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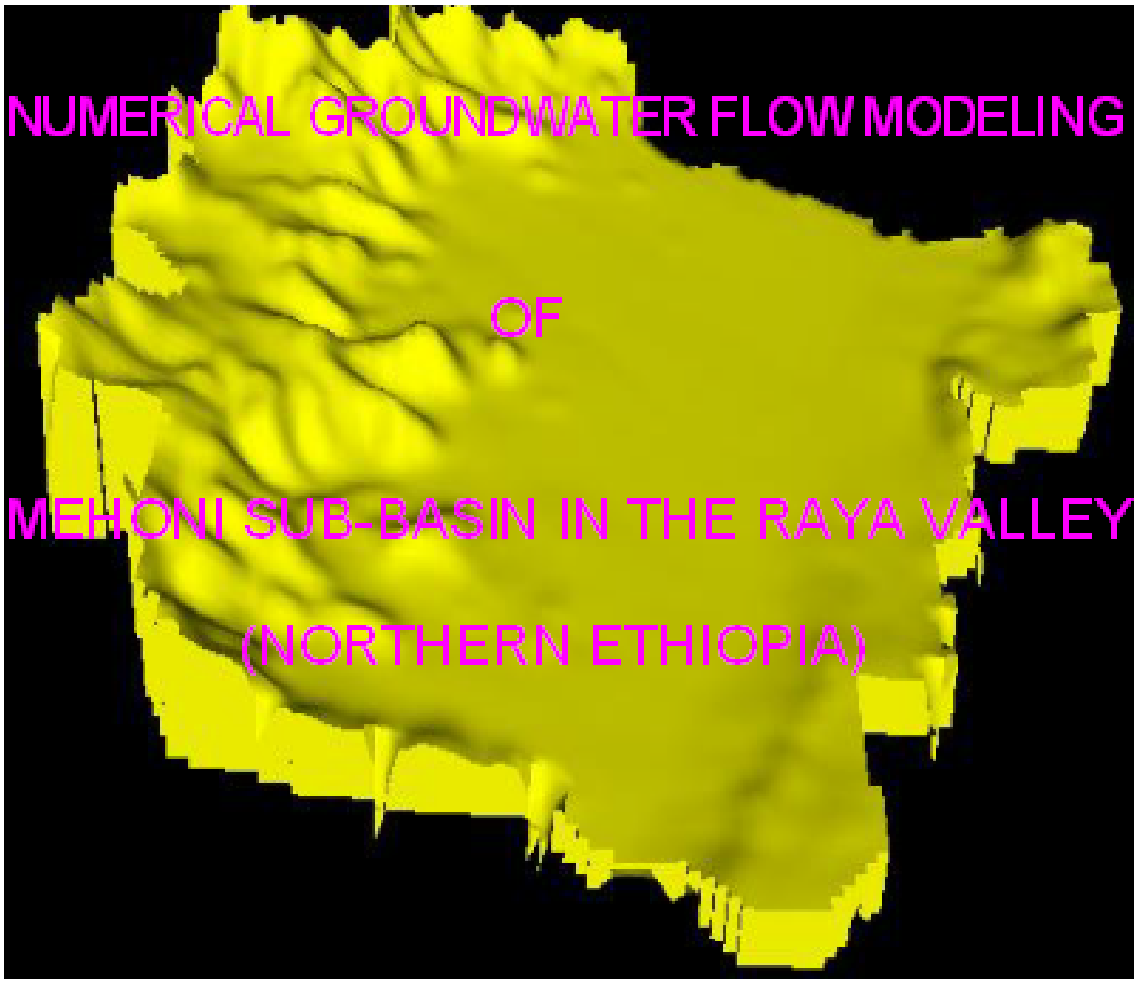

NUMERICAL GROUNDWATER FLOW MODELING OF MEHONI SUB-BASIN IN THE RAYA VALLEY (NORTHERN ETHIOPIA)

A THESIS
SUBMITTED TO THE SCHOOL OF GRADUATE STUDIES OF ADDIS ABABA UNIVERSITY
IN PARTIAL FULFILMENT OF THE REQUIREMENTS FOR THE DEGREE OF MASTER OF SCIENCE IN HYDROGEOLOGY

BY MUAUZ AMARE
FEBRUARY, 2007

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BY

MUAUZ AMARE

FACULTY OF NATURAL SCIENCE
DEPARTMENT OF EARTH SCIENCE

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Declaration

This thesis is my original work and has not been presented for a degree in any other university, and that all sources of the material used for the thesis have been duly acknowledged.

Dr. **Tamiru Almayohu** (Advisor)

Signature _________________

Muauz Amare

Signature _________________

Date and place of submission, March 20, 2007, Addis Ababa

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I would like to express deepest gratitude to my family particularly my Mom, my Dad and my brother Mulubrahn Amare for their inspiration, encouraging and helping me in so many ways.

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ABSTRACT

Mehoni sub-basin which is found in the Raya Valley is located in the southern part of Tigray Regional State of the Northern Ethiopia. The area covers 60% of the Raya valley with area coverage of 1183km^2.The sub-basin water resource is used for irrigation and domestic water supply and is covered by tertiary volcanic in western and eastern highlands and alluvial deposit in the valley fill. The western rocks are affected by tectonic movement .The N-S faulting is responsible for tilting of the rock formations with a general strike NE-SW and dipping to the SE direction.

As numerical groundwater flow models represent the simplification of complex natural systems, different parameters were assembled into conceptual model to represent the complex natural system in a simplified form. The conceptual model was put into the numeric model to examine system response.

Numerical groundwater flow was simulated in the model by the finite-difference method using MODFLOW, 1996 (McDonald and Harabaugh, 1988). The finite difference grid consisted of 1 layer, 116 rows and 111 columns. Two dimensional profile model was developed considering the system to be under steady state condition and assuming flow system view point. Four scenarios Increase withdrawals by three fold of the average existing withdrawals for whole sub-basin and local areas, assigned 25 wells with daily discharge 1000m^3/d and decrease recharge by 50% were used. Model calibration was carried out by trial and error calibration method using groundwater contours constructed from heads collected in 40 observation points. The calibration showed that about 97.5%of simulated heads were within the calibration target and the overall root mean square error for simulated hydraulic heads is about 8.52m. The poor fit at some points was due to numerous limitations associated with the model.

Model sensitivity analysis was conducted by considering the horizontal hydraulic conductivity and recharge because they are sensitive. A change in hydraulic conductivity by -55%,-45%,-25% and 55%, 45% 25% resulted in root mean square (RMS) 27.51, 19.293, 12.948, 8.499, 11.86

and 13.356 respectively. A change in recharge by the same amount as the horizontal hydraulic conductivity RMS head changes by 9.553 ,9.254, 8.854,8.290, 8.385, and 8.571 respectively.

The calibrated model was used to simulate the possible effect of increased groundwater pumping. Increase of three fold of withdrawals from average existing withdrawals in the whole sub-basin has resulted in the maximum and minimum decline in head by 2.33m and 0.56m respectively. Local increase of withdrawal by three fold of an average existing withdrawal indicated that the two (volcanic and alluvial) aquifer systems are not interconnected.

CHAPTER ONE
INTRODUCTION

1.1 Background

Though two-third of our world is covered by water, freshwater is not more than 2.7%. Even of this small amounts more than 77% is occur in the form of Ices capes & glacier which is not convenient to domestic consumption while the rest occur in the form of surface water and groundwater (Fetter, 1994) Water is elixir of life; without it life is not possible. To stay being the base of the base of life, it has to be existing in sufficient amount and acceptable quality.

However, due to climatic change, industrialization & high population growth, the amount of water available is decreasing and its quality is degrading. Since, especially, the now a days surface water is highly vulnerable for contamination and rarely available, most of water resource for domestic, agricultural and industrial is becoming groundwater both in developed and developing countries. This intensive utilization of groundwater is leading to negative environmental impact, such as land subsidence, salt water intrusion, water resource mining and so on.

On the other hand, groundwater exploration and exploitation is an expensive business which leads to high amount of loss in case of failures. Thus, it is found good to study and know the spatial and temporal distribution, occurrence and depth below ground level of groundwater basin or catchment's wise rather than site wise and use economically in a way the abstraction do not cause negative environmental impact and unnecessary wastage of money from unsuccessful groundwater development activities. To accomplish this task, groundwater modeling has got wide application in developed countries and recently it has gained good attention in the developing countries also.

The Mehoni sub basin in the Raya Valley is found in the southern part of Tigray regional State between 1382209_1428036northing and 553356_597420easting.

The study area approximately covers a total land area of 1183 km^2. The altitude ranges between 3600 m.a.s.l in the mountain ranges and 1400 m.a.s.l in the lowland plains.

The Raya Valley, which is part of a seriously interconnected valley of the Ethiopian rift system, has a total area of 1411 km^2 and is bounded by the plateau and mountains to the west and Chercher Mountains to the east.

The western highlands are mainly composed of Tertiary volcanic rocks. These rocks are highly disturbed by tectonic movements. The N-S faulting is responsible for tilting the rock formations with a general strike NE-SW and dipping to the SE direction.

Most of the streams that originate in the highlands flow south easterly, following the dip direction and joins the major river system, the Hum Shet (Sulula River) in the center of the valley.

The valley is composed of loosely compacted sedimentary basin fill deposits. It is believed to have a high potential for agricultural development, the main problem being seriously felt is scarcity of water. The dominant type of agriculture is rain fed agriculture with a practice of supplementary irrigation using surface water sources.

The existing rain fed agricultural practice has been seriously affected by recurrent drought. To resolve this problem the solution sought was to support with and shift the existing rain fed agricultural practice to modern irrigation with a conjunctive use of both surface and ground water resources. However, this solution will be sustainable and dependable if it is wisely managed. Owing to this fact, the main issue of this research project is describing the numerically simulated characteristics of groundwater flow system of the study area under steady state condition.

1.2 Previous Works

Hunting Technical services (HTS) during the period of 1974 to 1975 has studied the Mohoni area but it used very short time meteorological data and underestimated the available water resources in the area RVADP (1998). The study by the German consult in 1977 under the Kobo-Alamata Agricultural Development Project (KAADP) undertaken in the area between Kobo and Alamata towns covering part of the Alamata sub basin has concluded that the geological conditions in the east and west of the valley floor are similar and hence the sub surface water inflow and outflow into and from the valley are equal. The KAADP (1977) also reported that the flattening of ground water level in the eastern part of the valley indicates wide-spread ground water losses by evapotranspiration in this part of the valley.

But the reconnaissance study of the Raya Valley Agricultural Development Project (RVADP, 1998) has disproved the above conclusions by the fact that the western plateau and mountains and the valley fill basin are hydraulically independent ground water systems because the western plateau groundwater is discharged on the escarpments via springs rather than as sub-surface inflow into the valley fill basin where as the eastern mountains act as sub surface dam.

According to RVADP (1998), the well inventory in the eastern margins of the valley fill basin shows that the ground water level depth ranges 18-20 meters and the electrical conductivity values did not differ much from the western and central part of the valley (960 µs/cm). More over no direct groundwater discharge (as springs, swamp etc) and no ground water level fluctuation in the inventoried wells was observed. In general the conceptual ground water model of the KAADP is modified in the study of the RVADP (1998).

In the hydrogeology report the RVADP (1998), the Raya Valley is sub-divided into two sub-basins called Alamata sub-basin and Mehoni sub-basin. The average annual groundwater recharge and the static ground water reserve for the whole valley were estimated to be 85.6 Mm^3/year and 7150m^3 respectively. They have tried to show that the regional ground water flow directions of the Raya valley are in the west-east and north-south directions. The fan-like fields along the foot of the western escarpment were considered as groundwater potential. They also concluded that the groundwater is suitable for drinking and irrigation purposes but it is hard and needs softening to be used for industrial purposes.

A shallow examination and verification of the RVADP (1998) feasibility report was done by Isaak Gershanovich in the year 2000. He supposed the existence of large size fault of regional extent in the eastern part of the valley accompanied by ramified fracture systems that can be one possible ground water potential zone for future development. As proposed in this report, the deep position of the aquifers in the fractured rocks will be compensated by significant discharge wells. He assumed the valley fill sediments as fans from the western highlands that interweave each other and the existence of buried sediments of predominantly proluvium origin. He also commented that the density of information used to produce transmissivity map in a scale of 1:50,000 is not sufficient to map at this scale. He estimated the exploitable groundwater potential of Raya valley to be 130 Mm^3/year rather than 162 Mm^3/year (estimate of RVADP). His conceptual model for the Raya valley was a fan-like field where water bearing layers extend to a large distance under the plain sinking deeper and deeper towards the central and eastern parts of the valley.

According to Dessie Nadew (2003), (the main purpose of his work was to study and characterise the groundwater of Raya valley from geological, hydrological and chemical point of view) the mean annual rain fall and mean actual evapotranspiration of the valley are found to be 779 mm and 695 mm respectively. The annual evaporation from Lake Ashange was estimated to be 1204mm and surface water outflow 60 $Mm^3$from the Raya Valley. The annual recharge to the whole Raya valley was computed to be 129.3 Mm^3. Based on calculated transmissivity values potentiality of the aquifers of the valley are classified from high (> 500 m^2/day) to weak (< 0.5 – 5 m2/day). As in the other previous works, he also concluded that high potentiality aquifers are found in the western margin of the valley floor. He indicated that Ca – Mg – HCO3 or Mg – Ca – HCO3 type groundwater is common in the Raya valley. Different from other previous studies, he concluded that there are two separate ground water flow systems named as the Alamata sub-basin flow system and the Mehoni sub-basin flow system.

Generally, all the stated previous works has recommended that their works are just giving a general out look of the regional hydrogeological setting of the Raya valley and further more detailed local studies are mandatory so that a better hydrogeological knowledge of the area will be attained and there by a sustainable development and usage of the water resources of the Raya valley will be achieved.

Ermias Hagos (2005) has studied Hydrogeology of Mehoni sub-basin and Lake Ashenge catchment, in this work the mean annual precipitation and mean actual evapotranspiration of the Mehoni sub-basin 723.57mm and 687.24mm respectively. The annual recharge of the sub-basin was computed to be 66.5Mm^3. From geophysical and log data he concluded that the thickness of unconsolidated sediments is maximum (>250m); Sand and courser grains are dominantly found in the western part of the valley floor and along main river courses while silt and clay sediments dominate the central and eastern parts of the valley floor. He classified the hydrogeologic unit into alluvial aquifer and volcanic aquifer.the alluvial aquifer was also classified into the Quaternary colluvial and alluvial deposits (Qcol and Qal) and the Quaternary interfluvial, fan foot plain and valley bottom deposits (Qsmc). Accordingly, most of the Qcol and Qal aquifers were found to have moderate to high aquifer potentiality with range of transmisivities between 216 m^2/day and 4200 m^2/day; the Qsmc aquifers have low to moderate aquifer potentiality with range of transmisivities between 8.52 m^2/day and 436 m^2/day and the volcanic aquifers have weak to moderate aquifer potentiality with range of transmisivities between 0.8 m^2/day and 94.2 m^2/day. Most wells drilled into the Qcol and Qal aquifers have fairly good efficiencies and they are moderately to highly productive. He indicated the dominant water type is bicarbonate.

Lastly He concluded that most groundwater of the valley floor has shown minimum risks of toxic ions and other miscellaneous chemical effects. Therefore, it can be used for irrigation with little restrictions on use.

1.3 Objective of the Study

General Objective of the Study

The general objective of this research work was to develop a better understanding of the ground flow system of the Mehoni sub-basin in Raya Valley.

Specific objective of the thesis research:

- to conceptually model the aquifer parameters and the hydrologic condition of the study area. This includes:

- ❖ Estimation of the hydraulic properties major Hydro-geologic units
- ❖Define and describe different hydrgeologic boundaries

- ➢ to simulate the groundwater flow system of the study area numerically by assigning the conceptual model parameters.
- ➢ to calibrate the simulated groundwater head data with the observed head data under steady state
- ➢ to analyze the sensitivity of the model to which parameter or stress the flow condition is sensitive.
- ➢ to assign different scenarios for reasonable groundwater abstraction rate.

1.4 Methodology

Groundwater flow system of the study area was simulated numerically by using a computer code, the U.S. Geological Survey modular three-dimensional finite difference groundwater flow model, MODFLOW (Harbough et al., 2000). MODFLOW uses input arrays (grided data) that describe hydraulic parameters such as hydraulic conductivity and recharge, top and bottom elevation of the aquifers and boundary conditions.

Prior to the construction of the three dimensional groundwater flow model, a conceptual model of the system was developed on the basis of the secondary data collected from different institutions, previous works, field observations and groundwater literatures. By analyzing the collected data, the input parameter for the numerical model was estimated. Different boundary conditions were estimated from physical and hydrogeologic boundaries. Hydraulic parameters such as hydraulic conductivity and storage coefficient were calculated from pumping test data whenever it is available and from literature for the others depending on the aquifer type. Besides, hydraulic head map was established from the water level measured during field data collection and drilling reports. Groundwater recharge was modified from the previous studies and a withdrawal was obtained from daily record of the Town Water Supplies, and irrigation practices.

After the aquifer parameters and hydrologic stresses were put in to the numerical model, the model was calibrated using trial and error method by rearranging the parameters and stresses within plausible range to get the best fit between the observed and simulated heads. The calibration was done under steady state and transient conditions.

Prediction for future groundwater management condition was done by assigning different aquifer system conditions, hydrologic stresses, and pumping conditions under different scenarios. Finally, the best pumping condition was selected under normal hydrologic stress.

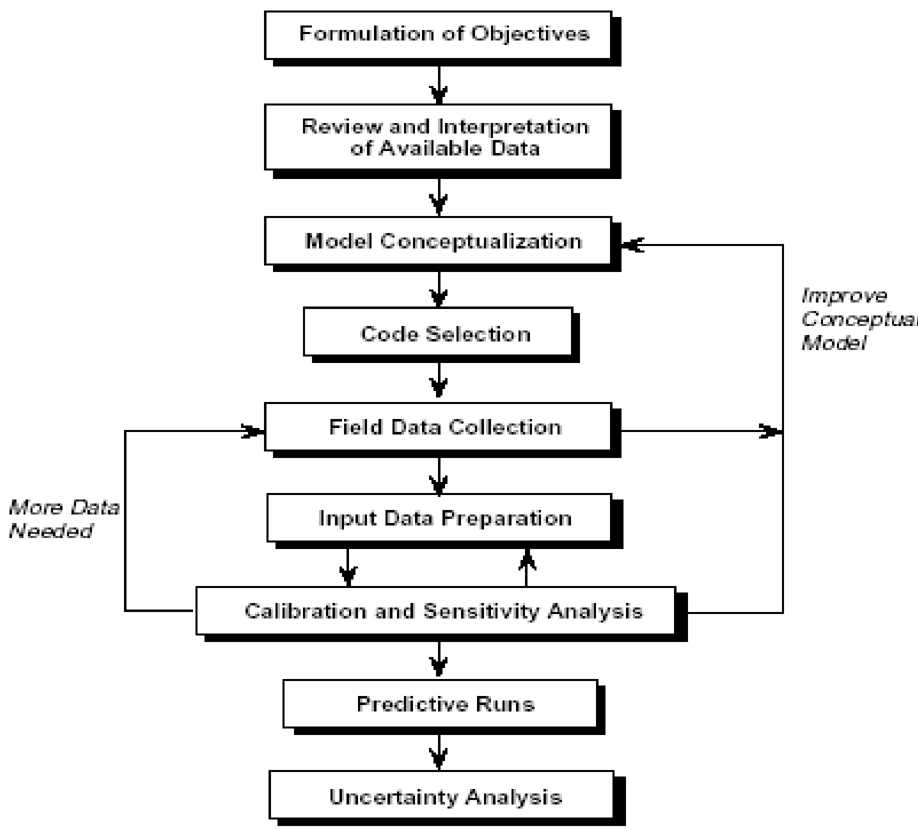


Fig 1.1 Model application process.

