Interbasin Groundwater Transfer in Upper Rift Valley Lakes and Upper Awash River Basins

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Addis Ababa University

June, 2015

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A thesis submitted to the School of Graduate Studies of Addis Ababa University in Partial fulfillment of the Degree of Masters of Science in Civil Engineering under Hydraulic Engineering.

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Advisor:

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Certification

The undersigned certify that he has read the thesis entitled**: Interbasin Groundwater Transfer in Upper Rift Valley Lakes and Upper Awash River Basins** and hereby recommend for acceptance by the Addis Ababa University in partial fulfillment of the requirements for the degree of Master of Science.

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Date

ABSTRACT

Groundwater models are used to predict the groundwater flow processes using mathematical equations based on certain simplifying assumptions. Numerical modeling method is used to investigate the possibility of groundwater transfer (interbasin groundwater flow) between Upper Awash River basin and Upper Rift Valley Lakes basin. Within the study the hydrogeology is investigated from previous studies, a conceptual groundwater model is produced and the numerical model is developed from this. The model conceptualization is done by idealizing the study area into three models related to the surface water divide and the main perennial Rivers and lakes. For these three basins the boundary conditions are identified and incorporated into the models. The boundaries are Upper Awash River, Meki River, Ziway Lake, and Koka Lake in addition to bottom and top surface of the aquifer in concern. The models are three dimensional steady state by finite element method. All the study basins are discretized according to finite element method into equal element dimensions. The elements are three dimensional. The finite element model used in the study was TAGSAC. The modeling process was towards estimating the hydraulic head in the model basins of the study area by a means of finding the hydraulic conductivity of different geological classifications and estimating surface recharge in terms of some percentage of the rainfall of the study area. Finally the models are calibrated with the measured hydraulic head for each model basins. The models results show that the groundwater divide and the surface water divide are not coincident. There is a clear hydraulic head difference which can make groundwater to flow from Upper Rift Valley Lakes basin towards Upper Awash basin in the middle of the study area, along Ziway- Koka Lake and from Upper Awash to Rift Valley Lakes basin at west and east part of it.

Key Words: Upper Awash River basin, Upper Rift Valley Lakes basin, Interbasin flow, Groundwater flow Modeling, Hydraulic head, TAGSAK

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TABLE OF CONTENTS

LIST OF FIGURES AND TABLES

LIST OF ACRONYMS

DEM: Digital Elevation Model

EPM: Equivalent Porous Medium

FDM: Finite Difference Method

FEM Finite Element Method

GIRD: Generation Integrated Rural Development

IGWT: Interbasin groundwater transfer

MER: Main Ethiopian Rift

MoWR: Ministry of Water Resources

NMSA: National Metrological Service Agency

NNE-SSW: North North East to South South West

NNW-SSE: North North West to South South East

N-S: North to South

REV: Representative Elementary Volume

RVLB: Rift Valley Lakes basin

SNNPR: Southern Nations and Nationalities Peoples' Region

UARB: Upper Awash River basin

URVLB: Upper Rift Valley Lakes basin

WFB: Wenji Fault Belt

1 INTRODUCTION

The Earth's hydrologic cycle consists of many varied and interacting processes involving all three phases of water. Groundwater flow is but one part of this complex dynamic hydrologic cycle. Saturated formations below the surface act as mediums for the transmission of groundwater, and as reservoirs for the storage of water. Water infiltrates to these formations from the surface and is transmitted slowly for varying distances until it returns to the surface by action of natural flow, vegetation, or man (Todd, 2005).

Groundwater, water below the ground surface, has recently become a major source of water supply in almost every sector. It makes up a significant part of world's water resource systems, supplying water for agriculture, industry and domestic use. Over-dependence on it for many purposes has led to its' over-exploitation, and this has led to much concern for groundwater assessment and management. In addition to that, expansion of human activities like industries, agricultural works using fertilizer and etc. causes dispersion of pollutants in the subsurface environment. The fate and movement of dissolved substances in soils and groundwater has generated considerable interest out of concern for the quality of the subsurface environment. These problems may not be limited in one certain basin but it may cross contaminate the neighbor basins through interbasin flow. Even if Ethiopian groundwater potential is not at such risk, in some areas it is nearly to happen/happening if it is not assessed and managed properly. But for a proper assessment and management of groundwater resources, a thorough understanding of the complexity of its processes is quite essential.

Almost every country is concerned about groundwater quality, and its usage within safe limits. On one side it helps to meet surface water deficits, while on other hand it also brings up more salts on the ground surface. Thus its use has created a vicious circle in which it is polluted as well as becomes a source of pollution. Though it is mostly free from sedimentation and pathogenic contamination but it is strong carrier of dissolved salts. The salt contamination is not only severe within the coastal aquifers but also is existing within inland aquifers especially where fresh groundwater is existing in a thin layer overlying a deep layer of highly saline groundwater. Effective planning and management of a groundwater system is an issue, which involves various crucial decisions, such as; determination of the quantum of water to be withdrawn periodically, location of pumping wells, rate of artificial recharge, maintenance of water quality in aquifer and the control conditions for aquifer boundaries.

Understanding the hydrological relation between basins has great importance in the study of watershed hydrology, management, and groundwater and pollution assessment. Groundwater movement beneath topographic divides is one component of the hydrological cycle that is typically ignored due to difficulty in analysis. Most watershed hydrology studies attempt to avoid interbasin groundwater transfer by selecting sites that are believed to be “tight”. While interbasin groundwater transfer (IGWT) is typically not addressed; it is an important hydrological process that has been found to occur throughout the world. Identifying and understanding of this hydrological process must be adapted in basin study and management for better and accurate measurement.

External forces which act on water in the subsurface include gravity, pressure from the atmosphere and overlying water, and molecular attraction between solids and water. In the subsurface, water can occur in the following: as water vapor which moves from regions of higher pressure to lower pressure, as condensed water which is absorbed by dry soil particles, as water which is retained on particles under the molecular force of adhesion, and as water which is not subject to attractive forces towards the surface of solid particles and is under the influence of gravitational forces. In the saturated zone, groundwater flows through interconnected voids in response to the difference in fluid pressure and elevation. The driving force is measured in terms of hydraulic head. The height of water measured in wells is the sum of elevation head and pressure head, where the pressure head is equal to the height of the water column above the screened interval within the well. Groundwater moves through the subsurface from areas of greater hydraulic head to areas of lower hydraulic head. The rate of groundwater movement depends upon the slope of the hydraulic head (hydraulic gradient), and intrinsic aquifer and fluid properties.

IGWT can be at least detected using head data. The potential for groundwater to move from basin to basin is related to the relative altitude and geological structure of the individual basin. Where the rocks that form the boundary between adjacent basins are sufficiently permeable, there will be flow into or out of the basin. Identifying the regional head and checking the geological structure of the two basins can give full information, about the possibility of IGWT. The detection of IGWT increases the importance of regional land use planning for areas overlying the IGWT system. This work will show the possibilities of interbasin groundwater between Upper Awash River basin (UARB) and Upper Rift Valley Lakes basin (URVLB).

Interbasin groundwater flow analysis have been an important research topic in the last few years. This is as a result of many geo-environmental engineering problems having direct or indirect impact on groundwater flow. Conventional analysis of groundwater flow is generally made by using the relevant physical principles of Darcy's law and mass balance. To predict the groundwater movement in the geological formation more accurately, many analytical solutions for partial differential equations exist but because of the difficulty in obtaining analytical solutions, numerical solutions are more generally used. Models are applied to a range of environmental problems mainly for understanding and the interpretation of issues having complex interaction of many variables in the system. They are not exact descriptions of physical systems or processes but are mathematically representing a simplified version of a system. This mathematical calculation is referred to as simulations. Groundwater models are used in calculating rates and direction of groundwater flow in an aquifer. By mathematically representing a simplified version of a hydrogeological condition, realistic alternative settings can be predicted, tested, and compared.

Groundwater models are used to predict/illustrate the groundwater flow and transport processes using mathematical equations based on certain simplifying assumptions. Because of the simplifying assumptions embedded in the mathematical equations and the many uncertainties in the values of data required by the model, a model must be viewed as an estimate and not an accurate replication of field settings. Numerical solutions are often more difficult to verify, so mathematical model error has to be kept as small as possible. Groundwater models, though they represent or approximate a real system, are useful investigation tools in many applications.

Groundwater models provide a relevant and useful scientific and predictive tool for predicting impacts and developing management plans. The development and evaluation of resource management strategies for sustainable water allocation, and for control of land and water resource degradation, are heavily dependent on groundwater model predictions. Regional scale groundwater flow modelling studies are commonly used for water resource evaluation and to help quantify sustainable yields and allocations to end-users.

1.1 Statement of the problem

The connection between basins is a relatively unexplored and potentially significant factor in further understanding of basin hydrology as well as the problem that can transfer from one to the other (basin to basin). Most hydrological studies in Ethiopia are based on the assumption that the groundwater divide and surface water divide are coincident. Such assumption leads wrong quantification and wrong prediction of different scenarios in the basins, regions and nationwide.

The most obvious effect of IGWT is to diminish surface water discharge from watersheds that lie in the recharge area of the basin (in which IGWT originates), and enhance discharge in basin where the regional aquifer discharges (receiving IGWT). This problem is happening in some areas and need to see the problem in different direction, of course incorporating the concept of IGWT into the study must be one option. As the country get growing the population and exploiting of water resource increase and need wise and accurate management. These issues again need the concept of IGWT. Thus in order to quantify, manage and plan groundwater resource efficiently the IGWT must be studied and incorporated in basin hydrology specially in developed basins like Awash River and Rift Valley lakes basins. Moreover understanding groundwater movement within and between basins is needed to assess the potential for internal and cross contamination.

1.2 Objective of the study

The main objective of this research is to identify the possibility of groundwater transfer between basins, Awash River and Rift Valley Lakes, using three dimensional finite element models.

In relation to this main objective there is a desire to include the following specific objectives:

- Investigating boundary conditions, develop a numerical groundwater flow models and determining the head potential of the model basins
- Identifying hydrogeological relation of the two basins near the surface water divide
- Sketching of the field groundwater flow direction

1.3 Structure of the thesis

This thesis is organized in six chapters. Chapter one deals with the general introduction, statement of the problem and objective of the study. Chapter two gives the literature review on interbasin groundwater flow and the physical characteristics of the UARB and URVLB that includes the description of surface topography, hydro metrology, geology and hydrogeology, and groundwater. Chapter three gives an overview on methodology of the thesis, it includes location of the study area, data collection, processing and analysis, groundwater modeling and model conceptualization. In Chapter four is the result and discussion part of the work namely well inventory, rainfall distribution, Aquifer geometry, hydraulic head and interbasin flow, and model calibration and results. Chapter five is about conclusion and recommendation. And finally in Chapter six list of the References are included.

2 LITERATURE REVIEW

In this part of the work literatures related to interbasin flow and the area under study (documents which can help to know geology, hydrology, hydrogeology and others) are reviewed. There are many studies in UARB and URVLB conducted in governmental level and personal researchers. Even if, there are some studies on hydrogeology of these basins, most of the studies are on surface water and limited studies on groundwater. Almost all of these studies both surface and groundwater resources are based on the generally accepted concept that most hydrologic systems are coextensive. The two basins are the most utilized basins due to urbanization and development of industries in this area. The UARB is characterized by one of Ethiopia's best aquifers providing water supply for different cities including the capital city, Addis Ababa (Yitbarek, 2009). But the aquifer in this area is very complex.

2.1 Interbasin flow

Interbasin flow is a process by which groundwater moves from one topographic basin to another through an intervening structural or topographic barriers (Mayo, Testing the inter basin flow hypothesis at Death Valley, 2004). Most hydrogeological studies are based on the assumption that the groundwater divide and surface water divide are coincident. But this is not always true. Interbasin flow may occur wherever potential and integrated permeability permits (Mayo, 2014).

Understanding the hydrological relation between basins has great importance in the study of watershed hydrology, groundwater management, and groundwater resource and pollution assessment. In Ethiopia neither government agencies nor the privet sector have fully addressed the complexity of groundwater use and management. Scientifically based studies, implementation and monitoring systems are required for the efficient and effective utilization and management of the resource (Halcro, 2008). Besides the difficulty of IGWT analysis many scholars believe in incorporating the concept of IGWT into groundwater management. The aggregate effect of groundwater seepage into or out of the basin as a whole may need to take into consideration when, for example, management policies are to be projected for several decades (Waltz, 1972).

Almost all groundwater studies in Awash River and Rift Valley Lakes basins did not consider the concept of IGWT in their model. Even Ethiopian Master Plan of water resource did not include the idea of IGWT in the assessment of groundwater potential of major basins of Ethiopia. The study was based on the assumption that the groundwater divide and surface water divide are coincident (WAPCOS, 1990). But simulation of groundwater movement may vary significantly depending on whether a model domain boundary is defined to allow flow across it, and such boundaries are defined by how the system is conceived by a modeler. In essence, a conceptual model largely predetermines the outcome of accompanying quantitative studies of groundwater movement. Interbasin flow is one conceptualization that can affect subsequent evaluation of a flow system in terms of aquifer dimensions, model boundaries, groundwater residence times, and aquifer sustainability relative to withdrawals (Mayo, 2014).

Even though the study of basins in Ethiopia by including the concept of IGWT is not well addressed this concept must be added in the study especially in highly populated basins. So among the major basins of Ethiopian Awash and Rift Valley basins are relatively the most populated and exploited. Some studies were done on the development of basins (Ministry of Water Resources, 2008). Recently some researchers are attracted towards this concept, and did some researches on the UARB interaction with the neighbor basins. Interbasin flow between Awash River and Blue Line River basins by Andarge Baye as PhD Thesis at University of Poitiers in 2009, Study of groundwater flow system and hydrochemistry between lakes region and Awash basin (Ziway-Koka Corridor) as MSc Thesis at Addis Ababa University in 2006 are some of the researches which can relate directly to this paper. Both the above mentioned researches are mainly based on water chemistry and lithology study.

Geochemical methods are commonly applied to study groundwater flow regimes (Anderson Mary p., 1992). But in study of groundwater flow between two basins based on geochemical methods there must be a consideration of flow from other neighbor basins (for Awash Koka case from Abay basin). According to Andarge Baye research conclusion Abay and Awash basin are connected. Based on the litho-hydrostratigraphic relationship correlated from the drilling data of the exploratory boreholes coupled with the respective geological structure scenarios, groundwater movement could be connected to each other through the permeable and porous scoraceous lower basaltic aquifer all the way from the Blue Nile Plateau to the study area. From the constructed relationship, the scoraceous Tarmaber and/or Amba Aiba formation together with the tectonic structures is therefore responsible in conveying the recharge from the adjacent Blue Nile plateau to the Upper Awash groundwater system (Yitbarek, 2009). In overcoming this problem groundwater models play an important role by taking available boundaries into consideration in the model, in the development and management of groundwater resources and in predicting effects of management measures. With rapid increases in computation power and the wide availability of computers and model software, groundwater modeling has become a standard tool for professional hydro-geologists to effectively perform most tasks.

The groundwater system is a very fragile one, because contaminants introduced in the system remain there for extremely long periods and move slowly towards outlets like wells or springs. A contaminant may well spend many years in a groundwater system and its movement may some cases be very difficult to detect or predict. Identifying the head in the study area can give good prediction of the possibility of IGWT (Belcher W.R., 2010).

The hydraulic gradients between spring and selected wells, and more generally the regional topographic gradient indicate the general direction of potential lateral groundwater movement in the regional system. Actual movement is dependent on the hydraulic conductivity of the rocks. A thorough understanding of the concept of groundwater heads is essential to identify and quantify the flow processes within an aquifer system (Rushton, 2003).

2.2 Surface topography

The Ethiopian Rift is part of the East African Rift system, which extends from the Kenyan border in the south up to the Red Sea to the north. Both Awash River basin (most part) and Rift Valley lakes basin lay along the Ethiopian Rift. The Awash River originates from Central West part of Ethiopia about 150 kms from the capital city, flowing 1200 Km long, and provides a number of benefits to Ethiopia. It is, relatively, the most utilized river basin and covers parts of the Amhara, Oromia, Afar, Somalia regional states, and Dire Dawa, and Addis Ababa City administrative states of the country. The river basin has a lowest elevation of 210 m and a highest elevation of 4195 m. Awash River basin has a catchment area of 112,696 km^2. The Awash is considered as the principal river of the Ethiopian section of the Great Rift Valley (Seleshi, 2007).

The Awash River basin is bordered on its western side by the Abay river (Blue Nile) basin, to the south west by the Omo-Gibe and Rift valley lakes basins and to the south east by the Wabi-Shebele river basin. The basin lies between longitude 7°52'12″N and 12°08'24″N and latitude 37°56'24″E and 43°17'2″E. Large part of the basin is characterized by high mountains and plateaus with elevations often above 2,000 meters above sea level (m.a.s.l) to the east and west (Ayenew, The distribution and Hydrogeological control of fluoride in the groundwater of central ethiopian rift and adjacent highlands, 2007). Upper Awash extends from its source to Koka Lake (the upper reaches of the Awash basin approximately 75 km southeast of Addis Ababa).

The Rift Valley basin has an area of 52,739 Km^2, covering parts of the Oromia, SNNPR regions. The basin is endowed with a number of lakes of varying size with high environmental significance (Seleshi, 2007). Within the Rift system extensive lakes were formed during the pluvial period with concomitant deposition of lacustrine sands, clay and diatomite (Ministry of Water Resources, 2008). The Rift Valley lakes basin is bordered on its western side by Omo Gibe River basin, on east side by Genale Dawa River basin, on the north side by Awash River basin.

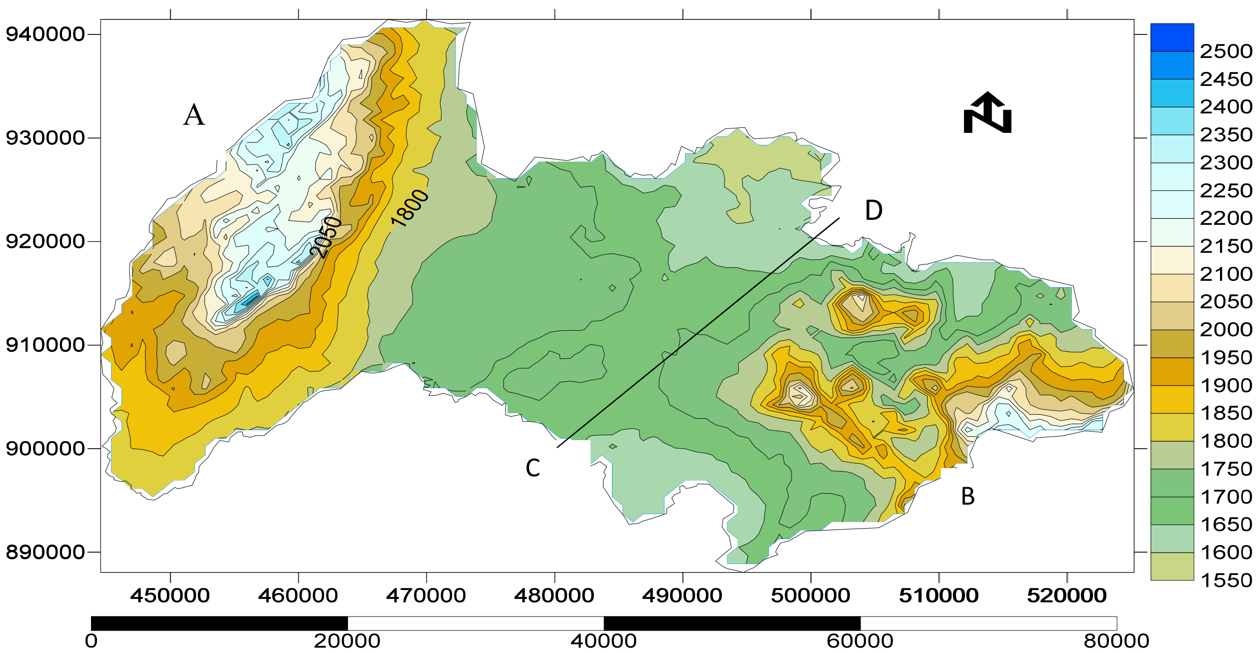

Figure 2-1Surface topography of Upper Awash River basin (UARB) and Upper Rift Valley Lake basin (URVLB)

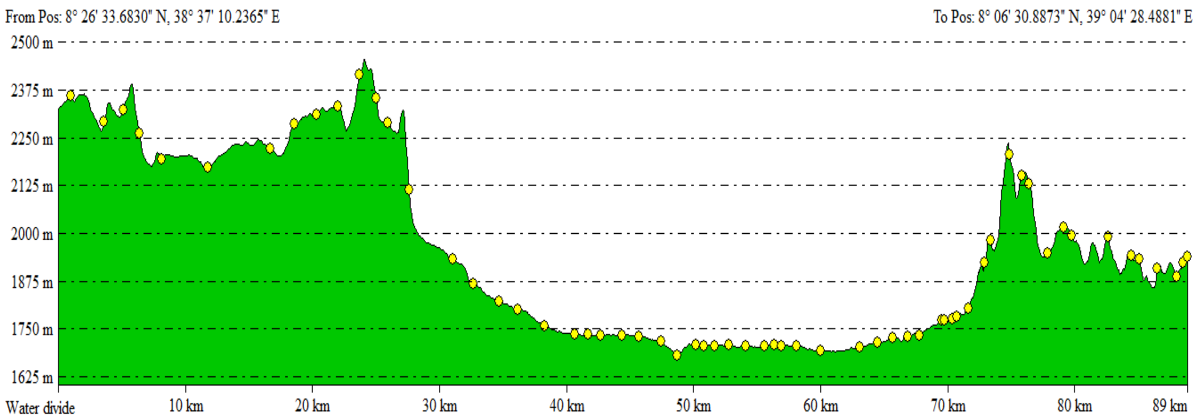

Figure 2-2 longitudinal profile of the study area along the surface water divide (Awash - Rift basin boundary from A to B in Figure 2.1)

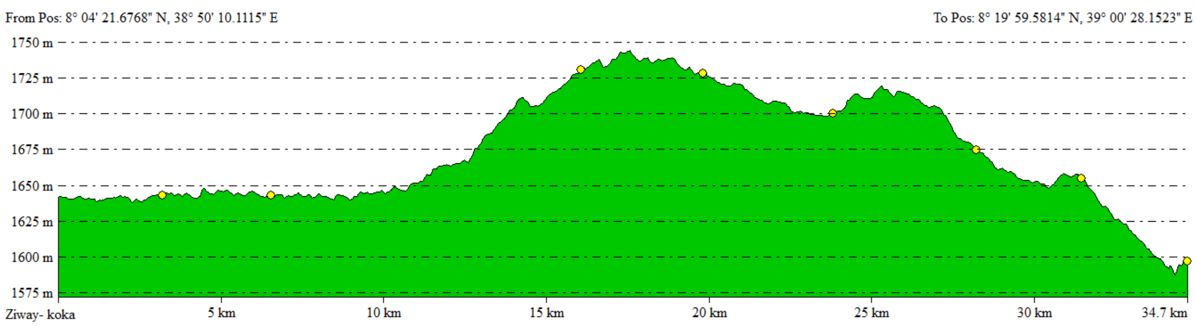

Figure 2-3 longitudinal profile of the study area from Ziway Lake to Koka Lake (from 'C' to 'D' in Figure 2.1)

2.3 Hydrometeorology

In the Rift region there exists a large topographic difference in the basin; as a result the climate is also variable. It is humid at the highlands and arid to semi-arid in the escarpment and Rift valley. The climatic characteristics of the area are determined largely by the annual movements of air currents across the country, Atlantic equatorial westerly, and the southerly and easterly Indian Ocean currents being the moisture sources for precipitations.

During October to February when northeasterly winds persist, long periods of dry weather are experienced. Between February and the ends of April the weather becomes more unsettled and a convergence of moist southeasterly air stream causes small rains, commonly referred to the "Belg Rains". The main rainfall is between June to September when moist winds from the Atlantic and

Indian oceans converge over the Ethiopian Highlands. Thus the year is characterized by a major rainy season from June to September, during which about 70% of the annual rain falls in the study area, followed by a relatively dry season until the end of January and the small rainy season from February to the end of April.

The climate is sub-humid in the central Main Ethiopian Rift (MER), semi-arid close to the Kenyan border and arid in the Afar. On the high plateau to the west of Addis Ababa the rainfall distribution shows a continuous increase from the spring rains to the summer peak rainfall. The distribution of rainfall over the highland areas is modified by orographic effects and is significantly correlated with altitude. In this study area there are a number of metrological stations, but ten of them are used for average rainfall estimation in the model basin.

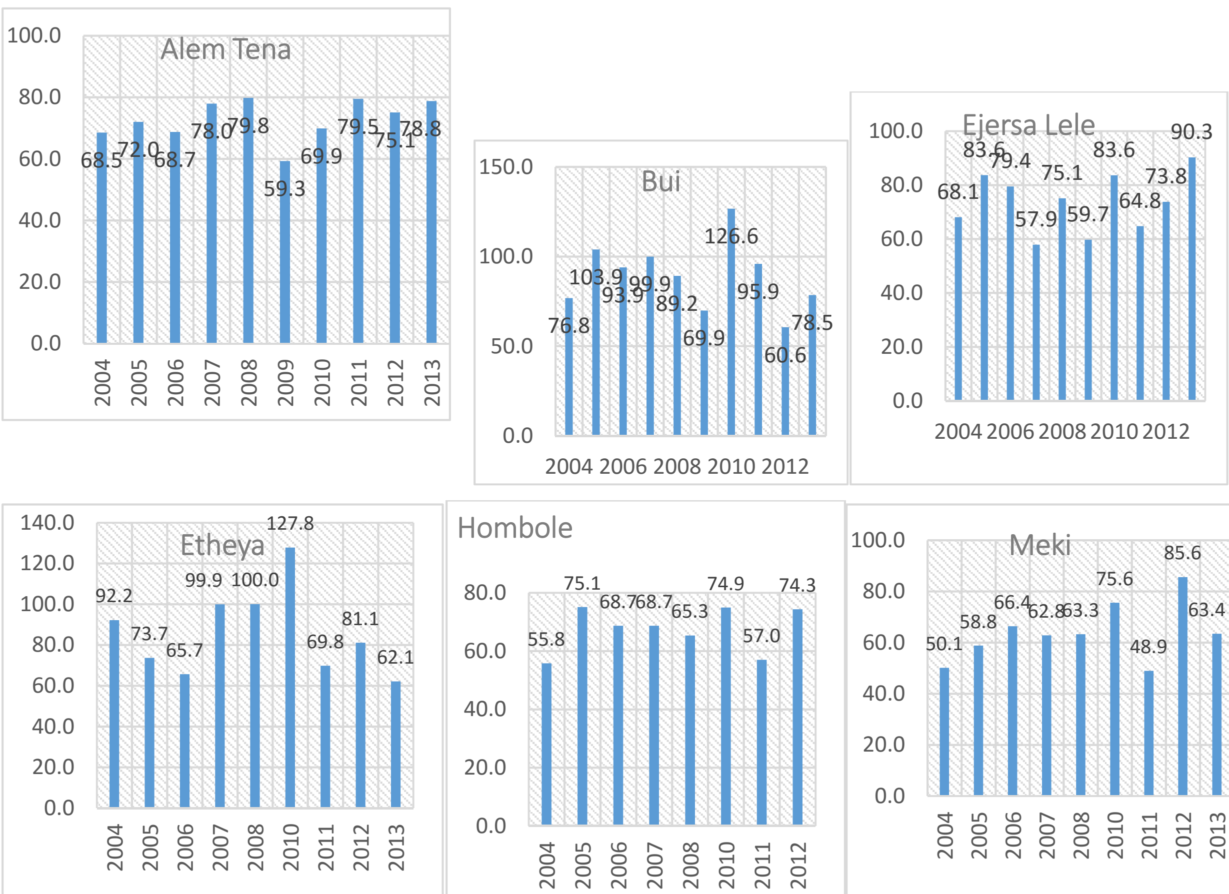

Figure 2-4 yearly average rainfall distribution of the study area for some selected stations (from NMSA)

To see the monthly variation and distribution of the study area rainfall for the year, 2013 of the above stations is shown below. The graph is in the order of Alem Tena, Bui, Ejersa Lele, Etheya, Hombole and Meki from left to right.

