters to which the model is sensitive and to understand the behavior of the system being modeled. The sensitivity analysis is performed by estimating the rate of change in the output of a model with respect to changes in the model parameters. A normalized sensitivity index is calculated for each parameter by the following formula:

$$S = \frac{P_{10} - M_{10}}{B} \tag{6.10}$$

where P10 is result of a simulation with variable 10% increased, M10 is result of a simulation with variable 10% decreased, and B is result of the baseline simulation.

Each parameter was increased and decreased by 10 %, and the simulation output was compared to the calibrated model output (i.e., recharge), and the sensitivity index (S) is calculated based on equation 6.10.

Table 6.3: Sensitivity analysis for Gofa Sefer and Abadir sites (the degree of sensitivity is defined as: if $S<0.05$-low, $0.05<S<0.2$-medium, $0.2<S<1$-high, and $S>1$ very high sensitivity).

Parameter	Gofa Sefer site		Abadir site	
	S_index	Sensitivity	S_index	Sensitivity
Maximum soil moisture	0.48	High	1.04	Very high
Residual soil moisture	0.35	High	1.78	Very high
Intial soil moisture	0.30	High	2.65	Very high
Soil moisture at field capacity	-1.53	Very High	-4.68	Very high
Maxiumum surface storage	0.00	Low	0.00	Low
Maximum interception loss	-0.08	Medium	-0.28	High
Saturated hydraulic conductivit	0.01	Low	0.01	Low
Unsaturate Rec. Constant	-0.05	Medium	-0.07	Medium
Saturate Rec. Constant	0.00	Low	0.00	Low
Storage coefficent	0.00	Low	0.00	Low

The sensitivity analyses of all parameters indicate that the model is sensitive to soil parameters, maximum interception loss, and unsaturated recession constant. This can be observed from Figure 6.12, where residual soil moisture is typified by very high sensitivity, while unsaturated recession constant is characterized by medium sensitivity for Abadir site. A change in storage coefficient and saturated recession constant does not affect the recharge amount, but significantly modifies the level of the simulated groundwater curve. The computed groundwater level is highly sensitive to small changes in the storage coefficient and the saturated recession constant. The model output is insensitive to the maximum surface storage and saturated hydraulic conductivity.

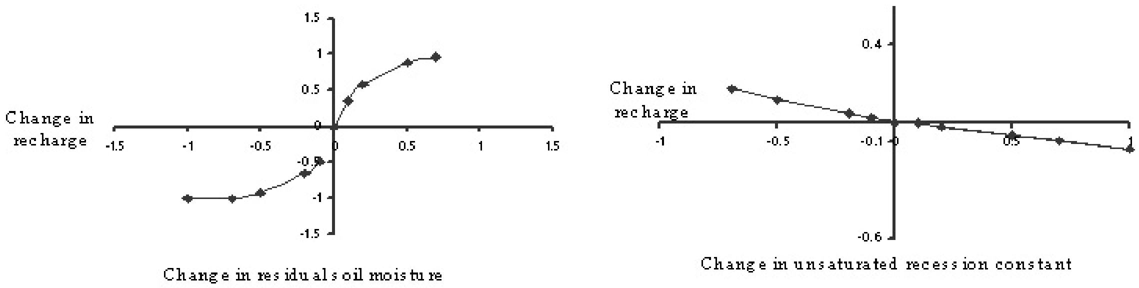


Figure 6.12: Sensitivity analysis of the EARTH model output (recharge) for change in initial soil moisture (left) and unsaturated recession constant (right) for the Abadir site.

6.3.4 Chloride mass balance method

Introduction

Since initially proposed by Eriksson and Khunakasem (1969), the chloride mass balance approach has been widely used for recharge estimation. It is based on the application of the mass conservation principle between the inputs of atmospheric chloride and subsurface chloride flux. Conservative tracers are concentrated in recharge as evaporation proceeds, so that the flux of tracer input at the surface equals the flux of tracers reaching the water table (Lerner et al., 1990). The chloride mass balance method is inexpensive, easy to use and a specifically suitable method for arid and semi-arid areas where chloride is significantly enriched in response to intense evaporation. The chloride mass balance approach has been used for estimating low recharge rates as low as 0.05 to 0.1 mm/year, mainly due to lack of other suitable methods (Scanlon et al., 2002).

Theoretical background

The chloride mass balance method estimates the total recharge in the saturated zone through both diffuse and preferential flow mechanisms based on the ratio of chloride concentration in rainwater to groundwater. The underlying assumptions for the application of chloride mass balance methods are: Chloride should behave conservatively under steady state conditions with no chloride sources and sinks, rainfall should be the only source of chloride to the groundwater system, and chloride should neither be taken up by vegetation nor leached from the aquifer formation (Eriksson and Khunakasem, 1969).

When the above-discussed assumptions are met, groundwater recharge in the system excluding the dry deposition term can be calculated as:

$$R = P.\frac{Cl_p}{Cl_{gw}} \tag{6.11}$$

where, P is precipitation (mm/yr), Clp is chloride content of rainfall (mg/l), R is recharge rate (mm/yr), and Clgw is chloride content of groundwater (mg/l).

The advantages of the chloride mass balance approach are the fact that it is not dependant on the method whether the recharge mechanism is preferential or diffuse, that the output reflects long-term averages and gives spatially integrated recharge estimates over areas up slope from the measurement point (Scanlon et al., 2002). The application of this method is limited to certain environments with no additional source or / and sink of chloride.

Results

Groundwater samples from 30 boreholes and spring sites in the watershed were analyzed for chloride concentration. Chloride composition of rainfall was not available for the study area, thus the average value of 0.47 mg/l Cl was adapted from the rainfall of the Akaki watershed (Demlie, 2006), which is located some 180 km south of the study area. The chloride mass balance approach yielded annual groundwater recharge values that ranged between 1.2 and 2.7 mm, accounting for only 0.2 to 0.5 % of the long-term mean annual precipitation.

The groundwater recharge values are too low and are unrealistic. Unrealistic recharge estimates can occur if the underlying assumptions for the application of the method are not satisfied in one or the other way. These low recharge values are estimated due to high chloride content of the groundwater system to be generated only through evaporative enrichment of the rainwater. The chloride content of the groundwater varied from 105 - 209 mg/l. The existence of another source of chloride beside the atmospheric chloride input to the groundwater system is evident. Chemical composition of the groundwater in the area is mainly controlled by hydrolysis of acidic volcanic rock, which is facilitated by CO_2 input from deeper sources, and significant rock-water interactions are to be expected in the area (chapter 4). According to Van Soest, et al. (2004) a possible source of chloride in volcanic areas can be the magma itself, chloride-rich crustal rock interacting with magma, or chloride-rich aquifer rock interacting with geothermal fluid.

The high amount of chloride in the groundwater system in the study area could be leached out from widely distributed acidic volcanic rock and / or related to hydrothermal out-gassing from deeper magmatic origin. A significant amount of chloride is believed to originate from deep thermal waters near volcanic centers (Chernet, 2001). In any possible ways that chloride might be originated, the presence of an additional source of chloride in the groundwater system other than precipitation limits the use of the chloride mass balance method as an independent recharge estimation method in the area.

6.4 Conclusions

Although quantitative information on recharge is critical for groundwater modeling and optimum resource management, reliable estimation in semi-arid and arid areas is neither straightforward nor easy (Lerner et al., 1990). In addition to temporal and spatial variability, of recharge in these regions, the absence of appropriate data which sufficiently represent the heterogeneity of the system further complicate the recharge estimation process. In semi-arid regions like Ethiopian rift valley, it is not only the amount of precipitation but also its temporal distribution that plays a crucial role in determining recharge amount. As explained by Larsson (1984), the amount of water that would be available to recharge groundwater system depends on annual total rainfall, intensity and duration of individual rainstorms, and on evaporation, which is a function of latitude, altitude and temperature.

The monthly potential evapotranspiration in the study area exceeds monthly rainfall amount for all months (Figure 6.3). Accordingly, it seems unlikely to have groundwater recharge in this area owing to the significant gap that exists between the available rainfall amount and the rate of evapotranspiration for each month. However, if the comparison is done on daily basis, surplus rainfall in excess of potential evapotranspiration is evident in many counts (Figure 6.1), which could be a possible source of recharge. A particular rainfall event may not cause recharge if it occurs over a longer period, but the same amount can generate recharge if it happens as heavy rainfall in short duration (Sophocleous, 2004). In dry areas, replenishment of groundwater often takes place during heavy rainfall events when there is surplus of rainfall to overcome the evapotranspiration. The importance of heavy episodes of unevaporated rain in recharging groundwater system was observed in the aquifer systems of Dijibouti, which is made up of lava and volcano-detrital deposits (Fontes et al., 1980). Here, the importance of rapid infiltration of heavy and unevaporated monsoon rains through fractured lavas was seen as an important recharge mechanism for the groundwater system of the Dijibouti region.

Recharge estimation is an iterative process, and it is sufficient to make approximate estimates and refine it with time (Lerner et al., 1990; Kamra, 2007). Recharge varies across the landscape, but finding a way to estimate and predict spatial and temporal variability and to regionalize point measurements remain a major problem in recharge assessments (Sophocleous, 2004). Field identification, quantification and development of statistical or other techniques for estimating recharge over an extended area remain a problem (Lerner et al., 1990). For lack of suitable means to quantify spatial distribution of recharge, groundwater modelers have traditionally assumed spatially uniform recharge rate across the water table equal to some percentage of average annual precipitation (Simmers, 1988).

The complexity of the recharge process and the absence of adequate temporal and spatial hydrological data for the study area limits the quantification of the recharge amount to the first estimate. Although the study area is characterized by a shortage of data, spatial qualitative hydrochemical and isotopic information from the previous parts of the study combined with temporal groundwater fluctuation data for a better understanding of the recharge process. Water table fluctuation, EARTH modeling and chloride mass balance were combined for estimating recharge of the study area.

Hydrochemical and isotopic composition of water bodies in the study area were used to discriminate groundwater origins and to understand groundwater flow paths (Chapter 4 and Chapter 5). The groundwater system in the western section of the watershed is characterized by relatively low EC and high HCO_3^- content compared to other anions. From the western section of the watershed toward the lake then to the eastern part, groundwater composition is hydrochemically evolved in a way that EC increases and SO_4^{2-} and Cl^- become significant. The presence of a high proportion of HCO_3^- in the western groundwater system could indicate a recharge area where the groundwater is continuously flushed by meteoric water and characterized by relatively quick circulating groundwater with low subsurface residence time. Significant amount of SO_4^{2-} and Cl^- in the groundwater in the eastern part of the watershed on the other hand illustrates the typical nature of a discharge zone where groundwater circulation is low and characterized by longer subsurface residence time.

The average isotopic composition of the groundwater in the study area is comparable to the weighted average isotopic composition of summer rains at the Addis Ababa station, indicating the importance of heavy summer rainfall in recharging the groundwater system. Moreover, the isotopic composition of the groundwater points out the insignificance of evaporative enrichment of rainwater before infiltration to the saturated zone. This suggests that preferential flow is important recharge mechanism in addition to the diffuse flow. The aforementioned two facts are in agreement with the output of the EARTH model. The model shows the importance of high intensity rains to overcome soil moisture deficits and to begin the recharge process. The significance of preferential flows as a recharge mechanism is also identified by the EARTH model, where groundwater level rises before the soil reaches field capacity.

Table 6.4: Annual recharge estimate by the water table fluctuation (WTF), EARTH modeling and the chloride mass balance (CMB) methods for sites.

Methods	Gofa Sefer well (East of the lake)	
	Recharge (mm/year)	% of Rainfall
WTF	42	6.6
EARTH	47	7.5
CMB	1.6	0.3

Methods	Abadir well (South of the lake)	
	Recharge (mm/year)	% of Rainfall
WTF	42.6	6.7
EARTH	54	8.6
CMB	1.5	0.3

The lack of enriched spring rains in the isotopically depleted groundwater system of the area suggests that its contribution to groundwater recharge is very small. However, model results indicate the presence of a recharging process during both rainy seasons. This could be due to the fact that isotopes normally reflect average recharge conditions for a considerably longer period of time than the model. Recharge could takes place during the spring rain seasons however, its amount could not be substantial enough to modify the isotopic signature of the groundwater system, which is characterized by isotopic composition comparable to that of the summer rainfall. Therefore, summer rainfall is the principal source of recharge to the groundwater system of the area.

The annual recharge amount was estimated for two locations using multiple methods, water table fluctuation, EARTH modeling and chloride mass balance. The study area is situated within the northern section of the Main Ethiopian rift valley, mainly characterized by shallow leptosol soil types, and covered by open bushy woodland. Most part of the study area has comparable geomorphologic and hydrogeologic features to those of the Gofa Sefer well monitoring site, taking the prevalent soil type, land use and land cover, aquifer properties and climate factors into consideration. Thus, the recharge value of seven percent of the long-term mean annual precipitation is taken as being representative of the study area.

7 GROUNDWATER MODELING

7.1 Introduction

Groundwater and surface water are not isolated components of the hydrologic system, but instead interact in a variety of physiographic and climatic conditions (Sophocleous, 2002). The hydraulic interaction of a lake with the surrounding aquifer system is a complex process, as it is dynamic and continuously changing with respect to space and time. The rate of exchange of water between the lake and surrounding groundwater system is governed by the position of the lake stage compared to the groundwater level, and by the type of material that separates them. Physical and chemical characterization of the surrounding groundwater system is essential for better understanding of the lake system, as the groundwater fluxes influence both water and nutrient budget of the lake.

Quantification of groundwater fluxes around a lake is imperative, and groundwater models are valuable and relevant tools for estimating groundwater flows to and from the lake. A groundwater flow model was developed for the watershed of Lake Beseka to compute the rate and direction of groundwater flow for the aquifer system of the area and for the hydraulically connected lake. MODFLOW, which incorporates the lake package, was used in this study to explicitly include the hydrological components of Lake Beseka into the groundwater flow model. The groundwater regime of the study area was simulated under steady and transient state conditions for the period from July 1976 to September 2007. For transient simulation, 375 monthly stress periods were set to represent the total simulation period of 31 years.

7.2 General Background

A model is an approximation of a very complex reality, which is used to simplify reality in a manner that essential features and processes relative to the problem at hand are represented (Konikow and Reilly, 1998). Bear et al. (1992) defined a model as a simplified version of a real world that approximately simulates the relevant excitation-response relations of the real world system. When the hydrological system is the model of interest, the hydrological cycle or part of it is the natural or real system that would be represented by the model. The natural system can be represented in either physical or mathematical models. A physical model is the tangible representation of the natural system, while a mathematical model is an explicit sequential set of equations, numerical and logical steps that convert numerical inputs into numerical outputs (Dingman, 2002). Physical models such as laboratory sand tanks simulate groundwater flow directly, whereas mathematical models simulate groundwater flow indirectly by means of a governing equation that represents the physical processes that occur in the system (Anderson & Woessner, 1992).

Due to limitation of hydrological data, hydrological models have been used as a means of extrapolating already available measurements in space to where measurements are not available, and in time when measurements are not possible (Beven, 2001). Although more interplay has occurred in the last decades, two classical hydrological models, deterministic and stochastic, were developed in two separate schools during the 1960s and 1970s (Abbott and Refsgaard, 1996). Deterministic models are based on conservation of mass, momentum and energy, and describe cause and effect relations (Konikow and Reilly, 1998). Stochastic models may also be based on physical processes, but they allow for some randomness or uncertainty in the possible model outcomes due to uncertainty in the input variables, boundary conditions or model parameters (Beven, 2001). Deterministic hydrological models can be classified into lumped conceptual models and distributed physically based ones. Lumped models consider water flow to take place between storage units, and treat the watershed as a single unit where parameters and variables represent average values over the entire watershed (Abbott and Refsgaard, 1996). Distributed models, on the other hand, calculate flow of water directly from the governing partial differential equation and account for spatial variability of properties in the model domain. Most groundwater models in use today are deterministic mathematical models (Konikow and Reilly, 1998), which are generally accepted as physical based ones.

Groundwater models form an integral part of decision-support systems, and are used in investigating important features of groundwater systems around the world, and can be applied in wide variety of hydrogeological conditions (Mandle, 2002). Groundwater models are being increasingly used, as it is generally recognized and accepted that numerical simulation can provide a sound basis for important decisions making processes (PDP, 2002). Middlemis (2001) noted that groundwater models are relevant and useful scientific and predictive tools for developing management plans.

Even though many of the groundwater model applications have included crude, approximate descriptions of the hydrogeological setting and groundwater flow, they serve the purpose to provide valuable information for decision making in connection with groundwater planning, management and protection (Abbott and Refsgaard, 1996).

7.2.1 Groundwater Flow Models

Hydrogeological investigations are uncertain, as it is not possible to observe subsurface geological features and underground groundwater flow systems directly. Like any other hydrological system, a groundwater flow system is dynamic, continuously changing with respect to time and space, which makes the situation even more complex. Despite all the advances that have been made in remote sensing, ground-probing radar and other techniques for exploring the subsurface, our knowledge of what goes on underground is still very limited (Beven, 2001). As the accessibility of a groundwater system is very limited, Kinzelbach et al. (2003) described groundwater models as indispensable tools for bringing all available data together to derive a sound holistic picture of the system on a quantitative base. Groundwater models provide a scientific means to synthesize existing data into numerical characterization of a groundwater system (Middlemis, 2001).

Groundwater models are created by translating conceptual understanding of the groundwater system into a mathematical framework into which appropriate existing data are incorporated (PDP, 2002). Middlemis (2001) defined a groundwater model as a computer-based representation of the essential features of a natural hydrogeological system by using the law of science and mathematics. Thus, the applicability or usefulness of a model depends on how closely the mathematical equations approximate the physical system being modelled (Mandle, 2002). Due to the complexity of the real physical world, simplifications are introduced to groundwater models in terms of assumptions to obtain a quantitative solution for a given problem. These assumptions relate, among other factors, to the geometry of the investigated domain, the way various heterogeneities will be smoothed out, the nature of the porous medium (homogeneity, isotropy), the properties of fluids involved, and the type of flow regime under investigation (Bear et al., 1992).

Due to these assumptions and many uncertainties in the input data, a model must be viewed as an approximation and not as an exact duplication of field conditions (Mandle, 2002). Anderson & Woessner (1992) indicated that modeling is an excellent way to organize field data, but it is only one component of the general hydrogeological assessment, not an end in itself. Thus, groundwater flow models are not an alternative for field investigation but invaluable tools to synthesize the existing hydrogeological understanding of the given system. Groundwater models can be used as a predictive tool to predict the future of groundwater flow systems or as a generic tool to investigate groundwater flow processes.

Groundwater flow models are used to calculate the rate and direction of groundwater movement for both aquifers and hydraulically connected surface waters. They simulate groundwater flow by means of the governing equation that represents the physical process in the system, together with equations that describe heads or flows along model boundaries (Anderson & Woessner, 1992). Typical outputs of groundwater flow models are hydraulic heads and flow rates, which are in equilibrium with hydrogeological conditions (hydrogeologic framework, hydrologic boundaries, initial and transient conditions, hydraulic properties and sources or sinks).