CHAPTER TWO

DESCRIPTION OF THE STUDY AREA

2.1 Study Area Location and Aerial Extent

The study area encompasses an area of about 1183 km^2 and is located in the northern eastern part of Ethiopia within 1382209_1428036northing and 553356-597420easting and specifically in the south eastern part of Tigray region. It constitutes about 60 % of the Raya Valley, which is considered as one of the interconnected Valleys of the Ethiopian rift system (RVADP, 1998). It is accessed by the Maichew-Alamata Asphalt road, Mehoni-Maichew, Mehoni-Alamata and Mehoni-Chercher-Alamata all weather roads and other branching dry weather roads. The Maichew Mountains in the north, Chercher Mountains in the east, the Central Ethiopian plateau in the west and the Chegwara ridge in the south surround the study area with a narrow southerly surface water outlet.

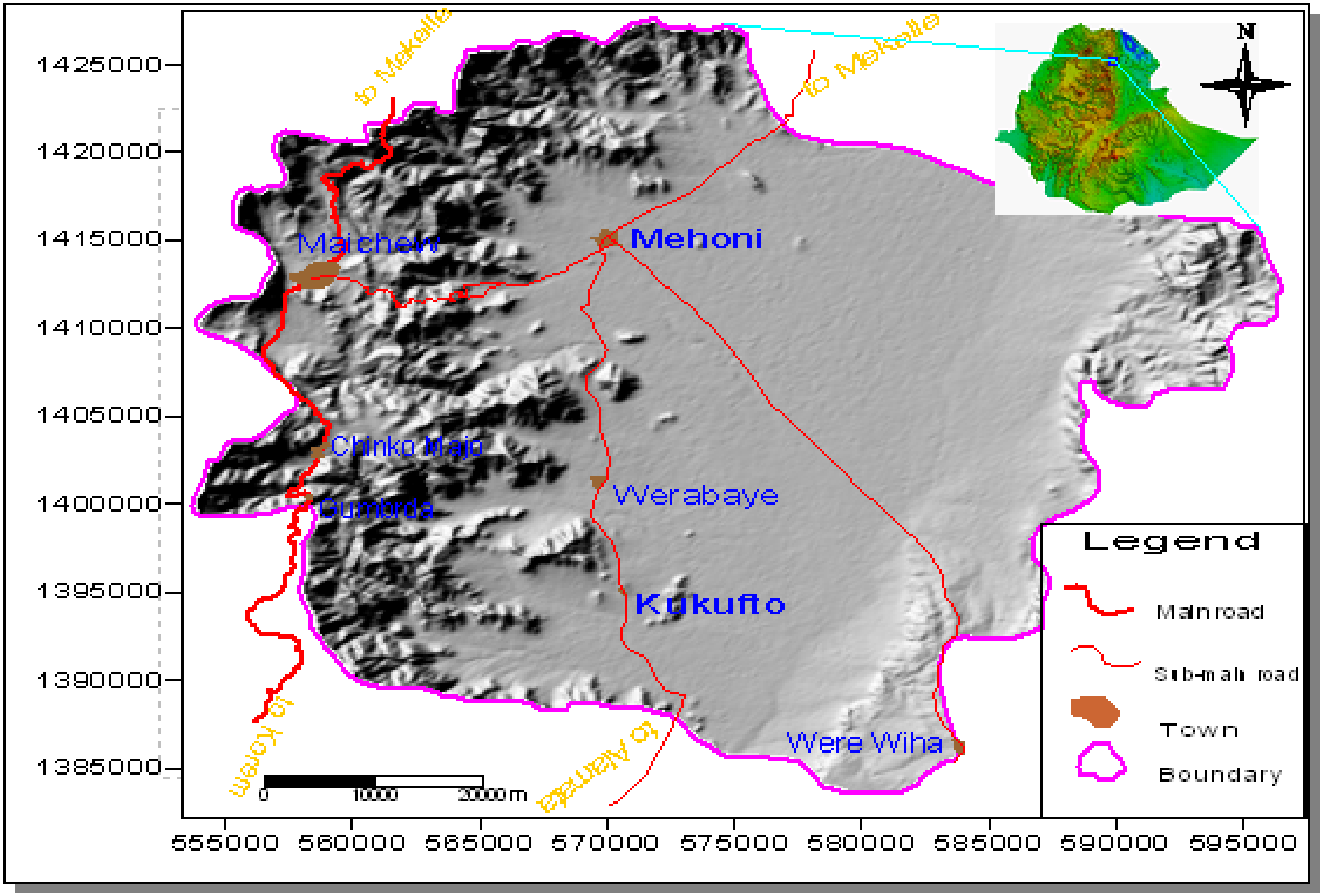


Fig 2.1 Location Map of the Study Area

2.2 Climate and Vegetation

Due to wide altitude variations, the study area is characterized by wide range of climatic elements. It is characterized by bimodal type of rainfall pattern with small rains in February to April and heavy rains in July to September. Its mean annual rainfall is 723.57 mm with mean daily maximum and minimum temperatures of 18.29 ^{o}c and 13.93 20^{o}c respectively for the western high lands and 23.44 ^{o}c and 19.64 ^{o}c respectively for the valley floor.

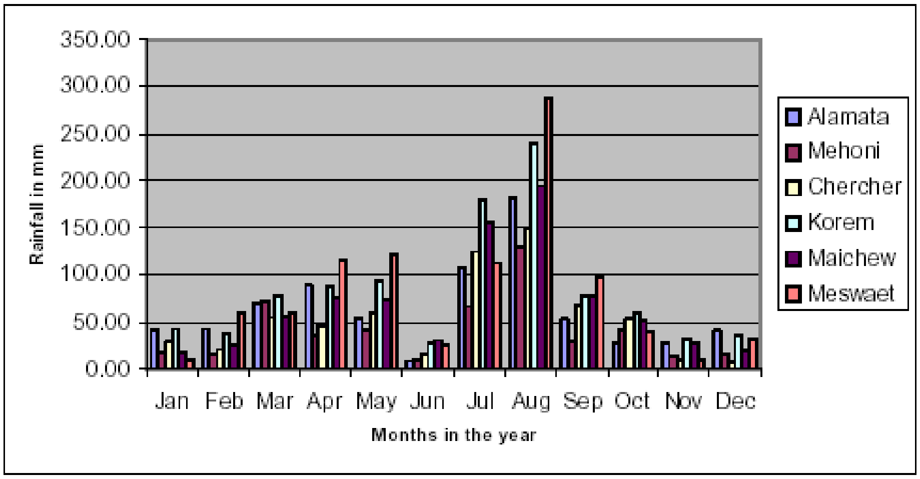


Fig 2.2. Monthly Rain Fall Distribution in Mehoni Sub-Basin

There are seven metrological stations in the whole basin and its surroundings with temperature readings. These station include Maichew , Alamata ,Korem , Chercher and Kobo whose data are collected from National Metrological Agency and Mehoni and Waja whose data are taken from existing literature. The duration of the data ranges from 26 years on Maichew to 5 years in Mehoni.

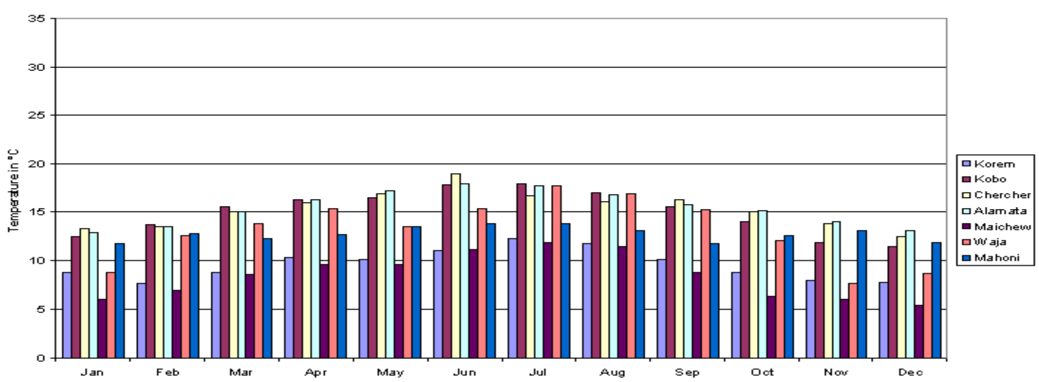


Fig 2.3. Mean Minimum Air Temperature in Different Stations

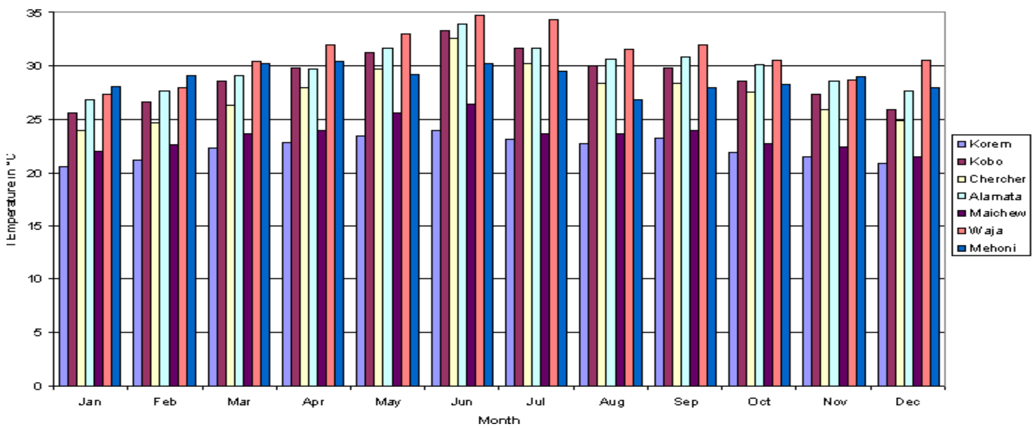


Fig 2.4. Mean Maximum Air Temperature in Different Stations

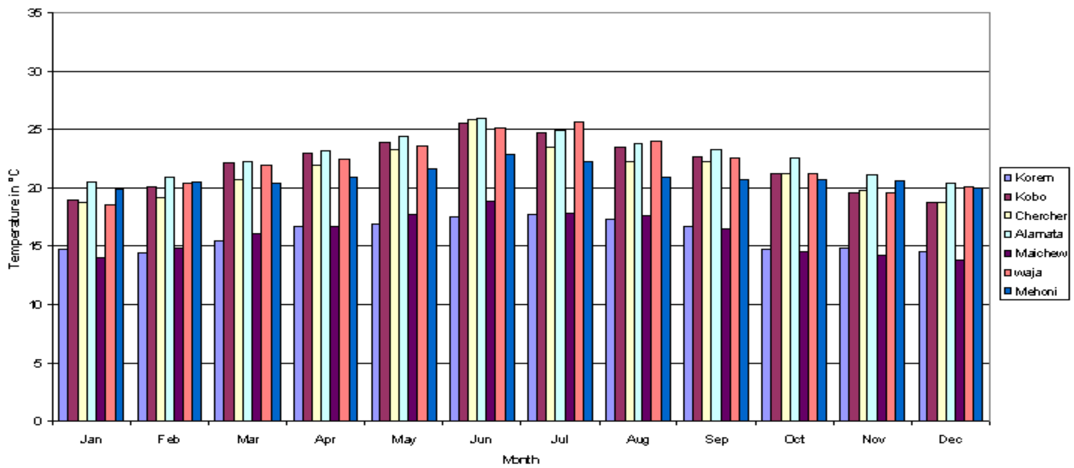


Fig 2.5. Mean Monthly Air Temperature (^{O}c) For Different Station in the Study Area

Stations	Type	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec	Mean
Korem	min	8.9	7.7	8.8	10.3	10.2	11.05	12.34	11.8	10.2	8.8	8	7.8	10
	Max	20.6	21.1	22.3	22.8	23.4	24	23.1	22.7	23.2	21.9	21.4	20.8	22
	Mean	14.7	14.4	15.5	16.7	16.9	17.5	17.7	17.3	16.7	14.7	14.8	14.5	16
Kobo	min	12.5	13.7	15.6	16.3	16.5	17.8	18	17	15.6	14	11.90	11.5	15
	Max	25.6	26.6	28.6	29.8	31.2	33.3	31.7	30	29.8	28.5	27.4	25.9	29
	Mean	19	20.1	22.1	23	23.9	25.5	24.7	23.5	22.7	21.2	19.6	18.7	22
Chercher	min	13.3	13.6	15	16	16.9	19	16.7	16.1	16.3	15.1	13.8	12.5	15
	Max	24	24.7	26.3	27.9	29.7	32.6	30.2	28.4	28.4	27.5	25.9	24.8	28
	Mean	18.7	19.1	20.7	22	23.3	25.8	23.5	22.3	22.3	21.2	19.8	18.7	21
Alamata	min	12.90	13.60	15.00	16.30	17.30	18.00	17.70	16.80	15.80	15.20	14.00	13.10	15
	Max	26.8	27.7	29.1	29.7	31.7	33.9	31.7	30.6	30.8	30.1	28.6	27.7	30
	Mean	20.5	20.9	22.2	23.2	24.4	25.9	24.9	23.8	23.3	22.6	21.1	20.4	23
Maichew	min	6	6.9	8.6	9.6	9.6	11.2	11.9	11.5	8.9	6.4	6.1	5.5	9
	Max	22	22.6	23.6	23.9	25.6	26.4	23.6	23.6	24	22.7	22.4	21.4	23
	Mean	14	14.8	16.1	16.7	17.7	18.8	17.8	17.6	16.5	14.6	14.3	13.8	16
Waja*	min**	8.8	12.6	13.8	15.4	13.6	15.4	17.7	16.9	15.3	12.1	7.7	8.7	13
	Max**	27.4	28	30.4	32	33	34.8	34.3	31.5	32	30.5	28.7	30.5	31
	Mean	18.5	20.4	22	22.4	23.6	25.1	25.6	24	22.5	21.2	19.6	20.1	22
Mahoni*	min	11.8	12.8	12.3	12.7	13.6	13.9	13.9	13.1	11.8	12.6	13.1	11.9	13
	Max	28.1	29.1	30.2	30.4	29.2	30.2	29.5	26.8	28	28.3	29	28	29
	Mean	19.9	20.5	20.4	20.9	21.6	22.8	22.2	20.9	20.7	20.7	20.6	20	21

Table 2.1 Mean Monthly Air Temperature in Different Stations

*data taken from RVDP report, Phase 1, volume II, Hydrology

** Extreme minimum and maximum

Name of stations, Rc, ID	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec	Mean annual Rainfall	recording years	number of years with >10 months of recording	Altitude
Korem	46.4	36.9	91.9	90.8	123.6	22.5	166.2	229.6	72.9	72.3	17.6	35	1005.7	19	15	2500
Rc	0.55	0.44	1.10	1.08	1.47	0.27	1.981	2.74	0.87	0.86	0.21	0.42				
ID	DM	DM	MC	MC	MC	DM	MC	HC	SR	SR	DM	DM				
Alamata	41.6	47.7	78.4	97	59.1	8.3	92.7	158.3	52.5	26.9	25.5	40.6	728.6	18	14	1580
Rc	0.68	0.79	1.29	1.60	0.97	0.14	1.53	2.61	0.86	0.44	0.42	0.67				
ID	SR	SR	MC	MC	SR	DM	MC	HC	SR	DM	DM	SR				
Maichew	10.5	30.5	49.7	82.8	64.1	26.2	168.3	193.7	68.8	36.2	31.4	19	781.2	31	18	2400
Rc	0.16	0.47	0.76	1.27	0.98	0.40	2.59	2.98	1.06	0.56	0.48	0.29				
ID	DM	DM	SR	MC	SR	DM	HC	HC	MC	DM	DM	DM				
Kobo	14.5	22.6	42.2	61.6	66.4	20.3	118.5	155.9	64.6	53.9	12.1	11.1	643.7	16	13	1470
Rc	0.27	0.42	0.79	1.15	1.24	0.38	2.21	2.91	1.20	1.00	0.23	0.21				
ID	DM	DM	SR	MC	MC	DM	HC	HC	MC	MC	DM	DM				
Waja	18.6	28	74.3	79.8	75.3	10.5	94.3	133.1	60.8	35.6	30.2	20.8	661.3	12	8	1450
Rc	0.33	0.51	1.35	1.45	1.37	0.19	1.71	2.42	1.10	0.65	0.55	0.38				
ID	DM	DM	MC	MC	MC	DM	MC	HC	MC	DM	DM	DM				
Chercher	28.4	21	54.8	45.7	59.8	17.9	124.7	172.6	68.4	53.2	9.9	7	663.4	8	6	1825
Rc	0.51	0.38	0.99	0.83	1.08	0.32	2.26	3.12	1.24	0.96	0.18	0.13				
ID	DM	DM	SR	SR	MC	DM	HC	VHC	MC	SR	DM	DM				
Mohoni∗[1]	23.4	20.4	85.6	55.8	35.5	10.3	58	121.4	27.9	35.3	6.4	8.2	488.2	4		1850
Rc	0.57	0.50	2.10	1.37	0.87	0.25	1.43	2.98	0.69	0.87	0.16	0.20				
ID	DM	DM	HC	MC	SR	DM	MC	HC	SR	SR	DM	DM				
Mesweat*	10.4	60.6	60.1	115.5	120.4	25.3	111.9	286.4	98.1	39.4	10.4	33.1	971.6	6		2400
Rc	0.14	0.83	0.83	1.59	1.66	0.34	0.16	3.94	1.35	0.54	0.14	0.46				
ID	DM	SR	SR	MC	MC	DM	DM	VHC	MC	DM	DM	DM				

Table 2.2 Mean Monthly Rainfall, Rainfall Coefficient (RC) and the Subsequent Designation (Id), For Rain Gauge Stations in Raya Valley (Modified From Desie Nadew, 2003)

* The stations are administered by local Agricultural Office

∗[1] Rainfall records from 1975 to 1981, but for basin mean rainfall calculation, the rainfall measurement from 1993 to 1996(RVDP, phase I, volume II) is used.

The vegetation in the area includes remnants of natural and planted trees, shrubs, and grasses. But it is continually being degraded due to rapid population growth and continual need for cultivation. Cactus tree also invades considerable land in the study area.