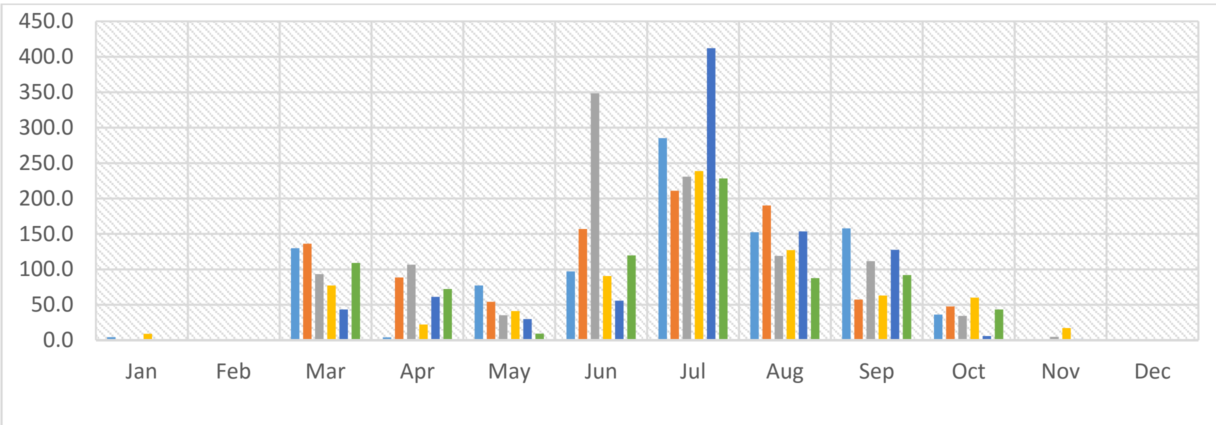

Figure 2-5 Monthly rainfall distribution of the study area for some selected stations (from NMSA)

2.4 Geology and hydrogeology

The Ethiopian Rift system is widely believed to have been formed by diverging lithospheric plates, which is active since the early tertiary times. The occurrence of groundwater is mainly influenced by the geology, geomorphology, tectonics and climate of the country. The variability of these factors in Ethiopia strongly influences the quantity and quality of the groundwater in different parts of the country. The geology of the country provides usable groundwater and provides good transmission of rainfall to recharge aquifers, which produce springs and feed perennial rivers. The difficulty of obtaining productive aquifers is peculiar feature of Ethiopia, which is characterized by wide heterogeneity of geology, topography, and environmental condition (Alemayehu, 2006). The formation of the Ethiopian Rift during the Miocene separated the eastern and western highlands. The Upper Awash basin is exclusively confined within the north-central plateau and the adjacent escarpment and Rift.

Meki river streambed deposits are typically composed of lacustrine sediments mixed with pyroclastic deposits such as tuff and ignimbrite, alluvial, debris flow or talus deposits and fan deposits. The lacustrine deposits vary from clayey silt to fine sand deposits and whereas the fluvial deposits vary from silt to cobble size and sometimes up to boulder size, thus it is well connected hydraulically to the underlying lacustrine sand aquifer.

2.4.1 Geological setting

The rock provide the framework in which groundwater occurs and moves. Groundwater may occur in interstitial openings or in fractures of the rocks. The opening may have been formed at the time the rocks were deposited or at a subsequent time by fracturing, weathering, or solution. The distribution and nature of these openings may relate generally to other physical and chemical characteristics of formations or group of rocks. Thus the general nature and distribution of the rocks in the region permit some inferences regarding the occurrence and movement of groundwater.

A number of geological investigations were made in this study area at different time. For this study geological setting of the area is referred from previous studies for both UARB and URVLB. The geological and geomorphological features of the region are the result of Cenozoic volcano-tectonic and sedimentation processes. Except some patchy Precambrian outcrops to the south and northern edge, the Rift is covered with Cenozoic volcanic and sediments. The Rift formation is associated with extensive volcanism.

The Rift valley basin is covered with Tertiary and Quaternary volcanic, sediments and lacustrine deposits. The highlands are dominantly covered with basic volcanic rocks (lavas and ashes, mainly of Tertiary age). In the Rift floor, volcanic rocks are dominantly acidic (silica-rich), including ash and pumice. The youngest volcanics are associated with obsidian flows, ignimbrite, pumice, rhyolitic flows and domes, pyroclastic surge deposits, scattered basaltic lava flows and spatter cones (Ayenew, The distribution and Hydrogeological control of fluoride in the groundwater of central ethiopian rift and adjacent highlands, 2007).

Thin strips of alluvium along rivers occur both in the highlands and the Rift. Extensive lacustrine deposit covers the Rift floor. The eluvial lateritic crust consists of clay, silt and fine sand; but it is dominantly silt and fine sand. Small alluvial fans associated with colluvial deposits occur at the foot of mountainous areas and fault scarps (Ayenew, The distribution and Hydrogeological control of fluoride in the groundwater of central ethiopian rift and adjacent highlands, 2007).

The various volcanic sequences and sediments are affected by different generations of tectonism. There are at least four sets of faults trending NNE–SSW, N–S, and NNW–SSE, and curved volcanic vent structures. The first two are dominant and extend for long distances following the axis of the WFB. Present day tectonic activity is dominantly confined along the WFB. The faults in the Rift and escarpment have increased the permeability of the rocks significantly. The complex spatial and temporal distribution of the volcanic rocks, their different reciprocal stratigraphic relationships with the Quaternary sediments, their wide compositional, structural and textural variability and degree of weathering and variable topographic position diversified the hydrogeological behavior of the region. (Ayenew, 2006).

2.4.2 Hydrogeology

Since the geology of Awash River basin is complex it is not easy to predict groundwater in the basin. The complexity of geology of the basin has a direct impact on its hydrogeological characteristics (Karimi, 2014). The hydrogeological complexity is also true in the Rift valley lakes basin, such that the groundwater contribution to the overall water resource balance varies considerably throughout the basin. The nature of the geological formations within RVLB and the intense tectonic disturbance that has affected them form a significant influence over the distribution and disposition of groundwater resources within the basin. There are complex relationships between groundwater recharge, flow, storage, discharge, and the surface water system. The frequent occurrence of groundwater as discreet bodies, which may not be readily identified, makes evaluation of the available groundwater resource extremely difficult (HALCROW and MOWR, 2009).

In the Rift, the localization of groundwater is strongly controlled by the axial faults. They have contrasting role in the movement and occurrence of groundwater. The majority of faults are conduits to groundwater flow. In places these open faults allow significant amount of preferential groundwater flow parallel and sub-parallel to the Rift axis. In the Rift valley, the direction of groundwater is strongly governed by the orientation of faults, which is often perpendicular to the regional groundwater contours in the highlands and escarpments. It was found that these axial faults govern strongly the subsurface hydraulic connection of the Rift lakes and the river-groundwater relations. Some rivers disappear in these open Rift faults as in the case of rivers coming from the Arsi highlands feeding lakes Langano and Shala (Ayenew, 2007).

Groundwater flows are of key importance in the Awash basin both as a major water supply for people and its impact on surface water flows. The highland's fractured volcanic cover is favorable for groundwater recharge processes. Thus, groundwater recharge from the highlands is substantial and it flows in a relatively shallow depth. Groundwater gradually percolates into the lower aquifers through large marginal faults before it reaches the Rift floor (Ayenew, 2001). In the Upper and Middle Valley the groundwater levels range between 30 and 70 m. The levels drop to lower than 200 m in some areas in the southern corner of the Awash valley. In the upper basin, upstream of the Koka dam, the Awash River is hydraulically linked to the aquifers (Karimi, 2014).

The aquifers in Upper Awash can be divided broadly into two categories; primary porosity aquifers and double porosity aquifers. The first category comprises aquifers related to Quaternary alluvial and lacustrine deposits and the second broad categories belongs to the basaltic volcanics and again subdivided in to upper and lower basaltic aquifers separated by less permeable, along fractured and weathered zones and/or impermeable otherwise, of acidic volcanics (Karimi, 2014).

2.4.3 Hydrology

The Ethiopian Rift is characterized by many perennial rivers, and a chain of lakes that vary in size, and hydrological and hydrogeological settings (Ayenew, 2007). The main source of water to the Rift lakes and rivers is the rainfall in the eastern and western highlands. The most important rivers are Meki, Bilate and Awash feeding lakes Ziway, Abaya and Abhe respectively. In addition, many seasonal streams feed the lakes. The primary lakes and rivers within the study region are Lake Koka, Lake Ziway, Meki River and Awash River. In the study area groundwater level is affected by fluctuation of perennial river stages. Water flows from the river into the aquifer and the groundwater becomes elevated when there is an increase in river stage with respect to the altitude of the groundwater level. A decrease in river stage with respect to the groundwater level causes water to flow from the aquifer into the river and result in the decrease of groundwater level. The extent of the change in the groundwater level elevation in response to river stage fluctuations depends on the magnitude of the change in the river stage, the length of time the river remains at the current river stage, the hydraulic properties of the aquifer material, and the distance from the river to the point of interest. In general, lake/river recharges an aquifer when the groundwater surface is below the bottom of the lake, but drain to the aquifer when the groundwater surface is above the bottom of the lake.

Due to the time required for the change to propagate into and through the aquifer the groundwater level increase or decrease in response to increase or decrease in river stage is more obvious in areas closer to the river. Therefore, the area of the aquifer that is affected by an increase or decrease in river stage depends on the length of time that the river stage remains at the new altitude. Changes in the groundwater level at distant locations from the river are the result of long term river-stage changes typically caused by seasonal high and low flows or long-term river-stage management.

The Awash River rises on the high plateau near Ginchi town west of Addis Ababa in Ethiopia and flows along the Rift valley into the Afar triangle, and terminates in salty Lake Abbe on the border with Djibouti. The Awash River basin has been divided in to three main sub-basins: Upper (upstream of Koka Dam station), Middle and Lower. The Upper Awash basin is located in central Ethiopia at the western margin of the MER.

2.5 Groundwater

Ethiopia has enormous surface water and groundwater resources, although the distribution is uneven in regional or national level. Very little has been done in this field and development of the water resources, particularly in areas of groundwater resources. Groundwater utilization has been limited to community water supply using shallow hand dug wells and unprotected springs. Limited deep boreholes were drilled in few rural areas, mainly in the Rift valley, in some peripheral semi-arid regions and in the vast highland volcanic terrain (Alemayehu, 2006). As an example we can take Meki town, there are a number of water wells but almost all of them are hand dug and shallow. And also in the rural area the farmers use hand dug wells for small scale irrigation purpose.

In the Rift valley at regional scale the flow of groundwater from the highlands in west to the Rift center is facilitated by the suitable geohy-drologic condition of the area. When viewed on a smaller scale, groundwater flow paths are affected by local structural setting. The existence of transverse faults in the Awash area, enables the Awash River to drain the western escarpment, flowing into the Rift and continuing its course on the Rift floor (Kebede, 2013).

Groundwater is recharged by direct rainfall to permeable ground, (generally greatest in the basin margins where rainfall is relatively high), via river systems and lakes and from overlying or adjacent groundwater bodies. Groundwater discharge is from springs, either into surface waters (supporting base flow), directly into lakes, to the surface and into adjacent (or overlying) groundwater bodies (HALCROW and MOWR, 2009).

2.5.1 Groundwater recharge

Recharge is defined as the process of downward movement of water through the saturated zone under the force of gravity or in the direction determined by hydraulic conditions. Groundwater recharge is a fundamental component in the water balance of any watershed, however, because it is impossible to measure it directly, numerous methods, widely ranging in complexity and cost, can be used for estimation of recharge. It 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 groundwater resources.

Groundwater in unconfined aquifers moves from topographically high areas (recharge) to topographically low areas (discharge). In a recharge area, the potential energy decreases with depth, which results in down ward movement of water. Between the recharge and discharge areas, groundwater flow is primarily horizontal. In a discharge area, the potential energy increases with depth, which results in upward movement of groundwater. Although commonly displayed on a two-dimensional surface, hydraulic-head distribution is generally a three-dimensional phenomenon i.e., hydraulic head varies vertically as well as laterally. If hydraulic head increases with increasing depth, groundwater flow is upward and, in general, indicates an area of discharge. If hydraulic head decreases at a given location with increasing depth within the aquifer, then groundwater flow is downward, indicating an area of groundwater recharge (Moore, 2002).

Precipitation, surface waters and irrigation losses can be sources of groundwater recharge. Recharge mechanism from these sources can be diffuse (direct) or preferential (localized/indirect). Diffuse recharge mechanism refers to the water added to the groundwater reservoir in excess of soil-moisture deficit and evapotranspiration. The preferential recharge also called localized recharge mechanism 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 sources (Lerner, 1990). Preferential recharge takes place via pathways such as macrospores opposed to diffuse recharge which takes place through the entire vadose porous medium

In humid areas with porous soils, 25% of annual rainfall may recharge the aquifer. In contrast, in desert regions recharge is very small to 1% of rainfall or less. Aquifers in these areas may contain very old water, which has accumulated over centuries or under different climatic conditions (Moore, 2002).

Most groundwater originates as recharge in upland areas, water that infiltrates from precipitation on the ground surface. Some water enters the subsurface by seeping out of the bottom of surface waters, a situation more common in arid climates than in humid climates. Groundwater discharges from the saturated zone back to the ground surface in low-lying areas, usually at springs or the bottom of surface waters. Since groundwater always moves towards lower head, these exit points are always at a lower elevation than the water table where groundwater enters the system as recharge. Recharge is highest in areas with wet climates and permeable soil or rock types. In permeable materials, the rate of recharge can be as much as half the precipitation rate, with little overland flow. On the other hand in low permeability materials only a small fraction of the precipitation becomes recharge. With massive clay soils, the recharge rate can be less than 1% of the precipitation rate (Fitts, 2002).

The highlands of Gurage Mountain gets large amount of recharge. This area receives high rainfall and discharges its surface and subsurface flow to the low-lying areas. Generally the western Rift escarpment of the catchment gets the highest recharge due to its high amounts of rainfall, characteristic mountain chains and escarpment slopes which have vertical to steep slopes and strongly dissected. This facilitates infiltration instead of runoff, and its fracture and joints directing stream channels feeding the aquifer in the foot of the escarpment.

2.5.2 Groundwater flow model formulation

Many techniques are presently available for describing groundwater flow in aquifers. Of these numerical models are best suited for analysis of basin wide groundwater flow because of flexibility and speed of analysis (Waltz, 1972). To model regional groundwater flow means to develop mathematical and numerical models of the aquifer system being studied and to use these models to predict the value of hydraulic head at points (and times) of interest (Istok, 1989). Because of the complex interdisciplinary interests in groundwater, models differ markedly in purpose, information requirements, assumptions, usefulness, and the mathematical schemes incorporated (one-, two-, and three dimensional models; steady-state, saturated flow models; transient, unsaturated flow models; fracture flow models) to study groundwater flow systems. Real groundwater problems are three-dimensional in space and time-variant (Rushton, 2003). The selection of the type of model to apply to a particular field problem mainly governed by particularly, the availability of field data of study area, accuracy required in the result and the boundary conditions

Groundwater modeling involves formulating a correct conceptual model through selection of those parameters and their values, which help to describe spatial and temporal variability within the groundwater flow system more accurately. The attraction of groundwater modeling is that it combines the subtlety of human judgment with the power of a digital computer. Most of the groundwater models are aimed at predicting the consequences of a proposed action; they can also be used as interpreters to assimilate the controlling parameters or as framework for systematizing and collecting data and formulating ideas regarding system dynamics; as in this study (predict interbasin groundwater transfer between basins as inferred from head potential). Here we simulate groundwater flow indirectly by means of governing equations or mathematical models that describe heads.

Numerical modeling involves approximation of the equations to describe systems that have variable properties and irregular geometry. Numerical procedures used to describe groundwater problems create system of equations that must be solved simultaneously to produce accurate results with greater efficiency. Either finite difference method (FDM) or finite element method (FEM) has been involved for development of most of the groundwater models; nevertheless, FEM is more favored than FDM since it can treat the complex boundaries and irregular geometry more effectively than the FDM. It is also a common practice to solve the spatial portion of the flow equation through FEM, and the time marching portion through the FDM (Istok, 1989). The domain discretization of FDM models is limited to an orthogonal grid of rectangular blocks, which is less flexible than the triangular or quadrilateral elements of FEM models. Because of this difference, the approximation of boundary conditions can be more accurate in an FEM model than in an FDM model (Fitts, 2002). In addition to that, the method has several advantages: (a) irregular or curved aquifer boundaries, anisotropic and heterogeneous aquifer properties, and sloping soil and rock layers can be easily incorporated into the numerical model, (b) the accuracy of solutions to groundwater flow is very good (exact in some cases), (c) the finite element method lends itself to modular computer programming wherein a wide variety of types of problems can be solved using a small set of identical computer procedures.

A model, in a general sense, is a representation which attempts to explain the behavior of some aspect of the prototype system. It is always less complex than the real system it represents. According to Wang and Anderson groundwater flow model formulation can be summarized as follows.

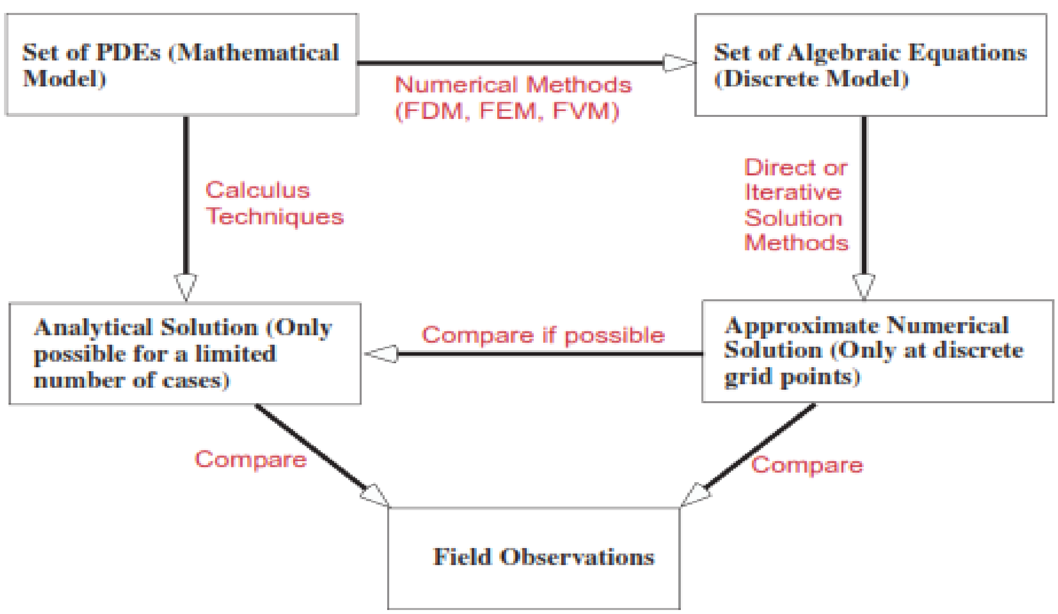

Figure 2-6 General flow chart of numerical groundwater modeling process after (Anderson, 1982)

Most groundwater models in use today are deterministic mathematical models. Deterministic models are based on conservation of mass, momentum, and energy and describe cause and effect relations. The underlying assumption is that given a high degree of understanding of the processes by which stresses on a system produce subsequent responses in that system. Deterministic groundwater models generally require the solution of partial differential equations. Exact solutions can often be obtained analytically, but analytical models require that the parameters and boundaries be highly idealized.

Some deterministic models treat the properties of porous media as lumped parameters (essentially, as a black box), but this precludes the representation of heterogeneous hydraulic properties in the model. Heterogeneity, or variability in aquifer properties, is characteristic of all geologic systems and is now recognized as playing a key role in influencing groundwater flow. Thus, it is often preferable to apply distributed-parameter models, which allow the representation of more realistic distributions of system properties. Numerical methods yield approximate solutions to the governing equation (or equations) through the discretization of space and time. Within the discretized problem domain, the variable internal properties, boundaries, and stresses of the system are approximated. Deterministic, distributed-parameter, numerical models can relax the rigid idealized conditions of analytical models or lumped-parameter models, and they can therefore be more realistic and flexible for simulating field conditions (if applied properly).

The number and types of equations to be solved are determined by the concepts of the dominant governing processes. The coefficients of the equations are the parameters that are measures of the properties, boundaries, and stresses of the system; the dependent variables of the equations are the measures of the state of the system and are mathematically determined by the solution of the equations. When a numerical algorithm is implemented in a computer code to solve one or more partial differential equations, the resulting computer code can be considered a generic model. When the grid dimensions, boundary conditions, and other parameters (such as hydraulic conductivity), are specified in an application of a generic model to represent a particular geographic area, the resulting computer program is a site-specific model.

2.5.3 Groundwater flow equation

In the finite element method deriving an integral formulation for the governing groundwater flow which leads to a system of algebraic equations that can be solved for values of the field variable (hydraulic head, pressure head) at each node in the mesh is the first step. Several methods can be used to derive the integral formulation for a particular differential equation. The method of weighted residuals is a more general approach that is widely used in groundwater flow modeling. In the method of weighted residuals, an approximate solution to the boundary or initial value problem is defined. When this approximate solution is substituted into the governing differential equation, an error or residual occurs at each point in the problem domain. We then force the weighted average of the residuals for each node in the finite element mesh to equal to zero. Galerkin's Method is the subset of the method of weighted residuals that is most commonly used to solve groundwater flow and solute transport problems (Istok, 1989). Before the approximating equation the general three dimensional groundwater flow equation for steady state is summarized from Istok as follows.

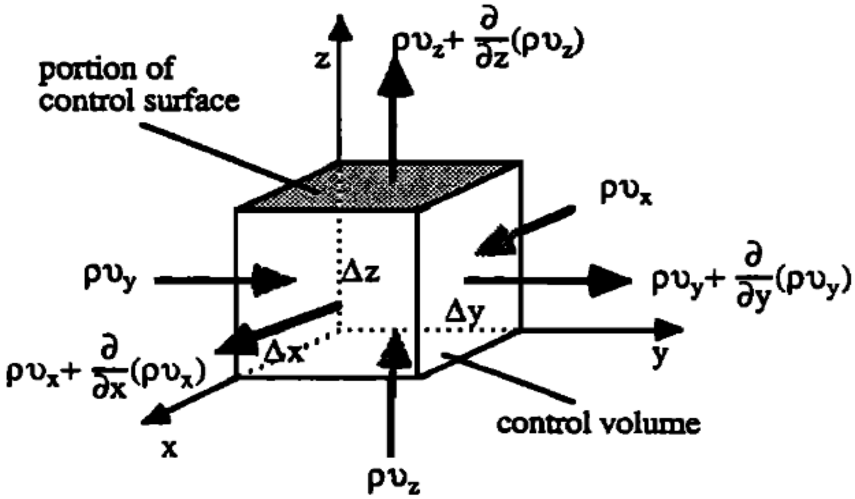

Figure 2-7 Representative control volume (REV)

The law of conservation of mass for steady-state flow requires that the rate at which fluid is entering the control volume is equal to the rate at which fluid is leaving the control volume (net rate of inflow = inflow - outflow = 0) for purposes of analysis, consider the rate at which groundwater enters the control volume per unit surface area to consist of three components $\rho\upsilon_x$,

$\rho\upsilon_y$, and $\rho\upsilon_z$, where ρ is the density of water and υ_x, υ_y and υ_z are the apparent velocities of groundwater flow entering the control volume through control surfaces perpendicular to the x, y, and z coordinate axes. Where a differential equation and known boundary conditions are given, an approximate solution may be obtained by applying a numerical method. Then using a Taylor Series approximation, the rate at which groundwater leaves the control volume in the x direction can be written as;

$$\rho\upsilon_x+\frac{\partial}{\partial x}(\rho\upsilon_x)\Delta x+\frac{\partial^2}{2!\partial x^2}(\rho\upsilon_x)\Delta x^2+\frac{\partial^3}{3!\partial x^3}(\rho\upsilon_x)\Delta x^3+\ldots \qquad 2.1$$

If we make the size of the control volume small enough, we can neglect higher-order terms (i.e. those involving Δ^2, Δ^3 etc.) and, because we have chosen a unit control volume ($\Delta x = \Delta y = 1$) the rate at which groundwater leaves the control volume is, $\rho\upsilon_x+\frac{\partial}{\partial x}(\rho\upsilon_x)$, and this is the same as Euler's method. The net rate of inflow in the x direction is then (net rate of inflow in x direction = rate of inflow in x direction - rate of outflow in x direction) i.e., $\rho\upsilon_x- [\rho\upsilon_x+\frac{\partial}{\partial x}(\rho\upsilon_x)] =-\frac{\partial}{\partial x}(\rho\upsilon_x)$ and in the same fashion the net rate of inflow in the y and z directions are $-\frac{\partial}{\partial y}(\rho\upsilon_y)$ and$-\frac{\partial}{\partial z}(\rho\upsilon_z)$, respectively. Because the net rate of inflow for the entire control volume must equal zero if the law of conservation of mass is to be satisfied, we can write as;

$$-\frac{\partial}{\partial x}(\rho\upsilon_x)-\frac{\partial}{\partial y}(\rho\upsilon_y)-\frac{\partial}{\partial z}(\rho\upsilon_z)=0 \qquad 2.2$$

If we assume that groundwater density, ρ is constant (i.e., the fluid is incompressible), we can use the product role of calculus to evaluate a typical term in equation 2.2, $-\frac{\partial}{\partial x}(\rho\upsilon_x)=-[\rho\frac{\partial\upsilon_x}{\partial x}+\upsilon_x\frac{\partial\rho}{\partial x}]$, But $\frac{\partial\rho}{\partial x}$ is zero since the fluid is assumed to be incompressible. Thus $-\frac{\partial}{\partial x}(\rho\upsilon_x) = -\rho\frac{\partial\upsilon_x}{\partial x}$ similarly for the y and z direction$-\rho\frac{\partial\upsilon_y}{\partial y}$, $-\rho\frac{\partial\upsilon_z}{\partial z}$ respectively. Because groundwater density appears outside the derivative it cancels from equation 2.2 and we have;

$$-\frac{\partial\upsilon_x}{\partial x}-\frac{\partial\upsilon_y}{\partial y}-\frac{\partial\upsilon_z}{\partial z}=0 \qquad 2.3$$

Now the apparent groundwater velocities in x, y and z direction are given by Darcy's Law

$$\upsilon_x = -k_x \frac{\partial h}{\partial x}, \upsilon_y = -k_y \frac{\partial h}{\partial y} \text{ and } \upsilon_z = -k_z \frac{\partial h}{\partial z} \qquad 2.4$$

Substitute equation 2.4 into equation 2.3 the three-dimensional form of the equation for steady-state groundwater flow through saturated porous media is written as;

$$\frac{\partial}{\partial x}\left(K_x \frac{\partial h}{\partial x}\right) + \frac{\partial}{\partial y}\left(k_y \frac{\partial h}{\partial y}\right) + \frac{\partial}{\partial z}\left(k_z \frac{\partial h}{\partial z}\right) \pm q = 0 \qquad 2.5$$

Where K_x, K_y, and K_z are the saturated hydraulic conductivity of the porous media in the x, y, and z coordinate directions, h is hydraulic head and q is the recharge. This general saturated groundwater flow equitation is applicable under either confined or unconfined conditions (Engineers, 1999).