The subsurface environment constitutes a complex, three-dimensional, heterogeneous hydrogeologic setting, which strongly influences groundwater flow processes (Konikow and Reilly, 1998). Groundwater flow equations that describe three-dimensional groundwater movement in a porous medium are developed from the application of Darcy's law and the continuity equation. Darcy's law indicates the rate of water that flows through porous media is proportional to the hydraulic head gradient, properties of the porous media, and properties of the water (Konikow and Reilly, 1998). The concept of the continuity equation is explained that the sum of all inflows into and out of the cell equals the rate of change in the storage within the cell (McDonald & Harbaugh, 1988). The water balance equation is mathematically combined with Darcy's law to derive the following governing equation that describes the three-dimensional subsurface groundwater flow process in porous medium (Anderson & Woessner, 1992).

$$\frac{\partial}{\partial x}(Kx\frac{\partial h}{\partial x})+\frac{\partial}{\partial y}(Ky\frac{\partial h}{\partial y})+\frac{\partial}{\partial z}(Kz\frac{\partial h}{\partial z})=Ss\frac{\partial h}{\partial t}-R^{*} \tag{7.1}$$

where Kx, Ky and Kz are components of the hydraulic conductivity tensor [LT^{-1}], h is hydraulic head [L], t is time [T], Ss is specific storage [L^{-1}], R^* is general sink / source term that is intrinsically positive and defines the volume of inflow to the system per unit volume of the aquifer per unit time [T^{-1}].

The governing equation can be solved either through analytical or numerical methods. Both methods solve the equation based on their approaches, assumptions and applicability. Analytical methods solve governing equations into exact solutions using classical mathematical approaches. These analytical solutions incorporate many assumptions, resulting in simplification of aquifer homogeneity, and into one- or two-dimensional groundwater flow (Anderson & Woessner, 1992). Because of these inherent simplifications using analytic models, it is not possible to account for field conditions that change with time and space (Mandle, 2002). Numerical methods, in contrast, allow a representation of the physical complexity, and yield approximate solutions through discretization of the space and time domains (Konikow and Reilly, 1998). They solve partial differential flow equations through numerical approximations using matrix algebra and discretization of the modeled domain (Middlemis, 2001).

Numerical modeling of groundwater flow in fractured aquifer presents a special challenge, as it is not that clear how to describe heterogeneity associated with fractures. Fractured media consist of solid rock with primary porosity cut by a system of cracks, microcracks, joints, fracture zones and shear zones, which create secondary porosity (Anderson & Woessner, 1992). Three common approaches are available to numerically model fractured aquifers: (i) equivalent porous medium approach (EPM), (ii) dual porosity approach (DP), and (iii) discrete fractures approach at the scale of individual fractures (DF) (Cook, 2003).

Fractured material is represented as an equivalent porous medium (EMP) by replacing the primary and secondary hydraulic conductivity distributions with a continuous porous medium that has so called equivalent or effective hydraulic properties (Anderson & Woessner, 1992). The EPM approach defines a fractured system as a single continuum or series of continua, where parameter values are affected by the presence of fractures, but fractures are not modeled explicitly (Stafford et al., 1998). This approach needs few parameters, and these parameters can be determined either from aquifer testing or calculated from detailed field descriptions of fractures (Cook, 2003). If a rock matrix has significant primary permeability, the dual porosity (DP) approach can be employed. This approach can be important when dealing with systems involving high matrix porosity, and when the fractured rock system is represented by one flow equation and matrix blocks are represented by another (Cook, 2003). Exchange of fluids between the two flow zones is considered.

The discrete fracture (DF) approach, on the other hand, is typically applied to fractured media with low primary permeability, as the model assumes that water moves only through the fracture network (Anderson & Woessner, 1992). Flow through each fracture is treated in these models as being equivalent to flow between two uniform plates with separation equal to the aperture of the fracture; the models allow an explicit characterization of fracture properties such as aperture, orientation, degree of fracture interconnection and length (Cook, 2003). Some of these parameters, particularly fracture aperture and degree of fracture interconnection, are difficult to measure, and may contain a high degree of uncertainty (Stafford et al., 1998).

According to NRC (1996), geology of fractured rock, scale of interest and purpose of the model can be considered as important factors when conceptual models are developed for fractured rock aquifers. These factors could play an important role when selecting suitable approaches to numerically represent fractured aquifers by one of the three aforementioned methods. Equivalent continuum approaches can be appropriate for answering questions that relate to averaged volume behavior; however, more detailed modeling approaches are required when dealing with solute transport, in which migration pathways are explicitly controlled by heterogeneity of fractured rock aquifer (Cook, 2003).

Long et al. (1982) as cited in Stafford et al. (1998)) determined features of fractured systems that increase the likelihood of EPM approach as being appropriate for a given system. These characteristics, according to this work, include a high density of fracture that creates connected fractures and thus connected pathways, and relatively constant fracture apertures to prevent only few fractures from controlling most of the flow. When fractures are few and far between, and the un-fractured block hydraulic conductivity is low, the EPM method may not be appropriate (Anderson & Woessner, 1992). In addition to the geology of fractures, the scale of interest can also play an important role in selecting the appropriate approach. As the scale increases, the more appropriate it is to employ equivalent porous media modeling approaches, where extensive regions of an aquifer are represented by uniform hydrogeologic properties (Cook, 2003). EPM approaches may adequately represent the behavior of regional flow systems, but poorly reproduce local conditions (Anderson & Woessner, 1992).

In the case where fracture densities are very high, and intermediate matrix units are very small, it may be possible to treat the system as one continuum where hydraulic storage and transmissivity are represented by a lumped sum accounting for both fractures and matrix blocks (Cook, 2003). When effective parameters can be defined, standard finite difference and finite element codes used for porous media may be applied to the EPM that represents the fractured system (Gerhart, 1984; as cited in Anderson & Woessner, 1992). The MODFLOW model (McDonald & Harbaugh, 1988) is an example of such implementation for groundwater flow simulations, and has been successfully applied to fractured rock aquifers (Cook, 2003). Dense fracturing and connectivity of fractures in the Lake Beseka watershed provide the base for applying the EPM approach. This approach is commonly implemented for modelling at watershed scale, where local matrix / fractures interactions are neglected.

The modular, three-dimensional finite difference groundwater flow model (MODFLOW) of the U.S. Geological Survey (USGS) is the most widely used numerical model in the world for describing and predicting groundwater flow systems. MODFLOW uses partial differential equation to describe three-dimensional groundwater movement of constant density through porous material (McDonalld and Harbaugh, 1988). It has a modular structure and uses the finite difference method with a block-centered approach to solve the governing groundwater flow equation (Konikow and Reilly, 1998). It is public domain and considered as the standard code for aquifer simulation. The code enjoys great popularity, as it has been verified against analytical solutions and has been widely used to successfully simulate a wide range of hydrogeological conditions across the world (Middlemis, 2001). While MODFLOW is well tested, it is limited in its treatment of groundwater- surface water interaction

(Council, 2005). It also lacks functionality to simulate surface water runoff and unsaturated zone flow processes.

Since its original development in the 1980's, the USGS has released four main MODFLOW versions, MODFLOW-88, MODFLOW-96, MODFLOW-2000 and MODFLOW-2005. The versions 88 and 96 were originally designed solely to simulate saturated groundwater flow through porous media. Since they were designed only to solve groundwater flow equations, additional related equations (transport flows, parameter estimation) had to be solved using separate computer codes (Chiang, 2005). After extension of the modular structure of MODFLOW to integrate these additional equations, the 2000 version was released, which has the capacity to solve the main groundwater flow equation and sets of related equations (Harbaugh, 2005). MODFLOW-2005 has a similar modular structure to that of the 2000 version regarding incorporation of transport flow equations and parameter estimation, but has different ways of handling internal data. Although input and output systems of the code were designed to permit flexibility, usability and ease of interpretation of model results can be enhanced by using one of several commercially available preprocessing and post processing packages (Konikow and Reilly, 1998).

7.3 Surface water – groundwater interactions

Groundwater and surface water are not isolated components of hydrologic system but a linked component of a hydrologic continuum, which interact in a variety of physiographic and climatic landscapes (Sophocleous, 2002). Surface water is commonly hydraulically connected to groundwater, but interactions are difficult to observe and measure, thus they have often been ignored in water management considerations (Winter et al., 1998). Surface-groundwater interaction is the interplay between water on and beneath the land surface, including the flow of surface water into the groundwater system and vice versa (Mace et al., 2007). Surface water bodies such as lakes, streams and wetlands form an integral part of most groundwater flow systems, and play an important role in the water and chemical budgets of the system.

Interactions between groundwater and surface water are complex, and to understand these interactions in relation to climate, landform, geology, and biotic factors, a sound hydro- ecological framework is needed (Sophocleous, 2002). Winter

(2000) also emphasized the importance of understanding the effect of physiography and climate on surface water runoff and groundwater flow system to fully recognize their interactions. He explained that the movement of water between groundwater and surface water is controlled by the position of the surface water within groundwater flow systems, geologic characteristics of surface water beds, and climate. In a local groundwater flow system where water moves from the recharge area to the next adjacent discharge area, the extent of interaction with surface water is expected to be higher than with the regional groundwater flow systems. This is due to the dynamic nature of the shallowest local groundwater flow system compared to the deeper one with longer flow paths.

The groundwater flow pattern is controlled not only by the configuration of the water table, which is commonly a subdued replica of the land surface, but also by the distribution of hydraulic conductivity in the rocks (Sophocleous, 2002). Groundwater seepage distribution toward the surface water and vice versa is controlled by the permeability of the bed at the interface, which, in turn, is influenced by the geological characteristics of the bed unit (lithology and structures). The geologic framework affects the flow paths through which groundwater flows, while the type of sediments at the interface between groundwater and surface water can influence the spatial variability of discharge to surface water (Alley et al., 2002). In addition to topographic and geologic effects, seasonal variation of climatic factors affects the rate and direction of surface-groundwater interactions by determining the amount of water that would be available to move toward or / and remove from the aquifer system. Nield et al. (1994) used numerical models to examine groundwater flows in a vertical section near surface water bodies, such as lakes and wetlands, and identified a number of flow regimes. These flow regimes are identified by their characteristics, which are controlled by regional water table gradients, recharge to the aquifer, water body length, aquifer anisotropy, and hydraulic resistance of the bottom sediments.

Although there has been some progress in recent years in the development of integrated surface-groundwater models, surface water and groundwater have been traditionally considered as two separate entities. As a result, separate models have been developed and used to evaluate surface and groundwater resources independently. In surface water models, the interaction between surface and groundwater is often represented in "black box" source /sink terms, while in groundwater models, surface water is represented as an infinite source of water (Genetti, 1999). The surface-groundwater interaction forms the most complex, sensitive and uncertain parts of a model, as the flow processes commonly involve consideration of the unsaturated and the saturated zone flow processes (Middlemis, 2001). To ease numerical simulation, simplifying assumptions are made from the groundwater perspective, where the unsaturated flow system is not addressed, and leakage from surface water to an aquifer is assumed to be instantaneous, suggesting that no head loss occurs in the unsaturated zone (Genetti, 1999).

In groundwater models, surface-groundwater interactions has been treated as boundary conditions varying from the simple specified head or flux boundaries to head-dependent flux boundaries. In the case of the specified head boundary, the head at the surface water is specified as the elevation of the water surface, and the flow across the boundary is computed using Darcy's law, whereas for specified flux boundary, the flow across the boundary is specified and the model calculates head value at the boundary. Despite their simplicity, both approaches suffer from the drawbacks that hydraulic conductivity at the interface cannot be determined precisely, and they do not take bottom elevation of surface water into consideration (Middlemis, 2001; Genetti, 1999). Thus, flow across the boundary interface can be potentially unlimited. The head-dependent flux boundary is a value-dependent boundary condition where flow across the surface-groundwater interface is computed based on relative water levels between the aquifer and the surface water, and conductance term at the boundary.

Because of the limitation of the model, surface-groundwater interactions have been treated by MODFLOW in a simple way. The river package is a widely used head-dependent boundary in MODFLOW for simulating flows between surface water features and groundwater systems. In recent years, more complex groundwater-surface water simulators have been used, which explicitly couple surface and groundwater processes such as groundwater flow, overland runoff, wetland flow, river flow, and lake flow. However, these simulators are significantly more complex than the river package of MODFLOW, and more data demanding (Lubczynski and Gurwin, 2005), which makes it more difficult to use (Council, 2005). The most widely known coupled surface-groundwater numerical models include the MIKE-SHE (from DHI, Inc.), the HMS-

MODFLOW (from HydroGeologic, Inc.), the ISGW (from SDI Environmental Services), and others.

7.3.1 Lake-groundwater interactions

An important aspect of lake hydrology is its interaction with the surrounding environment, as lakes interact with the atmosphere, subsurface, and other surface water features (Healy et al., 2007) (Figure 7.1). The hydrological regime of a lake is mainly governed by the position of the lake within the regional groundwater flow systems. Studies on lake-groundwater interactions from both mathematical simulations of idealized situation and field studies have shown that permanent lakes are almost always the site of groundwater discharge, i.e., smaller lakes for local flow systems and larger lakes for intermediate or regional systems (Dingman, 2002). Lakes interact with the surrounding groundwater system in three main ways, i.e., some receive groundwater inflows, some have seepage loss to the groundwater; and others receive from and lose to the groundwater system (Winter et al., 1998). The direction and magnitude of seepage between a lake and the adjacent aquifer system depends on the relation between the lake stage and the hydraulic head in the groundwater system, both of which can vary substantially in time and space (Merritt and Konikow, 2000).

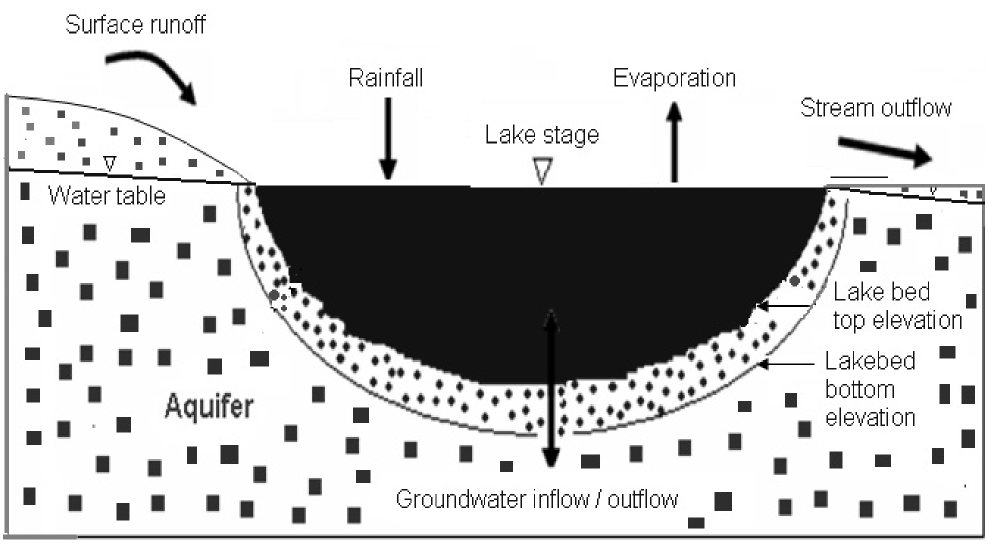


Figure 7.1: Water budget component of lake

Lakes have been traditionally simulated in standard groundwater flow models (MODFLOW) either using original MODFLOW boundary conditions (constant head, river, etc.) or using high conductivity and high storage aquifer cells. In early simulations, lake stages were specified as specified head or head dependent flux boundaries (Hunt et al., 2003). In the standard groundwater flow models (MODFLOW), original boundary conditions need prior knowledge of the lake stage to be used as an independent variable. The implicit assumption of this procedure is that the lake stage does not vary as a result of leakage through the lakebed or as a result of other stresses (Merritt and Konikow, 2000). Thus, the method neither provides a way to automatically update stages due to changing water fluxes to and from the lake (Council, 1999) nor recognizes the true nature of connections between the lake and surrounding groundwater system (Anderson, 2002).

The other traditional way of simulating lake-groundwater interactions is the "high hydraulic conductivity" technique where a storage value of 1.0 and a high hydraulic conductivity value are specified for lake cells in the model. In this technique, the lake stage is calculated using these high parameters (hydraulic conductivity and storage value) in the same equation to compute aquifer heads in the other part of the model domain. The principal difficulty of using this technique is that it has problems in representing lakebed leakance, stream-lake connection, and questions of solution stability (Merritt and Konikow, 2000).

The need for improvement in the way in which groundwater models handle surface water led to the development of specialized software packages for MODFLOW that address the dynamic exchange of water among groundwater, lakes and streams (Anderson, 2002). In line with this, sophisticated lake packages have been developed for MODFLOW by different authors: LAK1 (Cheng and Anderson, 1993), LAK2 (Council, 1999), and LAK 3 (Merritt and Konikow, 2000). All packages use the same underlying approach to solve lake stages and to compute water budgets; groundwater flows are calculated using the lake stage and the aquifer head in each cell adjacent to the lake (Hunt et al., 2003).

LAK1 (Cheng and Anderson, 1993) is the original lake package developed for use with MODFLOW to simulate lake-groundwater interactions. It computes lake stages as a transient response of precipitation, evaporation, stream flow and groundwater flux based on the given lakebed morphology and hydraulic conductivity. LAK 2 (Council, 1999) includes all features of LAK1 with an additional capacity of simulating steady state lake stage and a better description of stream outflows and stage-outflow relationships. In addition to having all LAK2 features, LAK3 (Merritt and Konikow, 2000) can be used with MOC3D to simulate solute transport, solute concentrations and fluxes. When using MODFLOW[9], lake packages are superior to other methods (constant head, head-dependent flux or high conductivity lake) because they allow explicit inclusion of surface water flow to and from the lake with powerful post simulation reporting features (Hunt et al., 2003).

The lake package[10] is a package that was developed for MODFLOW to simulate lake-groundwater interaction. It provides features for simulating lakes with variable and potentially unknown lake water levels resulting from environmental stresses. The package is designed not only to serve as boundary conditions like many standard MODFLOW packages (river, recharge, general head boundary) but also to calculate the water budget of the lake during the simulation, allowing the stage of the lake to vary in response to budget changes (Council, 1999). It provides four simulation modes to simulate lakes with known and unknown stages. The first two fixed stages modes (constant stage and interpolated stage) are used when lake stages are known and specified as independent variables to the lake package, whereas in the other two stage simulation modes (steady state and transient stage solution), lake stages are unknown and specified as dependent variables to the lake package.