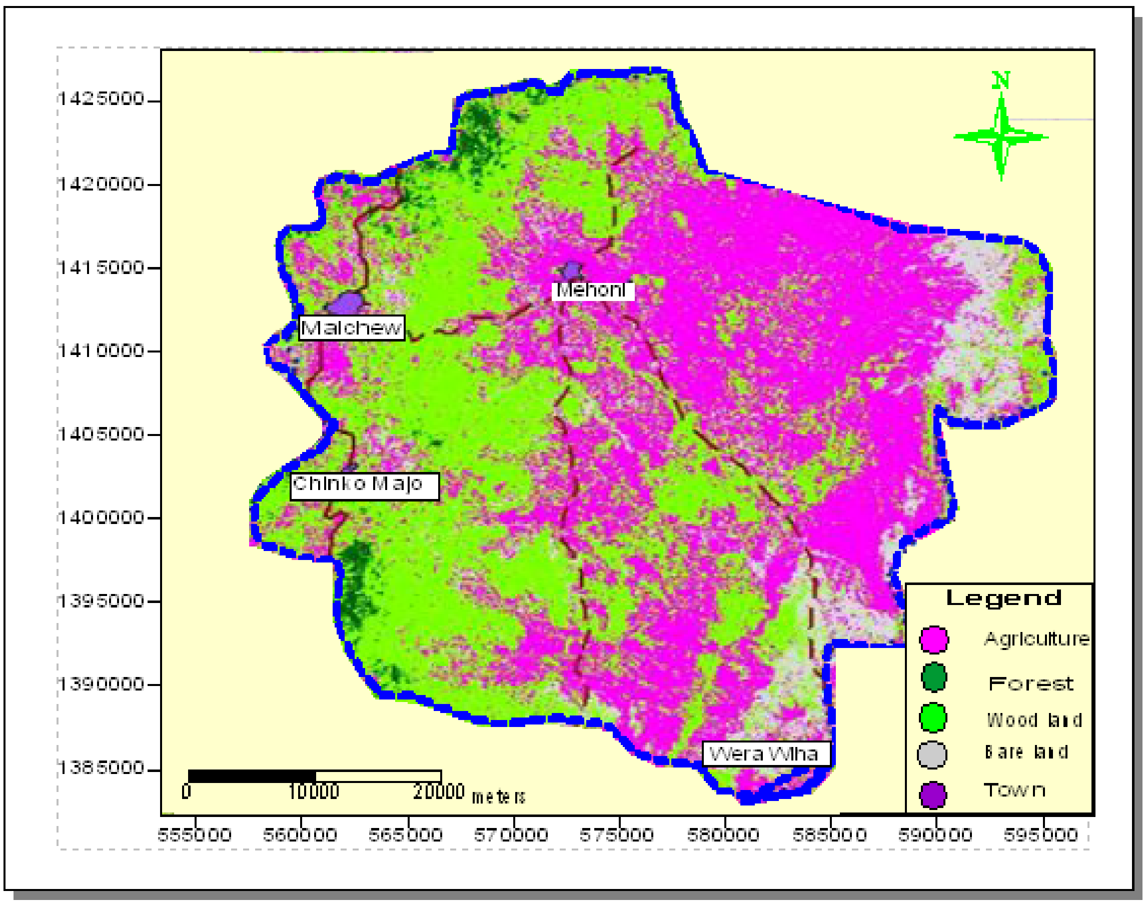


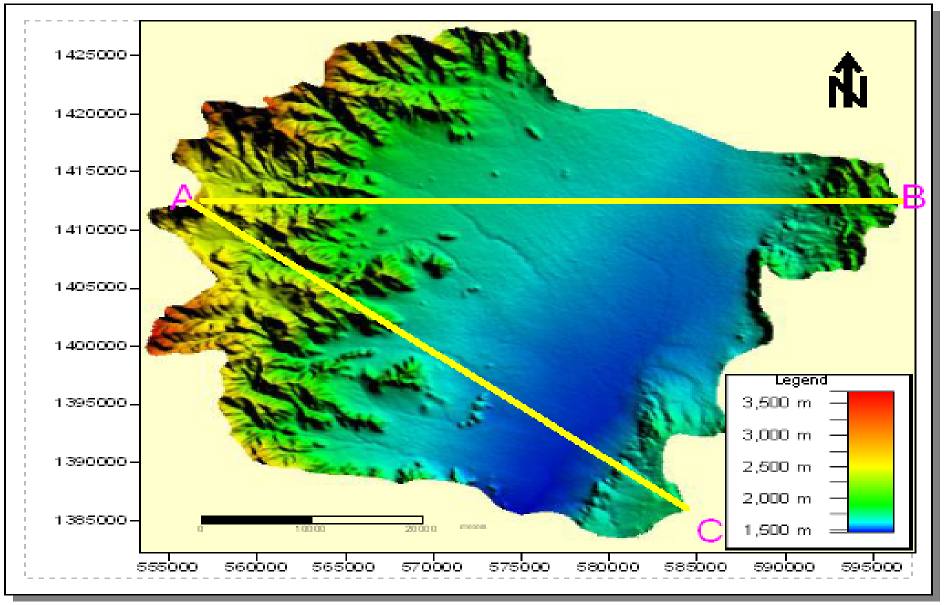
Fig2.6 Land Use and Land Cover Map (Modified From Desie Nadew, 2003)

The total projected population of Tigray regional state for the year 2005 is 4,223,014. The southern zone of Tigray where the study area is located contains a total population of 1,043331 with about 859,659 people living in rural areas and 183,672 people in the urban areas. The population size of the study area in particular is about 346,045 with about 81% living in the rural areas and the remaining 19% urban areas of Maichew, Mehoni, Chercher and other small towns (BPED, 1998).

2.3 Topography and drainage

2.3.1 Topography

Geomorphologically, the study area is characterized by an asymmetric graben and horst like structure bordered by fault scarps (figure 2.7). The graben is filled with Quaternary sediments that are mainly derived from weathering and erosion of the Tertiary Volcanic in the uplifted landforms. The minimum height in the study area is about 1400 m a.m.s.l where as the maximum height is found to be 3600m a.m.s.l. The weighted mean height of the basin is found to be 1955m a.m.s.l. The study area was classified into different physiographic units. RVDP, (1996) classified it into seven, Desies Nadew, (2003) in to four and Ermias Hagos, (2005) in to five. For this study the area is classified in to five land form units (Figure 2.9).



2.7 Digital Elevation Model of the Study Area

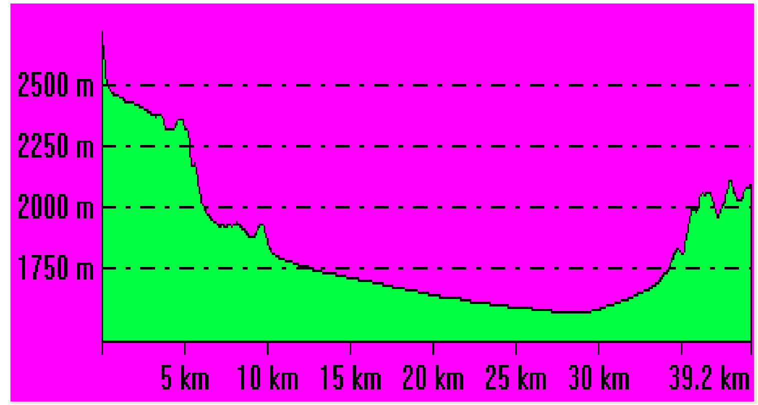


Fig 2.8 a Cross-section along A-B

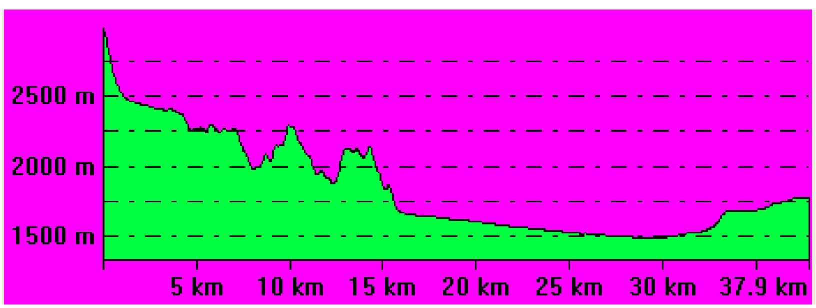


Fig 2.8b Cross-section along A-C

2.3.1.1 The Plateau or Crest Zone

The plateau zone comprises the eastern and western high lands. The western plateau is at higher elevation and it is more affected by tectonic disturbances such as faulting, fracturing, tilting and sliding. These disturbances resulted in either to deep fault scarps that fall in to the valley floor or steep like landform. The eastern plateau is characterized by rolling to dissected low hills. The plateau zone forms about 15.02% of the study area.

2.3.1.2 The Intermountain Basins

The intermountain basins are also results of sliding, faulting and slip movements of rock masses especially in the western highlands. They are low-lying depressions filled with sediments and water. These small basins are located around Maichew, Mekan and Chinko Majo area.In general, this physiographic unit forms about 1% of the study area.

2.3.1.3 The Fault Scarp or Free Face Zone

This zone constitutes the steeply dipping escarpments or free faces. At some places, this zone rises as high as 2000m from the level of the valley floor. Sometimes this zone is characterized by detachment of boulders and rock falls that are induced by gravity and rainwater. The fault scarp zone forms about 23.31% of the sub-basin.

2.3.1.4 The Pediment or Alluvial Fan Zone

This zone is found between the pediment zone and the flood plain. It is characterized by the accumulation of fan like alluvial and colluvial deposits. These deposits are mainly composed of coarse, angular and poorly sorted sediments. They are generally high permeable. That is why most rivers in the study area become influent or dry when they reach the pediment zone. This zone forms 15.72% of the sub-basin.

2.3.1.5 The Flood Plain

The flood plain constitutes the central part of the valley floor. It has more or less flat topography. The major river in the study area called Sulula River crosses this zone in a N-S direction. Fine sediments like clay, silt and fine sand are dominantly found in this zone. This zone is sometimes sprinkled by small basaltic domes. The entrenchment of river valleys is generally very shallow and streams do not form deep gorges. The discharge and flow velocity of water in these rivers and streams is so low that deposition of relatively finer sediments is favored. This zone is about 44.95% of the sub-basin.

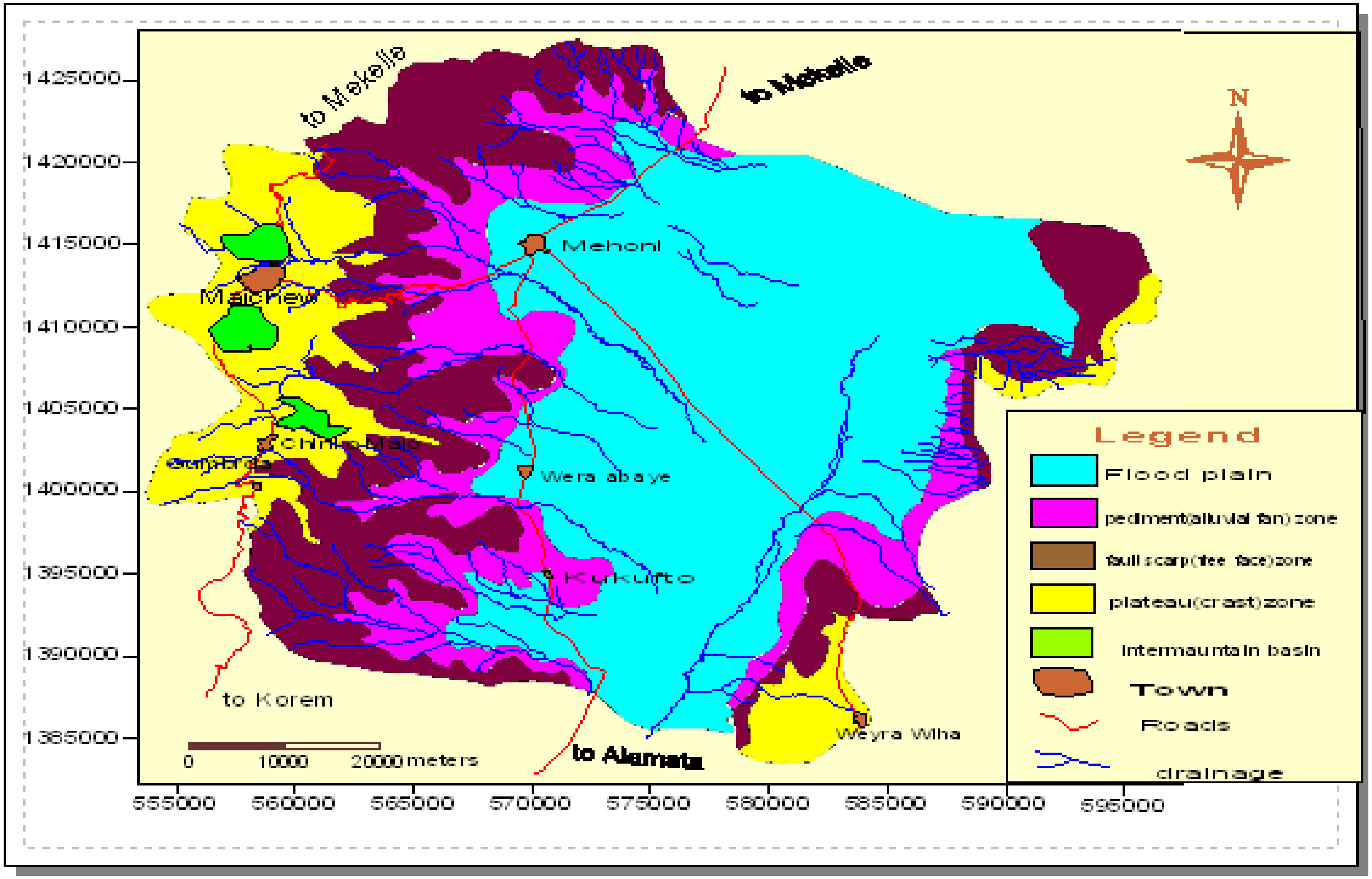


Fig 2.9 Land Form Classification

2.3.2 Drainage

The study area, as part of the Raya valley, is found in one of the closed catchments in the Denakil basin. The Denakil basin occupies 69,520 Km^2 in the northern region of rift funnel (Tesfaye Chernet, 1993) and it is bordered by the Tekeze and Mereb basins to the west, the Awash basin on the south and the Red sea costal basin on the north and east. In the study area, the most dominant drainage pattern is the dendritic pattern especially in the western part where most of the rivers of the area are concentrated. The major rivers that flow in the east-west direction from the western and eastern highlands are nearly parallel to each other showing structural control over individual rivers whereas Sulula River flows in the N-S direction. The only surface water out let from the study area is via the south-eastern part of the study area. Some of the major rivers in the study area are Beyru, Werabeyti, Haya, Fokisa, Guguf, Burka, and Mai-Anishti (Figure 2.10). They constitute separate drainage units emerging from the western and eastern highlands. Most of the rivers drain water only for a very short time during heavy rainfall events. But the valley floor has rather semiarid climate and its drainage system is sparse and the rivers are shallow.

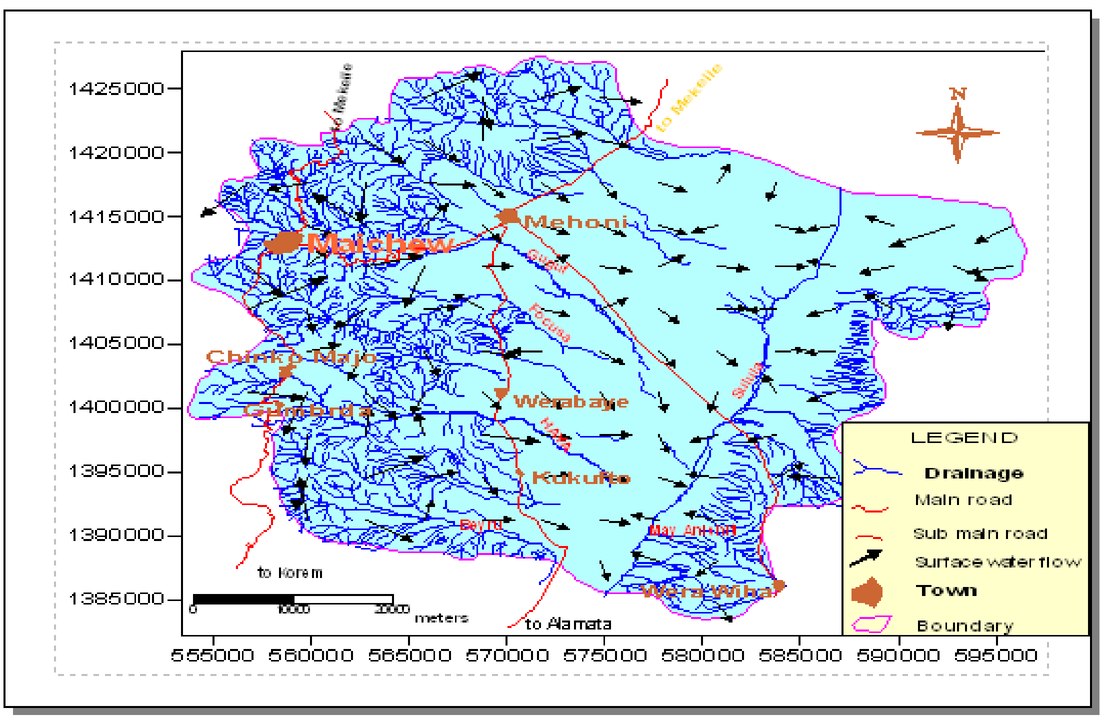


Fig 2.10 Drainage Map of the Study Area

CHAPTER THREE
GEOLOGICAL SETING

3.1 Regional Geological Setting

The African and Arabian plates are separated by the Red Sea and gulf of Aden Spreading centres. Exrension started 20 to 25 Ma, and oceanic crust appeared in the two rifts around 10 to 13 Ma or more recently. These two oceanic rifts are connected to the less developed East African rift at the Afar triple junction. This triple junction is the centre of a large volcanic province that is exposed in Ethiopia, Eritrea, Yemen, and the Republic of Djibouti. In northern Ethiopia and Yemen, volcanism started during the Oligocene with the emplacement of a thick continental flood basalt pile. These basaltic highlands now encircle the Afar depression. Most of these flood basalts were extruded over a short time period (possibly 1-2 Ma) 30 Ma and significantly predate the main extensional phases. However, there are evidences for older less extended flood magmatism further south in Ethiopia, Subsequent volcanism, from 22 Ma to present, occurs as large central vent volcanoes on the plateau or is associated with extension zone, which is localized in the rift axis. This zone is a focus of quaternary and recent volcanic activities. All of these volcanic units directly overlie the pan – African crust (composed of volcanic arcs accreted 600 – 700 Ma), late Archean crust (mainly in Yemen), or Mesozoic sediment (Pik et al., 1999).

The Ethiopian flood basalts (or traps) cover an area of about 600,000 km^2 with a layer of basaltic and felsic volcanic rocks. The thickness of this layer is highly variable but reaches 2km in some regions. The volcanic and shallow intrusive rocks have a total volume of about 350,000 km^3 (Mohr, 1983b; Mohr and Zanettin, 1988 cited in Dessie Nadew, 2003).

New structural and geochronological date (Cukstins et al., 2002; Wolfenden et al., 2004, In press) have shown that basin formation commenced between 29 and 26 Ma in the Southern Red Sea rift, while rifting in the northern most Main Ethiopian rift (MER) Initiated around - 11 Ma. Initial crustal extension within the Southern and central MER commenced between 18 and 15 Ma (Gidey Welde Gabriel et al., 1990; Ebinger et al., 2000). Thus, the MER, which is northen most sector of the

East African rift system. The East African rift system propagated northward from the Turkana rift at – 25 Ma. It is a region of anomalously thinned lithospoheric plate during Mesozoic time. The red Sea – Aden – MER triple junction is, therefore, a relatively young feature, developed only during the past 11 Ma, or 20 Ma after the flood volcanism (Dereje Ayalew et al., 2005).

As observed in the MER, magma chambers occur where there is a dense network of faults i.e., two or more trends of faults. (Dereje Ayalew et al., 2005,). There is a rift work migration of the locus of magmatism and faulting through time.

The felsic volcanic strata, capping the flood basalt sequences, are preferentially localized on or near the border fault, indicating that the age of the rhyolites actually dates the on set of rifting in the southern Red Sea and Northern main Ethiopian rift. This clearly shows that rifting provides a mechanism for generation of large volumes of felsic magmatism (Dereje Ayalew et., 2005).

The mineralogical and chmical composition of the flood basalts is relatively uniform. Most are aphyric to sparsel phyric, and contain phenocryst of plagioclase and clinopyroxene with or without olivine. Most have tholeitic to transitional compostions (Mohr, 1983a; Mohr and Zabettin, 1988; Pik et al., 1998 in Bruno Kieffer et al., 2003). Interlayered with the flood basalts, particularly at upper stratigraphic levels, are felsic lavas and pyroclastic rocks of rhyolitic, or less commonly trachytic compositions (Dereje Ayalew et al., 1999).

Except along its margins and in major river valleys, the entire Ethiopian volcanic plateau currently stands above 2000m in altitude. According to Jestin and Hucho (1992) and Menzies et al. (1992), the uplift took place concurrently with or very soon after eruption of the flood basalts (30 Ma) and since that time the high altitudes have been maintained. It is most unlikely that this uplift is due to the thermal or compositional influence of Oligocene plume, which should have dissipated by now. Instead, the prestn elevation of the plateau is normally attributed to dynamic support from a thermally anomalous up welling portion of the present upper mantle (Bruno Kieffer et al., 2003).