2.5.3.1 Hydraulic conductivity

Hydraulic conductivity is a characteristic proportionality constant K (L/T) in Darcy's Law, and as such is defined as the flow volume per unit cross-sectional area of porous medium under the influence of a unit hydraulic gradient. It is empirical constant to be measured in laboratory. Real subsurface materials always have a complex and irregular distribution of hydraulic conductivity. We often describe K distributions using the terms heterogeneity and anisotropy. In a heterogeneous material the value of K varies spatially, and in a homogeneous material K is independent of location. Anisotropy implies that the value of K at a given location depends on direction. Isotropy implies that K is independent of direction at a given location (Fitts, 2002). The hydraulic conductivity of a given medium is a function of the properties of the medium and the properties of the fluid. If hydraulic conductivity is dependent on the direction of groundwater movement the aquifer is anisotropic and the hydraulic conductivity is a second rank tensor.

$$\begin{vmatrix} k_{xx} & k_{xy} & k_{xz} \\ k_{yx} & k_{yy} & k_{yz} \\ k_{zx} & k_{zy} & k_{zz} \end{vmatrix} = K \qquad 2.6$$

The path lines that water travels are often very irregular due to the heterogeneous distribution of hydraulic conductivities in the subsurface. In a high-conductivity layer, water tends to flow parallel to the layer boundaries. But in a low-conductivity layer, water tends to take the shortest path through the layer, flowing nearly perpendicular to the layer boundaries (Fitts, 2002).

2.5.3.2 Assumptions of groundwater flow modeling

Modeling need some assumptions that can be posed by the researcher, Darcy's law is valid which means the groundwater flow in the study area is assumed to laminar. Fortunately, most underground flow occurs with NR <1(laminar, NR is Reynolds number) so Darcy's law is applicable (Todd, 2005). And also from Darcy law we take the assumption that groundwater movement is influenced by hydraulic conductivity and hydraulic head gradient. No significant change in water temperature or dissolved-solids concentration. Groundwater flow is assumed to be saturated and governed by the aquifer flow equation (Equation 2.5) which assumes Darcy's law as valid. Water is the only flowing fluid phase (i.e., the air phase is assumed to be inactive). The fluid is considered to be slightly compressible and homogeneous. The fractured medium maybe represented by a single continuum porous medium of spatially invariant properties. The porosity and saturated hydraulic conductivity are constant with time. Gradients of fluid density, viscosity, and temperature do not affect the velocity distribution.

Flow in fractured rock is difficult to analyze for several reasons. For one, flow occurs along discrete fractures, the distribution and properties of which are mostly unknown. It is generally not possible to map the location and orientation of the important water-bearing fractures in the subsurface, or to know their aperture (width) and roughness. Flow in some larger fractures is turbulent as opposed to laminar, so Darcy's law should not be applied to these (Fitts, 2002).

2.5.3.3 Limitations

It is critical to recognize the limitations of different modeling approaches. Unrealistic expectations and inappropriate applications of models can greatly reduce confidence in the use of numerical models. The equivalent porous media models developed in this study cannot be used to simulate local direction or rate of groundwater flow because major conduits are not explicitly represented in the models and because turbulent flow is not included. These models should be restricted to evaluation of regional groundwater flow issues. The model was discretized using a grid cell size of 300 meters element length. As a result of this discretization, the conditions within the node, such as groundwater level is reduced to one average value for the entire node. Therefore, the model is not suitable for analysis of site-specific problems or issues. Hydrologic parameters and aquifer unit geometry in portions of the model area are not well known at this scale. For instance, aquifer thickness and hydraulic conductivity can change at intervals smaller than the current model resolution.

Extremely cracked and current tectonic areas can have a widely variable hydraulic conductivity. Therefore groundwater level and flow in these areas may be simulated indirectly by increasing the hydraulic conductivity; the effects of these structures on the aquifer system cannot be appropriately addressed with the models. Calibration of the model could be improved by refining further the spatial discretization of some parameters, such as hydraulic conductivity or recharge; however, with scarcity of more field data, finer discretization is not reasonable.

The geological condition of the area is too complex and the model is assumed as a single layer, unconfined aquifer. Therefore, further refinement of the model would be possible with additional data which improve the accuracy of model prediction parameter to apply it for a detail analysis.

In addition to the above mentioned, limitations and uncertainties exist in any modelling study in regard to our hydrogeological understanding, the conceptual model design, and model calibration and prediction, as well as recharge estimation and simulation. The steady-state groundwater flow model of this study catchment provides a regional-scale simulation of groundwater flow in the aquifer system of the study area. The groundwater model results depends on the accuracy of the input data and has corresponding limitations in model precision, because the database management of the hydrogeology of the country is so poor and with limited vital parameter.

These limitations are best addressed by careful scoping of proposed modelling approaches at the outset and reviewing at various stages of the work. From the research objective point of view, the model is adequate for studying groundwater head potential towards understanding interbasin groundwater flow.

3 METHODOLOGY

There are three important steps in the computational modelling of any physical process: (i) problem definition, (ii) mathematical model, and (iii) computer simulation (Sherwin, 2005). The methodology adopted in this study follows literature review, data collection, organization and analysis of data as per the requirement of the model in use.

3.1 Location of the study area

Both UARB and URVLB are located in central Ethiopia at the western margin of the MER. The capital city, Addis Ababa, is located at the northern end of the basins. This study focuses on Upper Awash River basin and Upper Rift valley lakes basin bounded by Lake Koka, Awash River upstream of Lake Koka, Lake Ziway and Meki River. It covers a total area of 2201.6 km^2. But the area is divided into two sub-basins:

1. The area bounded by Lake Ziway, Meki River and the water divide with the Awash River basin called in here as Upper Rift Valley Lakes Basin (URVLB), and it covers 781.21 km^2 drainage basin area.
2. The Awash sub basin with a total drainage area of 1420.39 km^2 bounded by the surface water divide with RVLB basin, Lake Koka and Awash River upstream of Lake Koka, called in here as Upper Awash River Basin (UARB).

Digital Elevation Model (DEM), with resolution of 90mX90m is used. The drainage area of the model basins are delineated with a reference of the natural boundaries using Global Mapper 15.

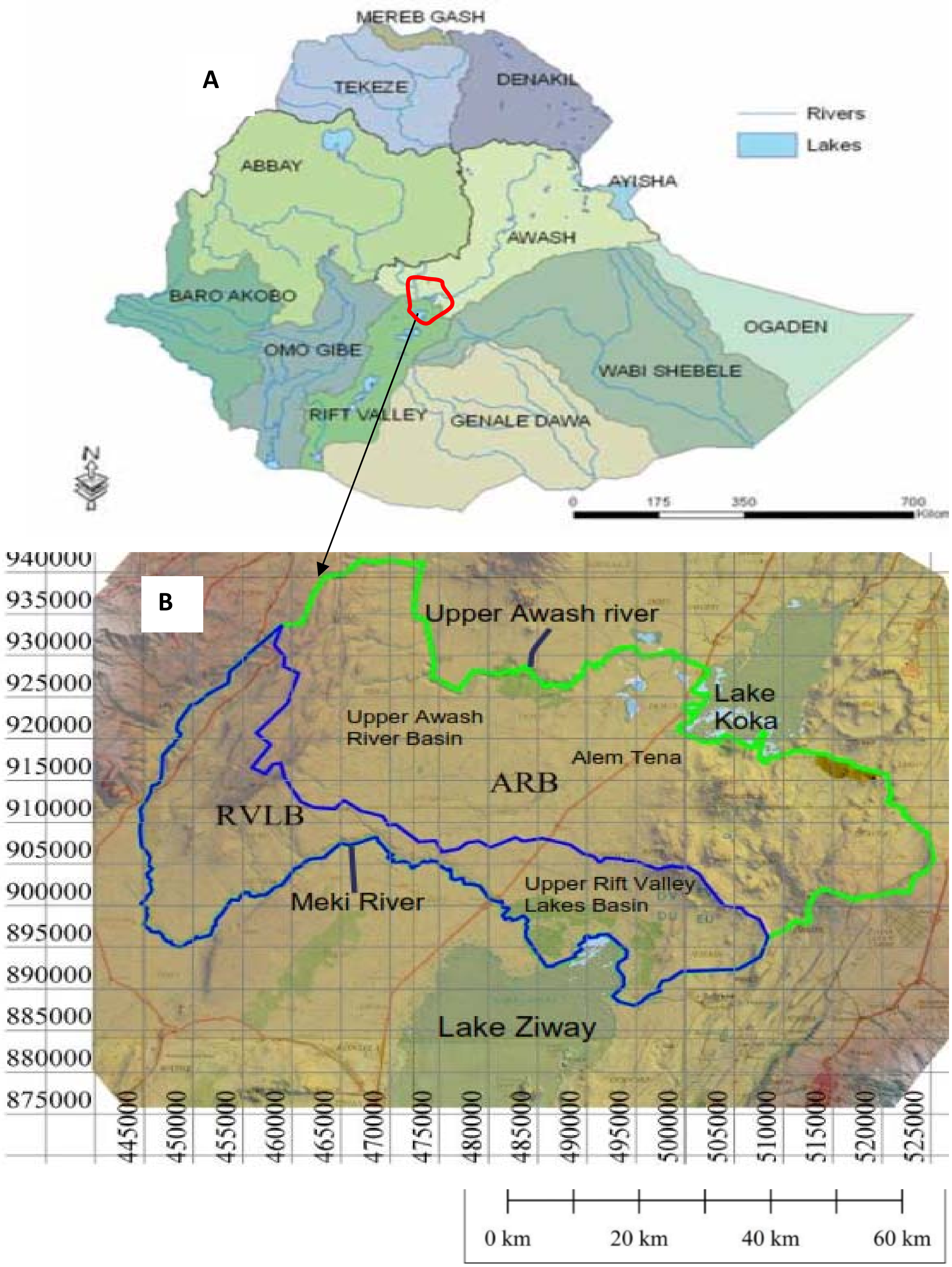

Figure 3-1 the three model basins of the study area ('A' is major basins of Ethiopia adopted from (Seleshi, 2007), 'B' is the study area showing the three model basins and boundaries)

3.2 Data collection

The study is to be based primarily on existing (collected by different organizations) geologic, hydro-geologic, Metrologic and topographic data. At this stage of the research the collection of data from different offices and field observation are done before desk work (selection, organizing, literature review on modeling books, previous work, and water well inventory and well completion reports). Recently the study of geology and hydrogeology in small and large scale become expanding by different governmental offices and researchers. Geological map of Ethiopia, monthly rainfall data, well inventory and surface water divide are collected from different offices viz. Ministry of Water Resources and Energy, Water Works and Design Enterprise, and National Meteorological Agency.

It also includes collection of Digital elevation model (DEM), Meteorological data and modeling soft wares like Surfer 10, Global Mapper 15, Matlab7.11.0 (R2010b). And where existing data are inadequate, supplemental field observation and borehole data collection such as UTM coordinates(x, y, z) using mobile handy GPS and static water level are made. The sampling area are Meki, Alem Tena and Buie.

3.3 Processing and analysis

To identify the possibility of IGWT between two basins the regional view of geologic, topographic, hydrologic condition of the two areas have to be studied well. This part of the work followed the modeling protocol like conceptual model development, defining model geometry and boundary, assigning the hydrogeological parameters, running the model and calibration.

Usually, the model takes the form of a set of mathematical equations, involving one or more partial differential equations. Even though the preferred method of solution of the mathematical model of a given problem is the analytical solution, for most practical problems, because of the heterogeneity of the considered domain, the irregular shape of its boundaries, and the non-analytic form of the various functions, solving the mathematical models analytically is not possible. Instead, we transform the mathematical model into a numerical one, solving it by means of selected computer program.

3.4 Groundwater modeling

Hydrogeological studies usually involve mathematical modeling of groundwater flow. Such models consist of a set of differential equations which govern the flow of groundwater. They have been in use since the late 1800’s, but have found widespread application with the increase in available computing power. In computer programming these models are implemented using different approaches among which the finite difference and finite element methods are most common (Anderson, 1982). In both these methods a system of nodal points is superimposed over the problem domain. The difference between the two methods is in the distribution of nodes. The finite difference nodes are in a regular grid order where nodes can be block-centered or mesh-centered. The finite element methods, on the other hand, can have an irregular distribution of nodes which are connected together to form triangular sub-areas called elements.

3.4.1 Discretization of the problem domain

Unlike analytical methods numerical methods yield approximate solutions to the governing equation through the discretization of space and time. The FEM uses concept of piecewise approximation technique to obtain solutions to a wide variety of problems. The domain of the problem that is the extent of the aquifer (model basin) to be simulated, is divided into a set of elements or pieces. So the first step in the solution of a groundwater flow by the finite element method is to discretize the problem domain. This is done by replacing the problem domain with a collection of nodes (or nodal points) and elements referred to as the finite element mesh. In the finite element analysis the precision of the solution obtained and the level of computational effort required to obtain a solution will be determined to a great extent by the number of nodes in the mesh. A coarse mesh has a smaller number of nodes and will give a lower precision than a fine mesh. However, the larger the number of nodes in the mesh, the greater will be the required computational effort and cost. The size and shape of the elements in a mesh is determined primarily by the size and shape of the problem domain, the number of different types of aquifer materials, and by the number of nodes in the mesh. The most common shapes for finite elements are triangles and trapezoids for two-dimensional flow, and triangular and trapezoidal prisms for three-dimensional flow (Fitts, 2002). In problems that have a complex geometry (e.g., caused by an irregular depth to bedrock) or geologic structure (e.g., due to the presence of faults) many elements will be required. Even if the problem domain contains curved boundaries or interfaces different types of elements we use single type of elements for simplicity.

By taking into consideration the coverage area of the problem domain and the computer in use for the three basins with same element length same element type are generated. The elements are triangular prism and the nodes have right-handed Cartesian coordinates (x, y, z), z-axis points in the vertical upward direction (the elevation of the nodes above sea level). The finite element mesh consists of several nodes for each model basin but each node is assigned a unique node number. Node numbers range from one to the number of nodes in the mesh; no "skips" in the node numbers are allowed and no two nodes can have the same node number in the one model. For the three study models the number of node are 10973x2, 19070x2 and 29200x2 for URVLB, UARB and Upper Awash-Rift basins respectively.

When drawing the finite element mesh, each element is assigned a unique element number. The element numbers begin with one and continue sequentially to the number of elements in the mesh for each model basin independently. Each element is described using six nodes; the nodal coordinates define the size and shape of the element. For this reason the node numbers for each element are listed. The material properties are specified for each element in the mesh. The properties for each material set are then listed once.

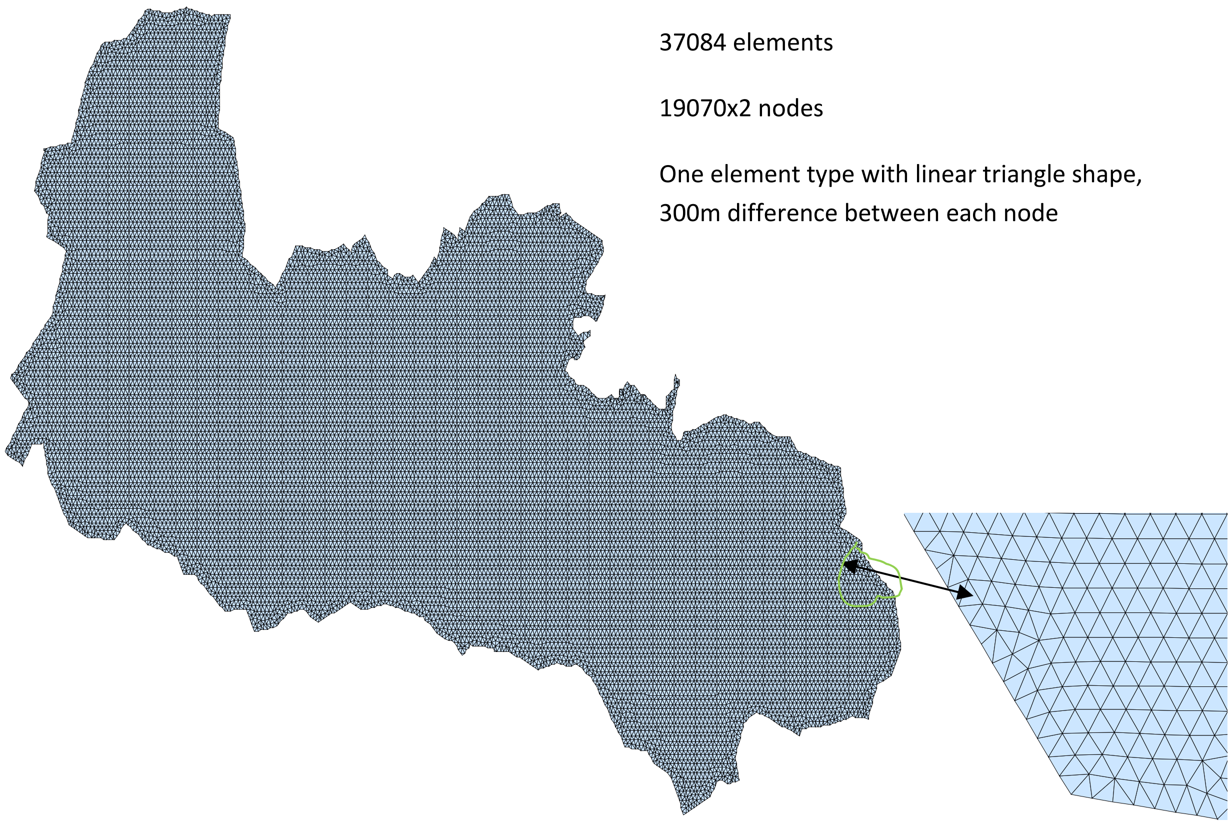

Figure 3-2 Upper Awash River basin, generated meshes

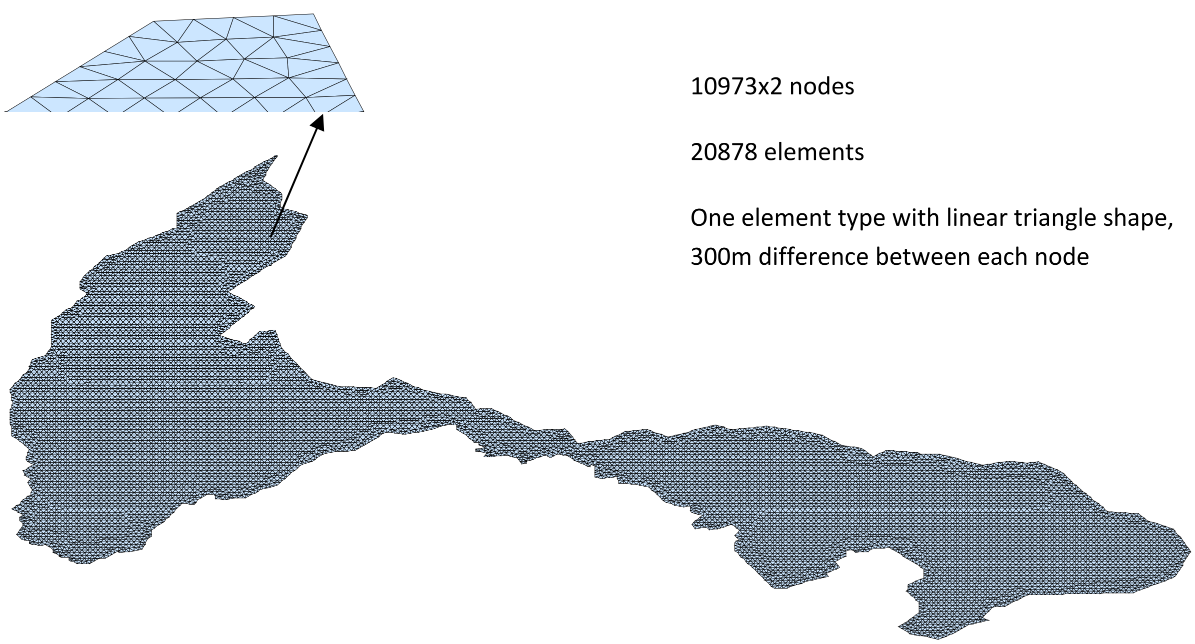

Figure 3-3 Upper Rift Valley lakes basin, generated meshes

Point values of the dependent variable i.e. head is calculated at nodes, which are the corners or vertices of the elements, and a simple equation is used to describe the value of the dependent variable within the element. This simple equation is called a basis function and each node that is part of an element has an associated basis function. The simplest basis functions that are usually used are linear functions. The solution to the differential equation for flow is approximated by a set of elements in which the dependent variable only varies linearly within the element, but the entire set of elements approximates the complex distribution of head.

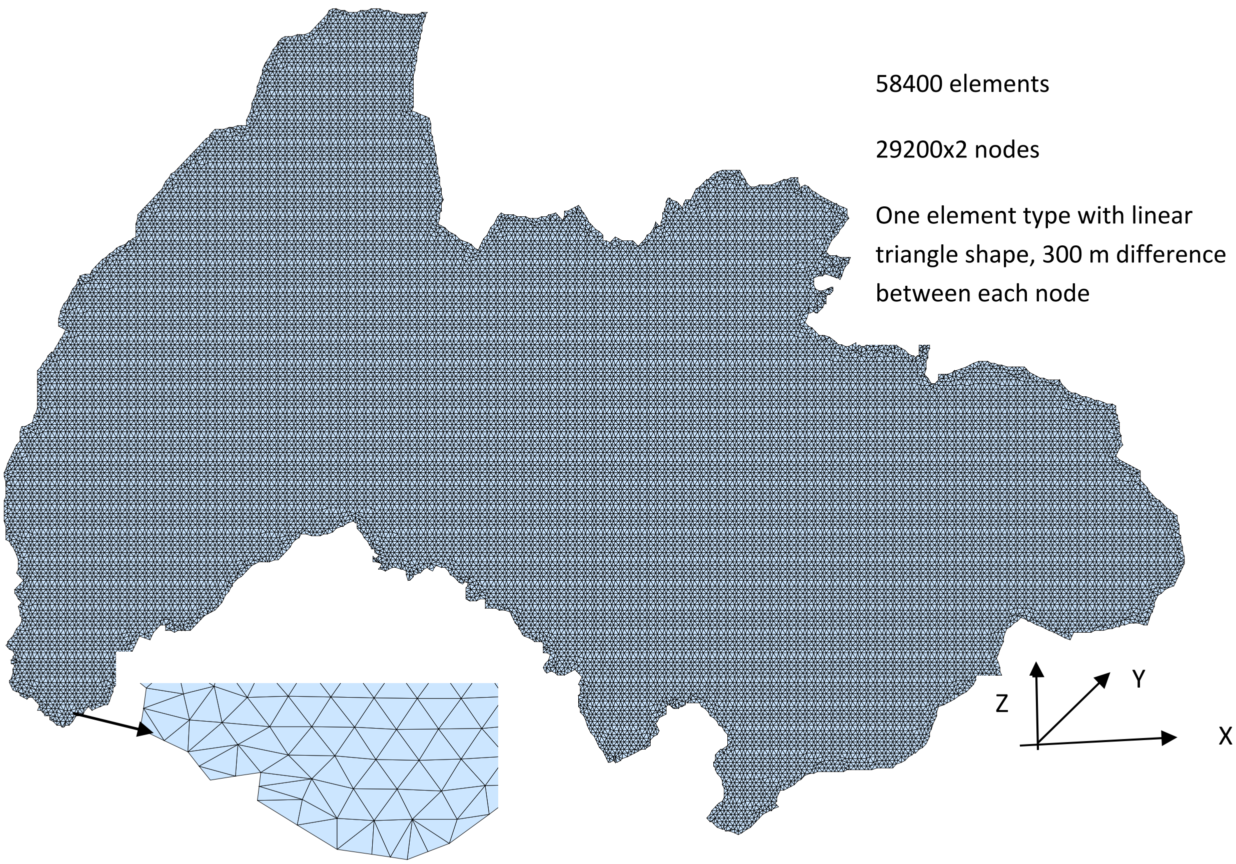

Figure 3-4 Upper Awash River basin – Upper Rift valley lakes basin generated mesh

3.4.2 Approximating equations

Driving an integral formulation for the governing groundwater equation which leads a system of algebraic equations that can be solved for hydraulic head is done by different methods. The method of weighted residuals is a more general approach that is widely used in groundwater flow modeling. In the method of weighted residuals, an approximate solution to the boundary or initial value problem is defined. When this approximate solution is substituted into the governing differential equation, an error or residual occurs at each point in the problem domain. We then force the weighted average of the residuals for each node in the finite element mesh to equal to zero. Galerkin's Method is the subset of the method of weighted residuals that is most commonly used to solve groundwater flow and solute transport problems (Istok, 1989). In Galerkin's Method the weighting function for a node is identical to the interpolation function used to define the approximate solution. And head (h) can be approximated as;

$$\hat{h}(e) = \sum_{i=1}^{n} N_i^{(e)} h_i \qquad 3.1$$

Where, $\hat{h}^{(e)}$ is the approximate solution for hydraulic head within element e, $N_i^{(e)}$ are the interpolation functions for each node within element e, n is the number of nodes within element e, and h_i are the unknown values of hydraulic head for each node within element e.