For the constant stage mode, the lake package behaves like the river package, as lake stages are held constant during stress periods. In the interpolated stage mode, the lake package carries out simulation in a way similar to the reservoir package, with lake stages being specified but varying linearly during stress periods. In the third simulation mode, i.e., the steady state stage solution, that lake stage is computed by the lake package that represents an equilibrium condition of the lake influxes (from precipitation, runoff, stream inflow and groundwater fluxes) and lake outfluxes (to evaporation, streams and groundwater fluxes). During the transient stage solution, lake

[9] An executable version of MODFLOW-96 with the lake Package (LAK2) can be downloaded from http://www.geotransinc.com/flake.html

[10] The phrase lake package will be used to specifically specify LAK2, which was developed by Gregory W. Council, HIS GEOTRANS, A Tetra Tech Company in 1999. LAK2 will be referred to as the lake package.

stages are computed as a function of time and volume (area), and the water budgets of the lake are updated each time when the stages change (Council, 1999). The stages of the lake are adjusted at the end of each time step, and separate water budget calculations are performed for each time step.

Quantifying the hydraulic relation between a lake and the adjacent aquifer requires a method for estimating the amount of water exchanged between the two water bodies by seepage through the material that separates them (Merritt and Konikow, 2000). The rate of exchange of water between the lake and aquifer is determined by the position of the lake stage with respect to the adjacent groundwater head and lakebed conductance (Council, 1999). In the lake package, seepage from the lake to the groundwater is possible when groundwater head falls below the lake stage, and the amount of leakage is controlled by the relative difference of the heads. Groundwater flows from the aquifer to the lake when lake stage drops below the lakebed top elevation and the groundwater head is above the top elevation of the lakebed (Council, 1999). The lake package needs an explicit specification of the lakebed top and bottom elevation (Figure 7.1), and either the hydraulic conductivity or the hydraulic conductance of the lakebed.

The flow across the lake-groundwater interface is calculated by the lake package from:

$$q_i = COND_i(S - h_{ijk}) \tag{7.2}$$

where q_i is flow (seepage) between lake and aquifer [LT^{-3}], S: lake stage [L], h_{ijk} is groundwater head in cell i, j, k (row, column and layer) [L], and $COND_i$ is hydraulic conductance of the lake bed which is calculated as:

$$COND_i = KA/(TOP_i - BOT_i) \tag{7.3}$$

where A is surface area of the lake cell (projected on horizontal plane) [$L^{2]}$, K is hydraulic conductivity of the lake bed [LT^{-1}], and TOP_i and BOT_i is top and bottom elevation of the lake bed [L].

The volumetric water budget of the lake is calculated by the following general equation:

$$\partial V / \partial t = Q_p + Q_{ro} + Q_{strin} + Q_{gw(S)} + Q_{e(S)} + Q_{strout(S)} \qquad (7.4)$$

where $\partial V/\partial t$ is rate of change of the lake volume with time, Q_p is precipitation, Q_{ro} is runoff, Q_{strin} is inflowing streams, Q_{gw} is groundwater, Q_e is evaporation and Q_{strout} is out flowing streams, which are specified to the lake package as negative values. Fluxes Q_g, Q_e and Q_{strout} are functions of the lake stage (S), while the others are independent.

Steady state lake stages can be calculated to determine the approximate average stage that would result from constant (or averaged) stresses, where the stresses are precipitation, evaporation, runoff, etc. (Council, 1999). The steady state lake stage represents the condition at which lake inflows and outflows are in equilibrium, and the rate of change of the lake storage (volume), i.e. $\partial V/\partial t$ in equation 7.4, is zero. In contrast, the rate of change of lake storage during the transient stage solution is different from zero, and it is computed and updated each time when the lake stage changes.

7.4 Model Application

7.4.1 Model construction

Model construction is the process of transforming the conceptual understanding of the study area into a mathematical framework to simulate groundwater heads and flows. The required outcome is an interactive model with features representing the hydrogeological framework, hydraulic properties, hydrologic stresses and boundary conditions (Mandle, 2002). The numerical groundwater flow model MODFLOW (McDonald and Harbaugh, 1988) was used in this study to model the watershed of Lake Beseka. Data preprocessing and post processing were carried out by using the PMWIN software (Chiang, 2005). Model construction included designing of the grid, setting of boundary and initial conditions, assigning aquifer geometry and properties and recharge values and time parameters.

Grid design

In numerical models, the continuous problem domain is replaced by a dicretized domain, which consists of an array of nodes and associated finite difference blocks or finite elements (Anderson & Woessner, 1992). In MODFLOW, the aquifer system is divided into a rectangular mesh of blocks which are organized by rows, columns, and layers. Each block is known as a "cell" and the center of each cell corresponds to a "node", where the head is calculated. Taking topography, size and hydrogeology of the lake watershed and available data into consideration, the model area was dicretized into 160 rows and 120 columns with a grid size of 250 m * 250 m. In designing the grid, the length-to-width ratio (aspect ratio) of the cells should be kept as close to one as possible to avoid numerical instabilities or errors (Konikow and Reilly, 1998). Active cells that fall within the lake watershed form an integral part of the numerical model. Lake Beseka is located in the eastern part of the watershed and is represented by 692 lake cells that have direct contact with the main groundwater flow model. However, during simulation, the number of lake cells varies with time.

Aquifer geometry

Aquifer geometry refers to the specification of the top and bottom elevation of the aquifer as well as thickness and areal extent of the water-bearing layer. The model layer is equivalent to the hydrostratigraphic unit, which comprises geological units of similar aquifer properties. Aquifer geometry information is used by the model to define aquifer properties (transmissivity and storativity), and to decide whether the given aquifer is confined or unconfined by comparing the top elevation of the layer with simulated heads (Middlemis, 2000). The bottom layer elevation is used by the model to identify when a cell is drained or goes dry. Existing lithological logs of boreholes, groundwater levels and geophysical data were analyzed to understand the hydrogeological setup of the study area. A combination of fractured and weathered basalt, welded tuff with pumice and sediments form the main water-bearing formation in the area. The lake watershed is characterized by a single unconfined hydrostratigraphic unit, which is represented by a two-dimensional horizontal model. The DEM of the study area was merged with the bathymetry of the lake and specified as the top elevation of the aquifer. The elevation of the bedrock derived from the lithological logs of boreholes and geophysical data was taken as the bottom elevation of the aquifer. The lake watershed is characterized by an average aquifer thickness of 25 m, varying from 9 m in the east to 46 m in the west.

Boundary and initial conditions

To obtain a unique solution of a partial differential equation corresponding to a given physical process, information about the physical state of the process should be supplied through boundary and initial conditions (Konikow and Reilly, 1998). Setting up boundary conditions is a critical step during groundwater model design, as boundaries have a profound effect on the computation of flow velocities and heads. Anderson & Woessner (1992) described boundary conditions as mathematical statements specifying the dependent variable (head) or the derivative of the dependent variable (flux) at the boundaries of the problem domain.

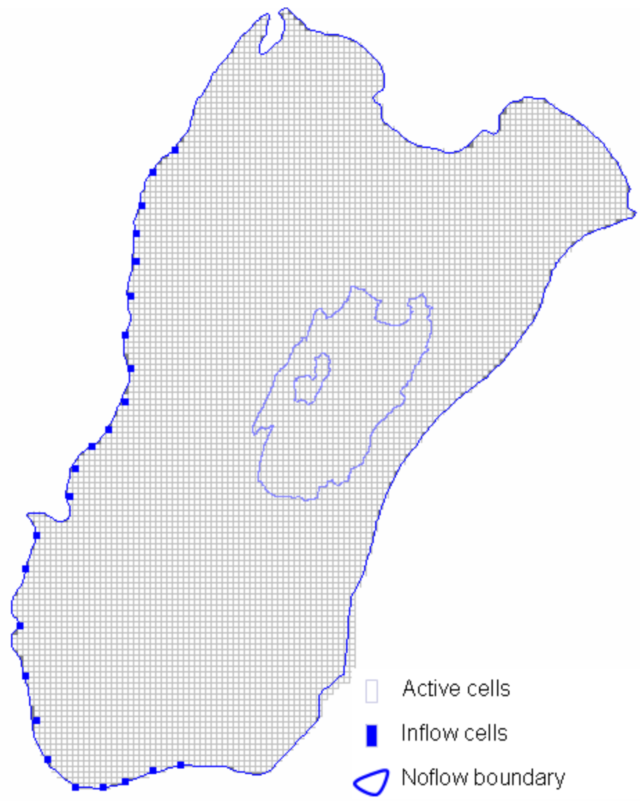


Figure 7.2: Boundary set up for watershed of Lake Beseka

Model boundaries describe the interface between the model domain and the surrounding environment, and are ideally based on actual physical or hydraulic boundaries (PDP, 2002). Physical boundaries relate to the physical presence of impervious geological formations, while hydrologic boundaries are invisible and formed along groundwater divides or streamlines (Anderson & Woessner, 1992).

The boundary of the northern and eastern part of the lake watershed was established as no-flow boundary. Fentale Mountain to the north and River Awash to the east of the lake watershed form pertinent physical boundaries. Hydrogeology, topography, hydrochemistry and isotopic information indicate that the groundwater watershed is wider than the surface lake watershed to the west and south of the watershed. Geological formations, fractures and faults are extended outside the lake watershed toward the west and south. In addition, hydrochemical and isotopic data indicate that groundwater that comes from the western part of the watershed plays an important role in the hydrology of the lake. The boundaries of the western and southwestern parts of the watershed were set as a combination of no-flow and specified flow boundaries (Figure 7.2). The specified flow boundary is represented by inflow cells that are simulated as injection wells using the well package in MODFLOW.

The initial conditions are values of dependent variables specified inside the boundary at the start of the simulation (Konikow and Reilly, 1998). For steady state models, initial conditions need only approximately match the natural system, because the solution for each dependent node can be found eventually through repeated iterations (Genetti, 1999). Transient models, on the other hand, need initial conditions closely matching natural conditions at the beginning of the simulation (PDP, 2002). Thus, it is a common procedure to use results of calibrated steady state conditions as the initial conditions for transient simulation, as the approach produces initial head data that are consistent with boundary conditions and parameters (Middlemis, 2001). For this study, field-measured static water table of wells were interpolated and used as initial conditions for the steady state simulation. Head distribution generated from the calibrated steady state simulation was used as the initial condition for transient simulation.

Aquifer properties

Aquifer properties refer to the hydraulic conductivity and storativity of an aquifer that govern groundwater flow rates and behavior of groundwater reservoir, respectively. Hydraulic conductivity values of water-bearing formations in the study area were derived from pumping test data collected from existing drilling contractor reports (Chapter 3), and these values were used as initial estimates for the model. The model domain was divided into hydraulic conductivity zones based on similarity in aquifer properties and geological structures. It was not possible to derive storage terms from existing pumping test data, as all pump tests were single-well tests. Thus, initial specific yield values were derived from literature and used as initial estimates.

Groundwater recharge

Groundwater recharge refers to the process by which the volume of water is percolated and crosses the water table to become part of the groundwater flow system. For lack of suitable means to quantify spatial distribution of recharge, groundwater modelers have traditionally assumed spatially uniform recharge rate across the water table equal to some percentage of average annual precipitation (Simmers, 1988). Prevalent soil type, land use and cover, and hydrogeological and geomorphological factors have been taken into account to determine representative recharge amounts from recharge values estimated in this study (Chapter 6), by using two estimation methods (water table fluctuation and EARTH modeling). The recharge value of 1 x 10-4 m/day was used as a representative recharge value. For stress periods of the last 15 months of the transient simulation, from July 2006 to September 2007, the output of EARTH model was directly used as a recharge estimate.

Lake Package

The lake package for MODFLOW allows the simulation of interactions of lakes with the groundwater flow system. The main routines of MODFLOW need to be modified to integrate the lake package into its modular program structure (Council, 1999). MODFLOW, which incorporates the lake package, was used in this study to simulate interactions of Lake Beseka with the surrounding groundwater flow system. The lake package needs specification of fluxes in terms of inflows and outflows to the lake domain for each stress period. Time series of monthly data on the rate of precipitation and evaporation of Lake Beseka were specified in [L^3 / T] units for 375 stress periods. The lake package calculates the precipitation flux as precipitation rate times total area of the lake, whilst evaporation flux is calculated as evaporation rate times the current wetted[11] lake area (Council, 1999). Surface runoff inflow rate was specified to the Lake Package as a flux [L^3 / T] for each stress period. Historic records from nearby meteorological station were used to prepare monthly time series climate data for the stress period of 31 years.

For each lake cell, the lake package requires specification of hydraulic conductance between the lake and groundwater cell, maximum area of the lake, and top and bottom elevation of the lakebed. The bathymetry of Lake Beseka (Chapter 2) was used to characterize the morphology of the lakebed. Top and bottom elevations of the lakebed were derived from the bathymetry and specified for 692 lake cells. Drilling during previous studies confirmed that the lakebed of Lake Beseka is mainly made up of fresh welded tuff. Therefore, hydraulic conductivity of the lakebed was estimated accordingly and used as an initial estimate for the model.

7.4.2 Model simulation

After assigning boundary conditions, hydrogeological framework and hydrological stress data of the study area, MODFLOW can be started. In steady state MODFLOW simulation, an equilibrium solution is desired for heads such that all inflows and outflows to the aquifer domain are in perfect balance (Council, 1999). The steady state solution provides one set of head field that represent long-term and average boundary conditions. Transient simulations, in contrast, produce sets of heads for each time step.

The groundwater regime of Lake Beseka area was simulated under steady and transient state conditions for the period of 31 years, from July 1976 to September 2007. Monthly stress periods were set for transient state simulations. The lake package can decrease the stability of MODFLOW, especially when lakes are directly simulated in the steady state mode (Council, 1999). To reduce this instability, first, interpolated groundwater level measurements of boreholes were run as initial heads with the lake represented as a fixed stage (fixed-stage mode). Then the resulting head field distribution was used as initial heads together with long-term average stresses of the lake (precipitation, evaporation and runoff) to attain a steady state stage solution. The

[11] Wetted lake area is the sum of areas of all cells having lakebed top elevation below the current lake stage

output of the steady state simulation was then used as an initial head for the transient state run. Inflow and outflow fluxes of the lake were specified for the transient model on a monthly basis.

7.5 Model calibration

7.5.1 Introduction

Models are subjected to many uncertainties, which can be related to assumptions used in the models, algorithms used to describe the physical processes, or temporal and spatial variability of input data. Groundwater models need field data to define parameters and hydrological stresses for each grid cell. Determination of spatial and temporal variability of parameters and hydrological stresses require extensive field testing, which is in most cases impractical due to economical and technical constraints. Models thus represent an attempt to solve a large set of simultaneous equations having more unknowns than equations (Konikow and Reilly, 1998). To demonstrate that groundwater models are realistic, field observations of aquifer response (such as groundwater levels) are compared to corresponding model-simulated values with the general objective of reducing the difference. Models are often used to infer reality by comparing the numbers that they produce with numbers obtained from some kind of measurement (Doherty, 2004).

Calibration solves a problem inversely by adjusting the unknowns (hydraulic properties and boundary conditions) until the solution matches the known heads (Genetti, 1999; PDP, 2002). Model calibration is the process of finding a set of parameters, boundary conditions, and stresses that produce simulated heads and fluxes that match field-measured values (Anderson & Woessner, 1992). Model accuracy is achieved by changing parameter values used in the model until satisfactory agreement between simulated and recorded variables is obtained (Abbott and Refsgaard, 1996). Calibration is necessary, but not sufficient to gain confidence regarding the model predictions, as model simulation can reproduce system behavior under certain set of conditions (Middlemis, 2001). Multiple calibration of the same system can yield different combinations of boundary conditions and aquifer properties due to non-unique characteristics of the calibrated models. Because of the large number of variables in the set of simultaneous equations represented in a model, calibration will not yield a unique set of parameters (Konikow and Reilly, 1998; Genetti, 1999. PDP (2002) suggested that non-uniqueness of calibrated parameters is partly caused by complexity of real system coupled with lack of sufficient data. Although a poor match provides evidence of errors in the model, a good match in itself does not prove the validity or adequacy of the model (Konikow and Bredehoeft, 1992).

Calibration can be carried out by manual trial and error and automatic optimization. The trial and error technique implies a manual parameter adjustment where initial parameter values are adjusted in a sequential model run to match simulated heads to observed ones. Because a large number of interrelated factors affect the output, the trial and error adjustment may become a highly subjective and inefficient procedure (Konikow and Reilly, 1998). The trial and error technique is also labor intensive. Automatic calibration, in contrast, involves the use of a numerical algorithm that finds the extreme of a given numerical objection function. The objective of automatic calibration is to obtain an estimate of system parameters that yield the closest match between observed data and model calculations (Konikow and Reilly, 1998). Automatic calibration is fast and less subjective.

7.5.2 Automatic calibration

Automatic calibration utilizes an objective function, such as minimization of the sum of the squared difference between observed and computed heads (residuals) to govern automatic iterative adjustment of values (Genetti, 1999). The objective of automatic calibration is to obtain the estimates of system parameters that yield the closest match (minimize deviations) between observed data and model calculations (Konikow and Reilly, 1998). (PEST (Parameter ESTimation)[12] is commonly used model-independent automatic calibration software, which undertakes parameter estimation using a particular model without making any change to the model. It takes control of an existing model and runs it as many times as necessary until discrepancies between model-generated numbers and corresponding measurements are reduced to a minimum (Doherty, 2004). For nonlinear models, parameter estimation is an iterative process. PEST uses a nonlinear estimation technique known as the Gauss-Marquardt-Levenberg

[12] PEST was developed by John Doherty in 1994

method, which has the strength of estimating parameters using fewer model runs than any other estimation method (Doherty, 2004). A detailed description of PEST software and its parameter estimation methods can be found in PEST manual, (Doherty, 2004).

To effectively take control of a particular model, PEST requires input files that contain the parameters that PEST has to adjust, and the output files that contain the model outcomes. To build an interface between PEST and any particular model, three input files should be prepared,

1. Template files: PEST needs template files to identify input files that contain parameters to be optimized.

2. Instruction files: PEST should be provided with instruction files on how to read the model output files and to identify model-generated observations.

3. Control files: Provide PEST with the names of the template and instruction files, the names of the corresponding model input and output files, initial parameter values, measurement values and weights, etc.

Konikow and Reilly (1998) indicated that discrepancies between observed and calculated responses are due to model errors that can possibly originate from inaccurate definition of the conceptual model, from numerical errors in the solution, or from a poor set of parameter values. The objective of the calibration is, therefore, to reduce these model errors until they become insignificant. Calibration results are evaluated by using qualitative and quantitative performance measures. Qualitative assessment (pattern matching) involves comparison of contour maps and hydrographs of measured and simulated head, while quantitative performance measure involves mathematical or statistical description of the residuals (Middlemis, 2001). The performance of any model should be evaluated by combining both qualitative and quantitative measures.

The water balance is one of the quantitative performance measures that is used to check solution accuracy or the total residual error in the calibration solution by comparing total simulated inflows and outflows (Anderson & Woessner, 1992). Mass balance calculations should always be performed and checked during the calibration procedure for assessing the numerical accuracy of the solution (Konikow and Reilly, 1998). The water balance error is expressed in percentage, representing the difference between total inflows and total outflows including changes in storage divided by total inflows or outflows. The error should be less that 1% for each stress period and for the entire model to ensure the calibration solution is numerically stable and acceptable (Anderson & Woessner, 1992; Middlemis, 2001; PDP, 2002).