The Ethiopian volcanic plateau is not a thick monotonous, rapidly erupted; pile of undeformed, flat lying tholeiitic basalts. Instead, it consists of a number of volcanic centers with different magmatic character and with a large range of ages.

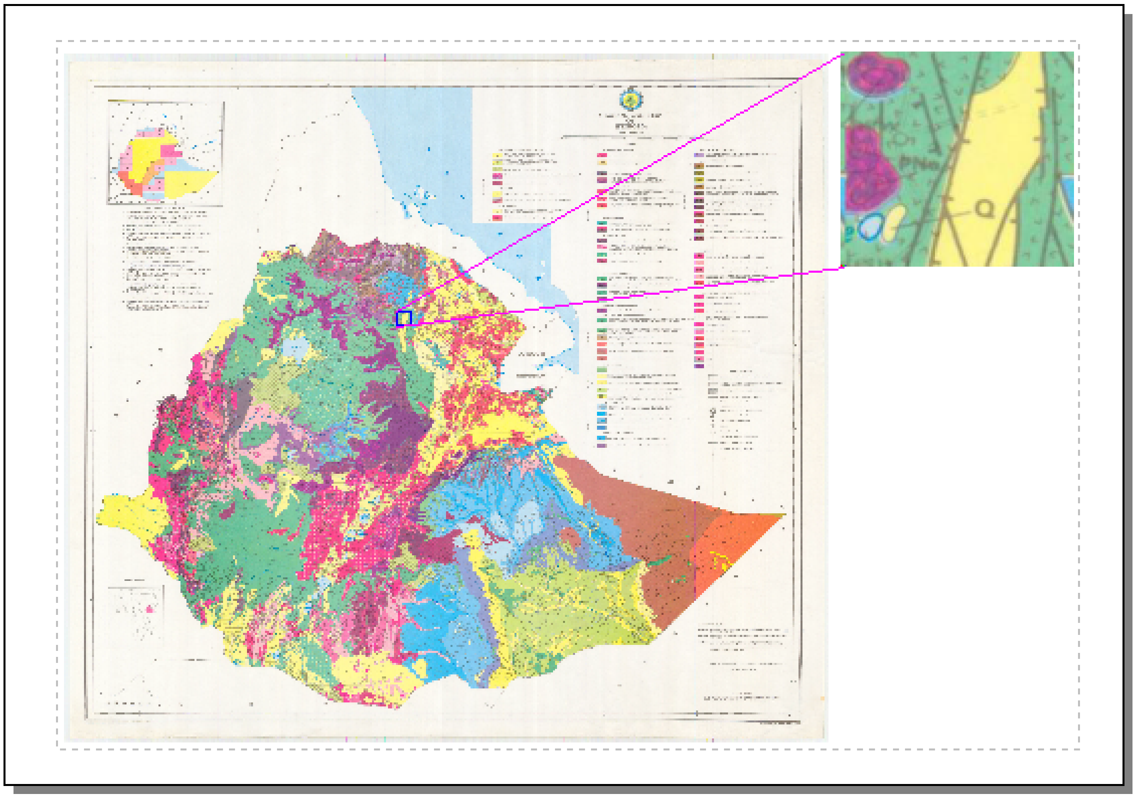


Fig 3.1 Geological Map of Ethiopia

3.2 Local Geologic Setting

The continental flood basalts of Ethiopia are commonly termed as the trap series to distinguish them from the post rift series. There are two views concerning the stratigraphy of this thick volcanic pile. The first one is classifying the flow into two distinct chrono stratigraphic units: pre Oligocene stage: Ashange basalt; and Oligocen – Miocene stage: Aiba basalt, Alaje fissural basalts and rhyolites and Tarmaber basalts (Zenetin et al., 1980, Berhe et al., 1987 cited in Dessie Nadew, 2003). The second prevailing view is considering this volcanic pile as a continuous lava sequence from the base to the top of the plateau (Pik et al., 1998 cited in Dessie Nadew, 2003).

The first approach is widely accepted and used. The geological map of Ethiopia published in 1996 depicts the study area to be mainly made of the pre-oligcene stage Ashange basalt with patches of overlying Aiba and Alaje formations on Maichew and Korem areas. The name Ashange basalt applies to strongly weathered basalts with out clear stratification which lie below the major pre – Oligocene (?) unconformity having a thickness of several hundreds of meters (Zebetin et al., 1980 cited in Dessie Nadew, 2003)

The surfacial geology of the study area consists of the following three general geologic units:

The Eastern highlands, includes vesicular, friable and scoracious with variegated color basalts. They are mostly slightly to moderately weathered and highly fractured. They are usually intruded by apilitic dykes that are also fractured and weathered. Most of the fractures and dykes have nearly N-S and E-W oriented strike and dip sub vertically. The basaltic formations in this part of the study area are generally highly disturbed where flow beds trend in many directions. Small part of the eastern highland is composed of sedimentary rocks which form steep cliffs towards the valley floor. This sedimentary formation consist mudstone at the top, sandstone in the middle and limestone at the bottom part of the cliff. The sedimentary beds are dipping to the east with a nearly N-S strike direction. This rock formation is also jointed with two major joint orientations (N 500W and N200E). These orientations are also common in fracture system of the basalt.

The western highland is dominantly composed of intercalated flow beds and masses of basaltic rocks. Exfoliation weathering, scoracious and vesicular appearance, friable nature and columnar jointing were the dominantly observed features in these basalts.

Some times large massive blocks of basalt are found standing as remnants of weathering. On the slopes and foot of the western highlands there are alluvio-colluvial deposits that comprise poorly sorted sediments. Some stains and deposits of white secondary minerals are observed along joints, fractures and on other weathering structures. There are also patches of basaltic dykes and sills with which are usually slightly to moderately weathered and friable but some times they are massive.

The basaltic flow beds on the Mountains around Maichew town are more or less horizontally lying. But in most other areas of the western highland they are disturbed and usually dipping in the east and southeast directions. Steep fault scarps are also clearly exposed on the northern escarpment areas of the western highland.

***Unconsolidated Sediments (Quaternary deposits*)**:The Mehoni sub-basin as part of the Raya Valley is considered as an intermountain trough-like structure that originated from local tectonic development attributed to the regional East African rift system (Gershanovich, 2000). Therefore the study area is one of the graben and horst lineaments with in the rift system. The graben is filled with sediments, which drift out of the basaltic rocks on the western and eastern highlands, mainly due to weathering and erosion of the rocks. The colluvial sediments on the sides and feet of the highlands, especially on the west, are very coarse grained and highly permeable. These deposits are largely composed of gravels and cobbles with sands and silts as matrix. The sediments are poorly sorted and sub-angular in shape because they may not have transported far distance from their origin. Adjacent to the colluvials there are alluvial deposits that are mainly transported by surface run-off along streams and rivers during the rainy seasons. The alluvial sediments are relatively fine, well sorted and sub-rounded to spherical shaped. They are dominantly composed of sands and silts with some proportion of fine gravel, especially adjacent to the streams and rivers in the study area. Finally, towards the centre of the valley floor, the sediments become dominated by finer proportions like clay, silt and sand. In general, borehole lithological logs and geophysical surveys in the valley floor has indicated that the valley fill is composed of intercalating layers of gravel, gravely sand, silty sand, clay, and silty clay with increasingly coarser materials towards the top of the succession (RVADP,1998) .The Rivers from the highlands flow in the transversal direction of the Tertiary volcanic succession. Therefore, sediments carried by these rivers should carry materials that are eroded from all the volcanic layers described in the previous sections. Although there is no complete information on the spatial distribution of the thickness of the valley fill sediments, the bore holes drilled in the valley floor and conducted geophysical explorations by different bodies has shown that the sediment thickness varies from few meters on the western and eastern escarpments to above 250m in the central valley floor. The deepest part of the unconsolidated sediments in the valley floor is thought as lacustrine deposit (German consult, 1977 cited in RVADP, 1998).

These Quaternary deposits can be considered as alluvial fan deposits that inter-wave each other and the gentle slopes of the valley floor might promote changes of stream courses from time to time. Therefore, buried streams of predominantly alluvial sediments can be expected in the valley (Gershanovich, 2000). Most intermountain valleys in the study area are also filled with colluvial and alluvial deposits. Such areas are found around the towns of Maichew and Shenko Majo.

The study area is basically composed of the Tertiary volcanic rocks. In previous studies of RVADP (Geo, 1996), Desie Nadew, (2003) and Ermias Hagos, (2005) of the study area a sort of rock classification was made by which is summarized as follow:

The major volcanic rocks are alkaline olivine basalts with minor rocks of tholeiitic affinity that seems to have come later, as dikes, sills and cap rocks.

The Amygdaloidal basalt rock type is medium to coarse grained with porphyroblasts of pyroxene and olivine. It is vesicular where white secondary minerals named as Zeolite fill vesicles. The amygdaloidal basalt is usually found in the lower zone of the volcanic succession overlying the alkali olivine basalt.

The alkali olivine basalt is also medium to coarse grained but with out vesicles. It is friable and highly susceptible to weathering and erosion. In some places where it is affected by local tectonics it shows jointing. Carbonates and Zeolite secondary fillings fill these joints. This rock mass is usually dissected dykes which are reddish brown colored and fine grained. These dykes are highly fractured and jointed due to fast cooling. In addition, interbeds of porphyritic basalts with speheroidal weathering are also observed in some places.

The Aphanitic (fine grained) basalt rock type is found in a form of highly jointed dykes with variable trends. The dyke material is some times coarser looking more like dolerite. It is generally dark colored and contains interbeds of spheroidally weathered rocks like in the A.O.B. These dyke intrusions were acting as feeders for the thick flood basalt accumulations (Mohr, 1988 cited in Dessie Nadew, 2003). As one goes up ward in the rock sequence the aphanitic basalt started to form a layer. Surface expression of this rock unit is in the form of pebbles and boulders.

The Agglomerates *(or Basic tuffs)*, unlike the other units, this unit do not form well-defined beds, instead mostly takes the form of big blocks and irregular masses within the basaltic rock. This rock imperceptibly grades into an amygdaloidal variety.

The volcano–sediments units, which are encountered by the RVADP (1996) in the study area on the way from Maichew to Mekelle, are ash and lapili tough forming thin horizontal layers. There are also interbeds of rhyolite and ignimbrite at the top.In general, the volcanic rocks become progressively acidic as one goes up the volcanic succession.

Sedimentary rocks*,* on the eastern highland there is a sedimentary rock succession of mudstone, sand stone and limestone from top to bottom respectively. These sedimentary rocks form steep cliffs that extend from Adi-bedera to Hade Alga. Their sedimentary beds dip 260 towards NE and strike N100W. The mudstone and sand stone formations exhibit variegated colors including reddish, yellowish and whitish. The sand stone is more or less equigranular medium sand. The limestone is siliceous type. It exists in form of beds or thin tabular sheets.

3.3 Geologic Structurs

The study area and its surroundings are structurally similar to the steep fault scheme of western Afar margin (Zanetin et al, 1978) (Fig 3.2A). Such structural set up was clearly visible in the aerial photographs and 3D model of the study area (Fig 2.7). Most lineaments that were traced from aerial photographs were oriented nearly in the N–S and E–W directions (Figure 2.4). The N-S striking faults are responsible for the easterly tilting and throw of block in that direction. But the E-W striking faults result to southerly tilting and minor throws of blocks. Most of the rivers and streams that flow in the E-W direction from the highland areas are indications of lineaments oriented in that direction. One fault line is also traced along Sulula River.

In general, the fracture pattern in the study area does not signify a single direction while the dominant strike directions are NNW-SSE and NNE-SSW (Dessie Nadew, 2003).

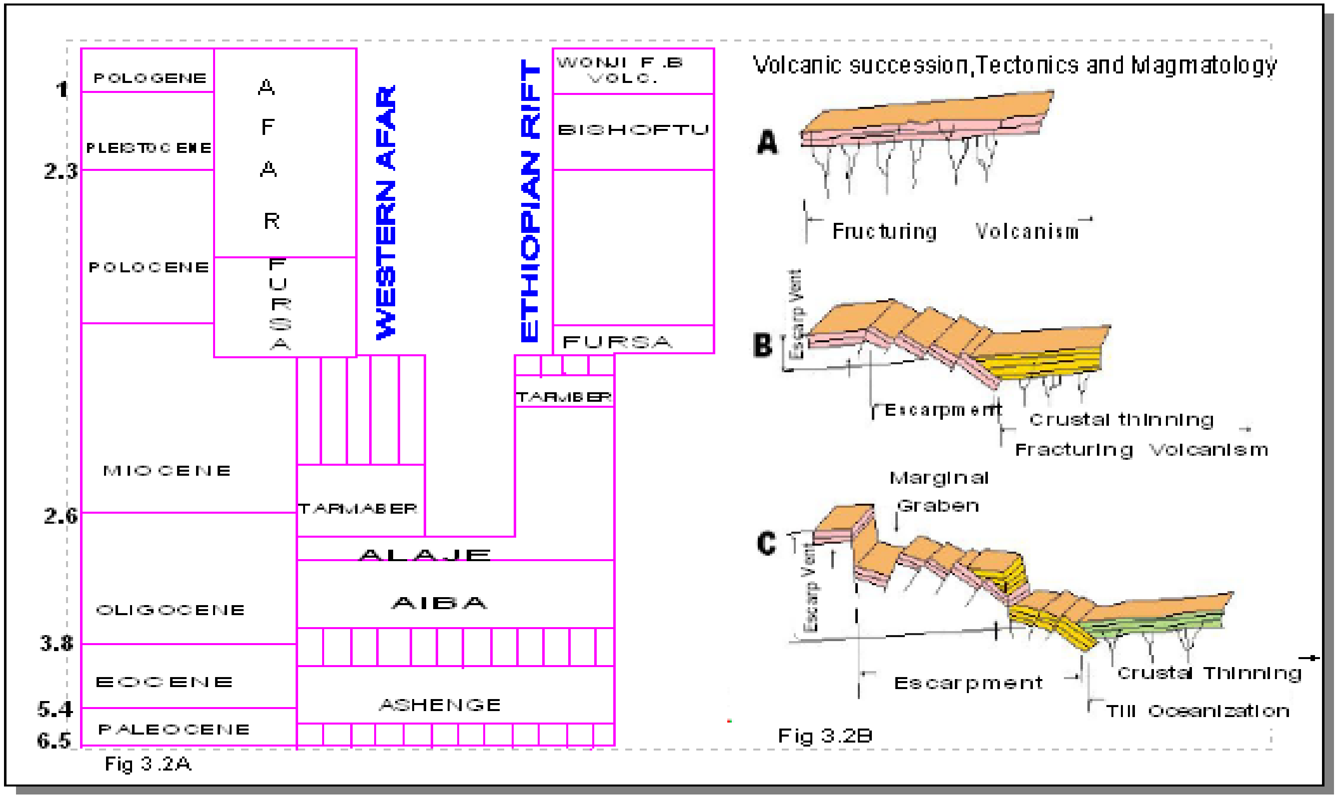


Fig 3.2A. Schematic Stratigraphic Section of the Tertiary Volcanic In Ethiopia (Zanettin Et.Al, RVADP, 1998)

Fig 3.2B. Tectonic and Volcanic Evolution of the Western Margin of Afar A to C

A) The fracturing, magamatic injection and Volcanism are restricted to increasingly narrower belts

B) The width and height of the escarpment increase while the Rift narrows

C)The narrowing of the Rift is opposed by thinning of the crust(Zeneti et.al RVADP,1998)

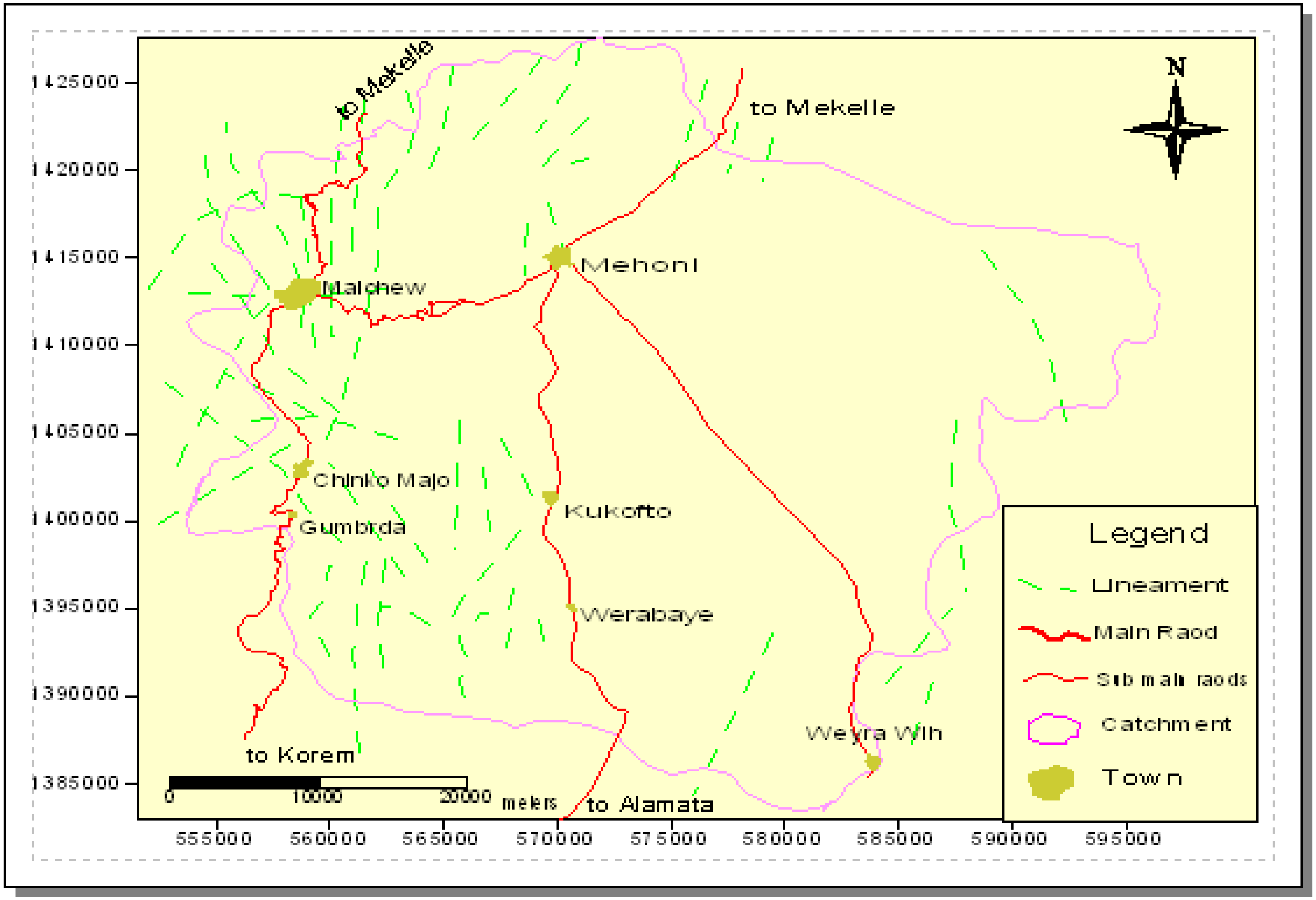


Fig3.3 Lineament map of the study area

CHAPTER FOUR

CONCEPTUAL MODEL OF THE GROUNDWATER SYSTEM

4.1 General

A conceptual model is a pictorial representation of the groundwater flow system, frequently in the form of a block diagram or a cross-section (Anderson and Woessner, 1992). Ground water flow models attempt to represent an actual ground water system with a mathematical counterpart, and the dimensions of the numerical model and the grid design depend on the nature of the conceptual model. The development of a conceptual model is the most important stage in ground water flow modeling work as it simplifies the field problem and makes the organization of the associated field data easier so that one can readily analyze the system. The initial stage of formulating a conceptual model is to define the study area (boundaries of the model) and the conceptual model should be a valid representation of the important hydrogeologic conditions. It is worth mentioning that before making any attempt of ground water flow modeling, the system should be conceptualized and all important data for the modeling work should be assembled in to the conceptual model.