When the approximate solution is substituted into equation 2.5, the differential equation is not satisfied exactly and an error or residual occurs at every point in the problem domain. And the contribution of any element e to the residual at a node i to which the element is joined is;

$$Ri(e) = -\iiint \mathbf{W_i^{(e)}}(\mathbf{x},\mathbf{y},\mathbf{z}) \left[\frac{\partial}{\partial x}\left(\mathbf{k_x}\frac{\partial \hat{h}^{(e)}}{\partial x}\right) + \frac{\partial}{\partial y}\left(\mathbf{k_y}\frac{\partial \hat{h}^{(e)}}{\partial y}\right) + \frac{\partial}{\partial z}\left(\mathbf{k_z}\frac{\partial \hat{h}^{(e)}}{\partial z}\right) \pm q^{(e)} \right] \mathbf{dxdydz} \qquad 3.2$$

Where, $W_i^{(e)}$ is the weighting function for node i and the limits of the integration are chosen to represent the volume of element e and R is the residual or error due to the approximate solution. In Galerkin's method we choose the weighting function for each node in the element to be equal to the interpolation function for that node, $W_i^{(e)} = N_i^{(e)}$. The residual varies from point-to-point within the problem domain. At some points it may be large and at other points it may be small (the sign of the residual also can vary from point-to-point).Therefore we cannot force R to be zero at certain specified points because the residual may then become unacceptably large elsewhere in the problem domain. In the method of weighted residuals, we force the weighted average of the residuals at the nodes to be equal to zero. If we also assume that values of saturated hydraulic conductivity in the three coordinate directions are constant within an element (but can vary from one element to the next), equation 3.2 can be written as;

$$Ri(e) = -\iiint_{v(e)} \left(W_i^{(e)}(x,y,z)\left[\left(k_x^{(e)} \frac{\partial^2 \hat{h}^{(e)}}{\partial x^2}\right) + \left(k_y^{(e)} \frac{\partial^2 \hat{h}^{(e)}}{\partial y^2}\right) + \left(k_z^{(e)} \frac{\partial^2 \hat{h}^{(e)}}{\partial z^2}\right) \pm q^{(e)} \right] \right) dxdydz \qquad 3.3$$

Where, $k_x^{(e)}$, $k_y^{(e)}$ and $k_z^{(e)}$ are the value of simulated hydraulic conductivity in the x, y and z direction within element e. Equation 3.3, using integration by parts it can be written as;

$$Ri(e) = -\iiint N_i^{(e)} \left[\left(k_x^{(e)} \frac{\partial N_i^{(e)}}{\partial x}\frac{\partial \hat{h}^{(e)}}{\partial x}\right) + \left(k_y^{(e)} \frac{\partial N_i^{(e)}}{\partial y}\frac{\partial \hat{h}^{(e)}}{\partial y}\right) + \left(k_z^{(e)} \frac{\partial N_i^{(e)}}{\partial z}\frac{\partial \hat{h}^{(e)}}{\partial z}\right) \pm q^{(e)} \right] dxdydz \qquad 3.4$$

The most general formulation for [K(e)] (element conductance matrix) can be written for the case of a three-dimensional problem being solved using elements with n nodes.

$$k^{(e)} = \iiint_{V(e)} \begin{bmatrix} \frac{\partial N_i^{(e)}}{\partial x} & \frac{\partial N_i^{(e)}}{\partial y} & \frac{\partial N_i^{(e)}}{\partial z} \\ \vdots & \vdots & \vdots \\ \frac{\partial N_n^{(e)}}{\partial x} & \frac{\partial N_n^{(e)}}{\partial y} & \frac{\partial N_n^{(e)}}{\partial z} \end{bmatrix} \begin{bmatrix} k_x^{(e)} & 0 & 0 \\ 0 & k_y^{(e)} & 0 \\ 0 & 0 & k_z^{(e)} \end{bmatrix} \begin{bmatrix} \frac{\partial N_i^{(e)}}{\partial x} & \cdots & \frac{\partial N_n^{(e)}}{\partial x} \\ \frac{\partial N_i^{(e)}}{\partial y} & \cdots & \frac{\partial N_n^{(e)}}{\partial y} \\ \frac{\partial N_i^{(e)}}{\partial z} & \cdots & \frac{\partial N_n^{(e)}}{\partial z} \end{bmatrix} dxdydz \qquad 3.5$$

Where v(e) is the volume of element e. If we combine equations 3.5 and equation 3.2 we can write

$$\begin{Bmatrix} R_1^{(e)} \\ \vdots \\ R_n^{(e)} \end{Bmatrix} = [\, k^{(e)}] \begin{Bmatrix} h_1 \\ \vdots \\ h_n \end{Bmatrix} \qquad 3.6$$

Then Equation 3.6 is written for each element in the mesh. These equations are then combined to obtain;

$$\begin{Bmatrix} R_1^{(e)} \\ \vdots \\ R_n^{(e)} \end{Bmatrix}_{px1} = \overset{glob}{[K]_{pxp}} \begin{Bmatrix} h_1 \\ \vdots \\ h_n \end{Bmatrix}_{px1} \qquad 3.7$$

And by setting the residuals equal to zero $[K]_{pxp}\{h\}_{px1} = \{0\}_{px1}$, for steady saturated three dimensional groundwater problem with Neuman boundary can be summarized as;

$$[K]_{pxp}\{h\}_{px1} = \{F\}_{px1} \qquad 3.8$$

Where ‘F’ is any specified flow rate at Neuman nodes.

In this study we have chosen a model called TAGSAC developed in Geosphere research Institute of Saitama University for solving the groundwater flow equation. It solves three-dimensional saturated/unsaturated groundwater flow equation in near field heterogeneous porous media using finite element method (FEM). TAGSAC tries to solve a three dimensional groundwater flow equation of the form $\frac{\partial}{\partial x}\left(K_x\gamma(\theta)\frac{\partial h}{\partial x}\right)+\frac{\partial}{\partial y}\left(K_y\gamma(\theta)\frac{\partial h}{\partial y}\right)+\frac{\partial}{\partial z}\left(K_z\gamma(\theta)\frac{\partial h}{\partial z}\right)=(C(\theta)+\alpha S_S)\frac{\partial h}{\partial t}-R^*$

Where

- h is hydraulic head
- $C(\theta)$ is specific moisture content
- Ss specific storage
- K hydraulic conductivity
- $\gamma(\theta)$ is Relative permeability
- θ is Saturation
- R* is source/sink term
- x, y, z are special orthogonal coordinates and t is time.

The numerical approximations of the groundwater flow equations describing fully three dimensional problems are obtained using the Galerkin finite element technique. In the TAGSAC approximation procedure, the flow region is first discretized into a network of finite elements, and an interpolating trial function is used to represent the unknown dependent variable (hydraulic head) over the discretized region. An integral approximation of the flow equation is then obtained using the Galerkin weighted residual criterion. Spatial integration is performed piecewise over each element. Upon assemblage of the elements and incorporation of boundary conditions, a system of nodal equations is obtained. For a steady-state simulation, these nodal equations are algebraic equations. TAGSAC has proofed to be applicable in a number of researches done all over the globe (Mebruk Mohammed, 2010).

3.5 Model Conceptualization

A conceptual model is a hypothetical simplified description of the groundwater system to be studied. Features often described in conceptual models include the following: (a) Relationship and extent of hydrogeologic units (hydrostratigraphy, hydrofacies). (b) Aquifer material properties (porosity, hydraulic conductivity, storativity, isotropy). (c) Potentiometric surfaces. (d) Water budget (inflows and outflows such as: surface infiltration, lateral boundary flux, leakage through confining units, withdrawals and injections). (e) Boundary locations (depth to bedrock, impermeable layer boundaries, etc.). (f) Boundary conditions (fluxes, heads, natural water bodies). (g) System stresses (withdrawal wells, infiltration trenches, etc.) (Engineers, 1999).

3.5.1 Aquifer material properties

Among the components of the environment, groundwater is one of the most sensitive. It occurs in a very wide spectrum of geological settings. Crystalline and metamorphic rocks, including granite, basic igneous rocks, gneiss, schist, quartzite, and slate are relatively impermeable. Water in these areas is supplied as a result of jointing and fracturing. The yield of water from fractured rock is dependent upon the frequency and interconnectedness of flow pathways. The more traditional notion of dealing with fractured rock aquifers is that as the scale of interest increases the more appropriate it is to employ equivalent porous media modelling approaches, where extensive regions of an aquifer are represented by uniform hydrogeological properties. In this approach, individual fractures are not explicitly treated in the model but rather the heterogeneity of the fractured rock system is modelled using a small number of regions, each of which is modelled as an equivalent porous medium. The hydraulic conductivity distribution are replaced with a continuous porous medium having equivalent hydraulic properties. An equivalent porous media approach makes the assumption that a representative elementary volume (REV) of material characterized by equivalent hydraulic parameters can be defined. Modelling results are only valid at scales larger than the REV. The theory of flow through fractured rock and homogeneous anisotropic porous media is used to determine when a fractured rock behaves as a continuum.

Equivalent continuum approaches may be appropriate for answering questions that relate to averaged volume behavior. In the equivalent porous media approach, hydraulic properties of the system are modelled using equivalent coefficients such as permeability and effective porosity to represent the volume averaged behavior of many fractures within a fractured rock body. Thus, the details of individual fractures need not be known. This is in direct contrast to the discrete network approach where the details of individual fractures are explicitly accounted for in the model simulation (Long, 1982).

Equivalent porous medium (EPM) is a concept that is used to model or simulate the flow of groundwater in fractured rocks. The concept is that is we take a large enough volume, the fractured geologic material will behave mathematically like a porous medium. A drawback with this approach is that it is often difficult to determine the size of the representative elemental volume of material from which to define the effective hydraulic property values. Hence, the equivalent porous material approach may adequately represent the behavior of a regional flow system but is likely to reproduce local conditions poorly (Hiscock, 2005).

3.5.2 Boundary conditions

Mathematical models of groundwater flow based on equations like Equation 2.5 are classified as boundary value problems. To obtain a unique solution of such equations additional information about the physical state of the process is required. This information is supplied by boundary and initial conditions. For steady-state problems, only boundary conditions are required

The two most basic types of boundary conditions in flow nets are constant head boundaries and no-flow boundaries. Constant head boundaries occur along the boundary of water bodies like lakes or reservoirs. In a flow net, a constant head boundary has a line of constant head along it and streamlines are perpendicular to it. No-flow boundaries occur at the interface between the aquifer and materials with markedly lower hydraulic conductivity. A no-flow boundary is a streamline and constant head lines are perpendicular to it (Fitts, 2002).

Boundary conditions indicate how an aquifer interacts with the environment outside the model domain. They include things such as heads at surface waters in contact with the aquifer, the location and discharge rate of a pumping well. For a distinct solution, at least one distinctive boundary condition is specified. There are three types of boundary conditions which are derived from the most common two.

i. Constant Head Boundary: This is a type of specified head boundary condition, in which the head is known and the source of water has a constant water level at the model boundary. This condition is used in modelling an aquifer that is in good interaction with a lake, river or another external aquifer. These are usually where the groundwater is in direct contact with surface water such as a lake or a river and drains interact freely with the aquifer. It is mathematically known as Dirichlet boundary.

ii. Constant Flux Boundary: This is a type of specified flux boundary condition also known as the second type of boundary condition, and mathematically known as Neumann's condition or recharge boundaries. Entering or leaving the aquifer is prescribed/constant flux. This boundary condition is used in simulating rainfall or distributed discharge for instance evaporation and also used in specifying known recharge to the aquifer owing to induced recharge

iii. No flow Boundary (across which minimal flow occurs): This is a very special type of the prescribed flux boundary and is referred to as no-flux, zero flux, impermeable, reflective or barrier boundary. No flow boundaries are impermeable boundaries that allow zero flux. They are physical or hydrological barriers which inhibit the inflow or outflow of water in the model domain. No flow boundaries are specified either when defining the boundary of the model grid or by setting grid blocks as inactive (i.e hydraulic conductivity = 0)

The finite element method can handle all manner of boundary conditions, including no-flow boundaries, specified head boundaries, specified flux boundaries, and leakage boundaries (Fitts, 2002).

3.5.3 Modeling approach and protocol

Modelling can be a very powerful tool when used in the right circumstances and when models are properly constructed. That is, the approach must be appropriate for the particular site conditions and the stated study objectives based on the data available or/and that can be collected. Accurate and reliable data must be available in general.

A groundwater model is a computer-based representation of the essential features of a natural hydrogeological system that uses the laws of science and mathematics. Its two key components are a conceptual model and a mathematical model. The conceptual model is an idealized representation of our hydrogeological understanding of the key flow processes of the system. A mathematical model is a set of equations, which, subject to certain assumptions, quantifies the physical processes active in the aquifer system(s) being modelled. While the model itself obviously lacks the detailed reality of the groundwater system, the behavior of a valid model approximates that of the aquifer(s). A groundwater model provides a scientific means to draw together the available data into a numerical characterization of a groundwater system. The model represents the groundwater system to an adequate level of detail, and provides a predictive scientific tool to quantify the potential head of the aquifer.

Groundwater investigations, and modelling studies in particular, involve both a science and an art. The scientific basis is important, and requires a sound knowledge of geology, hydrogeology, groundwater hydraulics, hydrology, surface-groundwater interaction and engineering, as well as sufficient spatial and time series data to describe the system. The art is manifest in the creative processes required for developing a groundwater model as a simple computer-based representation of a complex natural system. There is also an art in applying experienced judgment where data are lacking to sufficiently rationalize natural processes, and in effectively communicating the modelling study results. Best practice modelling is not primarily a question of understanding and implementing the appropriate mathematical techniques, but of understanding and implementing an appropriate modelling approach.

In this study, issues that require a greater understanding of groundwater behavior for evaluating management options and determining appropriate solutions, relates to the head potential of the aquifer(s) will be investigated. Regional groundwater system includes both the rocks and groundwater of the defined area. It includes the area of recharge and discharge, storage and transmission of water, and geologic units that control the occurrence and movement of groundwater. Identification of this regional groundwater system is based up on (a) the relative hydrologic property of the major rock groups in the area under concern, (b) the regional groundwater movement as inferred from potential hydraulic gradients, (c) the relative distribution and quantities of estimated recharge.

Since the model should be complex enough to represent well, the real system of the groundwater a three-dimensional steady state model is selected. Therefore in this study a description of a three-dimensional conceptual and numerical model of groundwater flow in the unconfined aquifer system is incorporated. While calibrating the model with the observed data at the boreholes a trial and error method has been adopted. In this method a trial set of effective hydraulic parameters and surface recharge are given to the model and the result is evaluated according to its fitness with the measured data.

Mathematical model simulates groundwater flow and/or solute fate and transport indirectly by means of a set of governing equations thought to represent the physical processes that occur in the system (Anderson Mary p., 1992). This paper approaches the problem of understanding the flow system of the area with the very deterministic and numerical model which approximates physical law with finite element method by conceptualizing into three study models,

1. Groundwater model of part of the Upper Awash bounded by water divide with RVLB basin, Lake Koka and Awash River upstream of Lake Koka.
2. Groundwater model of part of the URLB basin bounded by Lake Ziway, Meki River and the water divide with the Awash River basin.
3. Groundwater model of an area bounded by Lake Koka, Awash River upstream of Lake Koka, Lake Ziway, Meki River including in the watershed divide of Awash River basin and RVLB. By considering that the surface water divide do not exist for groundwater case.

The first two cases will tell that there is no basin interaction as long as the predicted and inventoried data are in good agreement (no significant difference) when compared with the result of the last case. But if the predicted data is lower or higher than the inventoried data this is a hint that water must be entering or leaving in these assumed or modeled basins. So to identify or to be sure that this exchange of water is between these two basins (upper Awash and upper Rift valley lakes) the third case will be done

The third case is in assumption that there is no barrier between these basins and there is free movement of groundwater between the basins. And if the accuracy of the prediction is better than the other two cases it could be suggested that there is IGWT between the two basins around the study area. The typical modeling approach for each of the above three cases shown below.

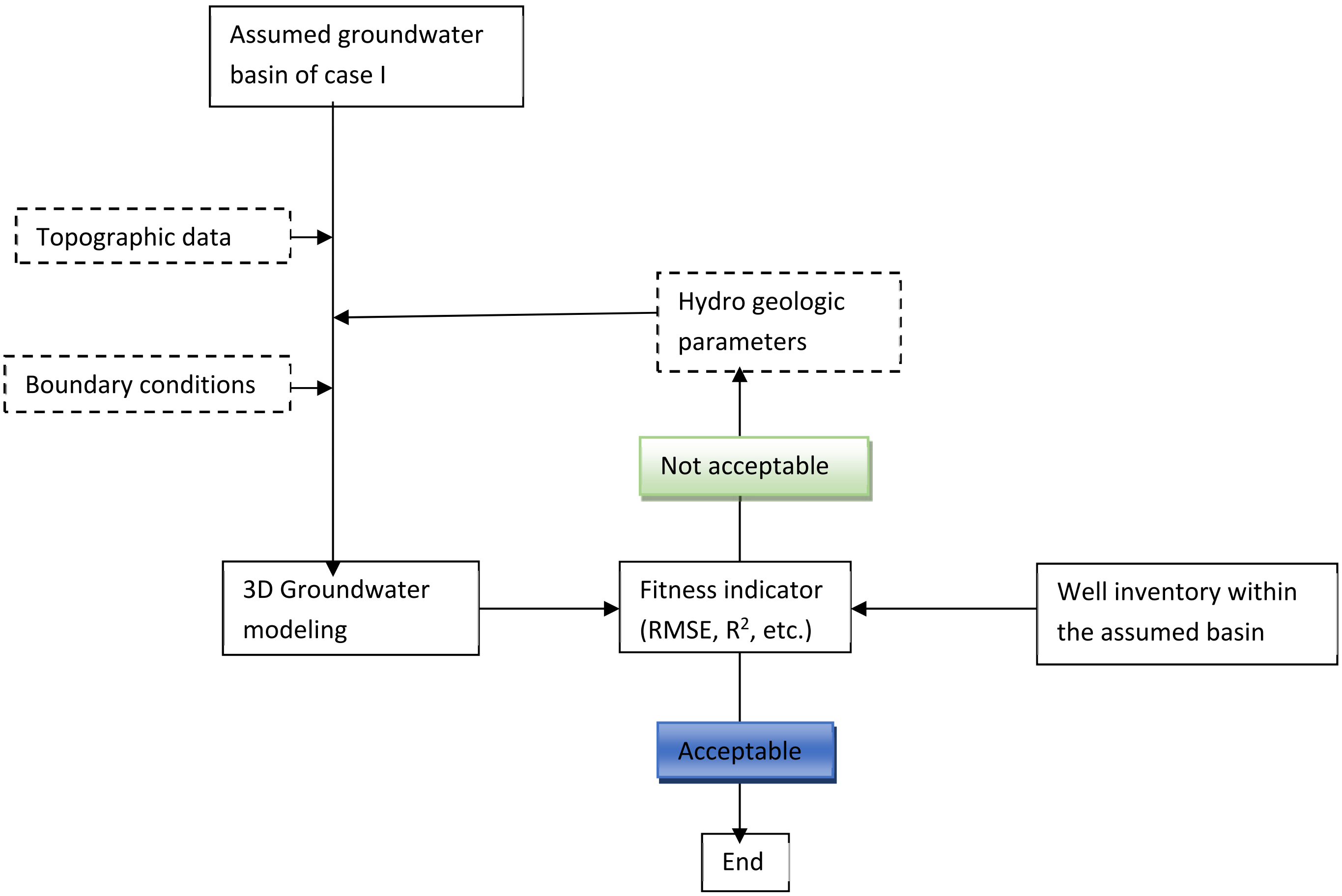

Figure 3-5 Modeling protocol

4 RESULT AND DISCUSSION

Based on the methodology and impute requirement of the model selected, all the necessary data are prepared and the hydraulic head is simulated. So in this part all the necessary results will be showed and discussed towards the objective of the research.

4.1 Well inventory

In both UARB and URVL, there are a lot of wells; shallow hand dug wells, deep wells and little of them springs. Almost all are secondary data, gained from offices. From these boreholes 917 are selected for analysis, 153 for Upper Awash basin in Appedx1 and 764 for Upper Rift valley lakes basin. The selection was based on the location or distribution in the area and their expected result in the model compared to the actual data (reliability). Based on this work some wells showed unexpected disagreement with the model result and we took our personal judgment on the accuracy of the data. Most of the wells are shallow and hand dug with minimum depth of 11meters and few are deep wells maximum depth of 295 meters. The information from these wells used in the model are; well location (Cartesian coordinate of each well) and static water level.

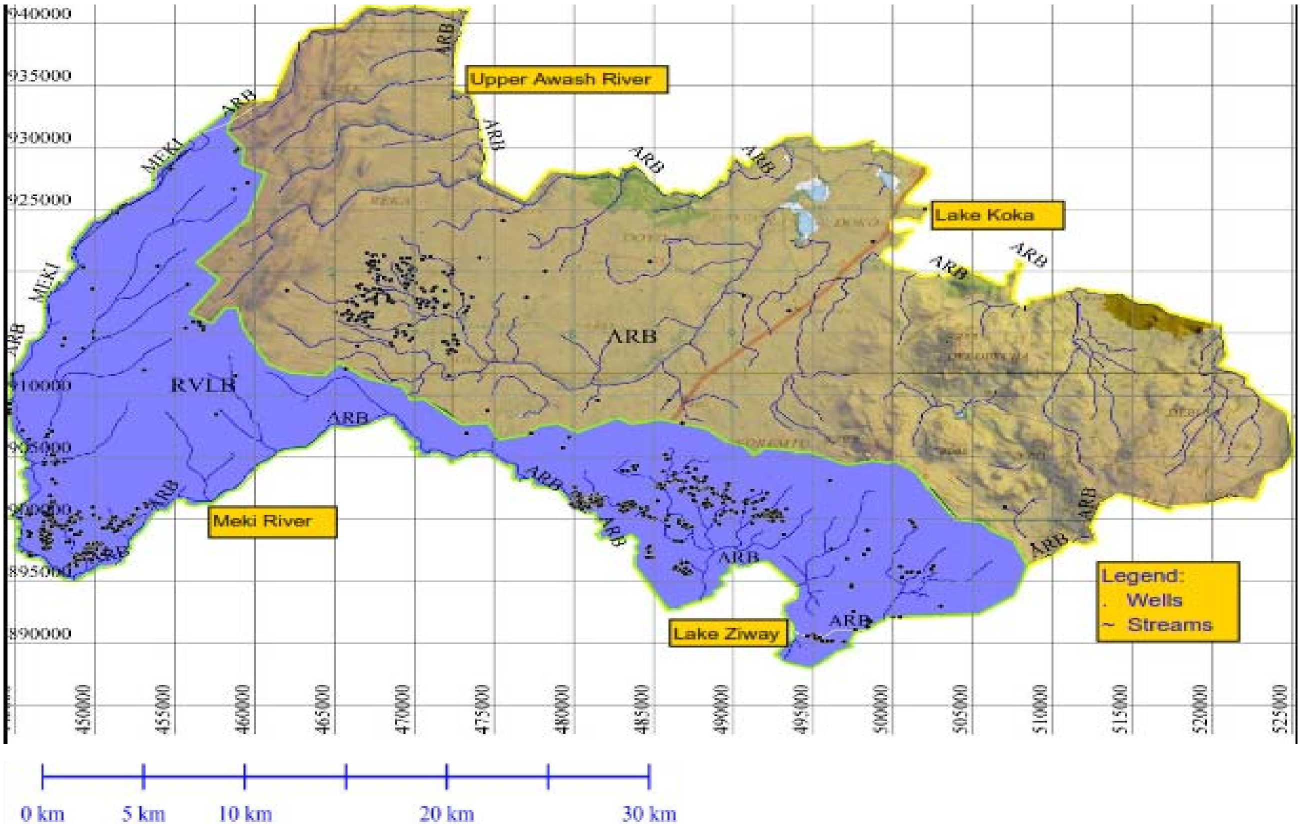

Figure 4-1 well location in the study area both in the URVLB and UARB

4.2 Rainfall distribution

To know the weighted average rainfall in the model basin the collected rainfall data from the National Meteorological Service Agency (NMSA) is analyzed by Thiessen polygon method. From NMSA we found 10 metrological stations in and very close to the model basin boundary which have full record except some months missing recordings. The monthly rainfall data of these selected stations are pre analyzed (missed data are filled, and averaged to its yearly) and analyzed using the selected method to know and use in the model to estimate or use as surface recharge of the aquifer. The procedure and theory of Thiessen polygon method is summarized as follows from the book called Ragunath. And the work with its results is shown in Figure 4.2 and Table 4.1.

Thiessen polygon method attempts to allow for non-uniform distribution of gauges by providing a weighting factor for each gauge. The stations are plotted on a base map and are connected by straight lines. Perpendicular bisectors are drawn to the straight lines, joining adjacent stations to form polygons, known as Thiessen polygons. Each polygon area is assumed to be influenced by the rain gauge station inside it, i.e., if P_1, P_2, P_3, ... P_n are the rainfalls at the individual stations, and A_1, A_2, A_3, ... A_n are the areas of the polygons surrounding these stations, (influence areas) respectively, the average depth of rainfall for the entire basin is given by;

$$\mathbf{P_{av}} = \frac{\sum \mathbf{A_i p_i}}{\sum \mathbf{A_i}} \qquad 4.1$$

Where, P_i mean annual rainfalls recorded at each rain gauge stations, P_{av} Average aerial depth of rainfall of the basin and $\sum A_i$=A Total area of the basin under concern. (Ragunath, 2006).

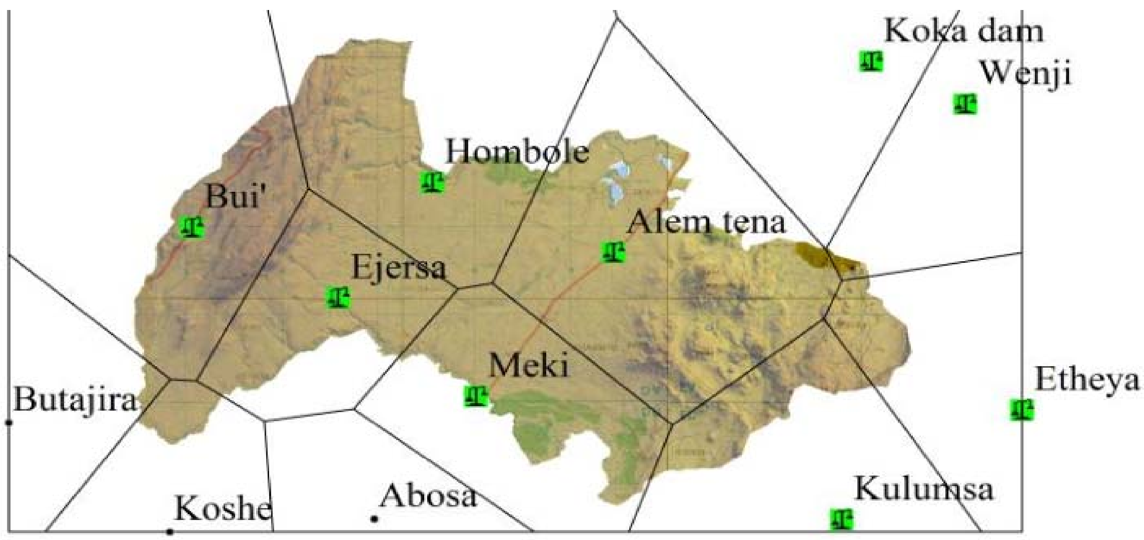

Figure 4-2 constructed Thiessen polygon for the study area

Table 4.1 Mean annual precipitation of the study area

No.	Station name	UTM-E(m)	UTM-N (m)	Mean rainfall(mm)	Enclosed A (sq km)	P*A
1	Alem Tena	494300	917000	875.36	653.78	572,290.25
2	Hombole	475500	925000	809.57	386.15	312,614.81
3	Ejersa Lele	465700	911800	883.55	297.69	263,023.01
4	Bui	450500	919850	1,074.21	290.15	311,682.42
5	Meki	480000	900500	766.34	289.19	221,618.67
6	Kulumsa	518000	886400	821.5	144.86	119,002.49
7	Etheya	536722	898971	1,029.70	64.76	66,685.68
8	Wenji	530822	933973	825.5	7.01	5,787.58
9	Koshe	448250	885000	944.36	40.33	38,088.87
10	Butajira	431500	897500	1,110.11	9.34	10,363.99
				Sum	2,183.26	1,921,157.76
				P_{av} = Sum(A*P)/Sum(A)		879.95

4.3 Aquifer geometry

The study area (aquifer) is selected in a way that to be bounded by the two perennial rivers namely Upper Awash River (a few kilometers lower from its origin to Koka Lake) and Meki River from its source to Ziway Lake. Lake Koka and Lake Ziway are part of the boundary of the model basin under study. In this steady state modeling the surface nodes near to the two rivers and the two lakes are taken as constant-head boundary conditions. The longitudinal profile of the rivers and the profile of the lakes edge is taken from DEM to represent the surface water level.