Lumped-sum comparison methods are the other commonly used quantitative measures employed to measure the average difference of residuals. However, these performance indicators provide lumped measures of calibration that do not indicate the spatial or temporal distribution of the error (Anderson & Woessner, 1992). These methods include mean error of residuals, absolute mean error to the residuals, and root mean squared error.

1. The mean error (ME) is the mean difference between measured heads (h_m) and simulated heads (h_s).

$$ME = 1/n\sum_{i=1}^{n}(h_m - h_s)_i \tag{7.4}$$

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

$$MAE = 1/n\sum_{i=1}^{n}\left|(h_m - h_s)_i\right| \tag{7.5}$$

3. The root mean squared error (RMS) is the average of the squared difference between measured and simulated heads.

$$RMS = \left[1/n\sum_{i=1}^{n}(h_m - h_s)_i^{\,2}\right]^{0.5} \tag{7.6}$$

The goodness-of-fit of calibration results can also be evaluated quantitatively by using numerical model performance measures such as correlation coefficient (R^2)

and coefficient of model efficiency (R_{eff}). The R^2 and R_{eff} measure how trends in the measured data are reproduced by the simulated output.

The correlation coefficient (R^2) is calculated by:

$$R^2 = \frac{\left[\sum_{i=1}^{n}\left(h_s - \bar{h}_s\right)\left(h_m - \bar{h}_m\right)\right]^2}{\sum_{i=1}^{n}\left(h_s - \bar{h}_s\right)^2 \sum_{i=1}^{n}\left(h_m - \bar{h}_m\right)^2} \tag{7.7}$$

Where h_s is simulated head value, $\bar{h}_s$ is mean of simulated head values, h_m is measured head value, and $\bar{h}_m$ is mean of measured head values.

The values of R^2 range from zero to unity. A value approaching unity is expected for perfect calibration, whereas a very poor calibration would have a value approaching zero.

After Nash and Sutcliffe (1970), the coefficient of model efficiency (R_{eff}) is calculated as:

$$R_{eff} = 1 - \frac{\sum_{i=1}^{n}(h_m - h_s)^2}{\sum_{i=1}^{n}(h_m - \bar{h}_m)^2} \tag{7.8}$$

Where h_s is simulated head value, h_m is measured head value, and $\bar{h}_m$ is mean of measured head values.

The coefficient of model efficiency measures how well simulated results predict the measured data by using mean of the measured data over the entire period of comparison. The value of coefficient of model efficiency ranges from negative infinity to unity (best). If R_{eff} equals unity, it means that the model prediction is a perfect fit with the measured data. When R_{eff} equals zero, it indicates that the simulation is as good (or as poor) as using the measured data. If R_{eff} is less than zero, it indicates that the model prediction is a very poor fit with the measured data.

7.5.3 Steady state calibration

Steady state is a condition at which system inflows and outflows are in equilibrium without considering changes in storage of the system with time. In fractured rock aquifers, pump-test parameters are usually uncertain and representative only for small areas around boreholes (Lubczynski and Gurwin, 2005). Thus, hydraulic conductivity instead of recharge is set as calibration parameter during steady state calibration of the model. Recharge is kept constant throughout the calibration period, hence it would be a mistake to attempt a simultaneous estimation of hydraulic conductivity and recharge for a steady-state model using measured water levels as the only observation dataset (Doherty, 2004). The PEST was used to optimize spatial variability of hydraulic conductivity in the lake watershed within the predefined variability range. Mean values of snapshot groundwater level measurements in 1998 and 2006 from 29 boreholes were used to calibrate the model under steady state conditions. The model domain was divided into five hydraulic conductivity zones based on similarity in aquifer properties and geological features. PEST ran the steady model 58 times to achieve the calibration values.

The result of the steady state calibration indicates that the hydraulic conductivity of the geological formations around the lake is very high compared to the surrounding formations (Figure 7.3). This could be related to the widespread occurrence of fractured and recent basaltic rocks in this part of the watershed. On the other hand, sediments at the northwestern tip and volcanics of the Kone complexes in the southwestern part of the watershed were calibrated at low hydraulic conductivity values.

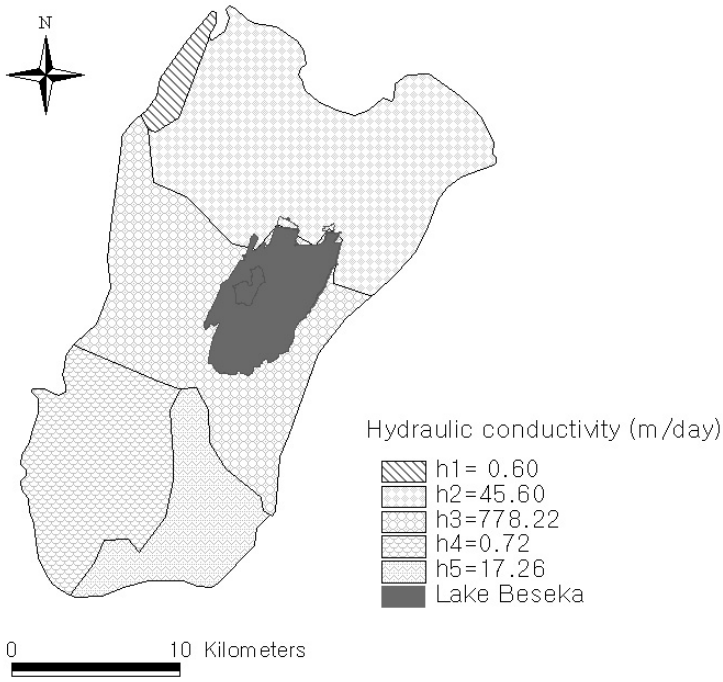


Figure 7.3: Calibrated hydraulic conductivity values for the watershed Lake Beseka (m/day)

The presence of fine-grained sediments in the northwestern corner of the watershed and undifferentiated lava flow of trachytic to rhyolitic volcanics of the Kone complexes could be the reason behind the low hydraulic conductivity values. The correlation coefficient matrix of the calibrated values indicates that there is a general lack of correlation between most of the parameters (Table 7.1). Some degree of correlation is observed between hydraulic conductivity of the Kone volcanic complexes (h4) and Dino ignimbrite (h5). This could be related to the small number of observation wells in those parts of the watershed.

Table 7.1: Correlation coefficient matrix of hydraulic conductivity in the lake watershed, where h1 is basaltic formation, h2 is Fentale ignimbrite, h3 is sediments at north western edge of the watershed, h4 is the Kone volcanic complexes and h5 is Dino ignimbrite.

	h1	h2	h3	h4	h5
h1	1.00	-0.44	0.09	-0.01	0.00
h2	-0.44	1.00	-0.40	0.06	-0.03
h3	0.09	-0.40	1.00	-0.22	0.12
h4	-0.01	0.06	-0.22	1.00	-0.79
h5	0.00	-0.03	0.12	-0.79	1.00

Acceptability of model calibration is commonly judged in relation to selected lumped quantitative performance measures. Lumped-sum comparison methods lump residual measurements into a single value. The mean error (ME) between simulated and measured heads of steady state calibration was estimated to be close to zero (Table 7.2). The mean absolute error (MAE) and root mean square error (RMS) are also characterized by low values, indicating that the model was well calibrated. The values of the correlation coefficient (R^2) and coefficient of the model efficiency ($R_{eff)}$ also indicate a good performance of the steady model.

Table 7.2: Summary of calibration errors for steady model

Model performance measures	values
Mean error (ME)	0.37
Mean absolute error (MAE)	1.17
Root mean square error (RMS)	1.66
Correlation coefficient (R^2)	0.81
Coefficient of model efficiency ($R_{eff)}$	0.93

The achievement of a best fit between values of observed and computed variables can be evaluated in a way that residual errors should have a mean that approaches zero and the deviations should be minimized (Konikow and Reilly, 1998). Analysis of the histogram of residuals demonstrates that residuals are randomly distributed around a mean value, which is close to zero (0.3), and that about 70 % of the residuals are found within $\pm$ 1 m difference between simulated and observed heads (Figure 7.4).

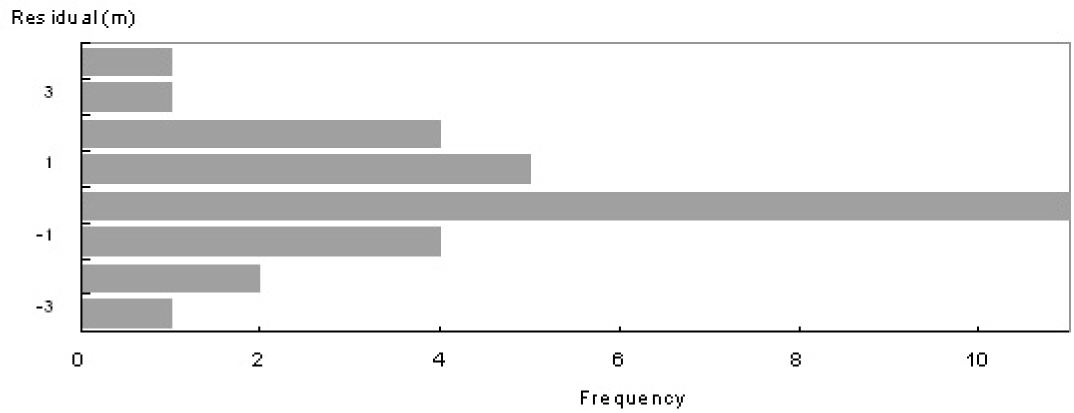


Figure 7.4: Histogram for residuals of groundwater heads (m) for steady state calibration

The aforementioned lumped measures evaluate average errors in the calibration but not the distribution of errors. The scatter plot of measured against simulated heads is another way of showing calibration fit, but it can also be used to detect trends in spatial distribution of errors (Anderson & Woessner, 1992). The scatter plot of the steady state calibration shows good fitness between measured and calculated heads (Figure 7.5). It indicates that there are no systematic errors in the spatial distribution of differences between modeled and measured heads.

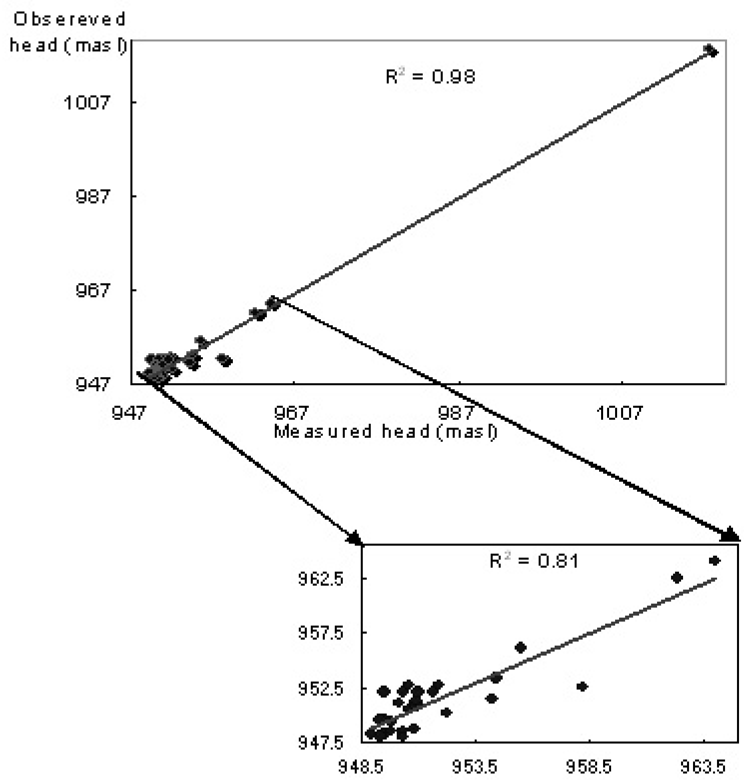


Figure 7.5: Scatter plot of measured vs. calculated heads of steady state calibration

7.5.4 Transient state calibration

Transient state is a condition at which system inputs and outputs are not in equilibrium, and net changes in the storage of the system is considered with time. Aquifer head distribution and calibrated parameters from the steady state calibration are specified for transient simulation as initial head distribution and parameters, respectively. Transient calibration was applied to calibrate specific yield, hydraulic conductivity of lakebed and inflow terms[13] to the watershed of the lake. Monthly lake level measurements of 22 years were used for calibration to match with corresponding computed lake stages of the transient simulation. Lake levels (1976-2007) were collected from Ministry of Water Resource (MWR). The period between July 1976 and December 1998 was set as calibration period for the transient simulation. This period was when continuous lake level measurements for Lake Beseka were available, and when the lake was under natural conditions (without pumping). The transient model was run 73 times in PEST to achieve calibrated parameter values.

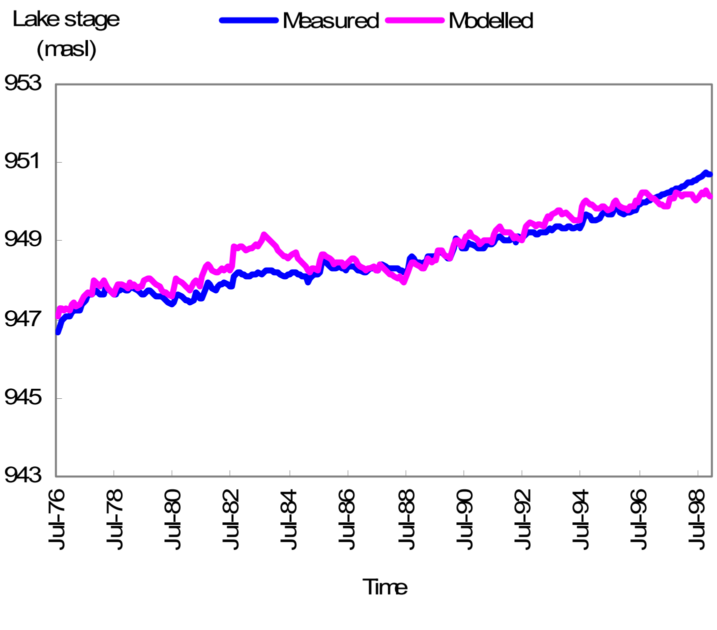


Figure 7.6: Plot of measured and modeled lake stages over the calibration period (July 1996 –December 1998).

Graphical evaluation of observed and simulated lake stages of Lake Beseka shows very good agreement in most parts of the calibration period (Figure 7.6).

[13] Inflow terms: The amount of water that laterally flows to the aquifer system of the lake watershed across the surface boundary.

Considerable difference between calibrated and measured lake stages can be observed during the period from 1982 to 1983. The two consecutive years were relatively wet years, and 1982 was the wettest year in the history of the area since the establishment of the meteorological station in 1966. Thus, the model tends to overestimate the lake stages for this wet period compared to observed measurements.

The piezometric surface of the transient simulation for the final stress period is plotted (Figure 7.7). The piezometric heads around Lake Beseka are sparsely spaced compared to the closely spaced head contours of the southern watershed area. This could be due to variations in the hydraulic conductivity of the geological formations that primarily determine the gradient of potentiometeric surface (Chapter 3). The piezometric map of transient calibration indicates a groundwater flow direction toward Lake Beseka from the western and southern part of the watershed.

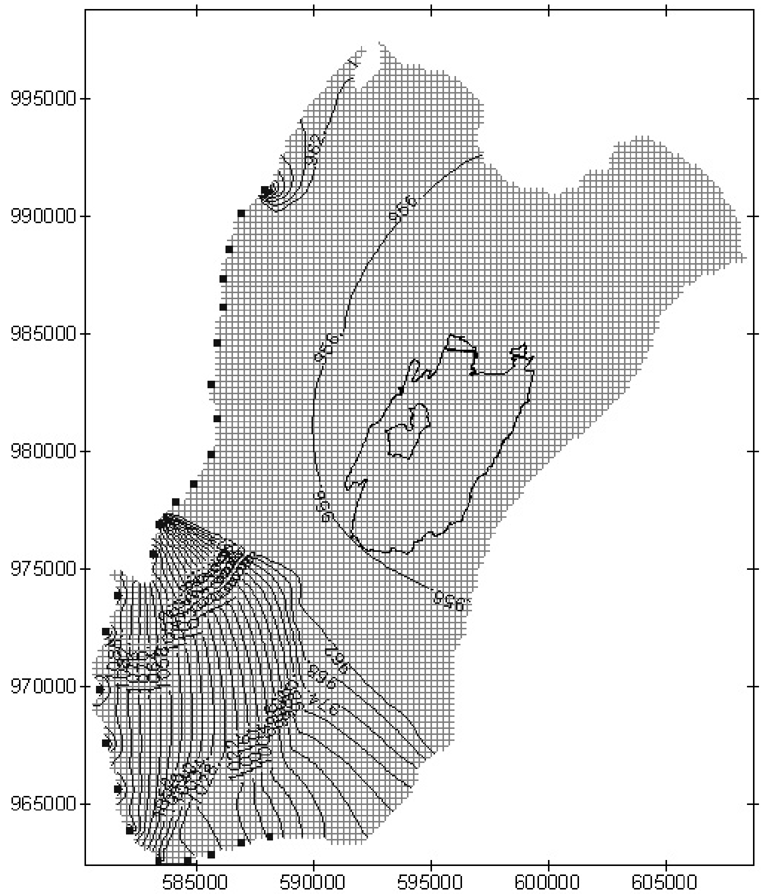


Figure 7.7: Piezometric map of transient calibration for the 270th stress period (December, 1998)

During the transient calibration, specific yield, lakebed hydraulic conductivity and inflow term to the watershed were calibrated simultaneously. Hydraulic conductivity of the lakebed was calibrated at the value of 0.15 m/day, lower and upper limits of 0.03 m/day and 0.27 m/day, respectively, in the 95% confidence limit. Specific yield is the least known parameter in the study area. The lake watershed was lumped into two specific yield zones, and initial values were specified based on literature. The first zone was assigned to the highly fractured, recent basaltic formation around the lake, whilst the second zone was specified for the remaining part of the watershed, which is dominantly covered by acidic volcanic formations. The results of transient calibration show that the calibrated specific yield value is higher for the recent basaltic formation compared to the surrounding acidic volcanic rocks (Figure 7.8).