In general the following aspects define the conceptual model:

- Type of aquifers.
- Number of aquifers considered during modeling.
- Relation among aquifers.
- Aquifer geometry.
- Location, magnitude and spatial distribution of the aquifers.
- Location and distribution of surface water bodies.
- Boundary conditions.
- Hydraulic properties of aquifers such as transmissivity,hydraulic conductivity, storativity, leakage coefficient, etc.
- Natural and/or artificial recharge zones.
- Pollution sources.

4.2 Hydrogeological Units

The surfacial geologic units described previously and the deposits at depth were differentiated into aquifer and confining units on the basis of areal extent and general water bearing characteristics. An aquifer is saturated geologic material that is sufficiently permeable to yield water in significant quantities to a well or spring, where as a confining units has low permeability that restricts the movement of ground water and limits the usefulness of the unit as a source of water supply. Permeabilities generally are higher in well sorted, coarse –grained deposits than in fine grained or poorly sorted deposited.

Igneous and metamorphic rock (granite, gneisess etc)yield reasonable (but still moderate amount of ground water when fractured by faulting or weathering Davis and Turk(1964) found that most of the interstitial openings in various types of jointed and faulted crystalline rocks occurred within a depth of about 30m.

Aquifers of unconsolidated or non indurate materials, such as alluvial, glacial or Aeolian deposits, are among the most common sources of groundwater. In alluvial or glacial deposits, buried valleys or old stream beds offer the best groundwater potential. These are ancient stream beds or valleys, primarily consisting of sand and gravel that have been covered by finer sediments (glacial till for example). Because of erosion, bed rock tends to be depressed below unconsolidated sediments of ancestral streams (Bouwer, 19780).

In the Mehoni sub-basin in the Raya valley, saturated coarse grained poorly sorted and angular sediment (quaternary colluvial and quaternary alluvial deposits units) and volcanic aquifer, which have considerable amount of secondary porosities resulting from the effects of extensive weathering, jointing, faulting and fracturing form high productivity aquifer where as poorly sorted fine sand, silt sand, and silty clay materials form low productivity aquifer.

Based on the lithologic descriptions from 26 well logs and 18 vertical electrical sounding geophysical survey results in the area, four principal hydrogeologic units are recognized by grouping lithologies of similar permeability. These are:

- Quaternary colluvial and alluvial deposits (Qcol & Qal)
- Interfluvial, fan foot plains and valley bottom quaternary deposits (Qsmc)
- Volcanic aquifer (Va)
- Bed rock

The quaternary colluvial (Qcol) deposits have characteristic coarse, poorly sorted and sub angular. By retarding the fast flowing run off over the steep mountain slopes they enhance infiltration and ground water recharge. Due to their extremely high porosity and permeability, the infiltrated water do not stay long in these deposits, rather it percolates down in to adjacent volcanic or alluvial aquifers. They are found on the gentle hill slopes and feet of the mountain especially on the western plateau and escarpments. These deposits result from fall and flowing of fragments derived from weathered rocks and subsequent erosion owing to gravitational attraction.

The Quaternary alluvial deposits (Qal) are found as alluvial fans and river bank deposits. The alluvial fans are found at the mouth and its surroundings of large river valleys that drain mainly from the western highlands. But, the riverbank deposits intrude far down into the center of the main valley of the study area. The Quaternary alluvial deposits are composed of medium to coarse sand and fine gravel with little proportions of very coarse and fine materials. The grains are usually rounded to sub-rounded and well sorted. These deposits were generally considered as the main aquifer media in the study area.

Most water wells for large-scale irrigation and water supply purposes in the valley floor were drilled into these quaternary colluvial deposits (Qcol) and Quaternary alluvial deposits (Qal) aquifers. Also water supply for people around Maichew town is pumped from wells drilled into quaternary colluvial deposits (Qcol) and Quaternary alluvial deposits (Qal) aquifers.

Larger part of the valley bottom unconsolidated deposit is comprised of the interfluvial, fan foot plains and valley bottom Quaternary deposits (Qsmc). This sub-unit of the alluvial aquifers is dominantly found in the central part of the valley floor. It is mainly composed of intercalations of medium to fine sand, silty sand, sandy silt and silty clay. Relatively larger proportion of fines is found in these deposits. But still there are layers and patches of sand and gravel within these deposits, which have fairly good transmisivities. Thickness of the Qsmc deposits may reach up to 250m or greater and there are two ideas about the material that may underlay these deposits. It can be thick lacustrine deposit, which is clay dominated or it is highly fractured bedrock, which is down thrown by the tectonic activities that form the valley.

The volcanic rocks particularly the Ashange basalts are the second important aquifers in the study area. These rocks, which form the western and eastern highlands bear considerable amount of secondary porosities, resulting from the effects of extensive weathering, jointing, faulting and fracturing. Therefore, these plateau basalts in and the intermountain sediments may render significant amount of groundwater. There are some productive boreholes, which tap groundwater from the volcanic aquifers. Their productivity varies from place to place depending on the availability of local recharge, degree and depth of weathering in the rocks, density and penetration depth of master joints and presence of regional fault zones. Degradation of permeability of the fractures and joints due to the clay rich weathering products of the basaltic rocks should also be considered. In general, the volcanic aquifers are important sources of potable water and small-scale irrigation in the highland and escarpment areas.

Bed rock unit underlain all the above described hydrogeologic units in the study area there area two ideas about this unit .it can be thick lacustrine deposit, which is clay dominated or it is highly fractured bed rock , which is down thrown by the tectonic activities that form the valley.

4.3 Hydraulic Characteristics

The hydraulic conductivity of the materials in an aquifer or confining unit is a measure of the ease which water can move through the material. It is the function of properties of both the matrix and fluid. Water in the regional flow system was assumed to have a uniform density and viscosity, and thus hydraulic conductivity only varies as the grain size, shape, sorting, and packing vary (Freez and Cherry, 1979). Because matrix properties may vary over short distance, the hydraulic conductivity also may vary over short distances.

Horizontal hydraulic conductivities generally were greater than the vertical hydraulic conductivities, as a result of the depositional history of the sediment. Horizontal hydraulic conductivities may be determined from single or multiple well aquifer tests. Source of information on the hydraulic characteristics of the sub surface material in the study area include water well (well log), aquifer test and literature.

Hydraulic characteristics for quaternary colluvial deposits (Qcol) and Quaternary alluvial deposits (Qal) unit were estimated from puming test data of 12 boreholes in the study area, which were recorded during drilling. The recorded data for these wells were used for calculating the transmissivity and horizontal hydraulic conductivity, because some of them have lithologic logs that contain discharge rate, time of pumping, draw down, static water level, well construction data and lithologic logs.

As observed from the records of the pumping test data of the above wells, recorded at different construction time, the wells are pumped under different pumping rate for different pumping time. Their yield ranges between 3l/s to 44l/s depending on the yield capacity of the wells for 4.5hr to 72hr pumping duration. The pumping test data for different wells were analyzed by plotting water level versus time using aquifer test software “Aquifer test 3.5” and the aquifer parameters (transmissivity and conductivity, if the aquifer thickness or water column or screen length is known) were obtain the hydrogeologic unit. From the Theis recovery analysis the transmissivity value for the aquifer unit 58.33% of the wells were greater than 900m^2/d that shows highly potential aquifer and 41.67% of the well indicate moderately potential aquifers with transmissivity ranging between 216 to 475 m^2/d (Sen., 1995).

The transmissivity computed by Theis recovery analysis for BH25, BH24 and BH66 from (RVADP, 1998) were 121, 1126.5 and 8.52 m2/d respectively. Wells analyzed by (Ermias Hagos, 2005) were BH57, BH58and BH59, their transmissivty values indicate moderate potentially aquifer.

The wells that were drilled into the basaltic rock aquifers of the study area are located in the western and eastern highlands and escarpments. Some of these wells are BH25, BH01, BH73, BH04, BH02, BH03, BH06, BH10, BH55, BH13, BH17, BH15, BH16 and Menora. BH25 was the only well with full pump test data among those found in the eastern basaltic areas.Their total depth ranges between 43.25m -100m. Their static water levels also range between 1.7m - 48.55m and their yields range between 0.7- 5.62 l/s.

As, it was observed from pre-existing transmissivity data and pump test analysis of 4 wells, the transmisivity values of the volcanic aquifers range between 0.66 m^2/day 81.25 m^2/day. The two extreme values were found from wells close to each other which may indicate a short distance variability of the degree and depth of weathering and fracturing of the basaltic rocks. Productivity of these volcanic aquifers is also highly dependent on the availability of enough local recharge. According to Sen's classification (table 4.1) the transmisivities found from those wells ranges from moderately potential aquifers -weak potential aquifers.
According to Sen (1995) aquifer potentiality is classified as table 4.1

Table 4.1 Aquifer potentiality

Transmisivity (m^2/day)	Potentiality
T>500	High
50<T<500	Moderate
5<T<50	Low
0.5<T<5	Weak
T<0.5	Negligible

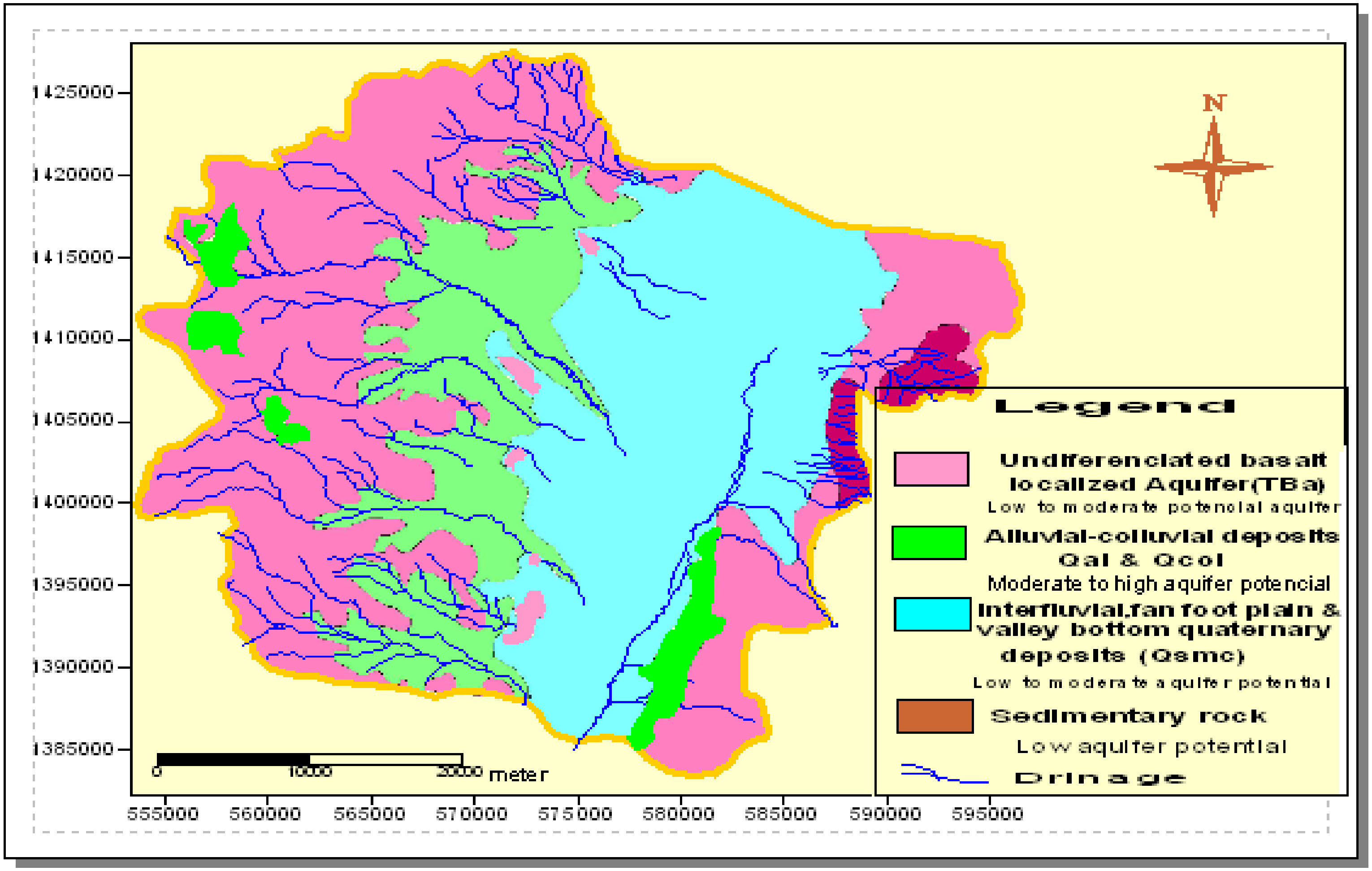


Fig4.1 Aquifer Potentiality Map of the Study Area

Hydraulic conductivity is a measure of the ability of a fluid to move through interconnected void spaces in sediments or rocks. It is a function of both the fluid and the medium. The higher the hydraulic conductivity of a rock/sediment, the higher the water yielding capacity of the rock is. The hydraulic conductivity of fractured rocks depends largely on the density of the fractures and width of their apertures.

Most of the wells with pumping test data of the study area are not uniformly distributed. Aquifer characterization made by Desie Nadew, 2003 and Ermias Hagos, 2005 using Sen (1995) aquifer potentiality. This aquifer characterization is difficult to quantify the hydraulic conductivity of the whole area even if the aquifer thickness, or screen length is known, Even though wells with known screen length are concentrated in small portion of the area ,it posed difficulty in mapping the hydraulic conductivity the area. Therefore, an assumption has been made to estimate the hydraulic conductivity in this study, by taking wells with transmissivity, static water level (SWL) and total water depth. Table 4.2 indicates estimation of hydraulic conductivity of wells by transmissivty divided by water column. Therefore the hydraulic conductivity map of the area prepared for the model input.

Table4.2 Hydraulic conductivity estimation from wells with transmisivity and water column

BH-ID	UTM		Total depth (m)	SWL (m)	Water column(m)	T (m^2/d)	K (m/d)
	X	Y					
BH-03	559938	1412807	87	8.7	78.3	3.68	0.046999
BH-07	558028	1415118	55.45	17.1	38.35	28.92	0.754107
BH-08	557183	1415300	80	21.2	58.8	2	0.034014
BH-09	558702	1409774	67.7	41	26.7	87.46	3.275655
BH-10	558853	1409498	54.4	24.58	29.82	0.66	0.022133
BH-11	556671	1414869	43.25	12.45	30.8	81.25	2.637987
BH-12	556892	1409954	37	12.13	24.87	19.49	0.783675
BH-13	569309	1392764	70.66	39.1	31.56	22.16	0.702155
BH-14	571531	1394926	75	35.5	39.5	11.4	0.288608
BH-18	574905	1419662	60	15.35	44.65	22.7	0.508399
BH-21	569881	1415748	66	36.72	29.28	71	2.424863
BH-24	576688	1406735	95	60.5	34.5	1126.5	32.65217
BH-25	577755	1402683	85	44.4	40.6	156	3.842365
BH-27	580035	141322	160	75.5	84.5	32.3	0.38224

		0					9
BH-28	576492	1415884	147	97	50	46.5	0.93
BH-30	570637	1403161	78	40.5	37.5	474	12.64
BH-31	573300	1388200	107	40.5	66.5	470	7.067669
BH-40	574600	1409500	147	46.74	100.26	1107	11.04129
BH-46	575500	1408500	144	43.9	100.1	1776.894	17.75119
BH-53	569802	1400032	82	28.9	53.1	4200	79.09605
BH-54	567162	1394661	114	77.48	36.52	1240	33.954
BH-55	567508	1400144	123	36.7	86.3	1250	14.48436
BH-56	571951	1400364	120	29	91	965	10.6044
BH-57	585118	1413417	140	64.2	75.8	419	5.527704
BH-58	577527	1396864	110	36.25	73.75	436	5.911864
BH-59	578474	1398916	120	38.92	81.08	98.9	1.219783
BH-64	583291	1399961	99	43.23	55.77	94.2	1.68908
BH-69	585553	1399786	100	41.65	58.35	42	0.719794

4.4 Aquifer System Boundary

For groundwater system of a given area, there are two types of aquifer boundaries which control the groundwater flow direction (Anderson and Woessner, 1992). These boundaries are; 1) Physical boundaries of groundwater flow system which are formed by physical presence of an impermeable body of rock or a large body of surface water. 2) Other boundaries form as a result of hydrologic conditions. These invisible boundaries are hydraulic boundaries that include groundwater divides and streamlines.

In groundwater flow modeling, boundary conditions influence the extent of the flow domain to be analyzed or simulated. The extent of the flow domain is initially determined by the extent of the area of concern and it is preferable if it is bounded by physically observable features. Moreover, it should be noted that correct conceptualization of boundary is important to select an appropriate mathematical representation in the model so that the effect of the boundary on flow can be correctly understood.

The Mehoni sub basin in the Raya Valley, aquifer system boundary was carefully assigned based on field visits made and existing works. The geographic boundaries of Mehoni sub-basin groundwater flow model approximately correspond closely with natural hydrologic boundaries across which groundwater flow was assumed to be negligible. This is true for the western, eastern, northwestern and northern boundaries of the system. The Maichew Mountains in the north, Chercher Mountains in the east, the Central Ethiopian plateau in the west and the Chegwara ridge in the south surround the study area with a narrow southerly surface water outlet. On these sides, major topographic divides were assumed to coincide with groundwater divides. It should be remembered that groundwater divide is not really a boundary in nature, but as groundwater on either side of the divide flows away from the divide and not across it, the divide itself acts as a no flow boundary. It is worth noting that the position of such a boundary changes based on stress applied to the system. In the case of Mehoni sub basin it was assumed that the effect of the stress applied does not go beyond the groundwater divides so that the divides might be consider as boundaries. A narrow southerly surface water outlet consider as general head boundary.

In addition, the upper boundary of the modeled sub-basin was assumed to be the water table and recharge was used to represent the boundary flux. Although not known to exact, the lower boundary was considered as a no flow because the aquifer was assumed to be underlain with an impermeable rock body.

4.5 Ground Water recharge and Discharge

In nature different types of recharges occur together, which lead to different recharge estimation techniques and this makes the quantification process difficult. In this study, direct recharge from precipitation was considered and it occurs in this area as there is a surplus of rainfall over evapotranspiration. The groundwater level of an area is directly influenced by the amount of water that enters the aquifer system, that is, areas with higher recharge amount have shallow ground water level and in general, groundwater flows from recharge areas to discharge areas.

The spatial distribution of recharge is affected by a number of natural and anthropogenic factors, some of which are climatic factors, geology, and vegetation cover, topography and land use/cover. Based on all these factors, the global net recharge estimated for the whole country ranges from 50-150mm (Tesfaye Cherinet, 1988).

The previous studies didn't classify the study area in to different recharge zonetion. For this conceptual model zoned in to three recharge zones: western, eastern and valley fill. The major source of recharge for the valley fill groundwater is recharge from runoff generated from the two flak highlands and recharge from precipitation with in the valley fill.