In the Upper Awash model basin, Awash River and Lake Koka, and in the Upper Rift valley lakes model basin Meki River and Lake Ziway are taken as constant head boundaries. In these models the surface water divide of Awash basin and Rift valley Lakes basin is taken as the no flow boundary (zero flux nodes). But during the third model basin the surface water divide of these basins is assumed to be not exist. Since the distance between nodes in all directions taken as 300 meters, the nodes 300 meters below the surface are considered as no flow boundary. And the nodes at the surface are recharge boundaries. The following figure (Figure 4.3) shows the location of all the boundaries of assumed basins.

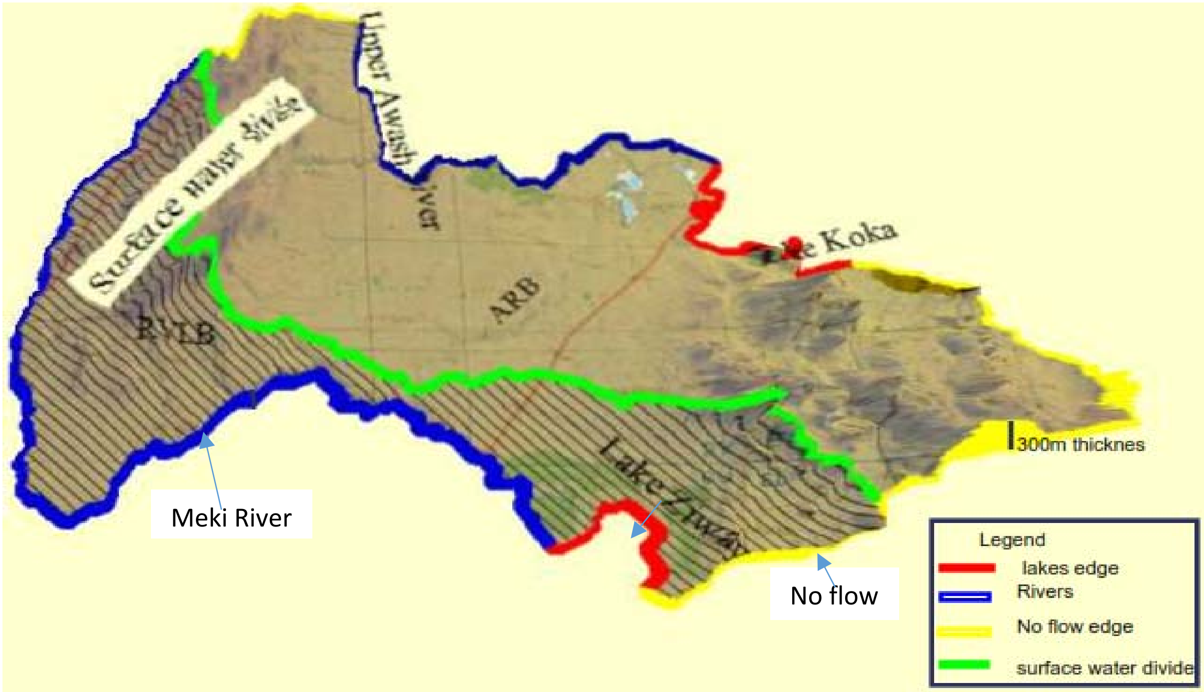

Figure 4-3 Aquifer geometry used in the model

4.4 Hydraulic head and interbasin flow

The basic unit of surface-water hydrology is the drainage basin, or catchment, which consists of all the land area sloping toward a particular discharge point. It is outlined surface-water boundaries, or topographic divides. In groundwater hydrology; we use the concept of a groundwater basin, which is the subsurface volume through which groundwater flow's toward a specific discharge zone. Groundwater divides surround it. The boundaries of a surface-water basin and the underlying groundwater basin do not necessarily coincide, although the water budget of the area must account for both ground and surface water (Fetter, 2001). Thus groundwater basin can be defined as a hydrological unit containing one large aquifer or several connected and interrelated aquifers. Porosity and hydraulic conductivity of porous material characterize the distribution and ease of movement of groundwater in geological formations with energy difference.

In addition to the outside forces acting on the groundwater, there is an energy contained in the groundwater that causes the water to move, hydraulic head. The total energy in groundwater consists of three components: pressure, velocity, and elevation of the water body (elevation head).

Because groundwater velocities are low, this energy component is essentially zero. Thus the potential energy at a given point (hydraulic head) is equivalent to the elevation at a point of measurement (elevation head), such as a well screen, plus the depth of the water column that rises in a well (pressure head).

In unconfined aquifer, the top of the saturated zone is at atmospheric pressure, which is constant across a site. Because the pressure head is constant across the site, this component is generally not taken into consideration when calculating the energy to drive groundwater to a point of discharge. Therefore, the height of the water column, z, represents the actual energy available to drive water through aquifer materials to a point of discharge like a well or spring. The differences in elevation head between point, at a higher elevation head, and a point, at a lower elevation head, force the groundwater towards the lower energy potential or towards the lower elevation head. The rate of groundwater flow towards a well or spring is proportional to the difference in elevation head between the source (high) and discharge (low) areas.

In this study area the surface water divide is drawn from previous study and from the model result the groundwater contour with flow direction is drawn on same map. From this map we can see that weather the surface water divide and groundwater divide are coincident or not by following the groundwater flow directions in Figure 4.6. Furthermore, to beef up this visual observation the geological and geographical relationships of the two basins are taken into consideration. And also the calibration results are cross checked by simulating the hydraulic head with the calibrated results of surface recharge and hydraulic conductivity of each model basins.

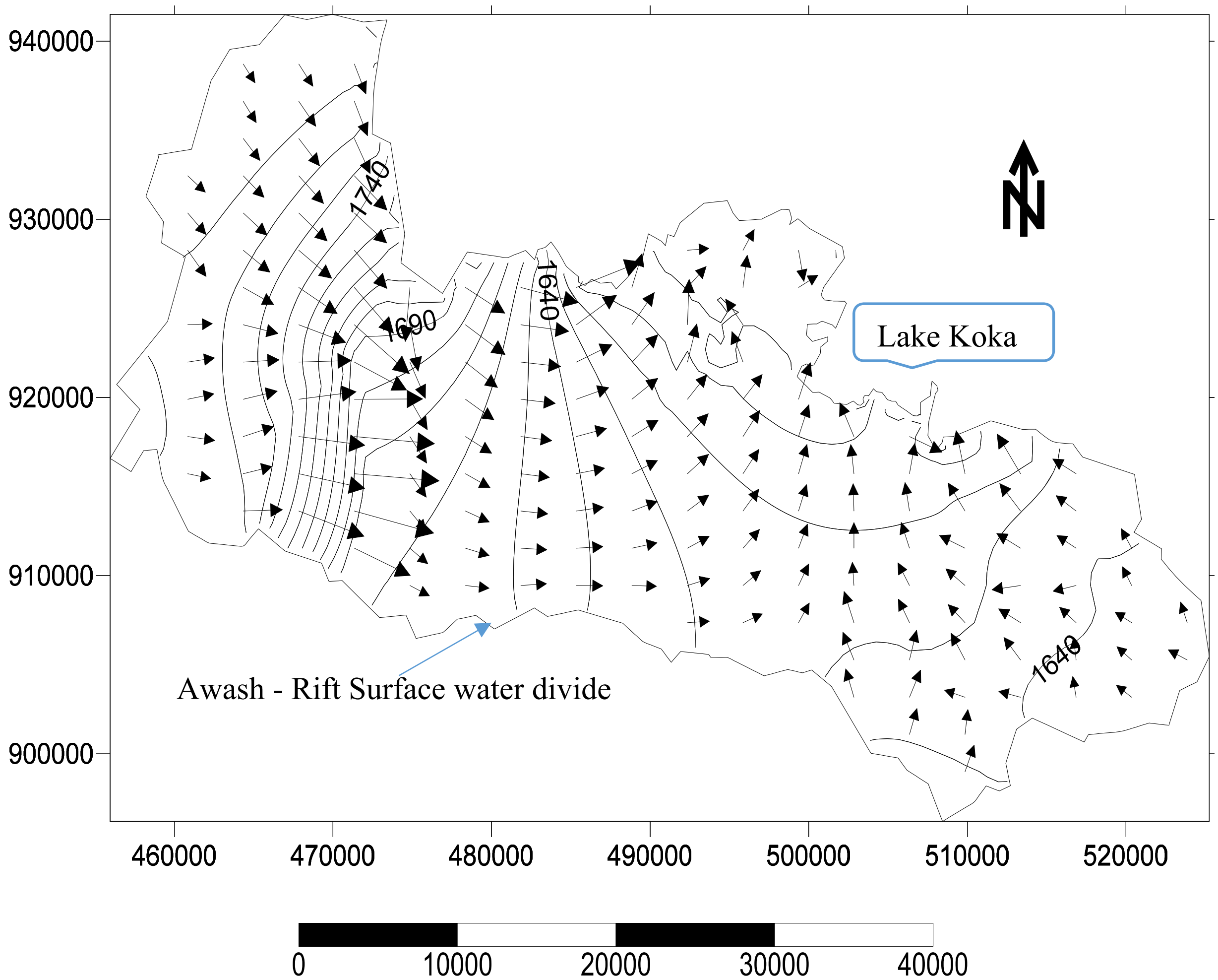

Figure 4-4 Contour of simulated head potential and groundwater flow direction of Upper Awash River basin

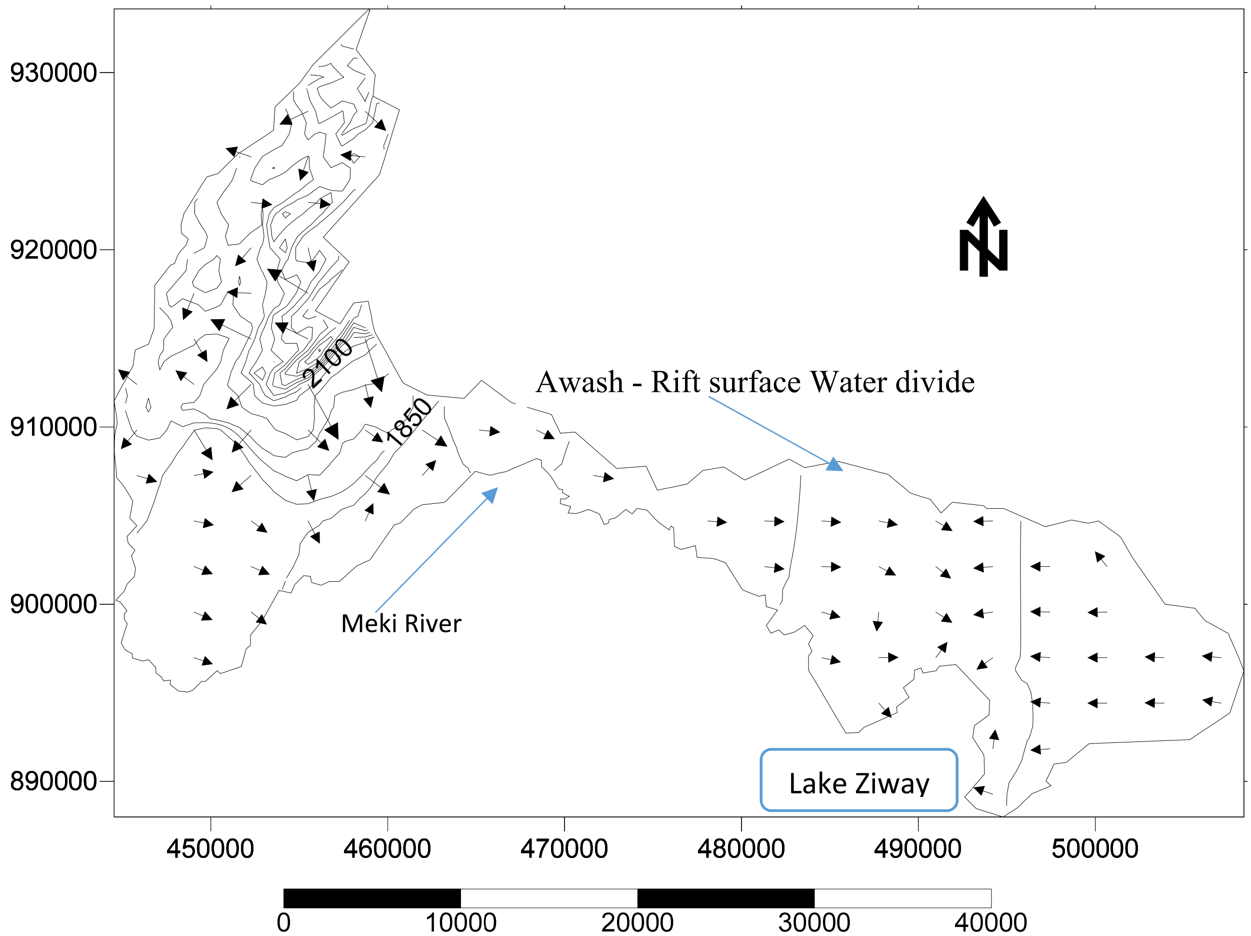

Figure 4-5 Contour of simulated head potential and groundwater flow direction of Upper Rift Valley lakes basin

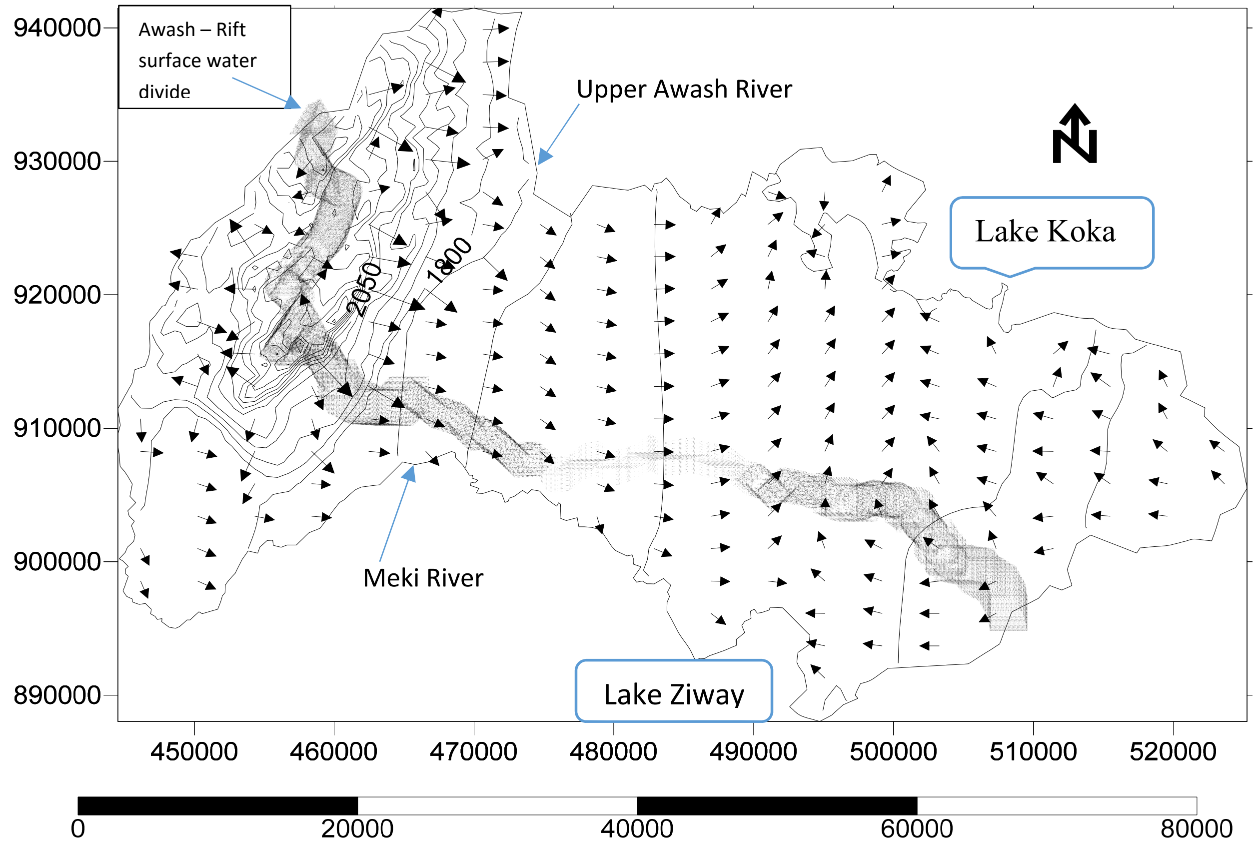

Figure 4-6 Contour of simulated head potential and groundwater flow direction of Upper Awash and Upper Rift Valley lakes basin as considered as one basin (Awash –Rift)

4.5 Model calibration and results

After all preparation of the imputes including, designing the model, assigning boundary conditions and input parameters, the groundwater regime of the three model basins were simulated under steady state for unconfined aquifer of the study area. To demonstrate that groundwater models are realistic, field observations of aquifer response are compared to corresponding model simulated values in model calibration, with the general objective of reducing the difference.

Calibration is the process of adjusting model parameters (material properties and boundary conditions) until (l) the model is consistent with the analyst's understanding of the groundwater flow system and with all available data, and (2) computed values of head closely match measured values at selected points in the aquifer (locations of wells and springs). The procedure is essentially an exercise in "trial and error" wherein a plausible set of model parameters are proposed, computed and measured values of head are compared, and model parameters are adjusted to improve the fit (Istok, 1989). A flow model is considered calibrated when it can reproduce, to an acceptable degree, the hydraulic heads of the natural system being modeled. This is accomplished by finding a set of values for the boundary conditions and aquifer properties. In other words, calibration methods solve a problem inversely by iteratively adjusting the unknowns until the solution matches the hydraulic heads.

Multiple calibrations of the same system are possible using different boundary conditions and aquifer properties. There is no one unique calibration that is “correct” for any model. Methods of calibrating can be grouped into two categories: manual trial-and error calibration and automated calibration. Even if automated methods are becoming increasingly usable and accepted due to its familiarity and easy of understanding in complex solutions trial and error is mostly used one. The calibration of a deterministic groundwater model is often accomplished through a trial-and-error adjustment of the model input data to modify model output. This method of calibration is labor-intensive. The modeler makes successive cycles involving interpreting prior results to determine where inputs need adjustment, making speculative adjustments to the input code, re-running the model and output software, and then comparing the computed results to the natural system. Hundreds of iterations are made before an acceptable calibration is achieved. Typically, the inputs being adjusted are hydraulic conductivity (or transmissivity), storage, leakage across a confining layer used as a boundary, flux to and from a surface water body, and designation of boundary conditions. A typical manual trial-and-error calibration process followed in this study includes the following steps: (a) Complete initial model development and assignment of properties (b) Identifying the parameters to be adjusted during calibration and the appropriate range for each. (c) Identify the locations and values for the target points forming the calibration set. Groundwater flow models are usually calibrated to a set of observed potentiometric head levels. (d) Iteratively run the modeling software and adjust input parameters, hydraulic conductivity for each geologic zone and surface recharge, until an acceptable match between observed and calculated values at the target points is achieved.

As a large number of interrelated factors affect the output, trial and-error adjustment may become a highly subjective and inefficient procedure. So other calibration method called automated calibration is used in combination with manual trial and error method. This method utilizes an objective function, such as minimization of the sum of the squared differences between observed and computed heads (residuals), to govern automatic iterative adjustment of values that would otherwise be adjusted manually. Automated calibration methods have some potential advantages over trial-and-error methods. They can provide a systematic approach to calibration, allowing for efficiencies within individual modeling jobs and a basis for comparison between different modeling jobs. Statistical measurements are available from some automated approaches that are not usually performed in trial-and error approaches.

In the model hydraulic head is computed for the three model basins by varying the hydraulic conductivity for the four geologic zones of the area and the surface recharge from the rainfall. After many trial and error works selection of a set of hydraulic parameters for geologic zones in Figure 4.7 and recharge for each model basin, the following parameter values are selected as the best among other combinations. Some values of the hydraulic conductivities are higher in the vertical direction than the horizontal one, this is because of the geological fault of the area.

Table 4.2 Geologic parameters of the study area for classified geological zones

Model basins	Geology no	Kxx(cm/s)	Kyy(cm/s)	Kzz(cm/s)	surface recharge
UARB	1	9.74E-02	9.74E-02	7.08E-02	10%
	2	8.85E-03	2.74E-02	7.97E-03	
	3	3.10E-03	1.77E-03	7.08E-03	
	4	5.75E-03	4.51E-03	1.77E-03	
URVLB	1	1.36E-02	3.40E-01	3.40E-01	10%
	2	1.70E-05	1.70E-04	1.70E-04	
	3	3.40E-03	5.10E-01	3.40E-01	
	4	1.19E-01	3.40E-02	3.40E-01	
Upper Awash- Rift	1	3.45E-02	2.30E-01	2.30E-01	16%
	2	2.99E-04	2.76E-04	2.30E-03	
	3	1.73E-03	4.60E-02	9.20E-02	
	4	1.96E-02	5.75E-03	1.15E-01	

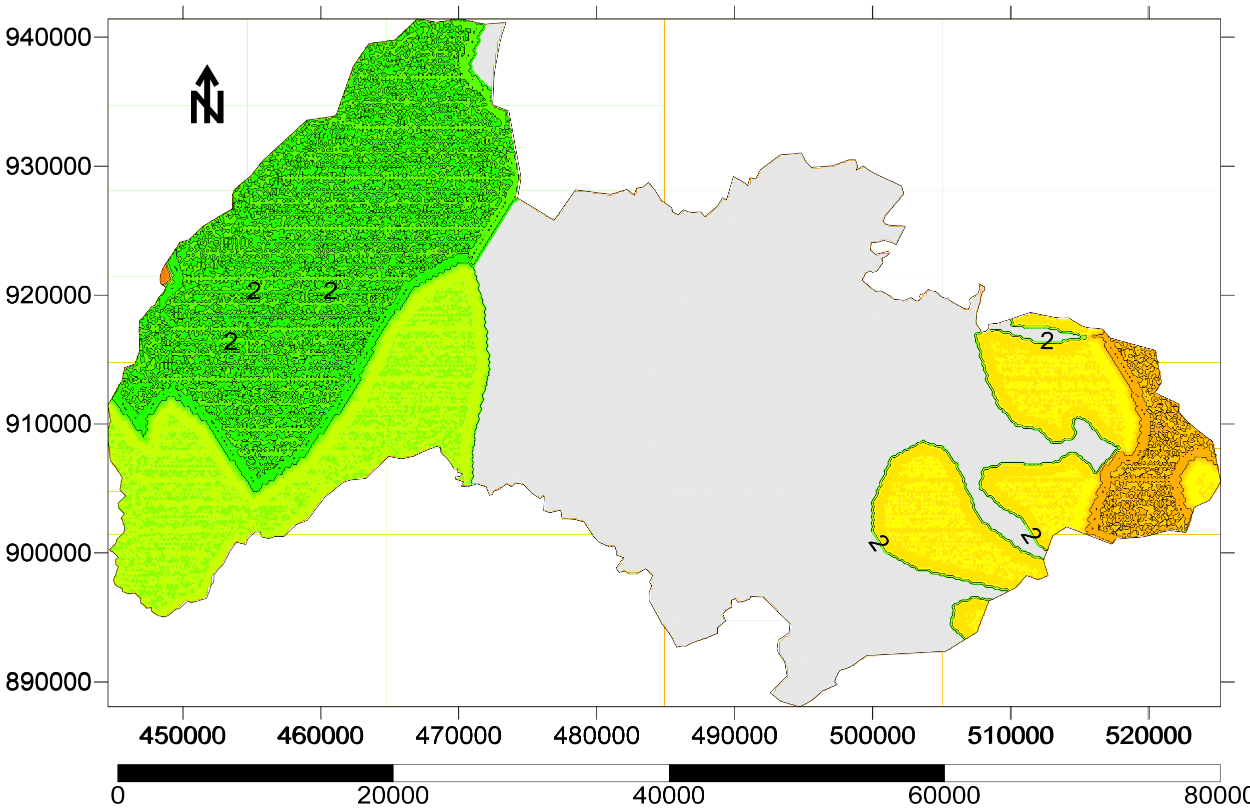

Figure 4-7 Geological classification of the area used in the model

Even though the geological classifications in Figure 4.7 are six only the first four of them are taken as calibrated geological parameters of the area. This is because of their small size relative to the other four and also their location in the model basin.

Observed target values (measurements) are plotted versus the values computed by the model. In an ideal calibration, the points will fall on a straight line with a 45 degree slope; i.e., the computed value equals the measured value. The degree of scatter about this theoretical line is a measure of overall calibration quality (Anderson Mary p., 1992). In Figure 4.8, 4.9 and 4.10 the best fit equation obtained between the modelled and measured head are made for the three model basin.

The R^2 value compares estimated and measured head, and ranges in value from 0 to 1. If it is 1, there is a perfect correlation in the between the modelled and measured values or there is no difference between the estimated and measured values. At the other extreme, if the coefficient of determination is 0, the regression equation is not helpful in predicting. In the following regression plots the intercept made zero.