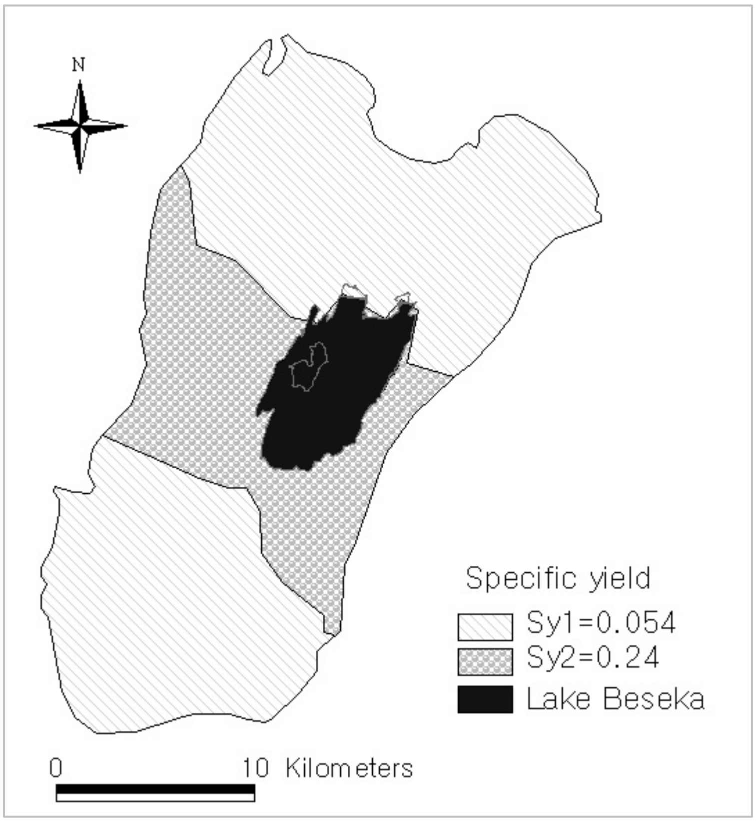


Figure 7.8: Calibrated specific yield values for the watershed of Lake Beseka

Groundwater inflows to the lake watershed across the northwestern and southwestern model boundaries were calibrated at an annual inflow rate of 30.45 Mm^3, with lower and upper limits of 24.4 and 36.5 Mm^3, respectively, in the 95 % confidence limit. The correlation coefficient matrix of specific yields, lakebed hydraulic conductivity and inflow terms (Table 7.3) shows a reasonable degree of independency among the calibrated parameter values.

Table 7.3: Correlation coefficient matrix of transient calibration, where Sy_1 and Sy_2 are specific yield values for basaltic and acidic volcanic formations, respectively; K_lk is hydraulic conductivity of the lakebed, and Inflow is amount of water that enters the lake watershed across the boundary

	S_{y1}	S_{y2}	K_lk	Inflow
S_{y1}	1.00	0.16	0.00	0.71
S_{y2}	0.16	1.00	0.25	0.76
K_lk	0.00	0.25	1.00	0.02
Inflow	0.71	0.76	0.02	1.00

Calibration is commonly evaluated by analyzing residuals or the difference between observed and simulated values. Lumped-sum comparison methods (ME, MAE and RMS), and R^2 and R_{eff} are used to evaluate the performance of the transient model (Table 7.4). The mean error between simulated and measured lake stages is close to zero. The mean absolute error and root mean square error are very low, indicating how well the transient model is calibrated. The values of correlation coefficient (R^2) and coefficient of the model efficiency (R_{eff}) indicate a good performance of the transient model as well.

Table 7.4: Summary of calibration errors for transient model

Model performance measures	Values
Mean error (ME)	0.18
Mean absolute error (MAE)	0.26
Root mean square error (RMS)	0.32
Correlation coefficient (R^2)	0.92
Coefficient of model efficiency (R_{eff})	0.88

Analysis of the histogram of 270 lake level residuals indicates that about 90 % of the residuals are plotted close to the mean, which is characterized by a value close to zero (0.18 m), whereas the remaining 10 % accounts for residuals found within ± 1 m difference between simulated and observed lake stages (Figure 7.9). The achievement of a best fit between values of observed and computed variables can be evaluated in a way that residual errors have a mean that approaches zero (Konikow and Reilly, 1998).

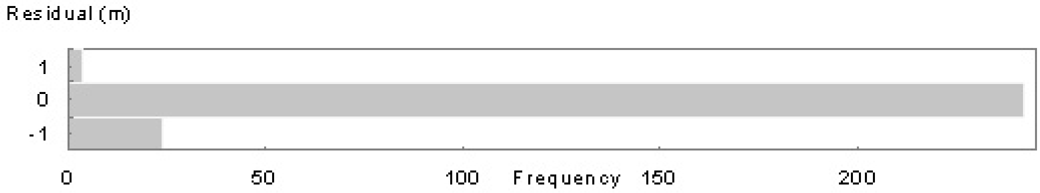


Figure 7.9: Histogram for residuals of lake stages (m) for transient calibration

The scatter-plot of measured lake stages against corresponding simulated lake stages demonstrates a good calibration fit (R^2 = 0.92), with points plotted on both sides of the line (Figure 7.10).

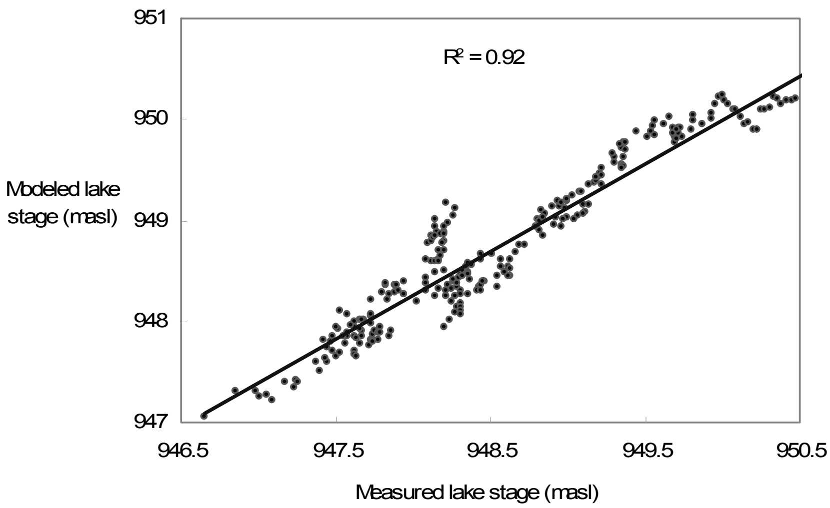


Figure 7.10: Scatter plot of measured lake stage vs. modeled lake stage

Water Budget

A water budget states the rate of changes in water stored in an area, such as a watershed, and is balanced by the rate at which water flows into and out of the area (Healy et al., 2007). Hence, water budgets are valuable assessment tools, as they provide a measure of the relative importance of each component to the total budget (Konikow and Reilly, 1998). Following statistical and graphical evaluation, the water budget of Lake Beseka and the groundwater regime of the watershed of the transient calibration were analyzed.

The long-term water budget of Lake Beseka indicates that water inflow to the lake is mainly contributed from groundwater seepage and precipitation flux.

Table 7.5: Long-term annual water budget of Lake Beseka

Flux	Inflow (m^3/year)	Outflow (m^3/year)
Preciptation	24,437,546.1	0
Runoff	7,672,641.1	0
Groundwater seepage	33,829,705.3	221,735
Evaporation	0.0	61,781,935.3
Total	65,939,892.5	62,003,670.5

Groundwater seepage to Lake Beseka accounts for 51 % of the total inflow, and the next important flux, precipitation, accounts for 37 % of the total input to the lake (Figure 7.11). The rest is contributed by surface runoff that flows into the lake. The main water loss mechanism for Lake Beseka is evaporation, which accounts for about 99 % of the total outflow. Model results show that there is no significant groundwater leakage from the lake to the surrounding groundwater system. This is due to the fact that Lake Beseka is a terminal lake, which receives water mainly from precipitation and groundwater inflows, and loses water only through evaporation. As difference of about 3 % is observed between total inflow and outflow of Lake Beseka, which is attributed to change in the storage of the lake for each year (Table 7.5). This is the main reason for the continuous expansion of the lake.

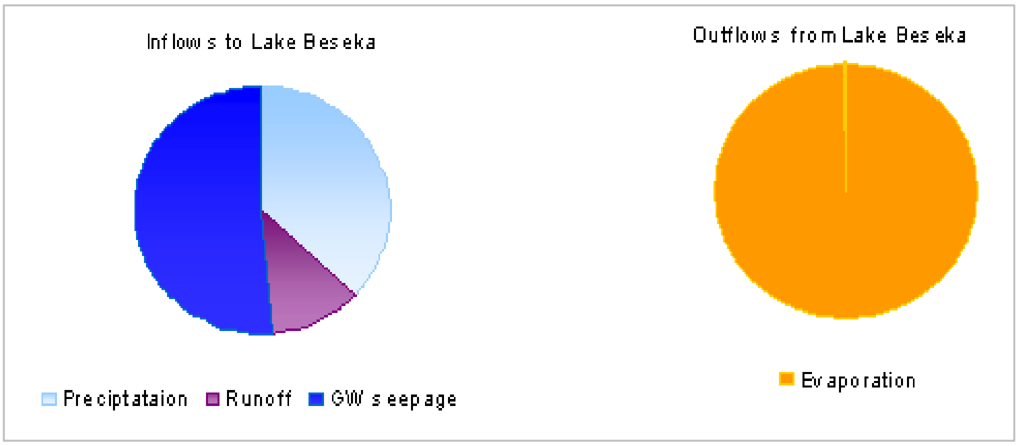


Figure 7.11: Pie chart showing the water budget of Lake Beseka

The water budget of the groundwater domain (Table 7.6) shows that groundwater that enters to the aquifer system of the lake watershed across the western and southwestern boundaries is the main source of water to groundwater regime in the area. This groundwater that flows across the surface boundaries into the lake watershed accounts for 62 % of the total inflow to the groundwater system in the study area.

Table 7.6: Long-term annual water budget of the groundwater domain

Flux	Inflow (m^3/year)	Outflow (m^3/year)
Recharge	17,410,851	0
Inflows	30,476,894	0
Storage	1,333,509	15,398,451
Lake leakage	221,735	34,029,204
Total	49,442,989	49,427,655

The recharge that seasonally replenishes the groundwater system of the lake watershed accounts for 35% of the total inflow to the aquifer system. Groundwater seepage to Lake Beseka is the main sink of water for the groundwater flow system in the area, which accounts for 69 % of the total outflow. Thus, Lake Beseka acts as a discharging point to the surrounding groundwater flow system. The rest goes to the groundwater reservoir.

Lake Beseka is located in topographically low lying area of the watershed, which makes it function as a discharge zone or sink for the regional groundwater flow system. Permanent lakes are usually areas of permanent groundwater discharge (Dingman, 2002; Sophocleous, 2002). The long-term interaction of Lake Beseka with the surrounding groundwater system was simulated by the transient model. The model output indicates that Lake Beseka has been gaining water from the surrounding groundwater system at an average monthly rate of 2.77 Mm^3. According to long-term groundwater seepage to the lake, July is the month with the highest groundwater seepage (3.04 Mm^3/month), while September is the lowest (2.36 Mm^3/month). This shows that that Lake Beseka receives the highest and lowest groundwater seepage from the surrounding aquifer system at the beginning and end of the rainy season.

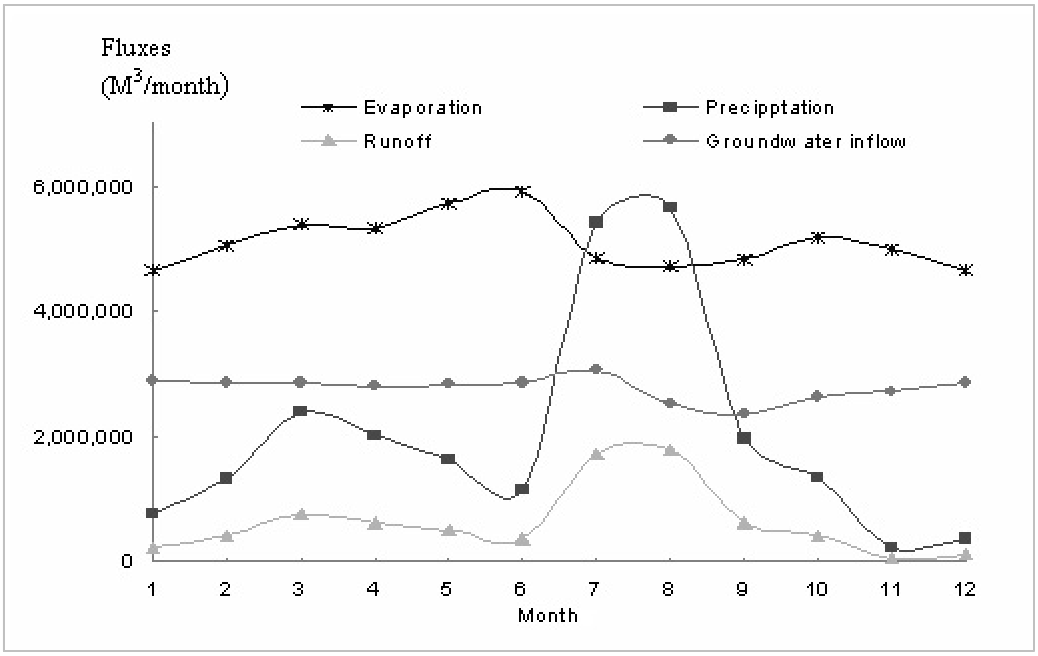


Figure 7.12: Long-term seasonal fluxes of Lake Beseka

The groundwater flow that discharges water directly into the lake can be observed along its western shore, where numerous hotspring are emerging and flowing into the lake. Despite the slight decline during August and September, seasonal fluctuation of groundwater seepage to the lake is not very significant. This can be verified by the hot springs that flow to the lake all the year along the western edge of the lake. Evaporation of water from Lake Beseka shows a seasonal variation. It has a peak value during the dry season in May and June, and the lowest value in August. Seasonal variability is also an important feature for precipitation and runoff fluxes of the lake, as these are influenced by seasonal migration of the ITCZ. Both precipitation and runoff fluxes are characterized by high input values during the rainy seasons (Figure 7.12).

The analysis of inter-annual fluxes of the lake illustrates that years which are characterized by relatively low precipitation inputs are characterized by higher groundwater inflows compared to the relatively wet years (Figure 7.13). This fact can also be observed in the seasonal fluxes of Lake Beseka, when groundwater inflow to the lake shows a slight decline at the end of the rainy season. Although there is a continuous flow of groundwater to the lake, groundwater leakage to the lake is more stimulated during relatively dry seasons, when storage and stage of the lake are reduced. During relatively wet seasons, when water is abundant in the lake storage for the given period of time, groundwater flows to the lake, but the groundwater leakage would be in slightly reduced manner. This is due to the fact that flows toward the lake from the surrounding groundwater system are driven by the head differentials at the interface of two hydrologic systems.

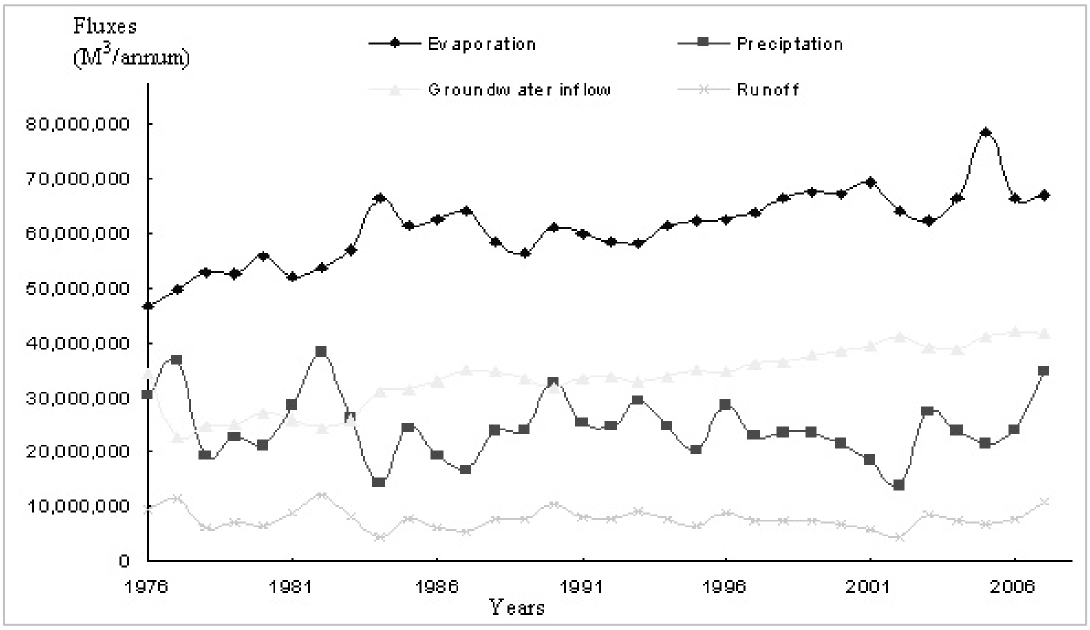


Figure 7.13: Inter-annual fluxes of Lake Beseka

The other interesting feature that can be observed from the analysis of the inter-annual fluxes of Lake Beseka is the increment of evaporation flux from the lake during the last 31 years (Figure 7.13). Evaporation from the lake has increased from an average monthly rate of 3.89 Mm^3 in 1976 to 5.59 Mm^3 in the year 2007. This could be due to the continuous expansion of the lake throughout the simulation period. The continuous growth of Lake Beseka could lead to the availability of more and more water with time for evaporation. The areal extent of Lake Beseka computed by the transient model was cross checked with the one that is independently derived from the bathymetric survey. The comparison shows a good agreement. The bathymetry of Lake Beseka conducted in December 1998 estimated the area of the lake to be 39.9 km^2. The transient model computed the area of Lake Beseka to be 42 km^2 for that specific period with a difference of only 5.3 %.

The storage of Lake Beseka was computed by the transient model for all 375 stress periods. Analysis of the temporal storage of the lake during the simulation period demonstrates that there has been a continuous expansion of Lake Beseka during the last 31 years. Storage has increased since 1976 by an average annual rate of 5.4 mcm.

According to the transient model, the volume of Lake Beseka has increased 1.5-fold its original volume at the beginning of the simulation period. The storage of Lake Beseka increased from its initial volume of 88 Mm^3 in July 1976 to 220 Mm^3 in September 2007. During this simulation period, a record change in the storage of Lake Beseka is observed during the wettest year in 1982, which was characterized by an annual storage increase of 23 Mm^3.