The annual groundwater recharge of the valley fill from runoff and precipitation is estimated by using Empirical (hydrologic) method. Recharge from runoff generated at the mountain area and flowing to the valley fills. The volume of runoff generated from the mountain area and flowing to the valley fill is calculated using runoff coefficient of the area to determine the runoff from precipitation by the formula

$$V_r=C_m* P_m* A_m$$

Where

V_r = Annual volume of runoff (m^2)

C_m = Runoff coefficient of the mountains area

P_m = mean annual rainfall at the mountains area (m)

A_m= area of the mountains area (m^2)

The average runoff coefficient of the mountainous area is assumed to be 0.2(RVDP, 1996) and the mean annual rainfall 723.57mm (Ermias, 2005). Area of the mountains is approximately $608x10^6 m^2$ then V_R is equal to $87.98x10^6$ m^3. The percent of runoff that recharges the groundwater of the valley fill is determined by V.V Drozda formula.

$$\alpha = \frac{C_{sw} - 20}{C_{gw} - 20}$$

Where

α = percent of runoff that recharges the groundwater of the valley fill (recharge coefficient)

C_{sw}= Total dissolved solids in the runoff (mg/l)

C_{gw}= Total dissolved solids in the groundwater (mg/l)

Substituting the hydro-chemical analysis data (from Ermias, 2005) of Haya River (C_{sw}=207mg/l) and the nearby borehole BH-63 (C_{gw}=982mg/l) then α is equal to 0.19

The recharge from the runoff is estimated by the formula

$R_{rm}=\alpha * C_m * P_m * A_m$ or $R_{rm}=\alpha *V_r$……………………

Where

R_{rm}=Recharge of runoff from the mountains and flowing to the valley. Therefore, the annual groundwater recharge of the valley fill from the runoff generated at the mountains area flowing to the valley fill is estimated to be $1.67x10^7$ m^3/year.

The estimation of recharge from precipitation in the valley fill is based on the studies undertaken by different organization using correlation of groundwater level fluctuation and rainfall for alluvial deposits in different parts of the world. Especially, the study undertaken in India by Sehgat (1973) for alluvial deposits in Panjub has the following relationship formula:

$$R_p = \beta (P_v - p_o)^n \times A_v$$

For this study, estimation of the values of β, P_o and n are assigned on literature review for similar geological, hydrological and hydrogeological conditions of groundwater basin. $\beta = 0.25$, P_O=380, and n=1(RVDP, 1996), the value of

$$R_p = 0.25(P_v - 0.380)A_v$$

Where

R_p=recharge of precipitation with in the valley fill (m^3/year)

R_v= mean annual precipitation with (624mmor 0.624m)

A_v=area of the valley fill (m^2), $575x10^6$

The annual recharge from the precipitation with in the valley fill is estimated to be35.1x10^6 m^3/year. Therefore, total annual recharge of the valley fill is the sum of recharge from runoff and precipitation and equal to 51.7751x106 m^3/year or 0.246mm/d.

For the western and easten high land, the recharge were computed from Ermias Hagos (2005). Accordingly the annual recharge of the total sub-basin was given 60.49 x10^6 m^3/year. So the recharge amount 0.27mm/d was used. These two values were used as an initial input parameter of recharge.

The ground water use of the Mehoni sub basin was steadily increasing from time to time as the population increased and groundwater used for irrigation purpose. The groundwater discharge from the modeled area is mostly by withdrawal from wells. Groundwater discharges from the system by evapotranspiration, where the water table is shallow. The total discharge by groundwater evapotranspiration is unknown but is presumed to be insignificant compared to groundwater pumpage from the system.

The pumping system of the groundwater from the sub-basin aquifer system is divided into two:

A) The shallow drilled wells are fitted with hand pumps and used for rural water supply. There are 13 shallow wells in the study area. Their yield ranges from 0.5 to 1.5 l/s and serves on average for seven hours a day for the whole years Based on this, it was assumed that the ground water abstraction by those shallow wells is 77.935m^3/d.

B) The deep wells, installed with submersible pumps, are used for public water supply and irrigation purposes. There are 34 wells drilled in the study area. Average groundwater abstraction from those wells assumed to be 13.89246 x10m^3/d.

The average body weight of one TLU (Tropical Livestock Unit) is 250 kg (FAO, 1984). On average an animal consumes 1 liter of water per day for each 10 kg of body weight, therefore, about 25 liters of water is required daily for each Livestock Unit. The total livestock population of this study area is estimated 31018TLU. Therefore, 775.45 m^3/d of water is a daily minimum requirement for the livestock; wells without beneficiaries were computed using yield average eight hours of work. Therefore average discharges of those wells were 4537.152m^3/d.

So the total discharge from the wells were15383.143m^3/d and was used as an initial discharge from the sub-basin.

BH ID	UTM			Beneficiaries		Rate (DPCPD)(m^3/d)		Pumpage (m^3/d)		Total (m^3/d)
	Easting	Northing	Elev.	Dom	Ls	Dom	Ls	Dom	Ls	
BH-25	577755	1402683	1555	243	2927	0.011	0.025	2.673	73.175	75.848
BH-53	569802	1400032	1623	300	1500	0.011	0.025	3.3	37.5	40.8
BH-54	567162	1394661	1657	110	1419	0.011	0.025	1.21	35.475	36.685
BH-55	567508	1400144	1672	350	1500	0.011	0.025	3.85	37.5	41.35
BH-56	571951	1400364	1585	195	1905	0.011	0.025	2.145	47.625	49.77
BH-57	585118	1413417	1573	200	3000	0.011	0.025	2.2	75	77.2
BH-58	577527	1396864	1571	139	1419	0.011	0.025	1.529	35.475	37.004
BH-59	578474	1398916	1518	150	903	0.011	0.025	1.65	22.575	24.225
BH-60	580815	1402660	1526	120	1500	0.011	0.025	1.32	37.5	38.82
BH-61	574151	1405561	1652	600	3000	0.011	0.025	6.6	75	81.6
BH-64	583291	1399961	1536	140	2200	0.011	0.025	1.54	55	56.54
BH-65	585553	1399786	1555	255	955	0.011	0.025	2.805	23.875	26.68
BH-66	574827	1392822	1525	354	3200	0.011	0.025	3.894	80	83.894
BH-67	572536	1389879	1540	285	2820	0.011	0.025	3.135	70.5	73.635
BH-69	585460	1403104	1524	95	970	0.011	0.025	1.045	24.25	25.295
BH-72	570147	1402326	1653	210	1800	0.011	0.025	2.31	45	47.31
									Total(m3/d)	816.656

Table4.3 Average daily groundwater abstraction of Raya Azebo Wereda

Table4.4 Average daily groundwater abstractions for irrigation purpose

BH ID	UTM			Pumpage	Average	Pumpage
	Easting	Northing	Elevation	(m3/h)	workinghr/d)	(m3/d)
BH-41	575889	1407832	1620	122	8	976
BH-42	573600	1408650	1650	135	8	1080
BH-43	574000	1406000	1631	108	8	864
BH-44	578000	1404000	1554	115.2	8	921.6
BH-45	574932	1408935	1639	127.8	8	1022.4
BH-46	575500	1408500	1627	114.48	8	915.84
BH-48	576046	1403659	1579	64.8	8	518.4
BH-50	577000	1407000	1598	129.6	8	1036.8
BH-51	577250	1409991	1622	157.68	8	1261.44
					TotalQ(m3/d)	8596.48

Table4.5Average water supply for Maichew town

BH ID	UTM			Q(m3/d)
	Easting	Northing	Elevation	
BH-01	558641	1413019	2405	58.5
BH-02	559360	1412779	2408	117
BH-03	559938	1412807	2400	20
BH-04	557946	1412790	2468.5	110.5
BH-08	557183	1415300	2480	150
BH-09	558702	1409774	2410	143
			T.Q(m3/d)	599

Table 4.6Average water supply for Mehoni,Weyra Wiha and Kukufto

Mehoni

BH ID	UTM			Q(m3/d)
	Easting	Northing	Elevation	
BH-62	568318	1413524	1786	300

WeyraWiha

BH ID	UTM			Q(m3/d)
	Easting	Northing	Elevation	
BH-27	580035	1413220	1600	300

Kukufto

BH ID	UTM			Q(m3/d)
	Easting	Northing	Elevation	
BH-53	569802	1400032	1623	156

Table4.7Average Groundwater Abstraction from Shallow Wells

BH ID	UTM			Beneficieries	DPCPD(m^3/d)	Pumpage
	Easting	Northing	Elevation			(m^3/d)
SBH1	560035	1412245	2411	350	0.011	3.85
SBH2	560059	1415224	2419	260	0.011	2.86
SBH3	558189	1415313	2459	550	0.011	6.05
SBH4	558162	1414084	2424	300	0.011	3.3
SBH5	555034	1409962	2536	200	0.011	2.2
SBH6	556664	1409286	2433	2500	0.011	27.5
SBH7	557419	1408707	2427	200	0.011	2.2
SBH8	588803	1409969	2410	400	0.011	4.4
SBH9	559258	1408895	2373	580	0.011	6.38
SBH10	558985	1409536	2390	225	0.011	2.475
SBH11	556933	1410150	2460	1000	0.011	11
SBH12	556814	1409443	2433	320	0.011	3.52
SBH13	555607	1409963	2539	200	0.011	2.2
					T.pumpage(m^3/d	77.935

Table 4.8 Average Discharge Rate of Wells from Their Yield

X	Y	Z	SWL	BHID	Yield(l/s)	Yield(m3/d
559903	1412019	2440	17.1	BH-06	3	86.4
558028	1415118	2455	21.2	BH-07	6.94	199.872
558702	1409774	2410	24.28	BH-09	0.78	22.464
558853	1409498	2390	12.45	BH-10	3.7	106.56
556671	1414869	2445	12.13	BH-11	1.8	51.84
556892	1409954	2470	39.1	BH-12	1.5	43.2
569309	1392764	1508	35.5	BH-13	2	57.6
571531	1394926	1590	47.6	BH-14	17	489.6
583324	1385988	1766	50	BH-15	2	57.6
583444	1385989	1775	50	BH-16	3.02	86.976
580500	1388250	1580	15.35	BH-17	4.56	131.328
574905	1419662	1710		BH-18	3.2	92.16
569675	1401249	1640	36.72	BH-20	4.5	129.6
569881	1415748	1890	58.7	BH-21	3	86.4
574204	1410936	1636	60.5	BH-23	5.7	164.16
576688	1406735	1600	44.4	BH-24	3.6	103.68
587750	1416000	1607	75.5	BH-26	2.46	70.848
576492	1415884	1700		BH-28	2	57.6
574336	1419297	1743	40.4	BH-29	0.91	26.208
570637	1403161	1650	40.5	BH-30	3.62	104.256
573300	1388200	1520	47.6	BH-31	2.98	85.824
580750	1419250	1660		BH-32	3.28	94.464
569929	1399050	1623	21	BH-34	3.6	103.68
567959	1410599	1719	26	BH-36	5.62	161.856
574668	1396412	1550	43	BH-38	4.22	121.536
571000	1394000		46.74	BH-39	22	633.6
576000	1400000	1537		BH-47	1.82	52.416
571089	1402927	1644	52.1	BH-52	27	777.6
581179	1397168	1513	34.9	BH-63	2.6	74.88
568477	1392062	1623	38	BH-68	3.45	99.36
586059	1406286	1549	43	BH-70	3.48	100.224
586575	1411430	1583		BH-71	2.2	63.36
					Total(m^3/d)	4537.152

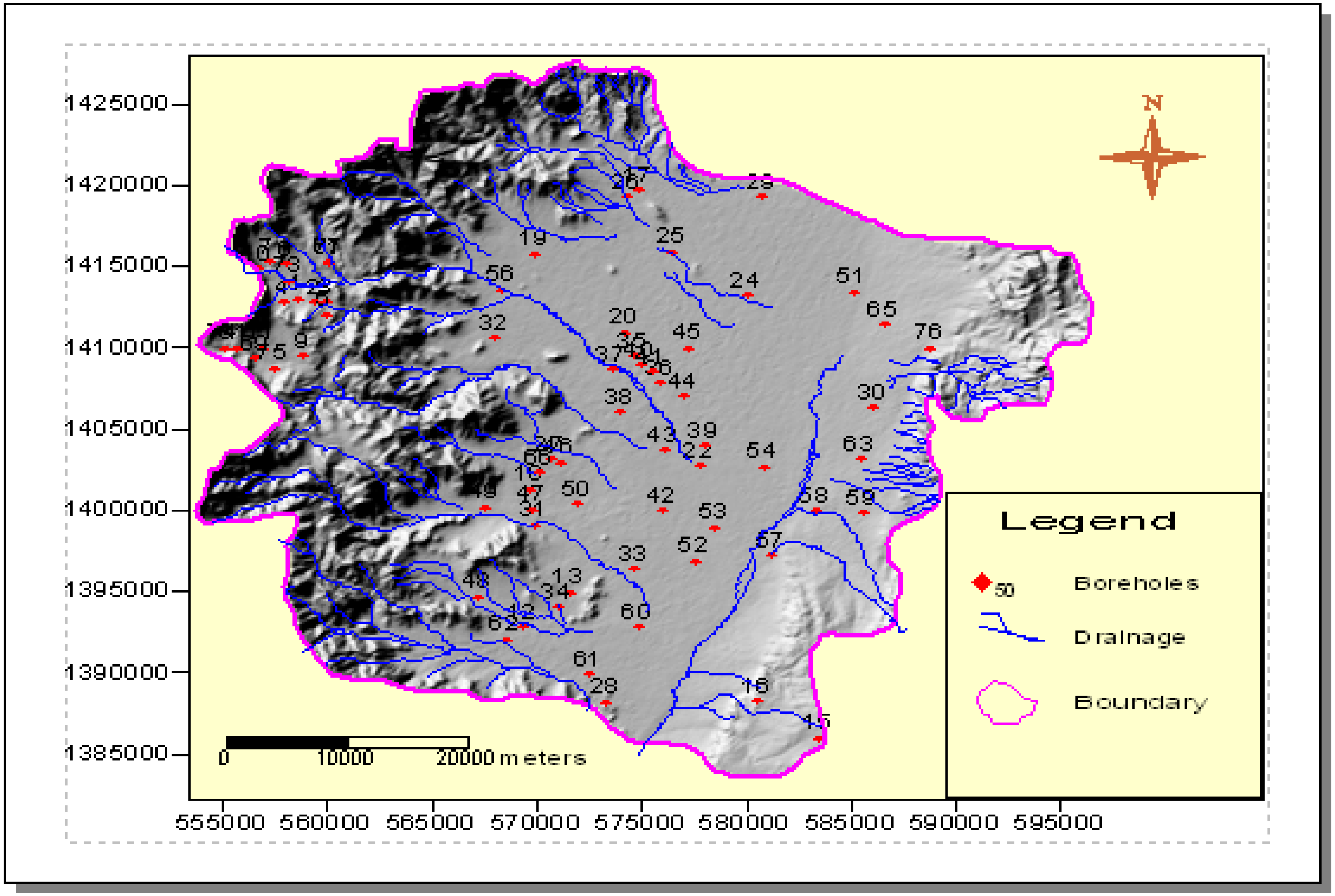


Fig 4.2 Groundwater Pumpage Location

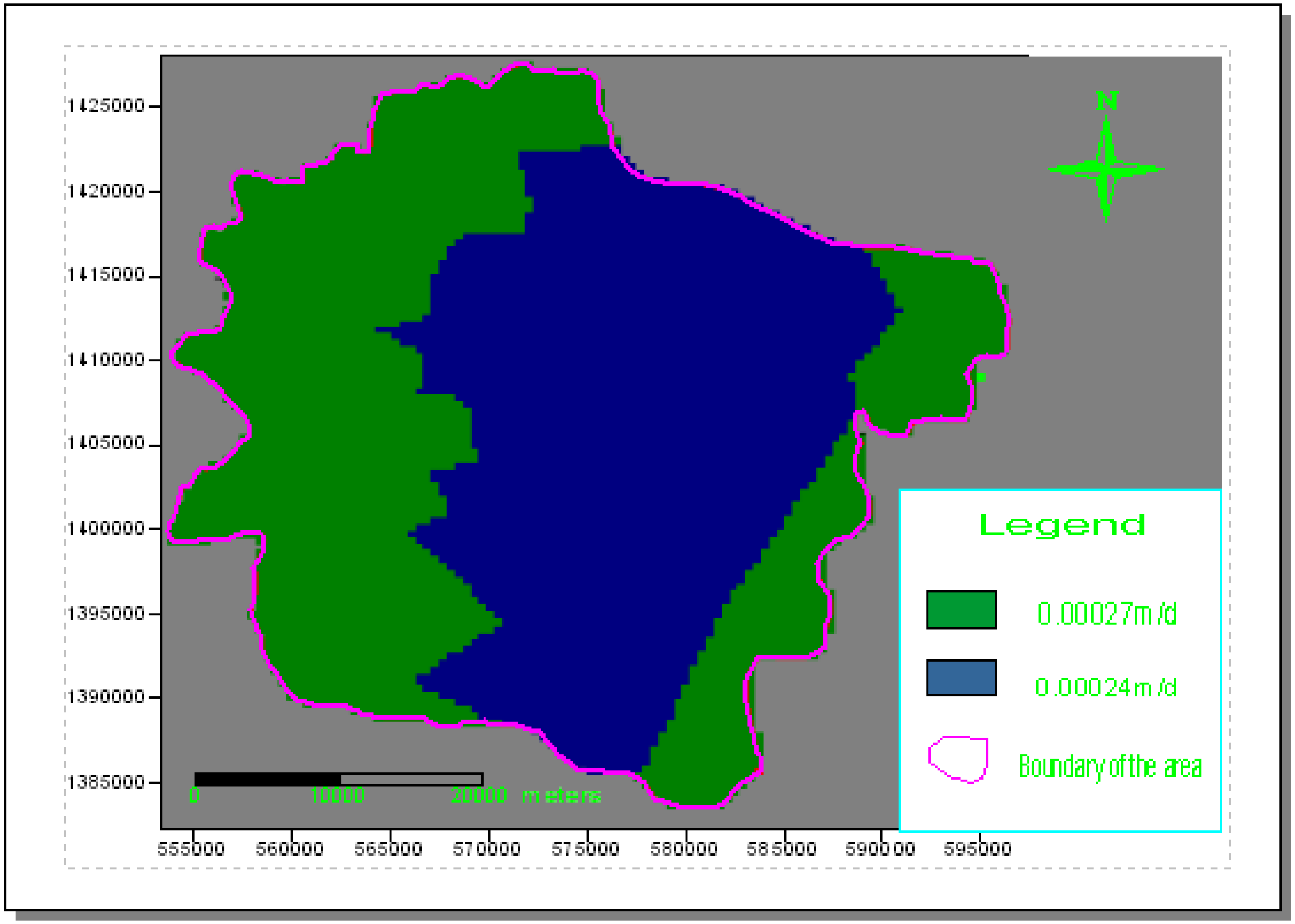


Fig 4.3 Model Derived Recharge Zonation of the Study Area.

4.6 Groundwater level fluctuation and movement

The seasonal groundwater level fluctuation in the western highland areas was greater than in the valley floor. In the wells that are located around Maichew town, it ranges between 1.1m and 18m. In the valley floor at BH63, seasonal fluctuation of 0.215m was detected. More importantly, there was one monitoring well fitted with a data logger, which had been continuously reading the groundwater level fluctuation in the area around kara-Adishawo (Well field 10) for the last 3 years and 6 months. This area is a place where irrigation activities were being practiced by using groundwater for the last 2 Years. Large-scale pumping from the ground water for the irrigation practices was started on 23/3/2003. Before the start of pumping the groundwater level was in a stable position with slight rise of about 0.2 meter in the first one-year reading But after the start of pumping the water level has been lowered by about 2 meters from its original position (figure4.2). In this period the water level had shown a declining trend with temporary rises in the rainy months of the year.