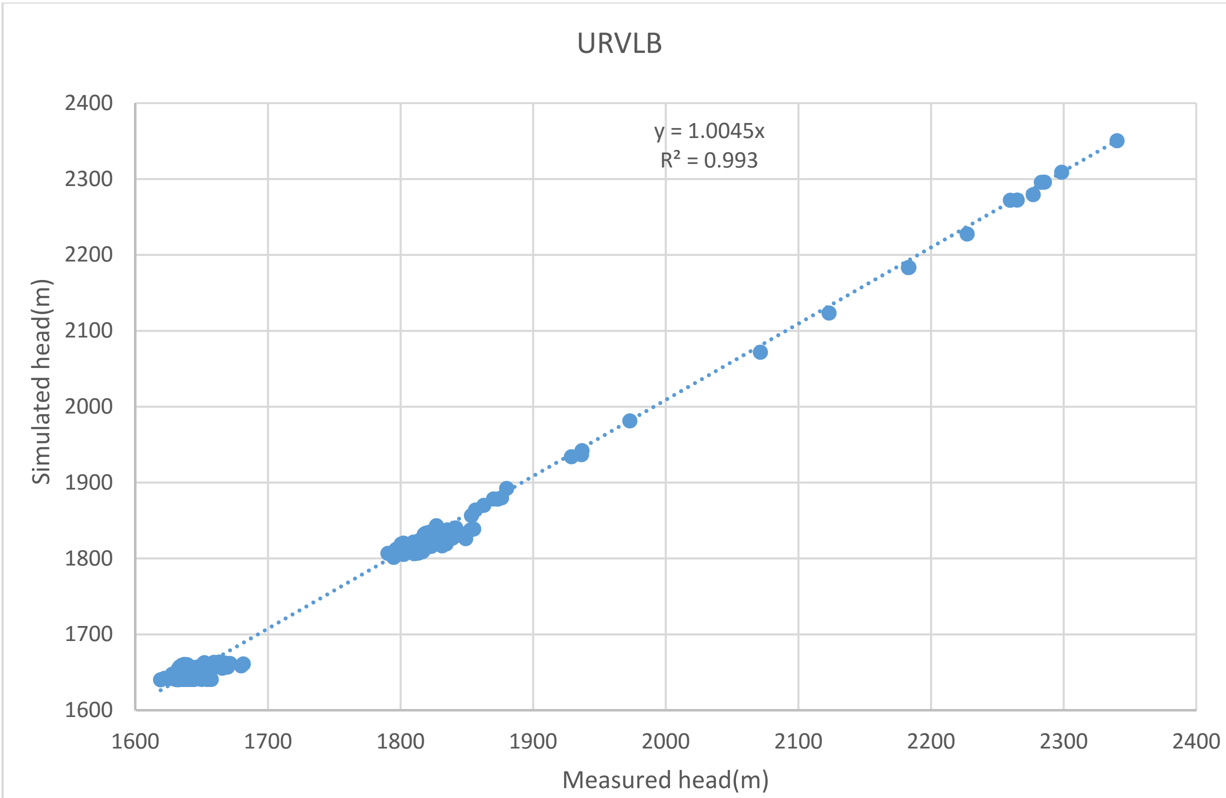

Figure 4-8 Plot (regression graph) of measured head versus modeled head for Upper Rift valley lakes basin

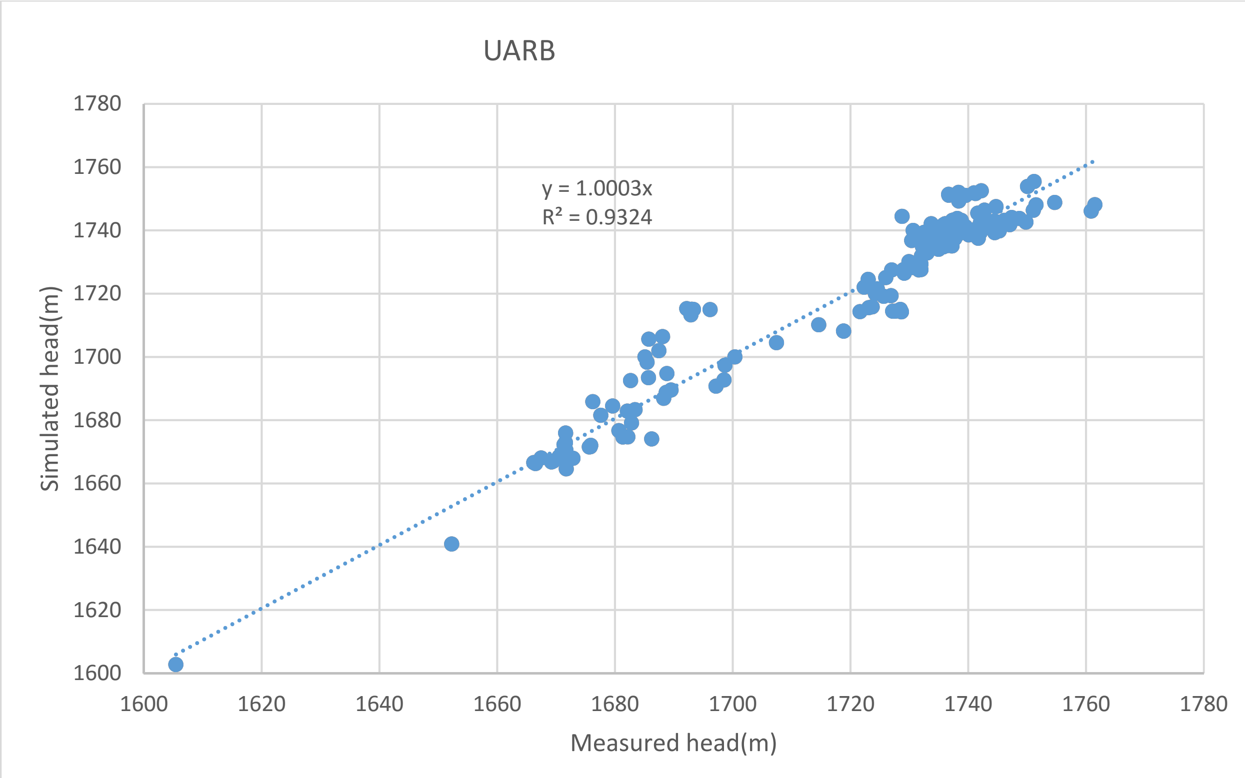

Figure 4-9 Plot (regression graph) of measured head versus modeled head for Upper Awash basin

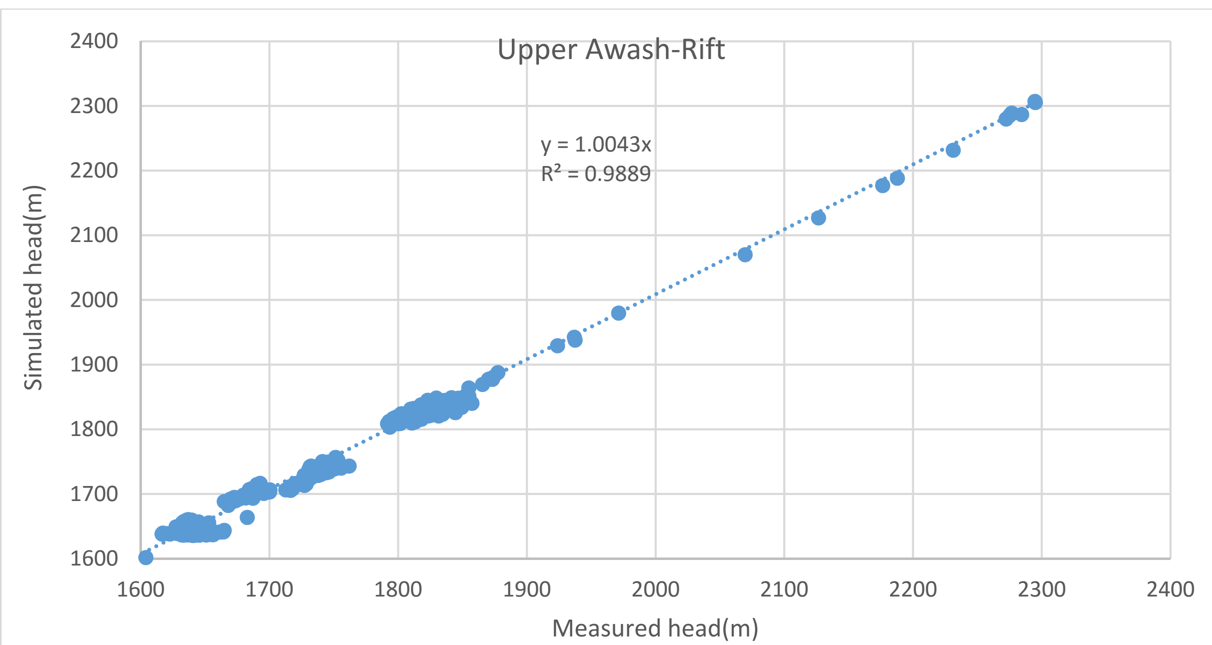

Figure 4-10 Plot (regression graph) of measured head versus modeled head for Upper Awash - Rift basin

Calibration results are evaluated by using qualitative and quantitative performance measures. Qualitative assessment (pattern matching) involves comparisons of contour maps and hydrograph of measured and simulated head, while quantitative performance measure involves mathematical or statistical description of residuals. The primary calibration target in groundwater modeling is hydraulic head (water level). Accordingly, in this study, steady-state calibration was made using static water level observations of 917 wells all in one. The effectiveness of calibration was evaluated by visual matching of measured groundwater level contours and lumped quantitative performance measures such as the mean error, the mean absolute error, and the root mean square error are done. The mean error (ME) is the mean of the differences between measured heads (h_m) and simulated heads (h_s):

$$ME=\frac{1}{n}\sum_{i=1}^{n}(h_m-h_s)_i \qquad 4.2$$

Where h_m and h_s are measured and simulated results and n is the number of data. Because both positive and negative residuals are resulted in the calculation, this value should be close to zero for a good calibration. In other words, the positive and negative errors should balance each other. The mean absolute error (MAE) is the mean of the absolute value of the differences between measured heads and simulated heads:

$$MAE=\frac{1}{n}\sum_{i=1}^{n}|(h_m-h_s)_i| \qquad 4.3$$

The MAE measures the average magnitude of the errors in a set of forecasts, without considering their direction. The root mean square (RMS) error is the square root of the average of the squared differences between measured heads and simulated heads:

$$RMSE=\left(\frac{1}{n}\sum_{i=1}^{n}(h_m-h_s)_i^{\,2}\right)^{0.5} \qquad 4.4$$

The RMSE is used as the basic measure of calibration for heads. Uncertainty in head measurements can be the result of many factors including, measurement error, scale errors, and various types of averaging errors, and accuracy of GPS in use both spatial and temporal. The maximum acceptable value of calibration criterion depends on the magnitude of the change in head over the problem domain. As a general calibration criteria RMSE equal to or less than 10 percent of the observed head range in the aquifer being simulated is better (Anderson Mary p., 1992). The mean error (ME) and the mean absolute error (MAE) are characterized by low values indicating that the model was well calibrated. The value of the correlation coefficient (R^2) indicates a good performance of the model. But due to the accuracy of GPS and the element dimensions in use RMSE slightly greater than 10 are taken as an acceptable errors.

Table 4.3 Summary of calibration errors

Models	Mean error (ME)	Mean Absolute Error (MAE)	Root Mean Square Error (RMSE)	R^2
UARB	0.544	5.492	7.605	0.932
URVLB	7.900	9.886	11.745	0.993
Upper Awash-Rift (combined)	7.567	10.713	12.702	0.989
In Upper Awash-Rift for UARB wells	1.261	9.774	11.582	0.759
In Upper Awash-Rift for URVLB wells	8.830	10.901	12.914	0.992

Simulation of head in each basin is also done by calibrated geological parameters of one basin to other basin for a matter of cross checking the agreement of one model basin with the other. And the RMSE values from these simulation: in UARB by using parameters of URVLB and Upper Awash- Rift we got 45.50 and 45.12, in URVLB by using parameters of UARB and Upper Awash –Rift 36.38 and 18.69, and also in Upper Awash-Rift 35.72 and 15.09 by parameters of UARB and URVLB. These results clearly indicate the parameters were calibrated for the specific region they were collected. And also the RMSE results show that URVLB and the combined basin (Upper Awash – Rift basin) have better agreement.

The above table (Table 4.3) shows that the Upper Awash-Rift basin model better describes the URVLB than the model solely developed for URVLB. However the Upper Awash-Rift basin model seldom describes the Upper Awash basin better than the model developed for UARB. This clearly indicate the influence of UARB on the URVLB i.e. their interaction.

Their interaction can also be seen from Figure 4.6; the figure shows both the surface water divide and groundwater flow direction lines, the groundwater flow shows crossing the surface water divide i.e. interaction of the two basins. Moreover, from the groundwater head grid groundwater shed is generated, this figure also agrees with Figure 4.6 and beef up the idea towards the interaction of the two basins in the modeled aquifer thickness.

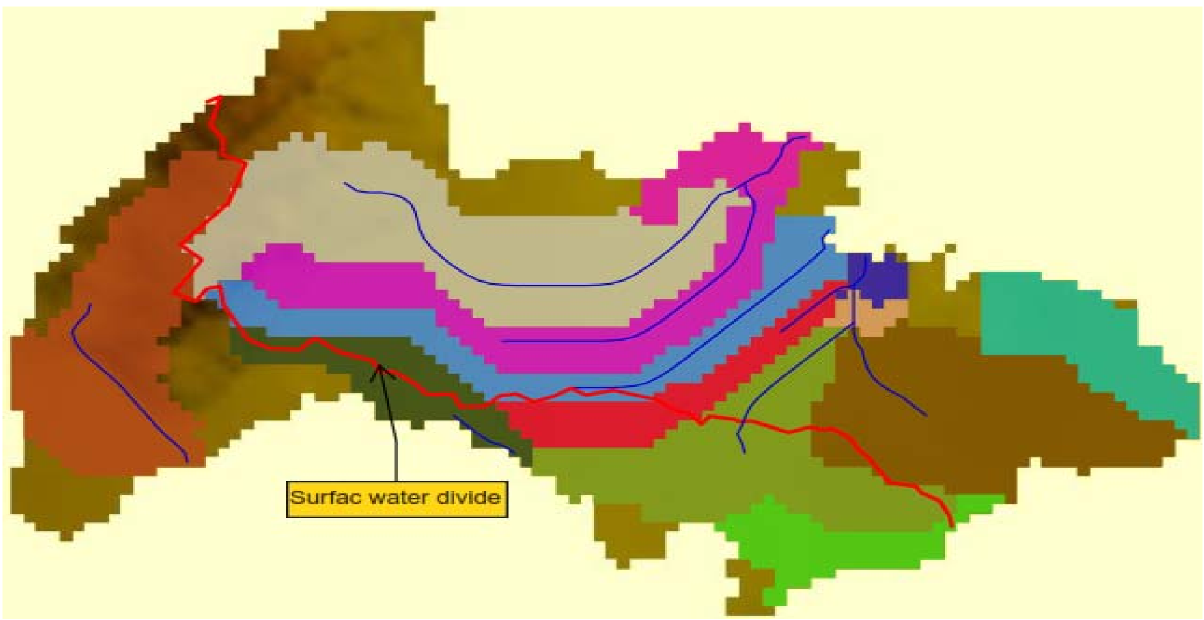

Figure 4-11 Groundwater sub basins of the combined basins (Upper Awash – Rift) from simulated head potential and Surface water divide of UARB and URVLB

In addition to the above calibration scatter plots are used in assessing the quality of calibration simulations. Error is the difference between the expected value and the observed value. Putting this together, the differences between the expected and observed values must be unpredictable. The idea is that the deterministic portion of the model is so good at explaining (or predicting) the response that only the inherent randomness of any real-world phenomenon remains leftover for the error portion. If there is explanatory or predictive power in the error, then predictors are missing some of the predictive information. Residual plots help to check this. A scatter plot can be created with observed value on the X-axis and residual on the Y-axis. In this case, the scatter of points should not have a pattern (it should be random). The residuals should fall in a symmetrical pattern and have a constant spread throughout the range. Both scatter plots for the calibration simulations of the three model basins are shown in Figure 4.12, 4.13 and 4.14 below.

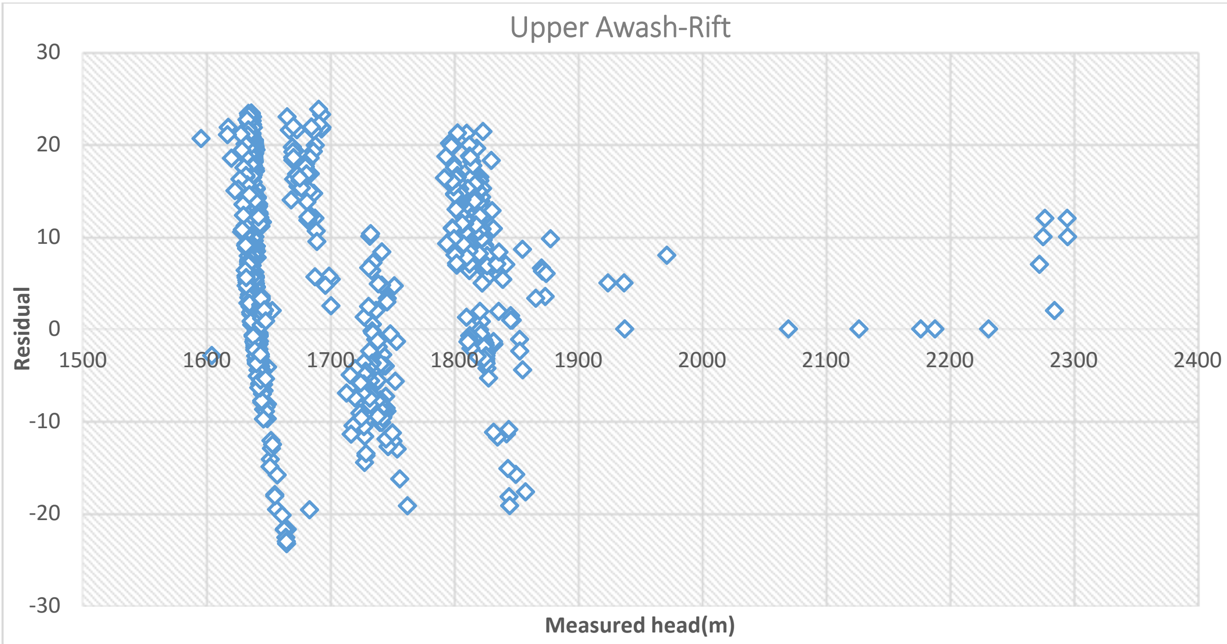

Figure 4-12 Scatter plot of Upper Awash - Rift basin.

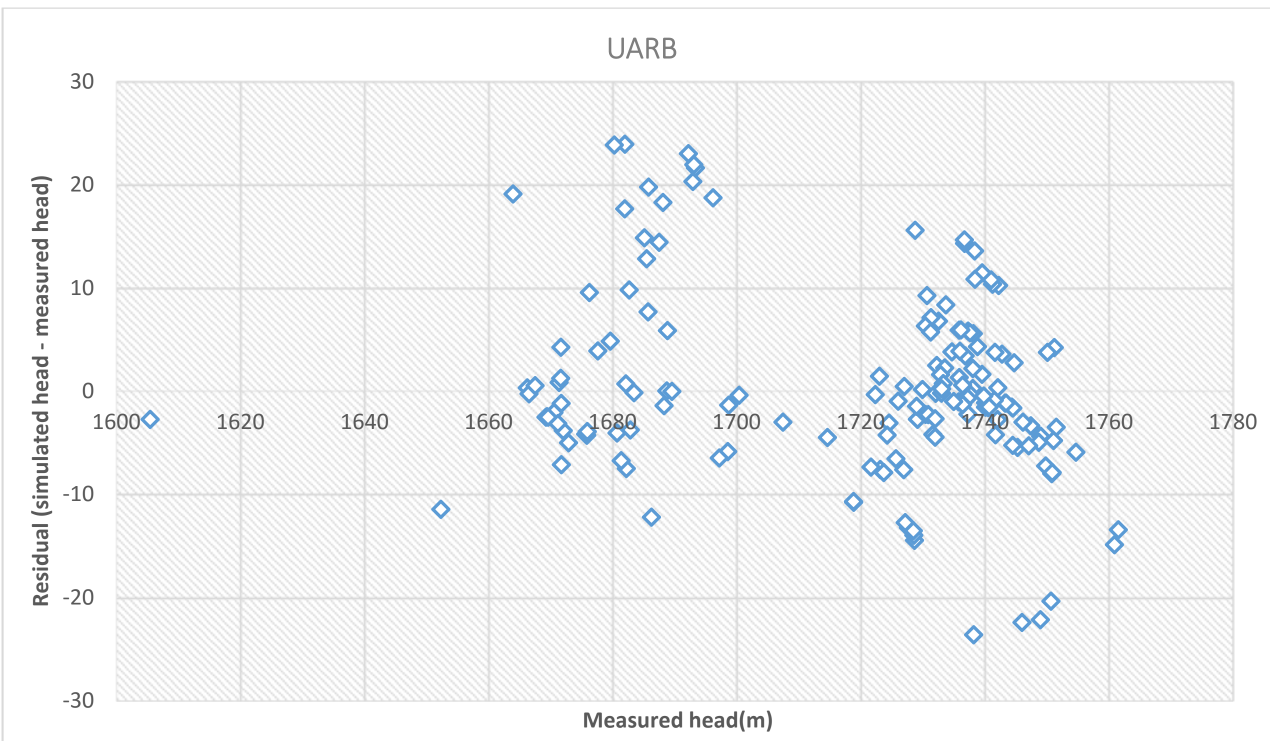

Figure 4-13 Scatter plot of Upper Awash basin

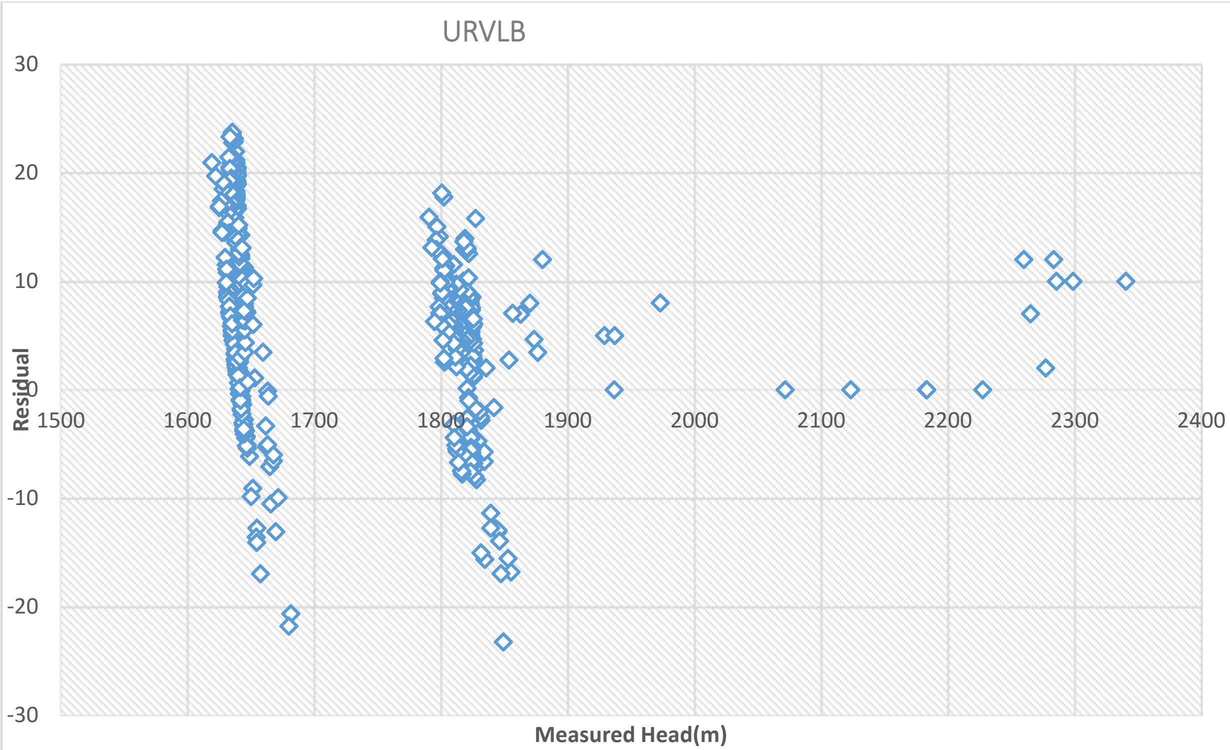

Figure 4-14 Scatter plot of the Upper Rift valley lakes basin

5 CONCLUSION AND RECOMMENDATION

5.1 Conclusion

For correct management of the existing water resources and their optimal use, groundwater potential and flow direction (groundwater basin) and suitable plans must be developed based on objectives compatible with the region. Three-dimensional models have been used extensively for groundwater potential assessment in Ethiopia and are generally adequate for predicting aquifer head changes. However, the three dimensional model described in this report was developed to identify the interbasin groundwater transfer between basins.

Groundwater flow is determined by differences in hydraulic head. Groundwater flows from areas with higher hydraulic head to areas with lower hydraulic head. In these assumed unconfined aquifers, the elevation of the water table surface is used to determine distribution of hydraulic head and indicate the direction of groundwater flow. The flow directions within the local flow system follows the topographic feature of the area. The study focuses on the groundwater head estimation in the conceptualized basins towards understanding and identifying interbasin groundwater flow in UARB and URVLB. And the study was by applying finite element numerical groundwater flow modeling software called TAGSAC. To decrease the uncertainties in the model, both field data related to wells and springs were collected and carefully selected among them by field visit and their result in the model compared to their expected result. These selected and tested wells with their related data were used to calibrate the model. The simulated finite element numerical flow model of this research was calibrated with evidence of groundwater level to fit the observed or measured groundwater levels collected in the three model basins.

To determine the hydraulic head, some constraints were applied regarding the aquifer property that needed as impute in the model. The following conditions are necessary and most likely to permit interbasin transfers in aquifers is given: Potential differences must exist between two or more basins. However, this is an insufficient condition. Adjacent basins that are unconnected are also likely to exhibit potential differences. An integrated fracture network must also exist. For flow to occur perpendicular to potential gradients, fracture heterogeneities must cancel out such that on some scale the rock is an equivalent porous medium. Fractures with aperture must exist. They have not been sealed by minerals nor are they closed due to lithostatic loads. Impermeable interbeds must be capable of sustaining and retaining open fractures or are bypassed by a tortuous flow system. Thus, interbasin flow is most likely to occur where flow is channeled by fracture systems with aperture.

Based on the objective of the thesis all factors are tested and investigated. The topography of the area (the two basins) as inferred from longitudinal profile from Ziway Lake to Koka Lake, Koka Lake is a little bit lower than Ziway Lake in elevation. And also from longitudinal profile of the surface water divide of the two basins, the middle part is lower elevation than West and East part of the area. The calibrated hydraulic head of Upper Rift valley lakes basin is higher than hydraulic head of Upper Awash Basin in the middle, Ziway-Koka lake side and in the rest part Upper Awash's head is higher. The plotted hydraulic head and groundwater flow direction supports this result. And from previous studies the geological feature of the basins near the boundary is similar and flow permitting, even if it is complex. Thus, from carefully selected and tested field parameters, and calibrated results of the model we can conclude that the surface water divide of the Upper Awash River basin and Upper Rift Valley Lakes basin does not exist for groundwater case. The model simulated groundwater flow direction is in both direction. In the middle part of it groundwater flows from Upper Rift Valley Lakes basin to Upper Awash River basin and from Upper Awash to Upper Rift Valley Lakes basin in the West and East of it.