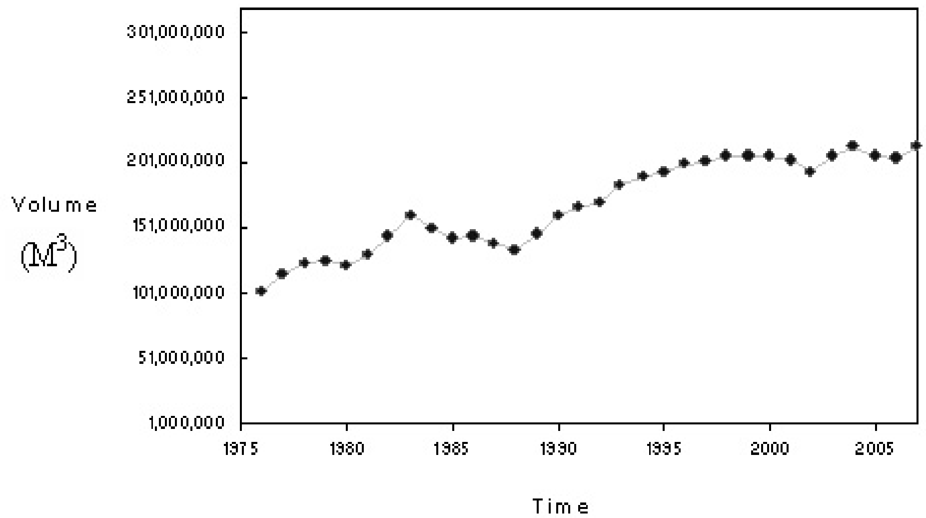


Figure 7.14: Storage values for Lake Beseka (1976-2007)

Although a constant increase in the lake storage is evident (Figure 7.14), the expansion of Lake Beseka has never been a continuous and smooth process. Its extension has been marked by a zigzag of crests and troughs in response to local meteorological conditions. The major crest was observed following the relatively wet years of 1982 and 1983 (Figure 7.14). This relatively wet period was characterized by an average annual precipitation of 727 mm. This is also reflected by the simulated lake stages, which were computed for the same period by the transient model. During this relatively wet period, plenty of water was available to enhance stage and storage of the lake, and the rate of evaporation was not high enough to significantly reduce the water in that period.

A decrease in the lake surface can be observed for the dry years of 1984-1987. This drought period was characterized by an average annual precipitation of 421 mm, and the lowest change in the storage of Lake Beseka was during this period. The other significant declining trend in storage of Lake Beseka can be observed in another drought period in 2002, where annual precipitation was only 308 mm. During these drought periods, the storage of Lake Beseka reduced in response to the substantial amount of water that had evaporated to the atmosphere from the surface of the lake. Little or no precipitation was available during those drought periods to compensate for the evaporation. These crest and troughs in the storage of Lake Beseka follow the pattern of extreme climatic conditions during the long-term expansion of the lake. This signifies the importance of variability in precipitation-evaporation fluxes in introducing fluctuation to the overall water budget of the lake.

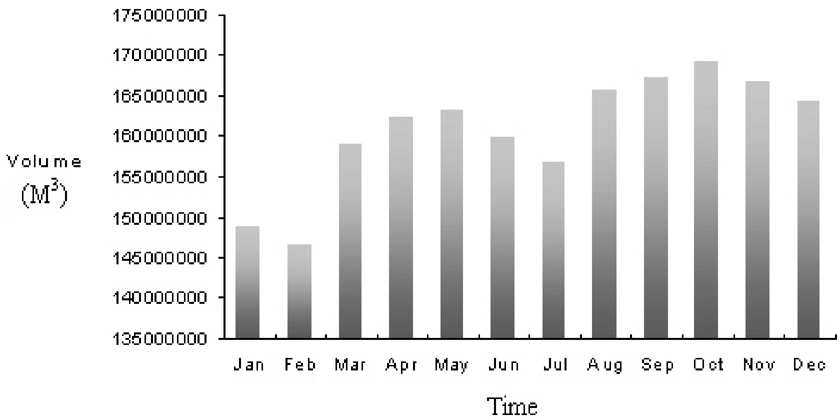


Figure 7.15: Seasonal fluctuation of storage of Lake Beseka in 1990

The analysis of the intra-annual storage of Lake Beseka demonstrates that the volume of the lake fluctuates seasonally according to the prevalent climatic conditions. As seasonal variation of groundwater seepage to the lake is not very significant, seasonal fluctuation is mainly governed by seasonality of precipitation-evaporation fluxes, and rainy seasons enhance the storage compared to the dry seasons, when evaporation is a predominate process, (Figure 7.15).

7.6 Sensitivity analysis

A sensitivity analysis is used to characterize the effect of changes in parameters or boundary conditions on the behavior of the calibrated model. The purpose of a sensitivity analysis is to quantify uncertainties in a calibrated model caused by uncertainties in the estimates of aquifer parameters, stresses, and boundary conditions (Anderson & Woessner, 1992). The sensitivity analysis is normally conducted by varying model input parameters and observing relative change in model response. The SENSAN (SENSitivity Analysis), a command program in PEST, was used in this study to conduct sensitivity analysis of calibrated parameter values of the steady and transient simulations in automate manner.

The analysis was conducted for the parameters of hydraulic conductivity, specific yield, hydraulic conductivity of lakebed, and for inflow terms. SENSAN can carry out multiple model runs without user intervention. It needs four input file types: template, instruction, SENSAN control, and parameter variation files. The first three input files provide SENSAN with structure details of a given sensitivity analysis, while the fourth provides the parameter values to be used in the model runs (Doherty, 2004). For a particular model outcome, SENSAN calculates model outcome sensitivities with respect to parameter variation from their base value as the difference between the model outcome and the pertinent model outcome base value divided by the difference between current parameter set and the parameter base values (ibid).

$$S = (O - O_b)/(P - P_b) \tag{7.9}$$

where O_b and P_p are model outcome and parameter base values, O and P are the model outcome and parameter values pertaining to particular model.

The calibrated values of hydraulic conductivity, specific yield, and hydraulic conductivity of lakebed as well as inflow terms were systematically altered, and the sensitivity of each parameter with respect to the model outcome was computed based on the aforementioned relation. The analysis was conducted by changing one parameter value at a time. The effect of changes the calibrated hydraulic conductivity values of the geological formations in the lake watershed on the aquifer heads was evaluated, and the result shows that the model is more sensitive at lower hydraulic conductivity values than at higher ones (Figure 7.16).

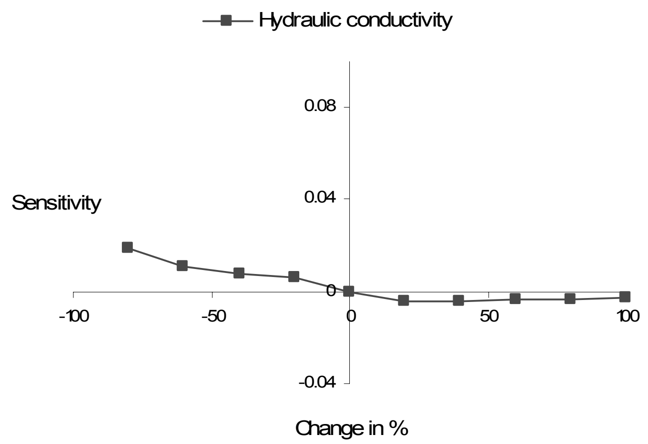


Figure 7.16: Sensitivity of hydraulic conductivity with respect to changes in the model outcome (groundwater head)

Sensitivity analyses of calibrated parameters for the transient solution were conducted and their effect on the model outcomes was evaluated. The analysis indicates that specific yield is the most sensitive parameter compared to hydraulic conductivity of the lakebed and inflow terms (Figure 7.17). It is even more sensitive at lower specific yield values. The inflow term is the least sensitive parameter, while hydraulic conductivity of the lakebed is intermediately sensitive between the two parameters.

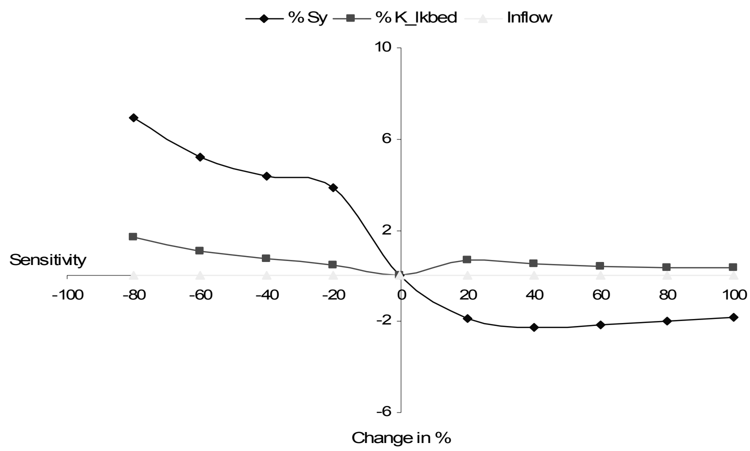


Figure 7.17: Sensitivity of specific yield, hydraulic conductivity of lakebed, and inflow terms with respect to changes in the model outcome (lake stage).

7.7 Conclusions

Annually, 30.45 Mm^3 of groundwater laterally flow to the aquifer system of the lake watershed across the western and southwestern model boundaries, accounting for 62%

of the total inflow to the groundwater regime. Recharge form the next important influx to the groundwater regime, accounting for the 35% of the total inflow to the aquifer system. The main water-outflow mechanism from the groundwater system is leakage to Lake Beseka, which accounts for 69% of the total outflow from the groundwater domain. Recharge of the groundwater takes place seasonally, and groundwater that laterally flows across the western and southwestern surface boundaries to the lake watershed forms a substantial part of the total inflow to the groundwater regime. On the other hand, groundwater leakage to the lake forms the major outflow from the groundwater system. The calibrations of the steady and transient solutions for groundwater system of the Lake Beseka area indicate that the fractured and recent basalt around the lake are characterized by higher hydraulic conductivity and specific yield values. The sensitivity analyses of the parameters signify that specific yield is the most sensitive parameter.

Groundwater seepage to the lake is the major contribution to the water budget of the lake, accounting for 51% of the total inflow. Precipitation is the next important flux to the hydrology of the lake, accounting 37 % of the total inflow. Lake Beseka is a terminal lake, which lacks both surface or groundwater outlets, and this fact is well supported by hydrochemical and isotopic evidences (Chapter 4 and 5). Consequently, evaporation is the only mechanism for water outflow from the lake, accounting for more than 99% of the total outflow. According to the water budget of the transient model, groundwater seepage to Lake Beseka forms an integral part of the lake inflow, while evaporation is the only water loss mechanism.

The piezometric map of the transient model indicates that groundwater flows from the western and southern part of the watershed toward Lake Beseka. Due to its topographical location in the watershed, Lake Beseka functions as discharge area or as a sink for the regional groundwater flow system. This can be seen from hot springs that directly flow to Lake Beseka at the western edge of the lake. The lake gains water from the surrounding groundwater system at an average monthly rate of 2.77 Mm^3. Generally, seasonality of groundwater seepage to the lake is insignificant, even though it shows slight declining trend at the end of the wet seasons. The slight decline in the groundwater seepage can be attributed to head build up in the lake during the rainy seasons, when plenty of water becomes available within a short period of time.

Precipitation and evaporation fluxes of Lake Beseka exhibit significant seasonal variation, and these two processes govern seasonality of the lake storage. During the wet season when precipitation is a predominant process, lake storage shows an increasing trend. During the dry season, in contrast, storage shows a declining trend due to high rate of evaporation. The rate of evaporation from the surface of Lake Beseka has increased during the 31 year simulation period (1976-2007). This can be mainly explained by the continuous expansion of the lake storage, which increases the availability of more and more water for evaporation with time. The storage of the lake has been increasing by an average annual rate of 5.4 Mm^3, and its rate of expansion has been affected by extreme climatic conditions. In the course of its expansion, the lake has experienced a significant decrease in size during drought periods when little or no precipitation was available to compensate evaporation. Thus, it is possible to conclude that the storage of Lake Beseka shows a seasonal fluctuation mainly in response to precipitation- evaporation fluxes.

8 SYNTHESIS AND PERSPECTIVES

8.1 General remarks

The main objective of this study was to understand the hydraulic interaction of Lake Beseka with the surrounding groundwater system and to identify and quantify the role of groundwater in the hydrology of Lake Beseka. To achieve this objective, physical and chemical approaches were combined. Lithological logs of boreholes, geophysical data and DEM were analyzed to set up a hydrogeological framework of the region. The flow system was analyzed by coupling a groundwater level map with the hydrochemical and isotopic composition of the water bodies in the region. The hydrochemical and isotopic data provided useful qualitative information for assessing the importance of groundwater to the water budget of Lake Beseka and gave a good picture in the circulation of the groundwater, flow directions and pathways. In the hydrogeological framework of the study area, a combination of groundwater flow map and hydrochemical and isotopic information was used to develop a conceptual model of the study area.

The temporal fluctuations of groundwater levels were assessed to determine the groundwater recharge processes, and groundwater recharge was estimated through multiple approaches. Finally, a numerical groundwater flow model was developed for the watershed of Lake Beseka to evaluate the interaction of the lake with the surrounding groundwater system, and to quantify the groundwater fluxes to and from the lake. This chapter tries to give a concise summary on the results of the study.

8.2 What is the role of groundwater to the hydrology of Lake Beseka?

The importance of groundwater for the water budget of Lake Beseka were appraised by combining results from groundwater contour maps, hydrochemical and isotopic information, recharge estimation and groundwater modeling.

Groundwater level map

A groundwater contour map was constructed based on groundwater level measurements of boreholes located inside and outside the lake watershed. The groundwater level is high in the south and northwestern part of the watershed and become lower toward Lake Beseka and the eastern part of the watershed. The map indicates that the water table is a subdued replica of the land surface, and groundwater flows from the south and northwestern part of the watershed toward the lake (Figure 3.3). It is also marked that groundwater flows in the study area are influenced by the presence of tectonic structures. Emanations of hot springs along tectonic structures are evident at the western and south western edge of Lake Beseka (Figure 2.7). Thus, it is not only topographic configuration but also geological and structural features that govern the direction of groundwater flow in the lake watershed.

Hydrochemical data

A hierarchical multi-element cluster analysis (HCA) of water samples from the study area identified four major hydrochemical facies. The first hydrochemical cluster of water samples from Lake Beseka is characterized by a Na-HCO_3-Cl to Na-HCO_3-CO_3-Cl type of water. The second cluster is identified by a Na-HCO_3 type of water; groundwaters from the western part of the watershed and the hot springs that emerge at the southwestern corner of Lake Beseka belong to this cluster. The third hydrochemical cluster is typified as a Na-HCO_3-SO_4-Cl or Na-HCO_3-Cl type of water, and groundwater from the eastern part of the watershed falls in this group. Water from River Awash, which is used in the study area for irrigation of Abadir farm, is characterized by a *Na-Ca-HCO3* type of water.

Silicate hydrolysis of volcanic rocks is the dominant process controlling the hydrochemical characteristics of groundwater, and has resulted in a Na-HCO_3 type of water. Although bicarbonate is the dominant anion in the groundwater system of study area, its concentration decreases and is replaced by chloride and sulfate anions along the groundwater flow direction. In the western part of the watershed, the groundwater is characterized by relatively low salinity and high bicarbonate concentration, ranging from 61 to 76 % of the total anions. Groundwater from the eastern part of the watershed has a significant concentration of sulfate and chloride, with bicarbonate concentrations ranging between 41 and 51 % of the total anions. The groundwater system at the eastern part of the watershed has a higher salinity, and its chemistry had evolved along its pathway from the western part of the watershed to the discharge area, and the proportion of chloride and sulfate increases at the expense of bicarbonate (Figure 4.4 and Figure 4.5). The importance of chloride and sulfate in the groundwater system in the eastern part of the watershed is attributed to a longer subsurface residence time and slow groundwater movement.

Lake Beseka is a saline and alkaline lake. The hydrological setting of Lake Beseka, especially the ratio of inflows to outflows, plays an important role in determining the hydrochemical characteristics of the lake. The hydrochemistry of the lake is affected by temporal variations, as the lake water has been diluted by relatively less-saline water inflows to the lake since the late 1960's (Figure 4.6). Comparison of the water chemistry of the lake with all water bodies in the area indicate that the lake has a similar hydrochemical signature to that of the groundwater system flowing from the western part of the watershed and to that of the hot springs that emerge at the southwestern edge of the lake. Groundwater that flows from the western part of the watershed and the hot springs that emerge at the southwestern corner of the lake are typified as $Na\text{-}HCO_3$ type water, while the lake is characterized by a $Na\text{-}HCO_3\text{-}Cl$ type of water. Salinity and concentration of chloride are higher in the lake compared to the groundwater in the western part of the watershed due to the dominance of evaporative enrichment in the lake water.

Irrigation water, on the other hand, exhibits a different hydrochemical signature from Lake Beseka, and is dominated by two cations (sodium and chloride) and typified as a Na-*Ca-HCO3* type of water. This fact is clearly observed in the ionic ratio, and Schoeller and Piper plots (Figure 4.7 left and right, Figure 4.3). The results of the hydrochemical investigation indicate that groundwater flows from northwestern and southwestern part of the watershed toward the east, and discharges into the lake in the form of emerged hot springs. This groundwater inflow governs not only the hydrochemical properties of Lake Beseka, but also the hydrological characteristics of the lake.

Isotopic data

The Local Meteoric Water Line (LMWL) was established for the study area based on the monthly isotopic composition of Addis Ababa rainfall for 37 years. Rainfall plots along a line with a slope of 7.1 and deuterium excess of 12.3 ‰. A seasonal isotopic variation is evident in Addis Ababa rainfalls (Figure 5.2, left). Summer rains are characterized by a relatively depleted isotopic composition compared to the spring rains. The weighted average isotopic value of summer rainfall (July and August) is -2.3 ‰ in $\delta^{18}O$, while the spring rainfall (March and April) is characterized by weighted average isotopic value of + 1.2 ‰ in $\delta^{18}O$. This seasonal variation can be attributed to the difference in the source of moisture and local meteorological processes.

The isotopic composition of water bodies in the region varies from isotopically depleted groundwater to enriched lake water (Figure 5.4). The groundwater system in the western part of the watershed, and hot springs located at the southwestern corner of the lake are characterized by an average isotopic composition of -2.8 ‰ in $\delta^{18}O$. This groundwater system has an isotopic composition comparable to Addis Ababa summer rainfall (July and August), which is characterized by weighted average isotopic value of -2.3 ‰ in $\delta^{18}O$. This isotopic similarity indicates the importance of summer rainfall in recharging the groundwater system of the area. The relative depletion of this groundwater system compared to the average summer rain could be related to selective recharge through heavy rainfall along tectonic structures. This depletion indicates that there is no evaporative enrichment of rainfall prior to recharge. This groundwater contain only little amount of tritium (< 0.8 TU), which suggests a sub-modern nature of the groundwater system, which had followed deeper circulation pathways from a high altitude source.

The groundwater system in the eastern part of the watershed is characterized by an average isotopic composition of -1.5 ‰ in $\delta^{18}O$, which is relatively enriched with respect to the summer rain but still depleted compared to the spring rain. Summer rainfall is the principal source of groundwater recharge in the area. The slight enrichment in isotopic composition can possibly be attributed to a mixing of depleted groundwater with slightly enriched river water. The tritium content of this groundwater is slightly significant (1.6-5.4 TU), which also suggests a mixture of modern and sub-modern meteoric water. Infiltration of water from the River Awash to the underneath groundwater system at the Abadir irrigation field could create favorable conditions for mixing.