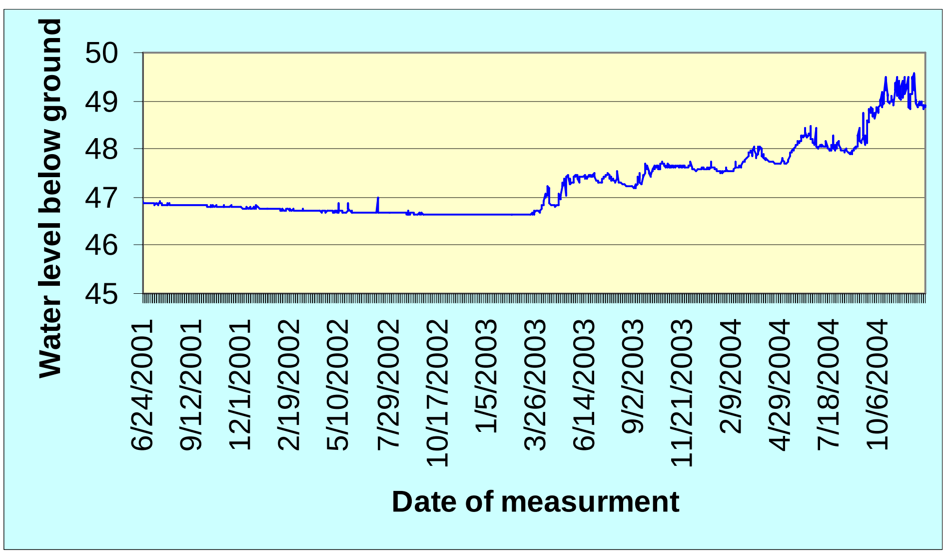


Fig 4.4 Groundwater Level Fluctuation Measured From Monitoring Well In Kara- Adishawo

The groundwater flow system in the study area was mainly treated by using groundwater level measurements taken from different points in the area. Water level measurements were taken from about 46 wells that are located in different parts of the study area. Most of these measured values of the static water level were taken either during construction of the wells or during pump test. But now it was almost impossible to take water level measurements from the wells because they are not fitted with observation pipes. Therefore, most of the static water level data in this work was collected from previously done literatures and from well completion and pump test reports. The static water level values and their measurement dates are listed in appendix 1. It was observed from the readings of the data logger and different time measurements taken from BH63 that the seasonal fluctuation of the ground water level in the valley floor was about 0.2 meter. This value was assumed as insignificant when using the water level data in this work for the evaluation of groundwater flow direction in the study area. Springs and rivers at different points in the area were also considered for the evaluation of the ground water flow systems. Finally, all this data was entered in to the SURFER 8 and contour map was prepared (Figure4.3) where each contour line represents the equipotential lines. Then the lines that are drawn perpendicular to the contour line are considered as flow lines and the groundwater flows from points of higher contour value to the lower ones (Figure 4.3). Based on this it seems that the groundwater in the Mehoni sub-basin flows from all directions towards the center of the valley floor with probable ground water out flow in the southeastern part of the study area and some localized systems that can be due to the control of the bedrock morphology.

Table4.9 Static water level used for groundwater contouring

X	Y	Z	SWL	BH ID	GWLE(m)
558641	1413019	2405	33.6	BH-01	2371.4
559360	1412779	2408	25.3	BH-02	2382.7
559938	1412807	2400	8.7	BH-03	2391.3
557946	1412790	2440	54.02	BH-04	2385.98
558028	1415118	2455	21.2	BH-07	2433.8
558702	1409774	2410	24.28	BH-09	2385.72
558853	1409498	2390	12.45	BH-10	2377.55
556671	1414869	2445	12.13	BH-11	2432.87
569309	1392764	1508	35.5	BH-13	1472.5
571531	1394926	1590	47.6	BH-14	1542.4
583324	1385988	1766	50	BH-15	1716
583444	1385989	1775	50	BH-16	1725
580500	1388250	1580	15.35	BH-17	1564.65
569675	1401249	1640	36.72	BH-20	1603.28
569881	1415748	1890	58.7	BH-21	1831.3
574204	1410936	1636	60.5	BH-23	1575.5
576688	1406735	1600	44.4	BH-24	1555.6
577755	1402683	1555	56.3	BH-25	1498.7
587750	1416000	1607	75.5	BH-26	1531.5

580035	1413220	1600	97	BH-27	1503
574336	1419297	1743	40.4	BH-29	1702.6
570637	1403161	1650	40.5	BH-30	1609.5
573300	1388200	1520	47.6	BH-31	1472.4
586046	1406271	1558	43	BH-33	1515
569929	1399050	1623	21	BH-34	1602
567959	1410599	1719	26	BH-36	1693
574668	1396412	1550	43	BH-38	1507
574600	1409500	1645	51	BH-40	1594
575889	1407832	1620	31.16	BH-41	1588.84
573600	1408650	1650	56	BH-42	1594
574000	1406000	1631	48	BH-43	1583
578000	1404000	1554	58.38	BH-44	1495.62
574932	1408935	1639	43.9	BH-45	1595.1
575500	1408500	1627	36.2	BH-46	1590.8
577250	1409991	1622	39.8	BH-51	1582.2
571089	1402927	1644	52.1	BH-52	1591.9
569802	1400032	1623	28.9	BH-53	1594.1
567162	1394661	1657	77.48	BH-54	1579.52
567508	1400144	1672	36.7	BH-55	1635.3
571951	1400364	1585	29	BH-56	1556
585118	1413417	1573	64.2	BH-57	1508.8
577527	1396864	1571	36.25	BH-58	1534.75
578474	1398916	1518	38.92	BH-59	1479.08
574151	1405561	1652	47	BH-61	1605
581179	1397168	1513	34.9	BH-63	1478.1
583291	1399961	1536	43.23	BH-64	1492.77
568477	1392062	1623	38	BH-68	1585
585460	1403104	1524	41.65	BH-69	1482.35
586059	1406286	1549	43	BH-70	1506

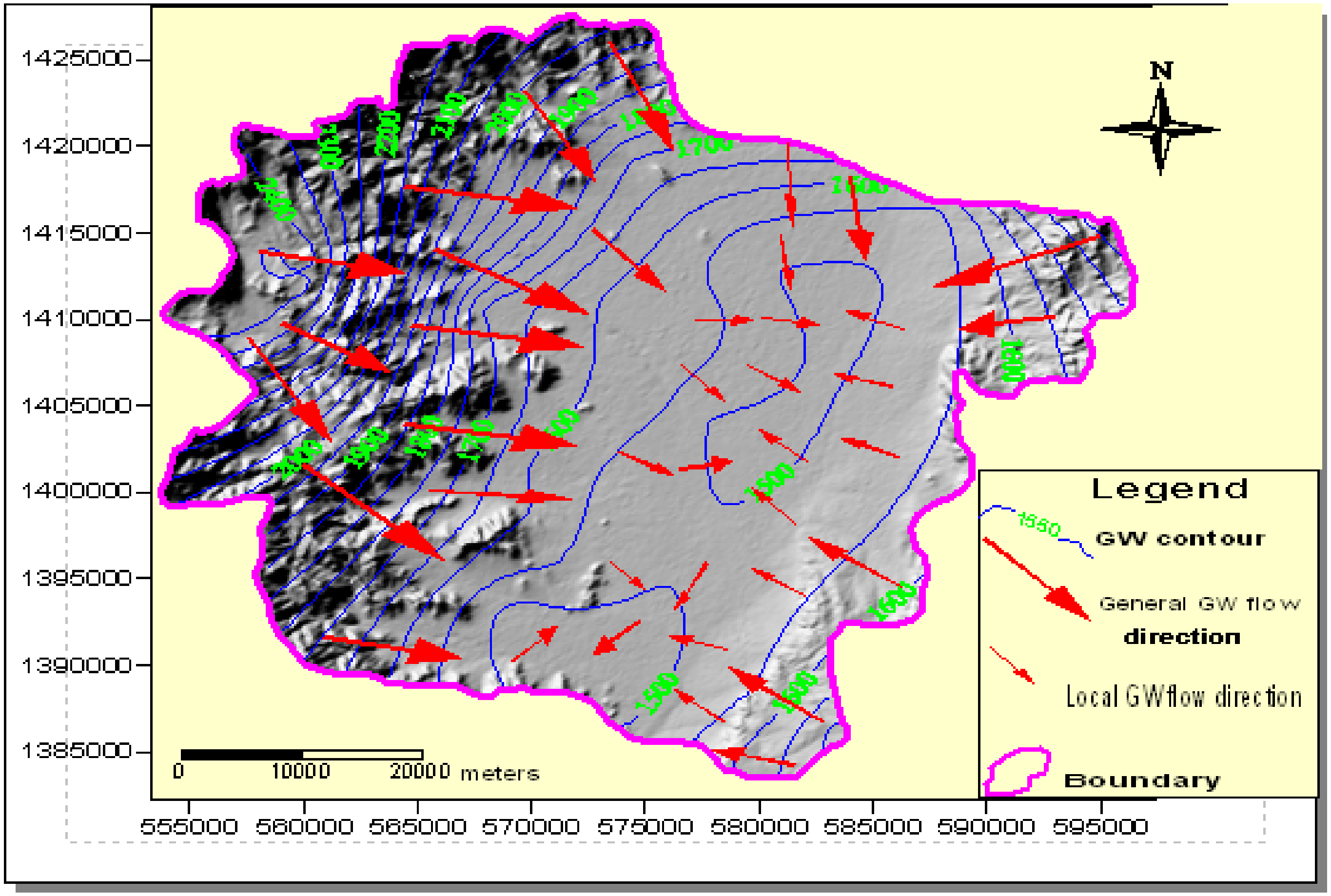


Fig4.5Groundwater equipotential line and Flow Direction Map of the Study Area

4.7 conceptual ground water balance

A steady state groundwater balance for the study area describes alluvial aquifer system and provides a conceptual framework for the groundwater flow model. In the steady state water balance, inflow to and outflow from the aquifer system are identified and quantified on an average annual basis. The water balance described here includes the components simulated in a groundwater flow model.

Inflow for the groundwater of the study area is the groundwater recharge from precipitation. Outflow includes withdrawals from rural town water supply and irrigation wells, groundwater evapotranspiration (not considered for balance) and subsurface outflow at southern part of the study area (not quantified).The estimated amount of groundwater recharge from precipitation is-$1.78 \times 10^5 m^3/d$.The outflow component of the study area is through groundwater withdrawals from shallow wells and deep wells are around $1.53831 \times 10^4 m^3/d$.

Under steady state conditions, inflow to groundwater system should be equal to outflow from the system as changes in storage on annual basis is considered to be negligible. The difference between inflow and outflow in this study may be due to under estimation of recharge or it can be attributed to exaggerated estimation of any component of the outflows. These values were changed during calibration process.

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CHAPTER FIVE

NUMERICAL SIMULATION OFTHEGROUND-WATER FLOW SYSTEM

5.1 General

A numerical groundwater flow model of the Mehoni sub basin in Raya valley is developed to better understand the aquifer system of the basin and to determine the long-term availability of groundwater by simulating groundwater condition at present and predict the future condition under different hydrologic and pumping scenarios for various groundwater management alternatives. The model was developed using assumptions and approximations to simplify the actual aquifer system. The model idealizes the complex geo-hydrologic relations of the actual system, thus, it is a simplification based up on the data and the assumptions used to develop it.

5.2 Modeling Approach

The ground water flow system was numerically simulated in three dimension using MODFLOW, a widely used modular finite –difference model that simulates the flow of ground water of uniform density(McDonald and Harbaugh,2000).MODFLOW - 96 is a modified version of MODFLOW(McDonald and Harbaugh,1988) that incorporates the use of parameters to define model in puts ,the calculation of parameter sensitivities and the modification of parameters value to match observed heads ,flows or advective transport using the observation ,sensitivity and parameters estimation process described by Hill and others (2000).

The model was calibrated in the steady-state mode with drilling report static water levels and the water level measured by Ermias Hagos, 2005. Hydraulic parameters were iteratively adjusted by trial and error between steady state of the model until satisfactory match is gained.

5.3 Model description

The MODFLOW is a versatile finite –difference modeling program used to construct numerical flow models of specific areas. A MODFLOW model consists primarily of a set of in put files that contain information on the physical properties of the modeled system such as the geometry boundary conditions, internal properties(such as the distribution of hydraulic conductivity and storage coefficient),and sources and sinks such as ground water recharge ,streams, and pumping wells. Once these files are created, the model program is run to solve a set of equations that describe the distribution of head at discrete points with in the system and the flow in response to that head distribution.

The steady–state model simulates the ground water flow system as it would exist in equilibrium with a given set of condition.

5.3.1 Governing Equation and Model Code

The movement of ground water through porous media is described by the following partial differential equation, which is based on Darcy's law and the conservation of mass (McDonald and Harbaugh, 1988): (1)

$$\frac{\partial}{\partial x}\left(K_{XX}\frac{\partial h}{\partial x}\right)+\frac{\partial}{\partial y}\left(K_{yy}\frac{\partial h}{\partial y}\right)+\frac{\partial}{\partial z}\left(K_{zz}\frac{\partial h}{\partial z}\right)-W=S_S\frac{\partial h}{\partial t} \quad \text{----------------------------------}1$$

Where

K_{xx}, K_{yy}, and K_{zz} are values of hydraulic conductivity in the x, y and z directions along Cartesian coordinate axes, which are assumed to align with principal directions of hydraulic conductivity (LT^{-1}),

h is hydraulic head (L),

W is a volumetric flux per unit volume and represents sinks and/or sources (T-1),

Ss is the specific storage of the porous material (L^{-1}), and

t is time (T).

Equation (1) describes the distribution of hydraulic head and flow throughout a continuous region. Derivations of equation (1) can be found in Freeze and Cherry (1979) and Anderson and Woessner (1992).

In order to simulate the unconfined aquifer of Mehoni sub-basin equation (2) is used which is the Boussinesq Equation.

$$\frac{\partial}{\partial x}\left(K_X h \frac{\partial h}{\partial x}\right) + \frac{\partial}{\partial y}\left(K_y h \frac{\partial h}{\partial y}\right) = S_Y \frac{\partial h}{\partial t} - W \quad \text{---}2$$

It is assumed that $T_x = K_x h$ and $T_y = K_y h$, where h is the saturated thickness, and S_y is the specific yield. Available data are limited to horizontal properties in aquifers and no relation can be established regarding the anisotropy of units. Thus, for this study, a hydraulic property within the layer is assumed isotropic. Consequently K_x and K_y are considered to be equal at any given location and K_x and K_y are replaced in this discussion by the single term K to describe horizontal hydraulic conductivity. Since the study is on a Steady State condition there is no change in head with time, therefore, this part of the right hand side of equation (2) becomes zero and it can be re-written as:

$$\frac{\partial}{\partial x}\left(K_X h \frac{\partial h}{\partial x}\right) + \frac{\partial}{\partial y}\left(K_y h \frac{\partial h}{\partial y}\right) - W = 0 \quad \text{---}3$$

Because of its continuity in space and time, generally equation (3) cannot be solved analytically for practical applications involving complex systems. As a result, numerical methods are employed where a set of spatially and temporally discrete points replace the continuous system described by equation (3) in a process of discretization. Equation (3) is then replaced by a set of simultaneous algebraic equations that describe the distribution of hydraulic head at each point, and flow through the system in response to this head distribution. These simultaneous equations are set up in matrix form and then solved. There exist several techniques to solve the set of simultaneous equations. In this model, the Slice Successive Over-relaxation (SSOR) method of Hill (1990) was used as a solver. Discussions of the finite-difference method, which is the numerical technique used in this study, can be found in Anderson and Woessner (1992).

Discretization

The numerical method used to approximate equation (1) requires that the modeled domain be divided into discrete volumes, called *cells*. The three-dimensional array of cells is known as the *model grid*. The center of each cell defines the point for which hydraulic head is determined. The head is taken to represent the average head within the cell. For transient models, time must also be divided into discrete intervals known as *time steps*. Heads and flows are calculated at the end of each time step.

Spatial Discretization

The numerical ground water flow modeling of Mehoni sub basin in raya valley,which has an areal extent of the modeled area is 45 km easting by 45.5 km northing and contains the entire study area shown in figure5.1. The model uses a uniform grid size of 400 m by 400 m and contains 7238 cells, 1 layer, 111 columns and 116 rows. The finite-difference model grid must be regular spacing, in order to facilitate data input from DEM & SURFER files; certain cells may represent areas outside the modeled area (which can have any shape). Such cells are considered inactive and the ground-water flow equation is not solved for them.

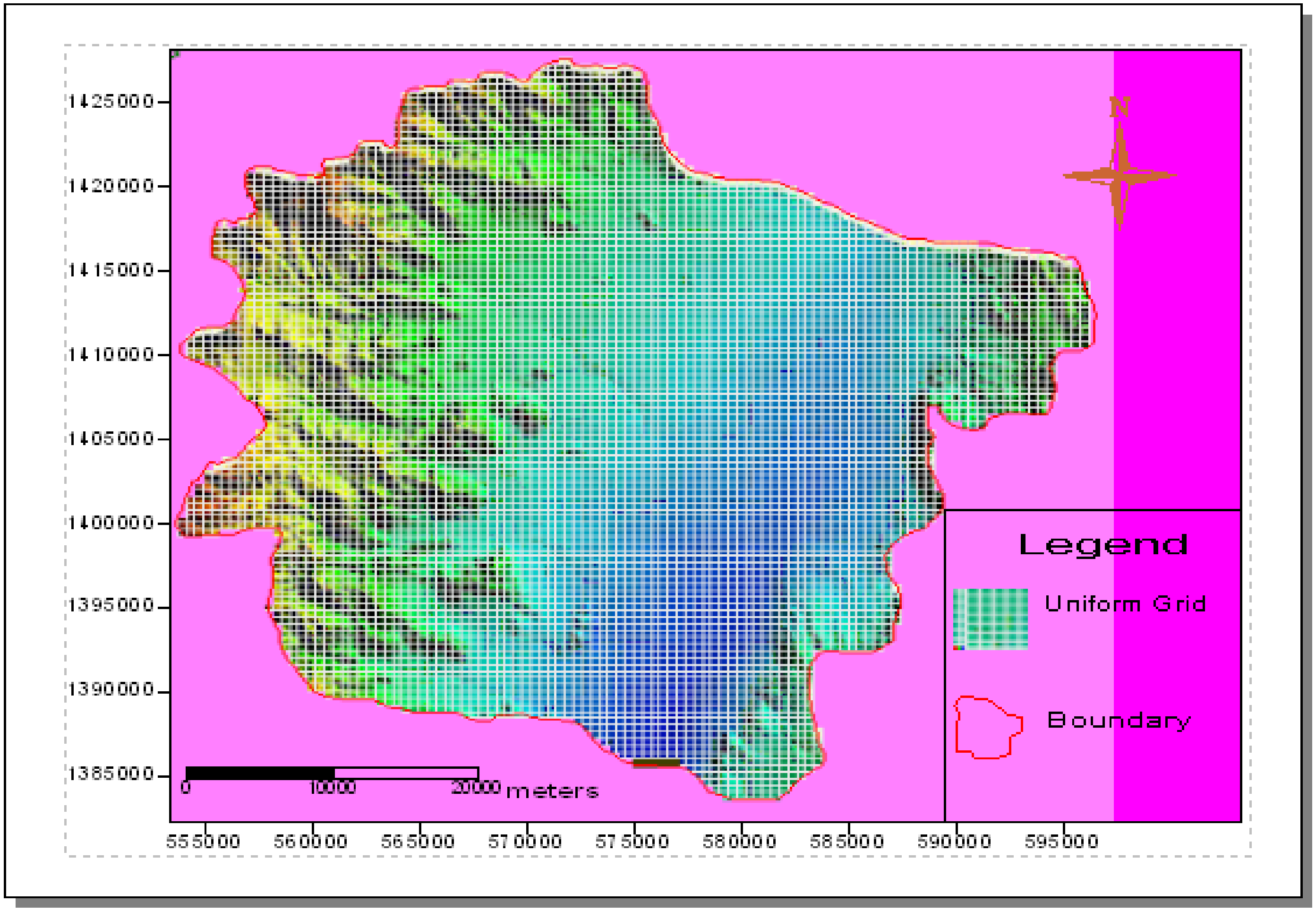

Fig 5.1 Model Area Descritization.

The modeled aquifer system was considered to be single layer and unconfined. Generally, the top layer elevation was considered to be the elevation of ground surface and in this case nodal values of ground surface elevation were interpolated from USGS digital elevation data. The interpolation was done at a resolution of 400m by 400m and then loaded into MODFLOW top elevation array. In this study, the aquifer thickness lies within the range from few meters to more than 250m in the valley fill except along the boundaries where ridges with high elevation are found. Elevated zones were simulated by giving relatively higher thicknesses at the cells in order to avoid drying of cells during simulations. Hence bottom elevation was set at 1400m.a.m.s.l. In fact, the thicknesses of the aquifers are very rough as it has not yet been determined exactly for the aquifers and was modified a bit in few areas during model calibration process. The relative thickness variation in most parts of the aquifer can be clearly observed in figure5.2 3D master shown below.