5.2 Recommendation

This study benefited from the hydrogeological data set obtained from different sources, mainly from the development of the area and availability of groundwater in shallow depth. However, the work also faced limitations with regard to both data quantity and quality as well as to the spatial and temporal coverage of hydrological and hydrogeological information. To better understand and increase the knowledge on the hydrogeological framework of a complex system, proper utilization of existing records and collecting new data having enough information in both temporal and spatial coverage is very crucial.

Groundwater monitoring has been given a very little and/or no attention in the country. But it is a very essential step towards evaluating and managing groundwater resources. Although numerous boreholes were constructed by different bodies (governmental, nongovernmental organizations, individuals, firms, etc.), a central groundwater database, either in a national and/or regional level is not established in the country to organize the records generated during the construction of these wells. Data are collected in a non-standardized and nonsystematic manner, most of the time not reliable to use for research and/or scientific purposes. Within the study area, there are no observation wells (non-pumping wells) that are continuously monitored and evenly distributed in the catchment. Monitoring of water levels, at least in each hydrogeological zone, would provide insight and valuable calibration points for future refinement of the model.

Bedrock interbasin flow may occur wherever potential and integrated permeability permit. However, as the spatial dimensions of an aquifer system increase, the probability of an integrated flow network being present in bedrock may decline. In the absence of compelling evidence to the contrary, the complex geology of regions like the Rift valley suggests that the first working hypothesis adopted should investigate local recharge through the mechanisms of mountain front recharge, mountain block recharge, and losses from both intermittent and perennial streams. From the hydraulic point of view, rock masses are heterogeneous, anisotropous and discontinuous media. As water flow in rocks occurs mainly along discontinuities, the exact knowledge of their distribution and their characteristic parameters is fundamental to find the features that describe the fluid flow.

Future studies should consider a variety of improvements to the existing models, including additional data collection, different conceptual model design, and other factors. The current model assumes that recharge is distributed uniformly throughout the study basin; however, this is unrealistic. Future modeling should consider focusing recharge at different points on the basis of field data, and sensitivity of model output to various distributions should be examined. Future studies should improve the reliability of the head data by accurately locating wells and measuring the surface elevation. A greater number and wider distribution of head measurements would also improve parameterization of the model. Future modeling should consider representing the aquifer as two layers that would allow vertical variation in hydraulic properties.

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APPENDIX 1 Wells used in Upper Awash River basin model

X	Y	Static_lev(m)	Depth(m)	Elevation	Scheme_type	Well Id.
465036	916680	65	66	1815.794	Borehole	2
465489	916508	62	63	1811.387	Borehole	3
465769	917544	64	65	1812.202	Borehole	4
465781	916637	62	63	1804.321	Borehole	5
465848	916860	63	64	1803.029	Borehole	6
465866	916066	63	65	1797.286	Borehole	7
465951	916212	64	65	1798.96	Borehole	8
466084	916509	62	63	1798.995	Borehole	9
466106	917544	64	65	1804.724	Borehole	10
466330	916016	49	50	1790.635	Borehole	11
466334	917358	62	63	1797.91	Borehole	12
466431	917259	52	53	1797.303	Borehole	13
466433	918795	38	40	1799.602	Borehole	14
466449	913966	49	50	1786.619	Borehole	15
466537	916375	49	50	1788.761	Borehole	16
466541	918624	38	40	1792.249	Borehole	17
466566	916177	49	50	1786.201	Borehole	18
466641	918933	38	40	1797.298	Borehole	19
466725	916544	49	50	1783.688	Borehole	20
466755	916653	50	51	1787.144	Borehole	21
466756	916157	49	50	1784.071	Borehole	22
466779	916625	50	51	1786.999	Borehole	23
466789	918030	52	53	1790.822	Borehole	24
466808	916131	49	50	1782.47	Borehole	25
466881	915903	49	50	1785.189	Borehole	26
466902	918560	38	40	1790.807	Borehole	27
466922	916235	49	50	1784.737	Borehole	28
466959	916353	44	45	1783.325	Borehole	29
466987	917612	52	53	1787.853	Borehole	30
467022	920660	47	48	1796.314	Borehole	31
467034	916656	37	38	1782.827	Borehole	32
467093	915494	44	45	1777.737	Borehole	34
467109	916333	37	38	1781.465	Borehole	35
467127	915593	37	38	1778.761	Borehole	36
467143	918396	38	40	1778.194	Borehole	37
467174	920541	47	48	1796.413	Borehole	38
467183	919934	46	47	1792.031	Borehole	39
467252	917792	42	42	1782.634	Borehole	40

X	Y	Static_lev(m)	Depth(m)	Elevation	Scheme_type	Well Id.
467309	920281	46	47	1788.998	Borehole	42
467321	917831	42	42	1782.488	Borehole	43
467357	920484	47	48	1789.085	Borehole	44
467360	916583	48	50	1777.634	Borehole	45
467363	916324	44	45	1779.708	Borehole	46
467430	917828	52	52	1783.324	Borehole	48
467440	920432	47	48	1782.603	Borehole	50
467448	918603	38	40	1778.783	Borehole	51
467522	917672	44	44	1779.143	Borehole	52
467526	920773	47	48	1796.548	Borehole	53
467536	918578	38	40	1777.881	Borehole	54
467551	917898	43	44	1782.408	Borehole	55
467552	920198	46	47	1789.027	Borehole	56
467558	917340	43	44	1780.6	Borehole	58
467599	920181	46	47	1789.541	Borehole	59
467617	917470	42	43	1775.474	Borehole	60
467653	918194	42	43	1778.289	Borehole	61
467656	920080	46	47	1787.67	Borehole	62
467702	919292	44	45	1781.194	Borehole	63
467721	919447	44	45	1781.69	Borehole	64
467749	919746	44	45	1786.167	Borehole	65
467751	920767	47	48	1793.254	Borehole	66
467756	920124	46	47	1785.996	Borehole	67
467819	918550	38	40	1777.364	Borehole	68
467828	919679	44	45	1785.015	Borehole	69
467868	919667	44	45	1784.943	Borehole	70
467978	919640	44	45	1783.515	Borehole	71
468041	918259	42	43	1772.414	Borehole	72
468049	919567	44	45	1780.83	Borehole	73
468088	921262	57	58	1788.345	Borehole	74
468092	919548	44	45	1783.044	Borehole	75
468099	919499	44	45	1782.865	Borehole	76
468160	918130	42	43	1773.954	Borehole	77
468259	918130	42	43	1772.347	Borehole	78
468281	919283	44	45	1779.486	Borehole	79
468309	918085	39	40	1771.367	Borehole	80
468355	917616	39	40	1770.261	Borehole	81
468356	917615	39	40	1770.224	Borehole	82
468360	919310	44	45	1777.517	Borehole	83
468369	919348	44	45	1777.043	Borehole	84

X	Y	Static_lev(m)	Depth(m)	Elevation	Scheme_type	Well Id.
468380	917870	42	43	1770.086	Borehole	85
468385	917871	39	40	1770.378	Borehole	86
468423	919292	44	45	1776.088	Borehole	87
468471	917747	42	43	1769.863	Borehole	88
468474	919237	44	45	1776.713	Borehole	89
468501	919379	44	45	1779.278	Borehole	91
468586	914006	58	60	1754.052	Borehole	92
468623	917466	39	40	1763.082	Borehole	93
468802	914769	58	60	1751.749	Borehole	94
468869	915293	58	60	1750.239	Borehole	95
468913	915031	58	60	1748.842	Borehole	98
469182	921023	55	56	1785.271	Borehole	101
469265	920755	55	56	1780.31	Borehole	102
469357	915132	55	56	1744.665	Borehole	104
469394	919027	39	40	1765.904	Borehole	105
469400	918676	39	40	1758.606	Borehole	106
469436	918634	39	40	1757.504	Borehole	107
469505	919069	39	40	1768.518	Borehole	110
469506	921049	59	60	1785.381	Borehole	111
469516	920606	54	55	1780.851	Borehole	113
469566	919126	39	40	1769.419	Borehole	116
469587	918860	39	40	1764.742	Borehole	118
469610	918955	39	40	1766.354	Borehole	119
469623	919162	39	40	1769.176	Borehole	121
469720	921069	59	60	1786.176	Borehole	125
469861	919358	48	50	1769.209	Borehole	131
470057	919064	48	50	1764.437	Borehole	140
470102	917540	57	58	1742.38	Borehole	141
470139	917740	59	60	1745.427	Borehole	142
470366	917359	57	58	1743.282	Borehole	144
470455	918632	44	45	1744.253	Borehole	145
470497	920960	59	60	1779.495	Borehole	146
470516	917586	57	58	1742.741	Borehole	147
470518	921008	59	60	1779.736	Borehole	148
470608	921167	64	65	1779.892	Borehole	149
470632	918360	44	45	1744.873	Borehole	150
470765	917179	54	55	1740.203	Borehole	151
470804	917359	53	54	1738.139	Borehole	152
470928	917515	51	52	1735.532	Borehole	153
470996	921168	73	75	1775.454	Borehole	154

X	Y	Static_lev(m)	Depth(m)	Elevation	Scheme_type	Well Id.
471096	918387	44	45	1737.3	Borehole	155
471220	918655	52	53	1739.041	Borehole	157
471233	917804	50	51	1740.871	Borehole	158
471339	918757	51	52	1741.807	Borehole	159
471443	917547	57	58	1735.882	Borehole	160
471826	914299	59	60	1728.186	Borehole	161
471847	919943	66	68	1749.633	Borehole	162
471902	918428	53	54	1736.728	Borehole	163
471912	917269	52	53	1729.762	Borehole	164
471920	917099	49	50	1732.031	Borehole	165
472016	914319	59	60	1731.738	Borehole	166
472104	913607	64	65	1730.073	Borehole	167
472109	914090	58	59	1731.859	Borehole	168
472134	917321	50	51	1731.207	Borehole	169
472144	913359	61	62	1728.322	Borehole	170
472166	911533	59	60	1729.879	Borehole	171
472304	913683	61	62	1732.779	Borehole	172
472340	919817	63	65	1741.511	Borehole	173
472404	920086	64	66	1745.858	Borehole	174
472461	914522	59	60	1730.235	Borehole	175
472494	914900	59	60	1734.108	Borehole	176
472551	913509	64	65	1729.648	Borehole	177
472673	914413	59	60	1729.748	Borehole	178
472675	914108	59	60	1728.96	Borehole	179
472977	918184	59	60	1731.113	Borehole	180
473150	916578	53	54	1731.963	Borehole	181
473153	916649	53	54	1730.69	Borehole	182
473270	916803	53	54	1731.317	Borehole	183
473372	917152	59	60	1729.992	Borehole	184
473596	917933	47	48	1734.361	Borehole	185
478200	920000	60	75	1727		190
493500	916800	62	93	1657		198
502000	925000	0	0	1606		200
507000	901000	128	160	1794	Borehole	201
508725	897993	160	190	1808.919	Borehole	202

APPENDIX 2 Wells used in Upper Rift valley Lakes basin model

X	Y	Static_lev(m)	Depth(m)	Elevation(a.s.l.)	Scheme_typ	Well Id.
444568	908532	12	15	1887.147	Hand dug well	203
444616	908834	13	15	1890.756	Hand dug well	204
444632	909323	10	12	1889.048	Hand dug well	205
444672	908532	10	11	1886.82	Hand dug well	206
444828	911727	12	48	1887.758	Borehole	207
445312	900101	3	14	1843.033		208
445418	907156	12	13	1875.665	Hand dug well	209
445562	897084	16	18	1837.601	Hand dug well	210
445822	899038	19	21	1840.001	Borehole	211
445844	906478	12	16	1876.342	Hand dug well	212
445870	898609	20	21	1845.503	Borehole	214
445889	897297	10	12	1827.42	Hand dug well	215
445893	897063	10	12	1828.531	Hand dug well	216
445951	898902	19	21	1846.986	Borehole	217
446004	897056	10	12	1830.674	Hand dug well	218
446541	899018	21	23	1843.77	Borehole	219
446599	898940	21	23	1842.499	Borehole	220
446606	898860	21	23	1843.741	Borehole	221
446628	898402	0	0			222
446649	899229	21	23	1844.737	Borehole	223
446657	898705	21	23	1844.049	Borehole	224
446681	898979	21	23	1843.36	Borehole	225
446691	898193	10	21	1839.997	Borehole	226
446697	899186	21	23	1844.085	Borehole	227
446701	898292	10	22	1843.566	Borehole	228
446711	898367	20	23	1845.62	Borehole	229
446750	897642	22	40	1841.854	Borehole	230
446760	899773	27	30	1846.722	Borehole	231
446771	899182	0	0			232
446781	899010	19	21	1842.598	Borehole	233
446789	899224	21	23	1846.651	Borehole	234
446791	899390	20.5	22	1847.135	Borehole	235
446805	904400	27	36	1856.199	Borehole	236
446829	899125	21	23	1845.166	Borehole	237
446839	900563	27	30	1845.977	Borehole	238
446847	899173	21	23	1843.838	Borehole	239
446858	898994	21	23	1843.901	Borehole	240
446865	900377	27	30	1849.868	Borehole	241

X	Y	Static_lev(m)	Depth(m)	Elevation(a.s.l.)	Scheme_typ	Well Id.
446866	898251	19	22	1843.613	Borehole	242
446872	897959	20	23	1839.583	Borehole	243
446878	897793	20	23	1838.104	Borehole	244
446896	904607	14	36	1853.858	shallow well	245
446916	898579	21	23	1843.635	Borehole	246
446922	898011	19	22	1841.153	Borehole	247
446938	898435	21	23	1843.048	Borehole	248
446947	906647	18	54	1870.37	Borehole	249
446950	898354	21	23	1843.347	Borehole	250
446963	898057	19	22	1841.29	Borehole	251
446968	897788	20	23	1840.067	Borehole	252
446983	898481	21	23	1843.909	Borehole	253
446986	898619	21	23	1844.232	Borehole	254
446995	898979	21	23	1846.097	Borehole	255
447011	899017	20	23	1846.052	Borehole	256
447028	897849	21	24	1842.751	Borehole	257
447042	906854	18	54	1870.489	Shallow well	258
447063	898545	21	23	1843.098	Borehole	259
447073	898520	21	23	1843.758	Borehole	260
447078	898941	23	25	1843.611	Borehole	261
447084	899047	21	25	1844.08	Borehole	262
447085	898127	20	23	1840.96	Borehole	263
447087	898484	21	23	1845.219	Borehole	264
447090	900172	27	30	1849.077	Borehole	265
447097	899986	24	27	1849.303	Borehole	266
447099	898608	21	23	1843.689	Borehole	267
447103	900098	29	32	1849.703	Borehole	268
447104	900361	29	32	1851.846	Borehole	269
447111	897757	21	24	1839.754	Borehole	270
447116	897800	21	24	1839.183	Borehole	271
447121	898829	19	23	1843.201	Borehole	272
447129	897854	21	24	1839.119	Borehole	273
447145	900945	34	43	1852.97	Borehole	274
447154	898631	23	23	1842.897	Borehole	275
447169	898993	19	22	1845.272	Borehole	276
447175	898957	23	25	1845.019	Borehole	277
447177	901885	0	0			278
447178	898780	19	23	1842.602	Borehole	279
447183	901074	26	35	1855.943	Borehole	280
447194	901149	32	40	1855.304	Borehole	281

X	Y	Static_lev(m)	Depth(m)	Elevation(a.s.l.)	Scheme_typ	Well Id.
447197	899555	23	25	1848.201	Borehole	282
447207	899507	23	25	1846.743	Borehole	283
447227	899421	26	28	1846.46	Borehole	284
447232	905062	18	21	1864.377	Borehole	285
447233	901362	29	37	1853.642	Borehole	286
447234	905193	19	22	1864.158	Borehole	287
447278	900417	29	32	1851.004	Borehole	288
447283	907116	22	23	1880.691	Borehole	289
447297	898293	20	23	1842.014	Borehole	290
447299	899869	29	32	1851.654	Borehole	291
447323	900059	31	34	1850.934	Borehole	292
447327	903220	30	42	1860.574	Borehole	293
447379	901426	26	35	1858.425	Borehole	294
447388	899815	24	25	1850.992	Borehole	295
447390	899568	28	30	1848.387	Borehole	296
447413	905158	19	22	1866.402	Borehole	297
447419	899628	25	28	1849.158	Borehole	298
447423	899629	21	23	1849.086	Borehole	299
447469	904449	39	52	1861.639	Borehole	300
447480	899778	0	0			301
447486	904370	23	25	1864.988	Borehole	302
447493	902918	30	42	1862.418	Borehole	303
447562	904659	30	52	1861.665	shallow well	304
447655	904660	26	30	1860.674	Borehole	305
447690	899511	28	30	1845.987	Borehole	306
447884	899514	26	28	1850.852	Borehole	307
447916	899234	32	33	1846.444	Borehole	308
448058	904643	38	40	1870	Borehole	310
448158	899502	32	35	1846.723	Borehole	312
448197	898183	21	25	1840.271	Borehole	313
448236	899492	32	33	1849.951	Borehole	314
448239	899493	39	42	1850.04	Borehole	315
448362	896251	8	22	1829.187	Borehole	316
448387	899842	28	30	1849.534	Borehole	317
448388	898193	22	24	1840.667	Borehole	318
448394	896300	10	22	1831.347	Borehole	319
448397	899923	37	40	1850.529	Borehole	320
448402	896358	11	22	1831.809	Borehole	321
448437	899180	32	33	1843.151	Borehole	322
448493	900142	37	40	1850.131	Borehole	323

X	Y	Static_lev(m)	Depth(m)	Elevation(a.s.l.)	Scheme_typ	Well Id.
448511	900251	37	40	1851.686	Borehole	324
448592	900291	37	40	1853.559	Borehole	326
448606	896477	8	21	1830.842	Borehole	327
448608	896401	6	21	1831.339	Borehole	328
448664	896386	5	21	1831.425	Borehole	329
448674	899951	37	81	1851.234	Borehole	330
448691	896977	0	0			331
448707	896436	7	21	1832.111	Borehole	332
448709	898787	21	23	1841.598	Borehole	333
448716	899781	34	37	1855.918	Borehole	334
448738	896933	10	22	1836.539	Borehole	335
448768	896979	9	22	1836.738	Borehole	336
448801	896707	9	21	1833.846	Borehole	337
448805	896350	10	22	1831.53	Borehole	338
448806	896799	9	22	1835.004	Borehole	339
448815	896988	9	22	1835.161	Borehole	340
448823	897111	9	21	1835.273	Borehole	341
448824	896294	9	21	1832.72	Borehole	342
448835	896615	10	22	1831.838	Borehole	343
448839	896746	10	22	1834.974	Borehole	344
448848	896649	10	22	1832.9	Borehole	345
448849	896963	10	22	1833.898	Borehole	346
448879	897172	10	21	1836.465	Borehole	348
448882	896963	10	22	1832.084	Borehole	349
448900	896731	10	22	1833.994	Borehole	350
448910	897084	9	21	1835.651	Borehole	351
448913	897116	9	21	1835.003	Borehole	352
448919	897053	9	21	1834.803	Borehole	353
448937	896993	10	22	1831.274	Borehole	354
448953	900604	44	92	1856.457	Borehole	355
448962	897008	9	21	1831.995	Borehole	356
448965	897181	10	21	1833.004	Borehole	357
448969	896697	9	22	1832.924	Borehole	358
448972	896644	10	22	1833.136	Borehole	359
448981	898759	31	33	1840.926	Borehole	360
448985	897139	9	21	1834.292	Borehole	361
448988	896860	8	21	1833.554	Borehole	362
448991	896617	10	22	1832.991	Borehole	363
449028	896746	10	22	1833.573	Borehole	364
449066	898968	0	0			365

X	Y	Static_lev(m)	Depth(m)	Elevation(a.s.l.)	Scheme_typ	Well Id.
449083	900047	41	45	1855.064	Borehole	366
449102	897063	10	22	1832.108	Borehole	367
449164	897152	9	22	1833.372	Borehole	368
449179	897048	12	22	1830.623	Borehole	369
449214	913791	5	22	1929.351	shallow well	370
449236	920325	8	158	1977.785		371
449267	897256	0	0			372
449288	896766	20	22	1833.898	Borehole	373
449349	897636	22	25	1832.519	Borehole	374
449374	897291	8	21	1830.973	Borehole	375
449446	897517	22	25	1833.515	Borehole	376
449460	897203	21	24	1831.895	Borehole	377
449489	897231	21	24	1829.795	Borehole	378
449509	897573	24	27	1832.803	Borehole	379
449521	897304	21	24	1832.051	Borehole	380
449528	898013	25	28	1832.172	Borehole	381
449539	897325	21	24	1832.715	Borehole	382
449596	897559	22	26	1834.395	Borehole	383
449600	897368	23	26	1832.486	Borehole	384
449601	897793	26	30	1835.565	Borehole	385
449605	897892	26	29	1832.841	Borehole	386
449616	896845	24	22	1828.014	Borehole	387
449624	896814	19	21	1828.532	Borehole	388
449626	897717	27	30	1832.959	Borehole	389
449655	897301	25	28	1829.495	Borehole	390
449658	897614	24	27	1833.417	Borehole	391
449686	897515	21	24	1832.9	Borehole	392
449691	897805	24	27	1832.438	Borehole	393
449723	897353	20	23	1830.151	Borehole	394
449734	897391	19	22	1831.542	Borehole	395
449753	896733	20	23	1826.859	Borehole	396
449757	897534	21	24	1833.027	Borehole	397
449775	897429	21	24	1831.281	Borehole	398
449814	918599	0	0			399
449837	900975	27.74	32	1871.609	Borehole	400
449872	898088	28	31	1836.497	Borehole	401
449876	898018	21	26	1830.565	Borehole	402
449877	897680	25	29	1831.26	Borehole	403
449879	897971	22	24	1831.039	Borehole	404
449880	900368	15	27	1862.976	Borehole	405

X	Y	Static_lev(m)	Depth(m)	Elevation(a.s.l.)	Scheme_typ	Well Id.
449881	897487	21	30	1830.895	Borehole	406
449892	915175	5	26	1951.588	shallow well	407
449928	914613	0	0			408
449988	898064	29	32	1832.344	Borehole	409
450017	898166	25	27	1833.869	Borehole	410
450055	897637	25	27	1831.523	Borehole	411
450072	897771	26	30	1830.041	Borehole	412
450074	897589	22	25	1832.166	Borehole	413
450096	897711	25	27	1831.46	Borehole	414
450105	897998	28	31	1833.432	Borehole	415
450220	898034	29	32	1829.734	Borehole	416
450237	897946	29	32	1829.237	Borehole	417
450250	900045	48	52	1846.166	Borehole	418
450401	898054	31	35	1829.135	Borehole	419
450410	897173	25	28	1827.207	Borehole	420
450455	899308	34	38	1834.029	Borehole	421
450643	897650	10	23	1826.638	Borehole	422
450686	899286	29	32	1833.11	Borehole	423
450810	899443	29	30	1832.993	Borehole	424
450944	899653	31	35	1834.922	Borehole	425
451020	899975	31	36	1835.781	Borehole	426
451060	899258	32	33	1832.204	Borehole	427
451089	900260	36	37	1836.112	Borehole	428
451189	900153	34	38	1834.574	Borehole	429
451218	898980	26	27	1828.403	Borehole	430
451247	899012	26	27	1828.928	Borehole	431
451265	899121	27	28	1827.873	Borehole	432
451280	900012	31	32	1831.087	Borehole	433
451303	900169	34	37	1833.376	Borehole	434
451396	899171	27	28	1826.338	Borehole	435
451431	899136	24	27	1823.073	Borehole	436
451527	899462	10	22	1825.136	Borehole	437
451681	899430	10	22	1826.604	Borehole	438
451762	899730	10	22	1826.544	Borehole	439
451980	899893	10	22	1825.009	Borehole	440
452104	899942	9	22	1824.024	Borehole	441
452146	900119	12	25	1823.789	Borehole	442
452152	899808	9	22	1820.746	Borehole	443
452165	900015	9	22	1821.361	Borehole	444
452177	899632	22	25	1820.788	Borehole	445

X	Y	Static_lev(m)	Depth(m)	Elevation(a.s.l.)	Scheme_typ	Well Id.
452179	899679	19	22	1820.188	Borehole	446
452242	899735	21	24	1822.45	Borehole	447
452262	899759	21	24	1822.575	Borehole	448
452284	899812	20	23	1821.069	Borehole	449
452289	899779	21	24	1822.984	Borehole	450
452291	900022	30	52	1823.779	Borehole	451
452324	899826	20	23	1819.803	Borehole	452
452338	900324	11	24	1824.176	Borehole	453
452383	899910	25	28	1816.716	Borehole	454
452438	900382	10	22	1822.824	Borehole	455
452614	900807	22	25	1823.653	Borehole	456
452637	900874	22	25	1824.069	Borehole	457
453276	901204	21.85	79	1814.091	Borehole	459
453905	920416	0	0			460
455654	915432	0	0			462
455782	918927	0	0			463
456111	915948	7	12	2269.587	Hand dug well	464
456444	915918	12	15	2277.663	Hand dug well	465
456545	915935	10	15	2288.009	Hand dug well	466
456574	915654	2	20	2288.856	Hand dug well	467
456727	915537	12	14	2308.697	Hand dug well	468
458677	926642	0	0			470
458790	929689	10	18	2298.419	Hand dug well	472
479685	906614	26	28	1708.634	Borehole	482
479789	901272	26	28	1660.529	Borehole	483
479811	901213	26	28	1660.14	Borehole	485
479925	901022	26	28	1662.131	Borehole	490
480189	901780	29	30	1663.405	Borehole	503
480211	901849	26	28	1665.03	Borehole	505
480331	901424	26	28	1665.864	Borehole	509
480387	901435	30	54			516
480448	901282	26	28	1665.374	Borehole	518
480482	900806	25	27	1659.414	Borehole	521
480508	901166	29	30	1663.944	Borehole	522
480529	901369	26	28	1662.236	Borehole	524
480533	901365	26	28	1662.436	Borehole	525
480547	901190	29	30	1663.113	Borehole	526
480552	901261	26	28	1663.649	Borehole	527
480564	901344	26	28	1664.676	Borehole	528
480565	901369	26	28	1664.226	Borehole	529