Water from Lake Beseka is isotopically enriched with respect to its inputs (rainfall, runoff and groundwater inflow) due to evaporative enrichment, and plot along local evaporative line is defined by a slope of 5.1 and intercept of 5.5. This line intersects the local meteoric water line of Addis Ababa rainfall at the input signal, which is characterized by an isotopic value of -3 ‰ in $\delta^{18}O$ and -9.3 ‰ in $\delta^{2}H$ (Figure 5.7, right). The input signal has a similar isotopic composition to that of the groundwater in the western part of the watershed and the hot springs that emerge at the southwestern corner of the lake. These waters are characterized by an average isotopic value of -2.8 ‰ in $\delta^{18}O$ and -10.7 ‰ in $\delta^{2}H$. On the other hand, the average isotopic composition of river water introduced to the study area for irrigation is -0.7 ‰ in $\delta^{18}O$ and + 3.2 ‰ in $\delta^{2}H$, which is in contrast to the isotopic composition of the input signal. This suggests that the groundwater system that flows from the western part of the watershed discharges to Lake Beseka in the form of hot springs, and mainly governs water inflows to the hydrology of Lake Beseka.

Recharge estimation and groundwater modeling

The rainfall distribution of the study area is bimodal (spring and summer rainfall) following the convergence of air flows from the Atlantic and the Indian Ocean (Figure 6.1, right). Comparison of daily rainfall with potential evapotranspiration indicates the presence of surplus daily rainfall that exceeds potential evapotranspiration; this surplus might be source of recharge to the groundwater system. The analysis of the temporal groundwater level indicates that groundwater responds to the rainfall events during both spring and summer rainy seasons. During the summer rainy season, groundwater starts to increase about three weeks after the onset of the rainy season, and reaches a maximum value after 90 to 100 days. Recharge is also reflected by the temperature of the groundwater system, which decreases following the arrival of rainwater to the water table (Figure 3.5, left). In this study, groundwater recharge was estimated through the water table fluctuation method, the chloride mass balance method, and EARTH modeling (Table 6.3).

A groundwater flow model was constructed for the area to estimate the rate and direction of groundwater movement for the aquifer system and hydraulically connected lake. The groundwater recharge that annually replenishes groundwater system of the study area was estimated to be 17.4 Mm^3. This recharge accounts for 35 % of the total groundwater inflow to the aquifer system of the lake watershed. Additionally, 30.4 Mm^3 of groundwater laterally annually flows to the groundwater system of the lake watershed across the northwestern and southwestern boundaries, accounting for 62% of the total inflow. Groundwater leakage to Lake Beseka is the main sink of water from the groundwater flow system. It accounts for 69% of the total groundwater outflow from the groundwater system. Groundwater seepage to Lake Beseka is computed as a 33.8Mm^3 per annum, accounting for 51% of the total inflow to the lake. Precipitation that directly falls on Lake Beseka was estimated to be 24.2Mm^3 per annum, accounting for 37 % of the total water inputs. Lake Beseka is a terminal lake without surface and groundwater outlets. Thus, evaporation is the only water loss mechanism for the lake, accounting for 99% of the total outflow from the lake.

The piezometric map of the model indicated that groundwater flows from the west and south part of the watershed toward Lake Beseka. Due to its topographic location in the watershed, Lake Beseka functions as a discharge zone for groundwater flow system of the region. The lake has been gaining water from the surrounding groundwater system at monthly average rate of 2.7 Mm^3. The model result indicated that the storage of Lake Beseka has been increasing by an average annual rate of 5.4 Mm^3 for the last 31 years. The expansion of Lake Beseka has been influenced by extreme climatic conditions such as droughts and severe wet periods. Storage of the lake has been affected by seasonal rise and fall as well, and its seasonality has been governed by fluctuation of precipitation – evaporation fluxes.

8.3 Conclusions

The water budget of Lake Beseka has changed since the late 1960's when relatively less-saline water started to flow to the lake. Three hypotheses can be forwarded to explain the sudden expansion of Lake Beseka. The first hypothesis is based on changes related to the climate of the region, which could lead to an increase in water inflow to Lake Beseka or reduce water outflow from the lake. The second hypothesis is the possible effect of excess water from irrigation practice south of the lake at Abadir farm. Possible changes in the hydrogeological regime of the area that might modify the amount of groundwater that flows to the region and then to Lake Beseka is the third hypothesis.

The analysis of climate parameters revealed that there has not been any climate change in the region that could have possibly triggered the growth of Lake Beseka. The precipitation and other climatic factors that govern evapotranspiration process have remained nearly the same in the study area since the establishment of meteorological measurements. Thus, this can be ruled out from being the possible explanation. The effect of excess irrigation water from the Abadir farm has been a widely accepted theory for the expansion of Lake Beseka in previous studies. However, hydrochemical and isotopic evidence indicate that Lake Beseka has a signature different to that of the irrigation water that has been used at the Abadir farm. Mixing of irrigation water with the underneath groundwater system could be possible, but there is no indication that this infiltrated irrigation water flows directly to the lake to increase water inflows of Lake Beseka thus to induce the lake growth. Furthermore, there is no evidence that shows that the amount of infiltrated irrigation water is significant enough to continuously provide water inflow to the lake to sustain the continuous expansion of the lake.

Hydrochemical and isotopic evidence indicate that groundwater flows from the western part of the watershed and discharges to Lake Beseka in the form of hot springs. This groundwater inflow from the western part of the watershed constitutes the major water inflow to the lake and forms an integral part of its water budget. This fact is well supported by groundwater modeling results, which estimated 51% of the total water inflow to the lake coming from groundwater seepage to the lake. This groundwater seepage to the lake was computed by the model to be annually 33.8Mm^3. It is evident from the recharge estimation that the recharge of the groundwater system takes place from infiltration of local precipitation in the area. However, this recharge from in the watershed of Lake Beseka is estimated to be only 17.4 Mm^3 annually. Thus, the groundwater recharge that seasonally replenishes the aquifer system within the lake watershed is not significant enough to explain the expansion of Lake Beseka, as more amount groundwater is flowing to the lake each year. The model estimated that

$30.45 Mm^3$ of groundwater laterally flows annually to the aquifer system of the lake watershed across the surface boundaries. The importance of Lake Koka as the possible source of groundwater to the watershed of Lake Beseka can be ruled out due to the fact that the water of Lake Koka is characterized by modern meteoric water with isotopic and hydrochemical signature more comparable to River Awash than that of Lake Beseka. Thus, it is most likely that lateral groundwater inflow to the watershed of Lake Beseka could be related to the mountain recharge.

The mountain recharge plays an important role as a source of recharge to the aquifer system of the Ethiopian rift valley. This is due to the fact that mountains are characterized by a higher amount of precipitation compared to the adjacent valleys because of their location in the windward side of the moisture sources and the orographic enhancement of rainfalls. The importance of a plateau-rift groundwater transfer as a source of recharge to the aquifer system of the rift has been discussed by different authors (Gizaw, 1996; Ayenew, 1998; Chernet, et al. 2001, Kebede, et al. 2007). In mountain-bounded lowland aquifers, the relative importance of the recharge component coming from the mountain as groundwater inflow compared to recharge from precipitation in the valley depends on the hydrogeology (nature and depth of bounding faults, the presence of transverse faults connecting the mountain with the valleys, the permeability of the mountain mass, etc.) of the interface between the mountains and the valleys (Kebede et al., 2007).

As described earlier, the amount of recharge from infiltration of local precipitation is small in the study area, and because of that more than 60% of the total groundwater inflow to the aquifer system of the region comes from outside the watershed of Lake Beseka. The Ethiopian Plateau, which is situated west of the rift valley and commonly referred as the North Western Plateau (NWP), could be a source for the groundwater system of the region. The isotopic composition of the groundwater in the western part of the watershed shows similar isotopic signals to waters from this plateau. The depleted isotopic composition and the sub-modern nature of this groundwater system indicate a high altitude source, which had followed deeper circulation pathways for a considerable time. The study area is located in a tectonic structural zone where NNE-SSW trending rift faults intersect with E-W trending Yerer - Tullu Wellel Volcano Tectonic Lineament (YTVL) structures. This zone is described as a hydrogeological window by Kebede et al. (2007) that channels groundwater from the mountains to the rift. The presence of this structural zone could transfer isotopically depleted groundwater from the central part of the western plateau to the region.

The importance of the groundwater that laterally flows to the aquifer system of the region is apparent. According to all evidences from hydrochemical and isotopic data, groundwater level map, and groundwater modeling, the groundwater that flows from the western part of the watershed to Lake Beseka forms a major contribution to the water budget of the lake. Thus, it is most likely that the expansion of Lake Beseka is related to changes in the amount of this groundwater, which first flows into the lake watershed and then to Lake Beseka. This could be due to modification of the hydraulic gradient of the groundwater in the region, which might increase the discharge of the hot springs that continuously flows into the lake at its western edge. Tectonic activities are frequent in this section of the rift valley, and abrupt tectonic movement can caused changes in the groundwater regime. Owing to the geologic setting of the study area in the seismically and volcanically active section of the northern Main Ethiopian Rift (MER), the tectonically induced modification of hydrogeological regime of the area, which changes the amount of groundwater that flows to the region, is to a high degree of certainty the explanation for the expansion of Lake Beseka.

8.4 Perspectives

The results of this study should be regarded as a single effort toward recognizing the importance of groundwater flux to the hydrologic system of Lake Beseka and understanding the intricate hydrogeologic environment of the rift. The area is situated in a remote section of the rift valley in Ethiopia. Consequently, this work faced limitations with regard to the quality and quantity of data, as well as to the spatial and temporal coverage of hydrological, hydrogeological, hydrochemical and isotopic data. The complexity of any system can be unveiled through proper utilization of existing records and by establishing time series data. With this regard, the following points are raised with hope that they will be addressed in the future groundwater studies not only in the study area but also in most part of the country.

- Groundwater monitoring is a very essential step towards evaluating and managing any groundwater resource. However, monitoring of groundwater levels is not established in most parts of the country. Observation wells are usually part of well design, but they are hardly used to monitor temporal groundwater fluctuation. A large number of abandoned wells are found in the rift, and they can be useful for monitoring groundwater heads. Automatic data loggers are not that expensive, easy to use, and can be easily installed in either pumping or abandoned boreholes. Moreover, automatic data loggers can easily be integrated with manual groundwater techniques to establish a network of groundwater level monitoring stations. However, the responsibilities for managing abandoned boreholes or for running groundwater monitoring tasks are not clear, neither at the regional nor at the national level.

- During construction of boreholes, well completion reports are prepared for individual wells. These reports should include a lithological description of the well, well design with casing set up, pumping test data and water quality data. These reports are normally prepared by the contractors and are submitted to the owner of the boreholes at the end of the project. Although a great deal of money has been spent on construction of boreholes by different bodies (individuals, companies, NGOs, governmental water authorities, etc.), a central groundwater bureau does not exist at the regional or national level to gather, organize and set up databases for these useful scientific records. As a result, well data exist only in a very random and non-systematic manner, and everything depends on the goodwill of people. Thus, obtaining these data becomes a matter of chance rather than being a matter of right and certainty.

- Semi-arid areas are characterized by a high rate of evapotranspiration, and the rainfall has sporadic nature. Thus, it is highly recommended to collect hydrological and meteorological records at the lowest possible time scale, and an estimation of recharge should be conducted at least on a daily scale.

- Data on the hydrochemical and isotopic composition of water has proven useful tools to obtain qualitative hydrological information. Hence, coupling of chemical methods with physical approaches is advisable to gain a better picture of the groundwater system.

9 REFERENCES

Abebe T, Mazzarini F, Innocenti F, and Manetti P (1998) The Yerer –Tullu Wellel volcano-tectonic lineament: a transitational structure in central Ethiopia and the associated magmatic activity. Journal of African Earth Science 26: 135 - 150

Abebe T (2000) Geological limitations of a geothermal system in a continental rift zone: Example the Ethiopian Rift Valley. Proceedings of the World Geothermal Congress, Kyushu - Tohoku, Japan, 2000

Abebe T, Manetti P, Bonini M, Corti G, Innocenti F, and Mazzarini M (2005) Geological map of the northern main Ethiopian rift. The Geological society of America, Boulder, CO

Abbott MB and Refsgaard JC (1996) Distributed hydrological modeling. Kluwer academic publisher, Netherlands

Aggarwal PK, Froehlich K, Kulkarni KM and Garcia – Agudo E (2004) Isotopes techniques in groundwater investigations, (Chapter 9) In: Groundwater Studies - An International Guide for Hydrogeological Investigation. Published by UNESCO, IHP-VI, Series on Groundwater No3, ISBN 92-9220-005-4

Alemayehu T, Ayenew T, and Kebede S (2005) Hydrogeochemical and lake level changes in the Ethiopian Rift. Journal of Hydrology, Volume 316, Issues 1-4

Allan RG, Perieira LS, Raes D, and Smith M (1998) Crop evapotranspiration – Guidelines for computing crop water requirements – FAO Irrigation and drainage paper 56. FAO – Food and Agriculture Organization of the United Nation, Rome

Alley WM, Healy RW, LaBaugh JW, Reilly TE (2002) Flow and Storage in Groundwater Systems. Science vol. 296

Anderson MP and Woessner WW (1992) Applied groundwater modeling. Academic Press, Inc. San Diego, California :381 pp

Anderson MP (2002) Groundwater – Lake interaction: response to climate change Vilas Country, Winsconsin. Final Report to the Water Resources Institute, University of Wisconsin, 16pp

Ayenew T (1998) Hydrogeological System of the Lake district basin, central Main Ethiopian Rift. ITC publication no 64, PhD thesis 259p. Enschede, Netherlands

Ayenew T (2004) Environmental implications of changes in the levels of lakes in the Ethiopian Rift since 1970. Reg. Environ. Change, Springer-Verlag, 192-204

Ayenew T (2007) The distribution and hydrogeological controls of fluoride in the groundwater of central Ethiopian rift and adjacent highlands. Environmental Geology Springer Berlin / Heidelberg, ISSN 0943-0105

Baalousha HM (2003) Risk assessment and uncertainty analysis in groundwater modeling. PhD thesis submitted to Rheinisch Westfälischen Technischen Hochschule Aachen, 145pp

Balek J (1988) Groundwater recharge concepts. In: I. Simmers (ed.) Estimation of Natural Groundwater Recharge. NATO Advanced Sciences Institutes Series C: 3-9

Bastow ID, Kendall JM, and Ebinger CJ (2005) Upper-mantle seismic structure in a region of incipient continental breakup: northern Ethiopian rift. *Geophys. J. Internat.*, v.162; p. 479 – 493

Bear J, Beljin MS, and Ross RR (1992) Fundamentals of Groundwater modeling. Groundwater issue, EPA/540/S-92/005, 11pp

Beekman HE and Xu Y (2003) Review of groundwater recharge estimation in arid and semiarid Southern Africa. In: Xu, Y., Beekman, H.E., (Eds), Groundwater recharge estimation in Southern Africa. UNESCO IHP Series No. 64, published by UNESCO Paris. ISBN 92-9220-000-3

Bendick R, McClusky S, Bilham R, Asfaw L, and Klemperer S (2006) Distributed Nubia –Somalia relative motion and dike intrusion in the Main Ethiopian Rift. *Geophys. J. Internat., v.*165; p. 303-310

Beven KJ (2001) Rainfall – Runoff Modeling. The Premier, John Wiley & Sons ltd., Chichester, England, 351pp

Boccaletti M, Bonini M, Mazzuoli R, Abebe B, Piccardi L, and Tortrici L (1998) Quaternary oblique extensional tectonics in the Ethiopian Rift (Horn of Africa). Tectonophysics 287, 97–116

Bredenkamp DB and Xu Y(2003) Perspectives on recharge estimation in dolomitic aquifers in South Africa. In: Xu Y, and Beekman HE (ed) Groundwaters recharge estimation in Southern Africa. UNESCO IHP Series No. 64, published by UNESCO Paris. ISBN 92-9220-000-3

Cheng X and Anderson MP (1993) Numerical simulation of groundwater interaction with lakes allowing for fluctuating lake levels. Groundwater. v. 31: 6, pp. 929-933

Chernet T (1982) Hydrogeology of lakes region, Ethiopia (Lake Ziway, Langano, Abiyata, Shalla and Awassa). Ethiopian Institute of Geological Survey, 97pp

Chernet T (1990) Hydrogeology of Ethiopia and water resources development. Report to Ethiopian Institute of Geological Surveys, Ministry of Mines and Energy, 198pp

Chernet T, Travi Y, and Valles V (2001) Mechanism of degradation of the quality of natural water in the lakes region of the Ethiopian Rift Valley. Water Reser.35, pp. 2819–2832

Chiang WH (2005) 3D – Groundwater Modeling with PMWIN. Second Edition: A simulation system for modeling groundwater flow and transport processes

Clark I and Fritz P (1997) Environmental Isotopes in Hydrogeology. Lewis Publishers, New York, USA, 327pp

Cook PG (2003) A guide to regional groundwater flow in fractured rock aquifers. CSIRO Land & Water, Glen Osmond, SA, Australia, 108pp

Conrad J, Nel J, and Wentzel J (2004) The challenges and implications of assessing groundwater recharge: A case study – Northern Sandveld, Western Cape, South Africa. Water SA Vol. 30 No. 5 (special edition)

Cornwell DG, Mackenzie GD, England RW, Maguire PKH, Asfaw LM, and Oluma B (2006) Northern Main Ethiopian Rift Crustal Structure from New High-Recision Gravity Data. Geological Society, London, Special Publications, v. 259, pp. 307 – 321

Council GW (1999) A Lake package for MODFLOW (LAK2), Documentation and user's manual version 2.2. HIS GEOTRANS, A Tetra Tech Company, 132pp

Council GW (2005) The surface / Vadose package for computing runoff, Evapotranspiration, and net recharge in MODFLOW. Proceedings of the 2005 Georgia water resources conference, April 25-27, University of Georgia, Athens, Georgia

Craig H (1961) Isotopic variations in meteoric waters. Science 133: 1702-1703

Custodio E (2004) Hydrogeology of volcanic rocks. In: Groundwater studies - an international guide for hydrogeological investigation (Chapter 16). Published by UNESCO, IHP-VI, Series on Groundwater No3, ISBN 92-9220-005-4

Demlie M, Wohnlich S, Gizaw B, and Stichler W (2006) Groundwater recharge in the Akaki catchment, central Ethiopia: evidence from environmental isotopes (δ18O, δ 2H and 3H) and chloride mass balance. Hydrol. Process. 21, pp. 807 – 818, Published online 12 October 2006 in Wiley InterScience

Dingman SL (2002) Physical hydrology, Second edition. Prentice-Hall, Inc., ISBN 0-13-099695-5, 646pp

Doherty J (2004) PEST, model independent parameter estimation, user manual 5th edition. Watermark numerical computing, 336pp

Domenico PA and Schwartz FW (1998) Physical and chemical hydrogeology. 2nd Edition, John Wiley, New York

Driscoll FG (1986) Groundwater and wells, 2nd. ed. Johnson Div., St. Paul, MN 55112, 1,089 pp

Eberts SM and George LL (2000) Regional ground-water flow and geochemistry in the Midwestern Basins and Arches Aquifer System in Parts of Indiana, Ohio, Michigan, and Illinois. U.S. Geological Survey professional paper 1423 – C, Reston, Virginia

Ebinger C and Casey M (2001) Continental breakup in magmatic provinces: an Ethiopian example. Geology, v. 29; no. 6, pp. 527-530

ELC - electroconsult (1987) Geothermal reconnaissance study of selected sites of the Ethiopian Rift System. Geological report for the Ethiopian Institute of Geological Surveys

Eriksson E and Khunakasem V (1969) Chloride concentration in groundwater, recharge rate and rate of deposition of chloride in the Israel coastal plain. J. Hydrol., Vol. 7, 178-197

FAO (1984) Agroclimatological data for Africa, vol1: Countries north of equator. Plant production and protection series, 22, FAO, Rome.