X	Y	Static_lev(m)	Depth(m)	Elevation(a.s.l.)	Scheme_typ	Well Id.
480569	901471	26	28	1665.839	Borehole	530
480570	901272	26	28	1663.921	Borehole	531
480574	901246	26	28	1662.851	Borehole	532
480585	901148	29	30	1664.42	Borehole	534
480594	901442	26	28	1666.339	Borehole	536
480596	901420	26	28	1666.39	Borehole	537
480601	901463	26	28	1666.51	Borehole	538
480603	901346	26	28	1666.544	Borehole	539
480606	901660	26	28	1665.636	Borehole	540
480608	901142	29	30	1664.885	Borehole	541
480609	901539	26	28	1664.112	Borehole	542
480610	901307	26	28	1665.477	Borehole	543
480621	901467	26	28	1666.874	Borehole	544
480626	901526	26	28	1664.245	Borehole	545
480628	901483	26	28	1666.255	Borehole	546
480637	901543	26	28	1663.582	Borehole	547
480641	901512	28	30	1664.888	Borehole	548
480644	901581	26	28	1662.725	Borehole	549
480646	901286	26	28	1664.906	Borehole	550
480658	901041	29	30	1665.171	Borehole	551
480659	901295	26	28	1664.969	Borehole	552
480660	901313	26	28	1665.01	Borehole	553
480663	901428	26	28	1666.992	Borehole	554
480665	901761	26	28	1663.797	Borehole	555
480667	901365	26	28	1664.917	Borehole	556
480668	901434	26	28	1667.037	Borehole	557
480681	900964	29	30	1663.329	Borehole	558
480686	901431	26	28	1665.986	Borehole	559
480689	901346	28	30	1664.458	Borehole	560
480714	901384	26	28	1664.637	Borehole	561
480715	901244	26	28	1665.03	Borehole	562
480717	901789	26	28	1666.169	Borehole	563
480719	901436	26	28	1664.631	Borehole	564
480720	901499	26	28	1665.37	Borehole	565
480725	901383	28	30	1664.778	Borehole	566
480728	901362	26	28	1664.795	Borehole	567
480730	901325	26	28	1664.947	Borehole	568
480733	901237	26	28	1664.992	Borehole	569
480737	901381	26	28	1664.841	Borehole	570
480749	901658	26	28	1665.599	Borehole	571

X	Y	Static_lev(m)	Depth(m)	Elevation(a.s.l.)	Scheme_typ	Well Id.
480759	901409	26	28	1664.57	Borehole	572
480760	901402	28	30	1664.622	Borehole	573
480761	901822	26	28	1666.327	Borehole	574
480762	901240	26	28	1665.385	Borehole	575
480771	901208	26	28	1664.616	Borehole	576
480772	901733	26	28	1663.541	Borehole	577
480773	901348	26	28	1664.209	Borehole	578
480776	901593	26	28	1663.984	Borehole	579
480780	901284	28	30	1665.369	Borehole	580
480783	901313	26	28	1664.379	Borehole	581
480795	901060	26	28	1662.232	Borehole	582
480799	901691	26	28	1663.33	Borehole	583
480800	901672	26	28	1663.663	Borehole	584
480801	901453	26	28	1664.948	Borehole	585
480803	901738	26	28	1662.772	Borehole	586
480815	901189	26	28	1665.676	Borehole	587
480816	901764	26	28	1663.065	Borehole	588
480817	901573	26	28	1662.897	Borehole	589
480819	901060	25	27	1662.011	Borehole	590
480822	901493	26	28	1663.129	Borehole	591
480826	901831	26	28	1663.77	Borehole	592
480827	901111	26	28	1664.358	Borehole	593
480829	901773	26	28	1662.554	Borehole	594
480831	901483	26	28	1663.769	Borehole	595
480835	900888	25	27	1663.748	Borehole	596
480839	901425	26	28	1666.048	Borehole	597
480842	901524	26	28	1662.944	Borehole	598
480846	900953	25	27	1662.094	Borehole	599
480847	901576	26	28	1662.667	Borehole	600
480852	901279	28	30	1663.798	Borehole	601
480859	901672	26	28	1662.82	Borehole	602
480861	901092	25	27	1664.402	Borehole	603
480864	901772	26	28	1660.507	Borehole	604
480865	901839	26	28	1661.321	Borehole	605
480869	901334	26	28	1663.411	Borehole	606
480874	901335	26	28	1663.379	Borehole	607
480882	901578	26	28	1662.784	Borehole	608
480885	901389	26	28	1664.191	Borehole	609
480886	901432	26	28	1664.523	Borehole	610
480890	901625	26	28	1662.859	Borehole	611

X	Y	Static_lev(m)	Depth(m)	Elevation(a.s.l.)	Scheme_typ	Well Id.
480891	901696	26	28	1662.83	Borehole	612
480892	901193	25	27	1665.802	Borehole	613
480894	901544	26	28	1663.284	Borehole	614
480900	901158	26	28	1665.295	Borehole	615
480904	901639	26	28	1663.847	Borehole	616
480916	901624	26	28	1663.851	Borehole	617
480921	901329	26	28	1662.9	Borehole	618
480927	901282	19	21	1663.257	Borehole	619
480932	900916	25	27	1661.817	Borehole	620
480936	901597	26	28	1663.138	Borehole	621
480940	900951	25	27	1662.178	Borehole	622
480944	900922	24	25	1661.716	Borehole	623
480952	901097	25	27	1662.939	Borehole	624
480953	901612	26	28	1663.232	Borehole	625
480959	901216	22	24	1664.886	Borehole	626
480966	901523	26	28	1662.477	Borehole	627
480976	901594	26	28	1662.921	Borehole	628
480979	901558	26	28	1662.641	Borehole	629
480983	901534	26	28	1662.334	Borehole	630
480990	901562	26	28	1662.415	Borehole	631
480992	901530	26	28	1662.118	Borehole	632
481004	901260	19	22	1664.855	Borehole	633
481009	901554	26	28	1661.874	Borehole	634
481017	900929	25	27	1658.466	Borehole	635
481018	901122	23	26	1663.436	Borehole	636
481024	901564	26	28	1661.976	Borehole	637
481040	900958	24	25	1659.124	Borehole	638
481064	901615	27	28	1663.107	Borehole	639
481072	901622	29	30	1663.323	Borehole	640
481079	901605	25	26	1663.129	Borehole	641
481085	901650	29	30	1664.016	Borehole	642
481089	901234	22	25	1664.509	Borehole	643
481091	901287	22	25	1663.552	Borehole	644
481096	901643	28	29	1664.066	Borehole	645
481098	901313	21	24	1663.752	Borehole	646
481100	901616	26	27	1663.727	Borehole	647
481115	901642	21	22	1664.318	Borehole	648
481117	901423	11	23	1664.743	Borehole	649
481128	901278	23	26	1662.674	Borehole	650
481134	901458	29	30	1664.305	Borehole	651

v	Y	Static_lev(m)	Depth(m)	Elevation(a.s.l.)	Scheme_typ	Well Id.
481138	901490	27	28	1663.841	Borehole	652
481140	901348	20	23	1664.139	Borehole	653
481159	901420	21	22	1664.315	Borehole	654
481172	901373	23	26	1664.003	Borehole	655
481203	901962	28	29	1659.935	Borehole	656
481206	901221	21	24	1663.155	Borehole	657
481208	901284	21	24	1663.868	Borehole	658
481213	901355	24	28	1664.033	Borehole	659
481216	901543	29	30	1664.516	Borehole	660
481220	901360	22	25	1664.004	Borehole	661
481266	901356	21	24	1664.454	Borehole	662
481271	901428	29	30	1665.584	Borehole	663
481283	901840	27	28	1663.808	Borehole	664
481288	901484	29	30	1664.911	Borehole	665
481291	901367	29	30	1664.967	Borehole	666
481292	901358	29	30	1664.736	Borehole	667
481295	901826	27	28	1664.248	Borehole	668
481300	901951	28	29	1663.053	Borehole	669
481307	901884	29	30	1663.919	Borehole	670
481315	901923	30	31	1663.991	Borehole	671
481319	901520	29	30	1663.069	Borehole	672
481326	901972	29	30	1662.316	Borehole	673
481328	901982	29	30	1661.861	Borehole	674
481334	901893	26	27	1664	Borehole	675
481336	901503	22	24	1663.362	Borehole	676
481340	901776	28	29	1663.962	Borehole	677
481341	901902	26	27	1663.962	Borehole	678
481342	901849	28	29	1663.989	Borehole	679
481350	901883	26	27	1663.901	Borehole	680
481351	901705	29	30	1662.67	Borehole	681
481353	901994	29	30	1661.607	Borehole	682
481358	901621	27	28	1663.201	Borehole	683
481364	901891	27	28	1663.873	Borehole	684
481366	901800	27	28	1663.686	Borehole	685
481374	901534	27	28	1663.291	Borehole	686
481381	901784	27	28	1663.248	Borehole	687
481382	901904	27	28	1663.72	Borehole	688
481391	901929	27	28	1663.65	Borehole	689
481394	901889	27	28	1663.342	Borehole	690
481401	901989	28	29	1663.519	Borehole	691

X	Y	Static_lev(m)	Depth(m)	Elevation(a.s.l.)	Scheme_typ	Well Id.
481407	901635	28	30	1662.149	Borehole	692
481413	901789	27	28	1661.986	Borehole	693
481414	901450	23	24	1664.518	Borehole	694
481417	901796	27	28	1661.821	Borehole	695
481418	901858	26	27	1661.91	Borehole	696
481423	901938	27	28	1663.381	Borehole	697
481426	902006	21	28	1662.203	Borehole	698
481437	901506	22	24	1664.104	Borehole	699
481439	901968	27	28	1663.231	Borehole	700
481444	901812	27	28	1661.253	Borehole	701
481449	901801	27	28	1661.436	Borehole	702
481459	901581	20	22	1665.351	Borehole	703
481464	901829	27	28	1661.079	Borehole	704
481481	901914	26	27	1662.582	Borehole	705
481482	901777	26	27	1662.763	Borehole	706
481485	901903	26	27	1662.115	Borehole	707
481486	901520	26	28	1665.942	Borehole	708
481487	901545	29	30	1666.273	Borehole	709
481490	901825	27	28	1661.191	Borehole	710
481491	901602	26	28	1665.798	Borehole	711
481496	901852	26	27	1661.008	Borehole	712
481498	901837	26	27	1660.998	Borehole	713
481501	901698	29	30	1663.536	Borehole	714
481502	901710	27	28	1663.899	Borehole	715
481511	901868	26	27	1661.151	Borehole	716
481519	901857	26	27	1661.089	Borehole	717
481522	901199	20	21	1662.288	Borehole	718
481535	901713	29	30	1664.009	Borehole	719
481540	901828	26	28	1660.866	Borehole	720
481545	901223	20	21	1662.599	Borehole	721
481556	901734	28	29	1663.739	Borehole	722
481564	901087	20	21	1656.134	Borehole	723
481565	901723	28	29	1663.556	Borehole	724
481567	901813	29	30	1660.902	Borehole	725
481577	901824	29	30	1661.002	Borehole	726
481584	901769	28	29	1662.901	Borehole	727
481618	901972	28	29	1661.184	Borehole	728
481655	901304	21	22	1665.307	Borehole	729
481660	901840	29	30	1662.202	Borehole	730
481662	901748	29	30	1663.23	Borehole	731

X	Y	Static_lev(m)	Depth(m)	Elevation(a.s.l.)	Scheme_typ	Well Id.
481664	901828	29	30	1662.753	Borehole	732
481669	901896	28	29	1661.457	Borehole	733
481709	901463	21	22	1664.497	Borehole	734
481732	901624	29	30	1663.121	Borehole	735
481733	901118	20	21	1652.96	Borehole	736
481778	901471	21	22	1663.163	Borehole	737
482795	901210	10	12	1646.313	Hand dug well	738
482931	900691	10	12	1647.685	Hand dug well	739
483003	901187	10	12	1647.4	Hand dug well	740
483047	903936	34	35	1666.374	Borehole	741
483062	900671	10	12	1646.242	Hand dug well	742
483065	901027	10	12	1644.533	Hand dug well	743
483087	901174	10	12	1646.313	Hand dug well	744
483091	900714	10	12	1645.96	Hand dug well	745
483092	900637	10	12	1646.113	Hand dug well	746
483134	901272	10	12	1649.676	Hand dug well	747
483196	901158	10	12	1647.532	Hand dug well	748
483201	901083	10	12	1647.565	Hand dug well	749
483263	901326	10	12	1650.026	Hand dug well	750
483341	900739	10	12	1644.836	Hand dug well	751
483353	901081	10	12	1648.03	Hand dug well	752
483360	903910	34	35	1668.664	Borehole	753
483375	900889	10	12	1642.702	Hand dug well	754
483402	901388	10	12	1649.352	Hand dug well	755
483417	901179	10	12	1651.303	Hand dug well	756
483453	901088	10	12	1650.174	Hand dug well	757
483459	901495	10	12	1650.725	Hand dug well	758
483536	901491	10	12	1650.296	Hand dug well	759
483541	901202	20	25	1651.06	Borehole	760
483580	901436	10	12	1649.799	Hand dug well	761
483584	901212	10	12	1652.522	Hand dug well	762
483621	900970	10	12	1645.662	Hand dug well	763
483643	901410	20	25	1646.578	Borehole	764
483656	901073	10	12	1647.4	Hand dug well	765
483670	900965	10	12	1644.556	Hand dug well	766
483719	901145	10	12	1645.008	Hand dug well	767
483723	904316	38	40	1664.042	Borehole	768
483741	904274	38	40	1665.194	Borehole	769
483804	901225	10	12	1649.693	Hand dug well	770
483833	904045	34	52			771

X	Y	Static_lev(m)	Depth(m)	Elevation(a.s.l.)	Scheme_typ	Well Id.
483931	904330	34	35	1664.358	Borehole	772
483933	904334	33	35	1664.495	Shallow well	773
484138	900957	10	12	1647.03	Hand dug well	774
484168	900862	10	12	1644.415	Hand dug well	775
484351	900521	10	12	1649.25	Hand dug well	776
484390	900301	10	12	1650.494	Hand dug well	777
484396	900515	10	12	1648.101	Hand dug well	778
484504	900653	10	12	1647.44	Hand dug well	779
484506	900944	10	12	1649.05	Hand dug well	780
484565	900982	10	12	1645.003	Hand dug well	781
484573	900746	10	12	1647.142	Hand dug well	782
484596	896941	11	12	1647.287	Hand dug well	783
484643	897679	11	12	1649.325	Hand dug well	784
484688	897349	11	12	1647.533	Hand dug well	785
484741	897834	11	12	1645.94	Hand dug well	786
484749	900172	11	12	1647.781	hand dug well (pump)	787
484754	896958	11	12	1647.53	Hand dug well	788
484941	896945	11	12	1644.944	Hand dug well	789
485049	900050	10	11	1645.795	Hand dug well	790
485226	902666	23	25	1664.334	Borehole	791
485237	902793	23	25	1663.065	Borehole	792
485241	901487	10	12	1646.984	hand dug well (pump)	793
485293	902621	23	25	1664.527	Borehole	794
485346	902761	23	25	1663.502	Borehole	795
485393	902817	23	25	1660.48	Borehole	796
485401	902829	23	25	1661.668	Borehole	797
485458	902755	23	25	1661.781	Borehole	798
485502	902509	23	25	1660.798	Borehole	799
485541	902678	23	25	1665.254	Borehole	800
485589	902630	23	25	1661.686	Borehole	801
485598	902647	23	25	1662.145	Borehole	803
485668	902793	23	25	1664.461	Borehole	805
485699	902799	23	25	1665.47	Borehole	806
485710	905212	58	60	1686.292	Borehole	807
485744	902815	23	25	1663.446	Borehole	808
485759	899947	10	12	1649.421	Hand dug well	809
485791	902815	23	25	1661.571	Borehole	810
485824	900078	10	11	1646.834	Hand dug well	811
485842	904853	54	76	1687.948		812

X	Y	Static_lev(m)	Depth(m)	Elevation(a.s.l.)	Scheme_typ	Well Id.
485861	900173	10	11	1645.484	Hand dug well	813
485924	900318	10	11	1645.688	Hand dug well	814
485930	903010	23	25	1663.257	Borehole	815
485943	899447	10	12	1647.209	Hand dug well	816
486025	903048	23	25	1664.771	Borehole	817
486068	903059	23	25	1664.63	Borehole	818
486100	903145	23	25	1664.249	Borehole	819
486194	903110	23	25	1665.911	Borehole	820
486290	896025	11	12	1646.882	Hand dug well	822
486310	903207	23	25	1661.385	Borehole	824
486341	898275	12	14	1643.42	Hand dug well	825
486377	896067	11	12	1646.891	Hand dug well	826
486385	896171	11	12	1644.456	Hand dug well	827
486445	896060	11	12	1644.806	Hand dug well	828
486489	903161	23	25	1666.357	Borehole	829
486511	903203	23	25	1666.399	Borehole	830
486535	898250	13	15	1642.849	Hand dug well	831
486570	899720	12	14	1641.656	Hand dug well	832
486583	899987	10	12	1644.878	Hand dug well	833
486605	895833	10	11	1647.165	Hand dug well	834
486627	895962	11	12	1643.903	Hand dug well	835
486644	902290	23	25	1658.358	Borehole	836
486657	899947	10	12	1645.521	Hand dug well	837
486667	903119	23	25	1665.1	Borehole	838
486684	898249	14	15	1643.799	Hand dug well	839
486721	903078	23	25	1666.403	Borehole	840
486740	896387	11	12	1640.854	Hand dug well	841
486752	902256	23	25	1660.348	Borehole	842
486755	898215	14	15	1644.711	Hand dug well	843
486767	902452	23	25	1662.915	Borehole	844
486768	896472	11	12	1642.464	Hand dug well	845
486783	903119	23	25	1666.95	Borehole	846
486830	895777	10	11	1644.879	Hand dug well	847
486837	899458	9	12	1646.51	Hand dug well	848
486907	895651	10	11	1640.923	Hand dug well	849
486915	902509	23	25	1662.986	Borehole	850
486938	903054	23	25	1668.015	Borehole	851
486945	895598	10	11	1641.364	Hand dug well	852
487008	899592	6	11	1644.275	Hand dug well	853
487023	896060	11	12	1644.519	Hand dug well	854

v	Y	Static_lev(m)	Depth(m)	Elevation(a.s.l.)	Scheme_typ	Well Id.
487024	899331	6	11	1642.062	Hand dug well	855
487057	904202	34	35	1685.326	Borehole	856
487069	896536	11	12	1641.694	Hand dug well	857
487070	899628	6	11	1646.043	Hand dug well	858
487074	899511	6	11	1648.291	Hand dug well	859
487111	899273	5	11	1640.087	Hand dug well	860
487170	903768	34	35	1677.151	Borehole	861
487226	895610	11	12	1642.314	Hand dug well	862
487269	895814	11	12	1644.731	Hand dug well	863
487275	895689	11	12	1647.507	Hand dug well	864
487288	902096	23	25	1661.342	Borehole	865
487313	895773	11	12	1642.602	Hand dug well	866
487339	895760	11	12	1643.481	Hand dug well	867
487358	902575	23	25	1664.601	Borehole	868
487406	899387	6	11	1646.476	Hand dug well	869
487429	902521	23	25	1664.481	Borehole	871
487474	901966	23	25	1660.872	Borehole	872
487485	901545	23	25	1652.557	Borehole	873
487504	901447	25	30	1651.834	Borehole	874
487534	903400	29	30	1669.6	Borehole	875
487597	903280	29	30	1669.576	Borehole	877
487606	901455	25	30	1653.031	Borehole	878
487614	902499	23	25	1663.078	Borehole	879
487616	903320	31	32	1671.716	Borehole	880
487617	902550	23	25	1663.73	Borehole	881
487626	904066	35	36	1687.325	Borehole	882
487638	902478	23	25	1663.844	Borehole	883
487649	903299	28	29	1670.824	Borehole	884
487737	903308	28	29	1669.044	Borehole	885
487738	902379	23	25	1662.292	Borehole	886
487766	903245	26	27	1670.783	Borehole	887
487785	902422	23	25	1664.475	Borehole	888
487807	903049	23	25	1670.988	Borehole	889
487856	902363	23	25	1665.129	Borehole	890
487861	902389	23	25	1664.698	Borehole	891
487929	902414	23	25	1664.686	Borehole	892
487936	902372	23	25	1663.994	Borehole	893
488097	901191	11	12	1652.225	Hand dug well	894
488119	903059	23	25	1672.931	Borehole	895
488150	901278	11	12	1653.766	Hand dug well	896

v	Y	Static_lev(m)	Depth(m)	Elevation(a.s.l.)	Scheme_typ	Well Id.
488218	903276	58	60	1680.91	Borehole	897
488222	902496	31	32	1669.085	Borehole	898
488255	903418	58	60	1686.269	Borehole	899
488267	902467	29	30	1669.362	Borehole	900
488289	900975	11	12	1654.204	Hand dug well	901
488336	901612	14	15	1662.453	Hand dug well	902
488337	902509	30	31	1668.493	Borehole	903
488370	902547	27	28	1668.598	Borehole	904
488394	901339	17	18	1658.391	Hand dug well	905
488448	901347	17	18	1657.847	Hand dug well	906
488539	902573	26	27	1668.721	Borehole	907
488590	902519	26	27	1668.624	Borehole	908
488632	901105	29	30	1651.979	Borehole	909
488649	902468	29	30	1668.269	Borehole	910
488735	902112	27	28	1668.576	Borehole	911
488810	902216	29	30	1668.558	Borehole	912
488835	902079	28	29	1666.332	Borehole	913
488836	899996	5	12	1649.098	Hand dug well	914
488944	902148	27	28	1665.34	Borehole	915
488961	902086	27	28	1667.594	Borehole	916
489207	900361	10	12	1651.601	Hand dug well	918
489270	901876	31	32	1666.837	Borehole	919
489362	901327	27	28	1664.992	Borehole	920
489369	900521	10	13	1654.3	Hand dug well	921
489405	901697	28.5	30	1665.515	Borehole	922
489420	900661	12	14	1653.621	Hand dug well	923
489459	900591	11.5	14	1656.829	Hand dug well	924
489527	901872	31	32	1666.106	Borehole	925
489542	902386	36	37	1675.096	Borehole	926
489582	902431	39	40	1676.715	Borehole	927
489657	902742	49	50	1684.064	Borehole	928
489664	902891	44	45	1685.822	Borehole	929
489727	901664	31	32	1665.368	Borehole	930
489841	900709	12	14	1661.596	Hand dug well	931
489975	901414	31	32	1665.878	Borehole	932
490183	900537	13	15	1656.135	Hand dug well	933
490616	900642	13	15	1659.967	Hand dug well	934
490629	900551	13	15	1654.992	Hand dug well	935
490708	901380	29	30	1668.74	Borehole	936
490756	901246	27	28	1669.002	Borehole	937

X	Y	Static_lev(m)	Depth(m)	Elevation(a.s.l.)	Scheme_typ	Well Id.
490837	901594	19	20	1670.581	Hand dug well	938
491150	899666	29	30	1651.168	Borehole	939
491167	902357	44	45	1683.471	Borehole	940
491235	901335	19	20	1673.19	Hand dug well	941
491241	900258	29	30	1657.876	Borehole	942
491286	900845	17	18	1666.118	Hand dug well	943
491289	900244	29	30	1657.337	Borehole	944
491403	901405	19	20	1674.934	Hand dug well	945
491471	900330	29	30	1661.534	Borehole	946
491610	900249	29	30	1662.839	Borehole	947
491760	900383	29	30	1662.885	Borehole	948
491788	900480	29	30	1668.21	Borehole	949
491789	901533	46	47	1675.51	Borehole	950
491900	900501	29	30	1668.213	Borehole	951
491908	901685	46	47	1677.954	Borehole	952
491972	900288	29	30	1664.696	Borehole	953
492009	900551	29	30	1671.731	Borehole	954
492023	900775	29	30	1673.88	Borehole	955
492029	900199	29	30	1662.705	Borehole	956
492043	900650	29	30	1672.063	Borehole	957
492055	900444	29	30	1667.432	Borehole	958
492077	902129	49	50	1686.958	Borehole	959
492103	900445	29	30	1669.196	Borehole	960
492106	900167	29	30	1662.732	Borehole	961
492113	900780	29	30	1673.21	Borehole	962
492130	900073	29	30	1662.956	Borehole	963
492169	900789	29	30	1672.154	Borehole	964
492177	900098	29	30	1662.708	Borehole	965
492205	900074	26	27	1662.7	Borehole	966
492218	900201	29	30	1666.154	Borehole	967
492231	899983	26	27	1663.43	Borehole	968
492233	900264	29	30	1666.502	Borehole	969
492335	900065	26	27	1663.234	Borehole	970
492403	900545	33.4	35	1672.002	Shallow well	971
492406	900552	29	30	1672.327	Borehole	972
492430	900119	26	27	1666.383	Borehole	973
492445	899820	26	27	1664.375	Borehole	974
492480	900421	34	35	1668.521	Borehole	975
492486	900417	27	28	1668.331	Borehole	976
492495	900638	34	35	1672.629	Borehole	977

X	Y	Static_lev(m)	Depth(m)	Elevation(a.s.l.)	Scheme_typ	Well Id.
492550	900346	27	28	1670.265	Borehole	978
492552	900582	34	35	1674.601	Borehole	979
492668	900727	34	35	1673.832	Borehole	980
492713	900560	34	35	1673.366	Borehole	981
492756	900189	34	35	1669.901	Borehole	982
492764	900235	35	80	1669.136	Borehole	983
492885	900449	34	35	1674.587	Borehole	984
492893	900356	34	35	1673.404	Borehole	985
492973	900192	34	35	1672.631	Borehole	986
493031	900145	34	35	1672.93	Borehole	987
493056	900183	34	35	1671.907	Borehole	988
493098	899869	34	35	1674.301	Borehole	989
493153	898768	34	35	1649.913	Borehole	990
493323	900829	36	37	1677.297	Borehole	991
494620	890557	10	12	1661.681	Hand dug well	992
494959	890665	12	13	1657.493	Hand dug well	993
494995	890692	10	12	1656.402	Hand dug well	994
495043	890430	10	12	1656.486	Hand dug well	995
495045	890719	10	12	1653.503	Hand dug well	996
495059	890541	12	13	1657.619	Hand dug well	997
495124	890549	12	13	1657.09	Hand dug well	999
495176	890385	10	12	1654.471	Hand dug well	1000
495311	890599	12	13	1654.491	Hand dug well	1001
495323	890585	11	12	1654.962	Hand dug well	1002
495349	890532	12	13	1654.568	Hand dug well	1003
495358	890485	11	13	1653.942	Hand dug well	1004
495383	890456	12	13	1654.868	Hand dug well	1005
495387	890501	12	13	1654.566	Hand dug well	1006
495496	890225	10	11	1654.736	Hand dug well	1007
495526	890140	10	12	1654.593	Hand dug well	1008
495567	890188	10	11	1655.947	Hand dug well	1009
495862	890176	25	30	1662.067	Borehole	1010
496131	897617	30	42	1675.382	Shallow well	1011
496181	890214	25	30	1668.971	Borehole	1012
497387	894554	22	23	1667.9	Borehole	1014
497516	892586	45	81	1694.227	Borehole	1016
497620	891108	24	25	1684.827	Borehole	1017
498185	897152	69	70	1689.825	Borehole	1018
498380	899093	55	60	1700.917	Borehole	1019
498383	891329	19	20	1685.718	Hand dug well	1020

X	Y	Static_lev(m)	Depth(m)	Elevation(a.s.l.)	Schseme_typ	Well Id.
498446	891962	19	20	1681.226	Hand dug well	1022
498631	891783	19	20	1681.537	Hand dug well	1023
500483	896177	65	75	1714.799	Borehole	1024
501594	895723	46	48	1698.233	Borehole	1028
502377	896000	43	45	1707.227	Borehole	1029
502415	895806	45	58	1703.089	Borehole	1030
502425	895932	43	45	1706.165	Borehole	1031