Fetter CW (2001) Applied hydrogeology, fourth edition. Prentice-Hall, Inc., ISBN 0-13-088239-9

Fontes J Ch, Florkowiski T, Saliege JF, and Zuppi GM (1980) Environmental isotope study of groundwater systems in the Republic of Djibouti. In Arid Zone Hydrology IAEA; 237-262

Freeze RA and Cherry JA (1979) Groundwater. Prentice-Hall, Englewood Cliffs, N.J., USA. 604pp

Gemechu D (1977) Aspects of climate and water budget in Ethiopia. Addis Ababa University press, 71 pp

Genetti AJ (1999) Engineering and design groundwater hydrology. Department of the Army U.S. Army Corps of Engineers, 83pp

Gibson IL (1970) A Pantelleritic Welded Ash – Flow tuff from the Ethiopian Rift Valley. Contr. Mineral and Petrology 28, pp. 89 -111

Gizaw B (1996) The origin of high bicarbonate and fluoride concentrations in waters of the Main Ethiopian Rift Valley, East African Rift system. J. African Earth Sci. 2, pp. 391-402

Glynn PD and Plummer LN (2005) Geochemistry and the understanding of ground-water systems. Springer-Verlag, Hydrogeol J 13: pp. 263–287

Görner A (2005) Non climatic lake level Changes in Ethiopia. TU Bergakadenie Freiberg, Faculty of Geoscience, Geotechnology and Mining, 9pp

Görner A, Gloaguen R, and Foltyn M (2006) Remote sensing for monitoring lake level changes in Eastern Africa. Proceeding of the 2nd Göttingen GIS and Remote Sensing Day

Griffioen J (2004)Groundwater quality. In: Groundwater Studies - An international guide for hydrogeological investigation (Chapter 2). Published by UNESCO, IHP-VI, Series on Groundwater No3, ISBN 92-9220-005-4

Gurwin J and Lubczynski M (2005) Modeling of complex multi-aquifer systems for groundwater resources evaluation-Swidnica study case (Poland). Hydrogeology Journal 13: pp. 627–639

Halcrow W (1978) The study of Beseka Lake Levels. Report for the Government of Ethiopia Awash Valley Development Agency, 83pp

Harbaugh AW (2005) MODFLOW – 2005, the U.S. Geological Survey modular Groundwater model - the Groundwater flow process. In: U.S. Geological Survey Techniques and Methods 6- A16, 253pp

Healy RW and Cook PG (2002) Using groundwater levels to estimate recharge. Hydrogeology journal, 10: pp. 91 -109

Healy RW, Winter TC, LaBaugh JW, and Franke OL (2007) Water budgets: Foundations for effective water-resources and environmental management. U.S. Geological Survey Circular 1308, 90 pp

Heath RC (1983) Basic Groundwater hydrology: U.S. Geological Survey Water-Supply Paper 2220, 86 pp

Hem JD (1992) Study and interpretation of the chemical characteristics of natural water. U.S Geological survey water supply paper 2254, third edition: 263pp

Hounslow AW (1995) Water quality data analysis and interpretation. Lewis publisher, 372pp

Hunt RJ, Haitjema HM, Krohelski JT (2003) Simulating groundwater – lake interactions: Approaches and insights. Groundwater, 41(2): pp. 227-37

Johnson AI (1967) Specific yield - Compilation of specific yields for various materials. U.S. Geol.Surv., Water-Supply Pap. 1962-D, U.S. Govt. Print. Off., Washington, DC 20402, 74 pp

Kamra SK (2007) A Critique on methodology for estimation of groundwater recharge in India: lessons from a visit to US. [Online]. Available at,
http://www.waterandfood.org/gga/Presentations%20from%20RW&S%20 Workshops.htm

Kebede E, Gebre Mariam Z, Ahlgren I (1994) The Ethiopian rift valley lakes: chemical characteristics of a salinity-alkalinity series. Hydrobiologia, Springer Netherlands, volume 228 no 1, 12pp

Kebede G (1987) Hydrogeology of Nazreth area , NC 37-15. Report to Ethiopian Institute of Geological Surveys, Ministry of Mines and Energy, 106pp

Kebede S (2004) Environmental isotopes and geochemistry in investigating groundwater and lake hydrology: cases from the Blue Nile basin & the Ethiopian Rift. PhD thesis, University of Avignon, France

Kebede S, Travi Y, Asrat A, Alemayehu T, Ayenew T, and Tessema Z (2007) Groundwater origin and flow along selected transects in Ethiopian rift volcanic rocks. Hydrogeology Journal (2008) 16: pp. 55-73. Published on line: 2 August 2007

Keir D, Ebinger CJ, Stuart GW, Daly E, and Ayele A (2006) Strain accommodation by magmatism and faulting as rifting proceeds to breakup: Seismicity of the northern Ethiopian rift. Journal of Geophys. Res., 111, B05314, doi: 10.1029/2005JB003748

Keranen K, Klemperer SL, and Gloaguen R (2004) Three-dimensional seismic imaging of a protoridge axis in the Main Ethiopian rift. Geological Society of America, v. 32; no 11, pp 949 – 952

Keranen K and Klemperer SL (2007) Discontinuous and diachronous evolution of the Main Ethiopian Rift: Implications for development of continental rifts. Earth and Planetary Science Letters 265 (2008) 96–11

Kinzelbach W, Bauer P, Siegfried T, Brunner P (2003) Sustainable groundwater management - problems and scientific tools. Episodes, Vol. 26, no. 4.

Kolja R, Aly I, El-Kadi, and Gingerich SB (2007) Estimating hydraulic properties of volcanic aquifers, using Constant-rate and variable-rate aquifer tests1. Published by Journal of the American Water Resource Association, Volume 43, Issue 2, pp 334-245

Konikow LF and Bredehoeft JD (1992) Groundwater models cannot be validated. Advance in Water Resources, 15 (1): pp.75-83

Konikow LF and Reilly TE (1998) Groundwater modeling. In: The Handbook of groundwater engineering [J.W. Delleur, ed.], CRC Press, Boca Raton pp. 20:1-20.40

Korme T, Acocella V, and Abebe B (2004) The role of pre-existing structures in the origin, propagation and architecture of the faults in the Main Ethiopian Rift. Gondwana Research, 7(2), pp. 467–479

Külls C (2000) Groundwater of the North-Western Kalahari, Namibia - Estimation of recharge and quantification of the flow systems. PhD thesis submitted to University of Würzburg, 166 pp

Kurz T, Gloaguen R, Ebinger C, Casey M, and Abebe B (2007) Deformation, distribution and type in the Main Ethiopian Rift (MER); a remote sensing study. Journal of African Earth Sciences, v.48, Issues 2-3; pp. 100 – 114

Lacaux JP, Delmas R, Kouadio G, Cros B, and Andreae MO (1992) Precipitation chemistry in the Mayombé forest of equatorial Africa: Journal of Geophysical Research 97: pp. 6195-6206

Larsson I (1984) Groundwater in hard rocks. Project 8.6 of the International hydrological programme. Published by UNESCO, ISBN 92-3-101980-5, 221pp

Lerner DN, Issar AS, and Simmers I (1990) Groundwater Recharge: A guide to understanding and estimating natural recharge. International contributions to Hydrogeology, volume 8, 345pp

Le Turdua C, Tiercelin JJ, Gibert E, Travi Y, Lezzar KE, Richert JP, Massault M, Oise M., Gasse F, Bonnefille R, Decobert M, Gensous B, Jeudy V, Tamrat E, Mohammed MU, Martens K, Atnafu B, Chernet T, Williamson D, Taieb M (1999) The Ziway – Shala lake basin system, Main Ethiopian Rift: influnence of volcanism, tectonics, and climate forcing on basin formation and sedimentation. Palaeogeography, Palaeoclimatology, Palaeoecology 150 pp 135–177

Lloyd JW (1986) A review of aridity and groundwater. Hydrological processes 1, pp.63 – 78

Lubczynski MW and Gurwin J (2005) Integration of various data sources for transient groundwater modeling with spatio – temporally variable fluxes – Sardon study case, Spain. Journal of Hydrology 306, pp.71–96

Lubczynski MW and Obakeng OT (2004) Monitoring and modeling of fluxes on Kalahari- setup and strategy of the Kalahari monitoring project: Serowe study case, Botswana. In: Water resources of arid areas: proceedings of the international conference on water resources of arid and semi arid regions of Africa WRASRA, August 3-6th, 2004, Gaborone, Botswana. pp. 271-278

Lubczynski MW and Roy J (2005) MRS contribution to hydrogeological system parameterization. Near Surface Geophysics 3, pp.131–139

Mace RE, Austin B, Angle ES, and Batchelder R (2007) Surface water and Groundwater – together again? State Bar of Texas, 8th Annual changing face of water rights in Texas, June 28 – 29, 2007, San Antonio

Mackenzi GD, Thybo H, and Maguire PKH (2005) Crustal velocity structure across the Main Ethiopian Rift: Results from 2-dimensional wide angle seismic modeling. Geophysical Journal International, 162, pp.994–1006

Maguire PKH, Keller GR, Klemperer SL, Mackenzie GD, Keranen K, Harder S, O'Reilly B, Thybo H, Asfaw L, Khan MA, and Amha M (2006) Crustal Structure of the Northern Main Ethiopian Rift from the EAGLE controlled – source survey, a snapshot lithospheric Breakup. Geological Society, London, Special Publications, v.259; pp. 269 -292

Mandle RJ (2002) Groundwater modeling guidance. Groundwater modeling program, Michigan Department of Environmental Quality, 54pp

McDonald MG and Harbauch AW (1988) A modular three-dimensional finite-difference groundwater flow model. Techniques of water resources investigation of the United Geological Survey, Book 6, 588pp

Merritt ML and Konikow LF (2000) Documentation of a computer program to simulate lake-aquifer interaction using the MODFLOW groundwater flow model and the MOC3D solute-transport model. U.S. Geological Survey, Water resources investigations report 00-4167, 146pp

Middlemis H (2001) Groundwater flow modeling, guideline for Murry – Darling Basin Commission, Aquaterra Consulting Pty Ltd, 125pp

Mohr PA (1960) Report on a geological excursion through southern Ethiopia. Bulletin of the Geophysical Observatory of Addis Ababa 3, pp. 9–20

Mohr PA (1962) The Ethiopian rift system. Bull. Bulletin of the Geophysical Observatory of Addis Ababa, 5, pp. 33-62

Mohr PA (1967) The Ethiopian rift system. Bulletin of the Geophysical Observatory of Addis Ababa 11, pp.1–65

Mohr PA (1967) Major volcano–tectonic lineament in the Ethiopian rift system. Nature 213, pp. 664–665

Mohr PA (1968) Transcurrent faulting in the Ethiopian rift system. Nature 218, pp.938–941

Mohr PA (1987) Patterns of faulting in the Ethiopian rift valley. Tectonophysics 143, pp.169–179

Mook WG (2000) Environmental isotopes in the hydrological cycle principles and applications. Vol 1 UNESCO, Paris. 280pp

Mook WG, Gat JR, and Meijer HAJ (2000) Environmental isotopes in the hydrological cycle. Principles and applications, technical documents in hydrology. Vol II UNESCO, Paris. 114pp

Mook WG, Rozanski K, and Froehlich K (2000) Environmental isotopes in the hydrological cycle. Principles and applications, technical documents in hydrology. Vol III UNESCO, Paris. 117pp

Murray (1996) Review of groundwater resource evaluation based on recharge. In: Groundwater resource assessment II – Recharge literature review report 3aA (2006). Department of water affairs and forestry, South Africa.

Nash J and Sutcliffe J (1970) River flow forecasting through conceptual models part 1 – a discussion of principles. Journal of Hydrology, 10, pp. 282-290

NRC (National Research Council) (1996) Rock fractures and fluid flow: Contemporary understanding and applications. National Academy press, Washington DC. 551 pp

Nield SP, Lloyd RT, and Barr, AD (1994) A framework for quantitative analysis of surface water – groundwater interaction: Flow geometry in a vertical section. Water Res. Res. 30: 2461 – 2475, South Africa

PDP (Pattle Delamore Partners) (2002) Groundwater Model Audit Guidelines. Ministry for the Environment, Mantu Te Taiao, 225pp

Rooney TO, Tanya F, Yirgu G, and Ayalew D (2005) Structure of the Ethiopian lithosphere: Xenolith evidence in the Main Ethiopian Rift. Geochimica et Cosmochimica Acta, Vol. 69, No. 15, pp. 3889–3910

Rozanski K, Araguas-Arguas L, and Gonfiantini R (1996) Isotope patterns of precipitation in East Africa. In: Johnson, T.C., Odada, E. (ed). The Climatology, Paleoclimatology, Plaeoecology of the East African Lakes, Gordan and Breach, Toronto, pp 79-93

Sandwidi JP (2007) Groundwater potential to supply population demand within the Kompienga dam basin in Burkina Faso., PhD thesis submitted to University of Bonn, 160pp

Scanlon BR, Healy RW, and Cook PG (2002) Choosing appropriate techniques for quantifying groundwater recharge. Hydrogeology Journal (2002) 10:pp.18–39

Simmers I (1988) Estimation of natural groundwater recharge. NATO ASI Series C: V.22, D. Reidel publishing Company, 510pp

Sophocleous MA (1991) Combining the soil water balance and water – level fluctuation methods to estimate natural groundwater recharge: practical aspects. Journal of Hydrology, 124, pp. 229-241

Sophocleous MA (2002) Interactions between groundwater and surface water: the state of the science. Hydrogeology Journal 10, pp.52–67

Sophocleous MA (2004) Groundwater recharge. In: Groundwater, [Eds. Luis Silveira, Stefan Wohnlich, and Eduardo J. Usunoff], in Encyclopedia of Life Support Systems (EOLSS), Developed under the Auspices of the UNESCO, Eolss Publishers, Oxford ,UK, [http://www.eolss.net]

Stafford P, Toran L, and Mckay L (1998) Influence of fracture truncation on dispersion: A dual permeability model. Journal of contaminant hydrology 30: pp.79-100

Sulebak JR (2000) Applications of Digital Elevation Models. "White paper" developed as part of the DYNAMAP project, Oslo

Swanson SK, Bahr JM, Schwar MT, and Potter KW (2001) Two-way cluster analysis of geochemical data to constrain spring source waters. Elsevier Science B.V., Chemical Geology 179Ž2001, pp. 73–91

Telford RJ (1998) The palaeoenvironmental record of Holocene environmental change in the Ethiopian rift valley. Unpublished Ph.D. Thesis, University of Wales

Tessema Z (1998) Hydrochemical and water balance approach in the study of high water level rise of Lake Beseka. MSc. Thesis [unpublished], 90 pp., University of Birmingham, UK

Thyne G, Guler C, and Poeter, E (2004) Sequential analysis of hydrochemical data for watershed characterization. Groundwater, Vol. 42, No. 5

UNDP (1973) Geology, geochemistry and hydrology of hot springs of the East African rift system within Ethiopia, United Nations, New York

UNEP (2008) "Africa: Atlas of Our Changing Environment." Published by Division of Early Warning and Assessment (DEWA), United Nations Environment Program (UNEP), Nairobi, Kenya

UNESCO (2006) Groundwater resources of the world: transboundary aquifer systems, 1:50, 000000 scale. UNESCO, Paris

Usher BH, Pretorius JA, and Tonder GJ (2006) Management of a Karoo fractured-rock aquifer system – Kalkveld Water User Association (WUA). Water SA Vol. 32 No. 1

Van der Lee J and Gehrels JC (1990) Modeling aquifer recharge, introduction to lumped parameter model EARTH. Hydroloical report, Free University of Amsterdam, The Netherlands

Van der Lee J and Gehrels JC (1997) Modeling of groundwater recharge for fractured dolomite aquifer under semi arid conditions, recharge of phreatic aquifers in (semi) arid areas. IAH Book no. 19

Van Soest MC, Evans WC, Mariner RH, Schmidt ME (2004) Chloride in hotsprings of the Cascade volcanic arc – the source puzzle. In: Water – rock interaction, volume 1: proceeding of the eleventh international symposium. Wanty & Seal II (ed). Taylor and Francis group, London.

WHO (1984) Guidelines for drinking water quality. Report: World Health Organization, Geneva, 139pp

Williams FM, Williams MAJ and Aumento F (2004) Tensional fissures and crustal extension rates in the northern part of the Main Ethiopian Rift, Journal of African Earth Sciences, vol. 38, pp. 183-197

Williams MAJ, Williams FM, and Bishop PM (1981) Late Quarternary history of Lake Beseka, Ethiopia. In: Palaeoecology of Africa and the surrounding islands and Antarctica, A.A. Balkema, Cape Town, pp. 93-104

Winter TC (1981) Effects of water table configuration on seepage through lakebeds. Limnol. Oceanogr., 26(5), lQ81, pp. 925-934

Winter TC, Harvey JW, Franke OL, and Alley WM (1998) Groundwater and surface water a single resource. US. Geological survey Circular 1139, Denver, Colorado, 87pp

Winter TC (2000) Interaction of groundwater and surface water. Proceedings of the groundwater / surface water interactions workshop. United States Environmental Protection Agency, EPA/542/R-00/007: pp. 15 – 20

WoldeGabriel G, Aronson JL, and Walter RC (1990) Geology, geochronology, and rift basin development in the central sector of the Main Ethiopian Rift. Geological Society of America Bulletin, vol. 102, pp. 439- 459

Wolfenden E, Ebinger C, Yirgu G, Deino A and Ayalew D (2004) Evolution of the northern Red Sea rift: Birth of a triple junction. Earth and Planetary Science Letters, vol. 224, pp. 213- 228

Wood RB and Talling JF (1988) Chemical and algal relationships in a salinity series of Ethiopian inland waters. Hydrobiologia 158, pp. 29–67

WWDSE (Water Works Design and Supervision Enterprise) (1999) Study of Lake Beseka, Ministry of Water Resources, Addis Ababa, Ethiopia, Main Report, Vol 1, pp. 203

Xu Y and Beekman HE (2003) Groundwaters recharge estimation in Southern Africa. UNESCO IHP Series No. 64, published by UNESCO Paris. ISBN 92-9220-000-3

Zemedagegnehu E and Egizabher R (2004) Determination for the cause of rising water levels in Lake Beseka and design of remedial measure. Ethiopian Engineers Association Journal, August issues

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