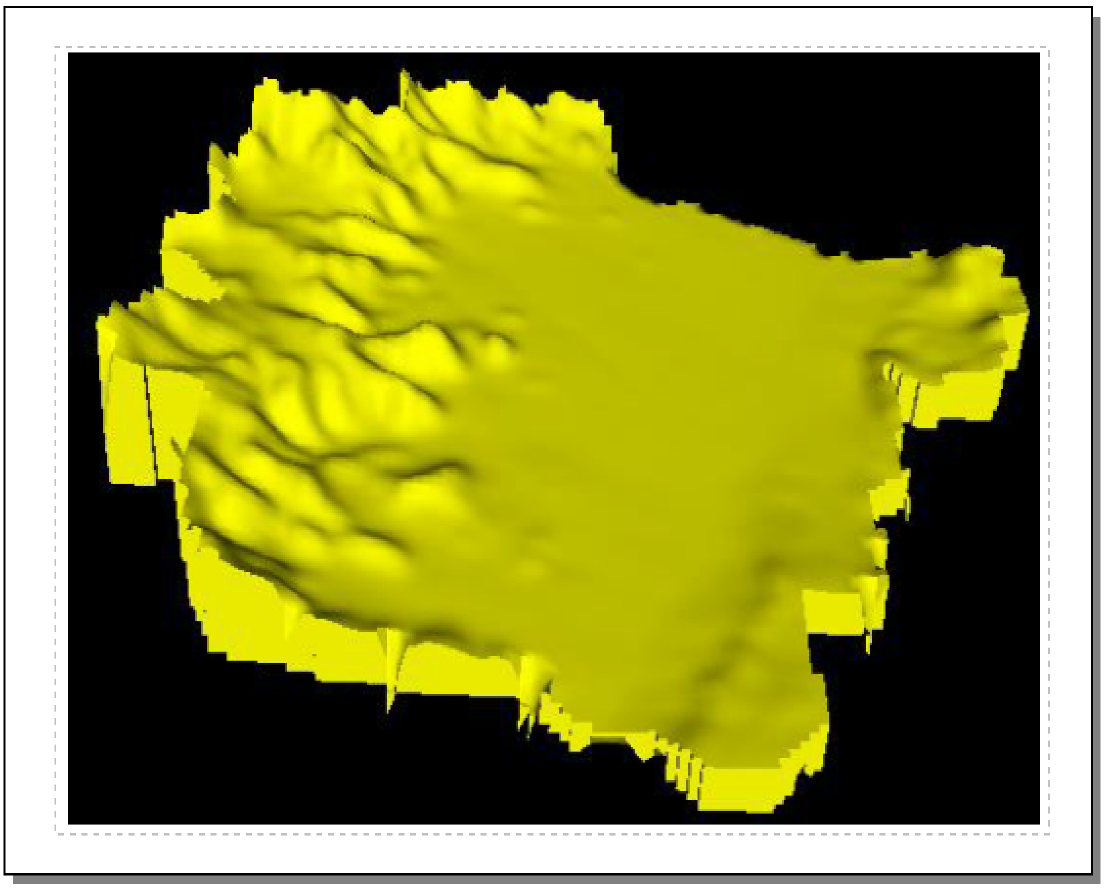

Fig 5.2 Geo-Spatial Model of the Area

5.3.2 Boundary Condition

Boundary conditions are mathematical statement specifying the dependent variable (head) or the derivative of the dependent variable (flux) at the boundaries of the problem domain (Anderson and Woessner, 1992). The choice in the type and location of model boundaries is important, as this may affect the simulation results. Ideally, model boundaries represent actual hydrologic boundaries, but this objective can not always be met. If model boundaries do not represent actual hydrologic boundaries, it is important that they are located far enough away from the area of interest so they do not affect the simulation results (Marijke and Smith, 2002).

The ground-water flow system in the Mehoni sub-Basin is controlled, to a large extent, by the hydrologic and geologic boundaries of the system. Boundary conditions define the geographic extent of the flow system as well as the movement of ground water into and out of the system, such as flow to or from streams.

Boundary conditions may be three types: no-flow boundaries, specified-flux boundaries, and head-dependent flux boundaries. Geologic or hydrologic barriers to ground-water flow were simulated using no-flow boundaries. The contact between the permeable ground-water flow system and nearly impermeable bedrock is an example of a no-flow boundary. Known or estimated hydrologic fluxes, such as recharge and well discharge, are represented using specified-flux boundaries. A head-dependent flux boundary is one across which ground water moves at a rate proportional to the hydraulic-head gradient between the boundary and the ground-water system. Streams are usually represented as head-dependent flux boundaries because the movement of ground water to or from a stream is proportional to the difference between the head in the aquifer and the stage of the stream (which was not used in this study).

The areal boundaries of the model area (fig5.3) were either no flow boundaries or head dependent flux boundaries. The no flow boundaries of the study area were chosen to correspond as closely as possible with natural hydrologic boundaries across which ground water flow can be assumed negligible, such as ground water divides, or can be reasonably estimated. Major topographic divides are often considered no flow boundaries because topographic divides are typically coincident with ground water divide. Ground water on either side of a ground water divides flows away from the divide and not across it, so the divide itself acts as no flow boundary.

The lateral boundaries of the study area flow model generally represented as no flow boundaries, with the exception of a narrow southerly surface outlet ,where it was represented as head dependent flux boundaries is assumed at the locality of topographically low surface water divides and ,where simulated with the general-head- boundary (GHB) module of MODFLOW (McDonald and Harbaugh,1988)

The top boundary of the model includes both no flow and head dependent –flux boundary cells. The no flow boundary is areally applied ground water recharge. Recharge was specified and simulated with (RCH) module (McDonald and Harbaugh, 1988).the bottom boundary of the model is a specified no-flow.

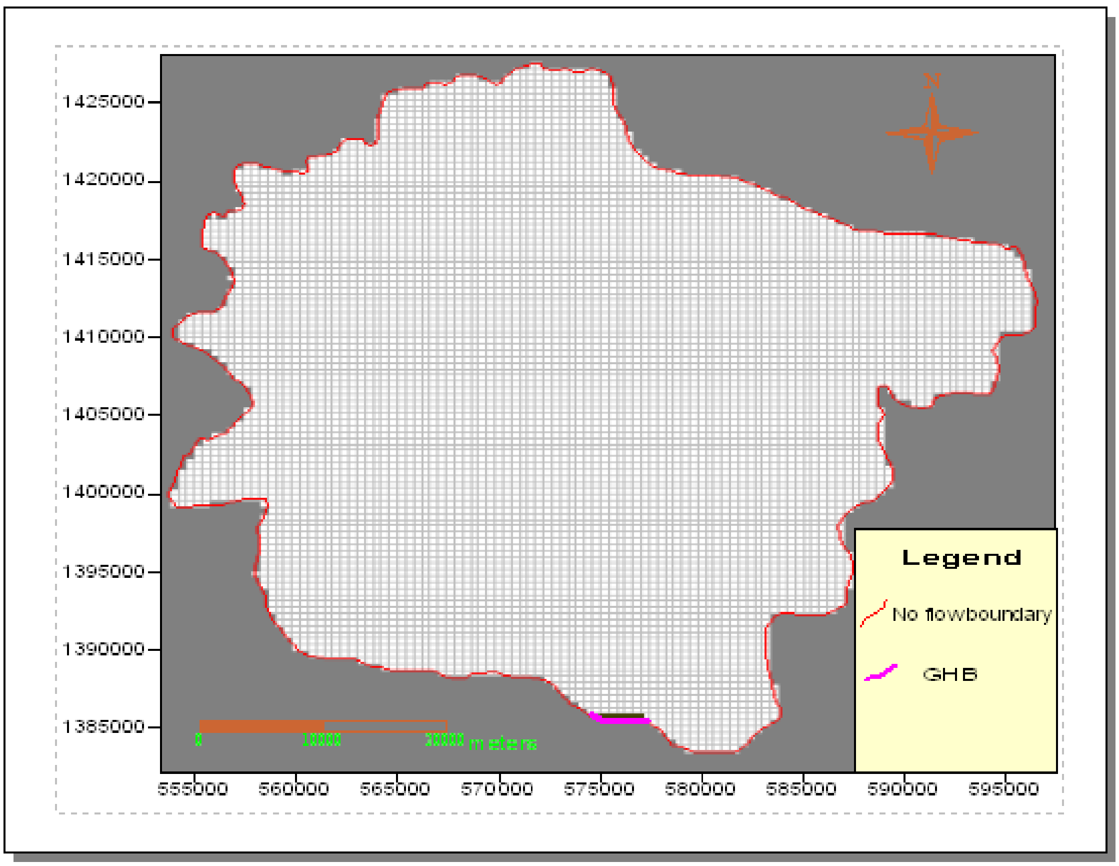

Fig 5.3. Boundary Condition of the Area

5.4 Model Input Parameters

Simulation of groundwater flow and fluxes requires specifying aquifer system properties and stresses. Aquifer system properties can vary considerably both horizontally and vertically and thus, cannot be precisely represented in a numerical model. The aquifer system properties specified for each active cell in the model are estimates of the average conditions in the area represented by the cell. Similarly, stress applied to the system (recharge and discharge) are estimates for the area represented by each cell. The initial aquifer system properties of the study area were conceptually modeled from different groundwater literatures and analysis of pumping test. Areal recharge estimation was obtained from Ermias Hagos (2005) for two flak of the study area and for the alluvial deposit or the valley fill was calculated as in the previous discussion, and pumpage was estimated, as described in the conceptual model, from daily pumping records and estimation of hand-dug wells and shallow drilled well discharges. Selected properties and stresses were modified during model calibration.

5.4.1 Initial and Prescribed Hydraulic Heads

Water table surface used to define initial head in the model are based on the DEM land surface data source for each model cell. The land surface value of the DEM data are adjusted to the water table depth by subtracting the depth to the water table from land surface at measured points of water level from Ermias and from drilling reports. By analyzing the elevation difference between the land surface and the water level for 72 shallow and deep wells, it was assumed that the level varies depending on the topography of the area. The difference increases as topography increases. Based on this, the initial and prescribed hydraulic head for the model was assigned for each cell of the model layers by the following methods:

- areas where land surface elevation ranges from 2700m – 3600m hydraulic head was assigned by land surface minus 37m,
- areas where land surface elevation ranges from 2700m – 2000m hydraulic head was assigned by land surface minus 39m,
- Areas where land surface elevation ranges from 2000m – 1400m hydraulic head was assigned by land surface minus 54m.

5.4.2 Hydraulic Conductivity and Transmissivity

Groundwater flow within the model layer was assumed to be horizontal. Hydraulic conductivity and transmissivity are properties that, in conjunction with the horizontal hydraulic gradient, control horizontal flow of groundwater. Hydraulic conductivity is a measure of the water transmitting properties of aquifer material; coarse and well sorted materials have a higher hydraulic conductivity than fine and poorly sorted materials. Transmissivity is the product of hydraulic conductivity and saturated thickness and represents the water transmitting properties of the saturated section of the aquifer.

The hydraulic conductivity of the study area was estimated from aquifer test data of 28 boreholes drilled in the study area at different times. Based on the pumping test analysis, the aquifer hydraulic conductivity area ranges from 0.022 m/d-79m/d.

The horizontal hydraulic conductivity of the area was applied to each active model cells by zoning the similar hydraulic conductivity areas based on the hydraulic conductivity value of table 4.2 and map produced by SURFER 8 software and contoured for equal hydraulic conductivity areas.

There is a wide range of transmissivity variation. This ranges, over 600 m^2/day in the North West and western part, and is less than 50m2/day in the eastern and larger part of the central basin. All the wells with a higher transmissivity values are located either in the western margin of the valley or along one of the main rivers, i.e. Guguf, Fokisa, Haya, Larger portion of the eastern part have a very low transmissivity value (< 100m^2/day).

Lithological logs of some of the wells with a higher transmissivity values indicate the aquifer material being composed of mainly coarse sand and gravel and in some cases of pebbles. In most cases, these wells are located in the western edge of the sub basin. This is due to the fact that all the streams/rivers that originate in the western highlands would deposit their coarser and highly permeable sediments in the western part of the basin fill as soon as they reach the plain. Even though there is a general trend of west-east decrease in transmissivity values, there are irregularities as a result of distribution of materials which in turn depends on the meteorological, hydrological and tectonic conditions and events in the geological past.

5.5 Model Stresses

Hydraulic heads in the groundwater flow system respond to stresses on the system, which corresponds to recharge and discharge. As noted earlier, recharge to the groundwater system of the study area is from the precipitation for the two flanks and precipitation and runoff from the mountains area for the valley fill and discharge from the system is through groundwater pumpage, and minor groundwater evapo-transpiration where the water levels are near the surface.

5.5.1 Recharge

Recharge to the model is only from natural recharge which is from the infiltration of precipitation which was left from evapo-transpiration and surface runoff. In general, precipitation recharge varies spatially with land surface permeability, which is a function of soil characteristics and land use. The spatial distribution of recharge rate in the study area was determined by averaging the different estimation systems, depending on the land use land cover data of the area by surface water balance calculated by different researchers and inflow outflow methods. By this estimation, the western and eastern part of the sub-basin gets 98.55 mm/yr of groundwater recharge and the valley fill get groundwater recharge is around 90mm/yr, as estimated before. Recharge was applied to the top active cell of the model as a spatially varying, specified flux to the upper most active layer.

5.4.2 Ground Water Pumping

Ground–water pumping rates were specified in the model by two different methods representing public water supply and wells for irrigation purpose. Public wells with drawing a total of 900m^3/d and wells for irrigation and rural water supply purpose withdraw 14483.14m^3/d were assigned to the appropriate location and hydrgeological units (Table4.3-4.8and fig 4.2). Ground water pumpage for the model was simulated using well package of the MODFLOW depending on the geographic coordinates of the wells.

CHAPTER SIX
MODEL CALIBRATION AND SENSITIVITY ANALYSIS

6.1 Model Calibration

Model calibration is the process where by model parameter structure and parameter values are adjusted and refined to provide the best match between measured and simulated values of hydraulic heads and flow parameters are adjusted with reasonable limits from one simulation to the next to achieve the best model fit. Model fit is commonly evaluated by visual comparison of simulated and measured heads and flows or by comparing root mean square (RMS) errors of head and flow between simulations.

Basically, calibration can be achieved in two ways. These are the forward and inverse problem solutions. In an inverse solution method one determines values for a given parameter structure and hydrologic stress using a mathematical technique, such as nonlinear regression (Cooley & Naff, 1990; Hill, 1992, 1998) from information about head distribution (Anderson and Woessner, 1992). This technique is sometimes called parameter estimation & it finds the set of parameter values that minimize the difference between simulated and measured quantities such as hydraulic heads and flows; where as in the forward problem system parameters such as hydraulic conductivity and hydrologic stresses are specified and the model calculates the head distribution.

The Mehoni sub basin ground-water flow model was calibrated using trial-and error method in adjusting initial estimates of aquifer properties (hydraulic conductivities) recharge and discharge to get the best match between simulated hydraulic heads and measured water levels and selected water budget items. The study area ground-water model was calibrated steady-state simulation of ground-water flow.

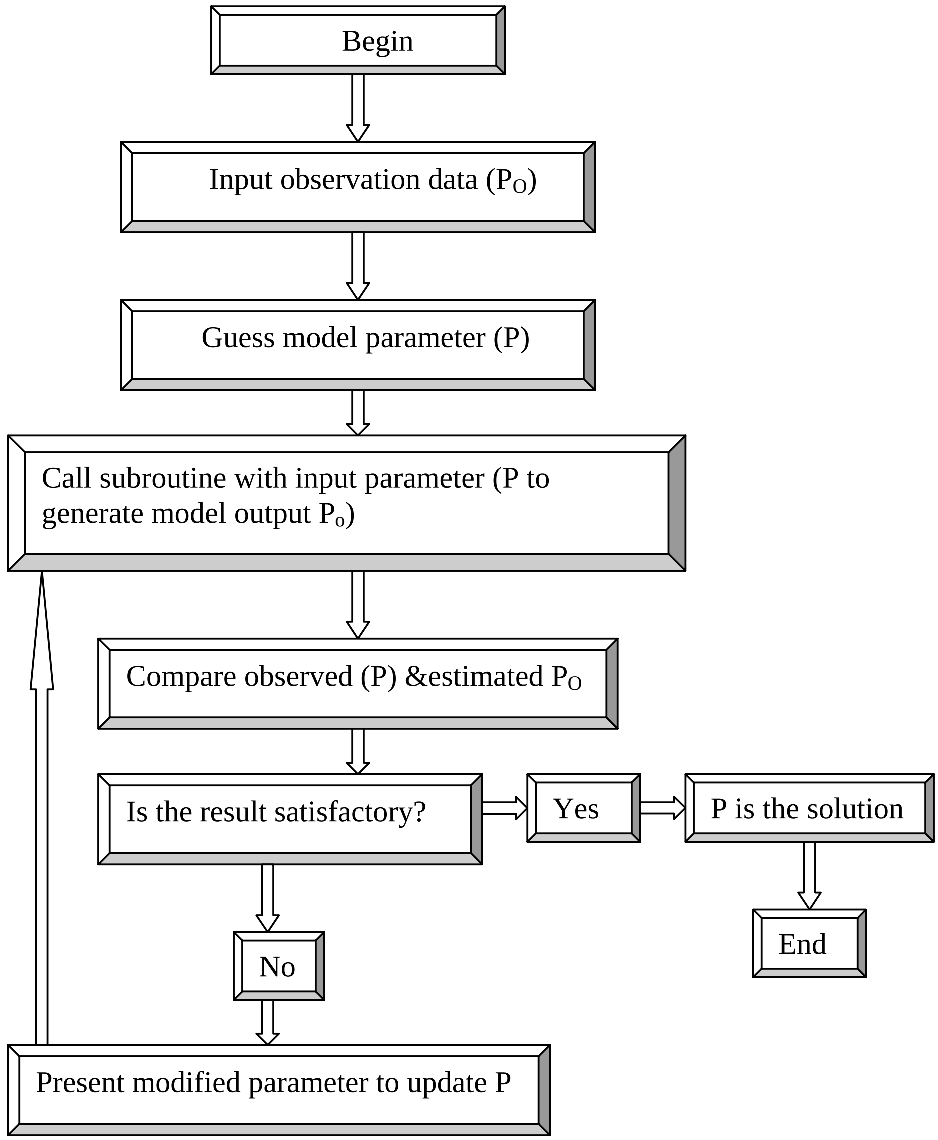

Fig 6.1Trial and Error Calibration Procedure Nezheng Sun, 1994

Steady-State Simulation

Steady-state flow condition exists when inflow is equal to out flow and aquifer storage does not exist. Hydraulic heads for steady-state conditions are sensitive to the amount of water that recharges to and discharges from the groundwater system; the hydraulic conductivity of the aquifer system and aquifer thickness.

The steady-state flow simulation of the study area based on the water level measurement of 65 wells during the construction, measurement taken by Ermias, (2005) and 10 wells in June, 2006 was made to provide initial conditions.

Criteria were established to evaluate the simulation result in relation to measured data. These criteria generally involve: (1) comparing simulated water level contours with observed water level contours; (2) Comparing measured water levels or drawdown in the well; (3) Comparing the simulated water budget and input data. The calibration criteria were set for the steady-state simulation depending on the existing data source, available time for the research work, and considering different limitations.

Based on the above conditions, the first calibration criterion for the steady state simulation was that the simulated potentiometeric surface and hydraulic gradients generally match those of the estimated average potentiometric surfaces (contours). The simulated steady state potentiometric surfaces (contour) generally are similar to the estimated average potentiometric surface (fig6.2), indicating satisfaction of this criterion.

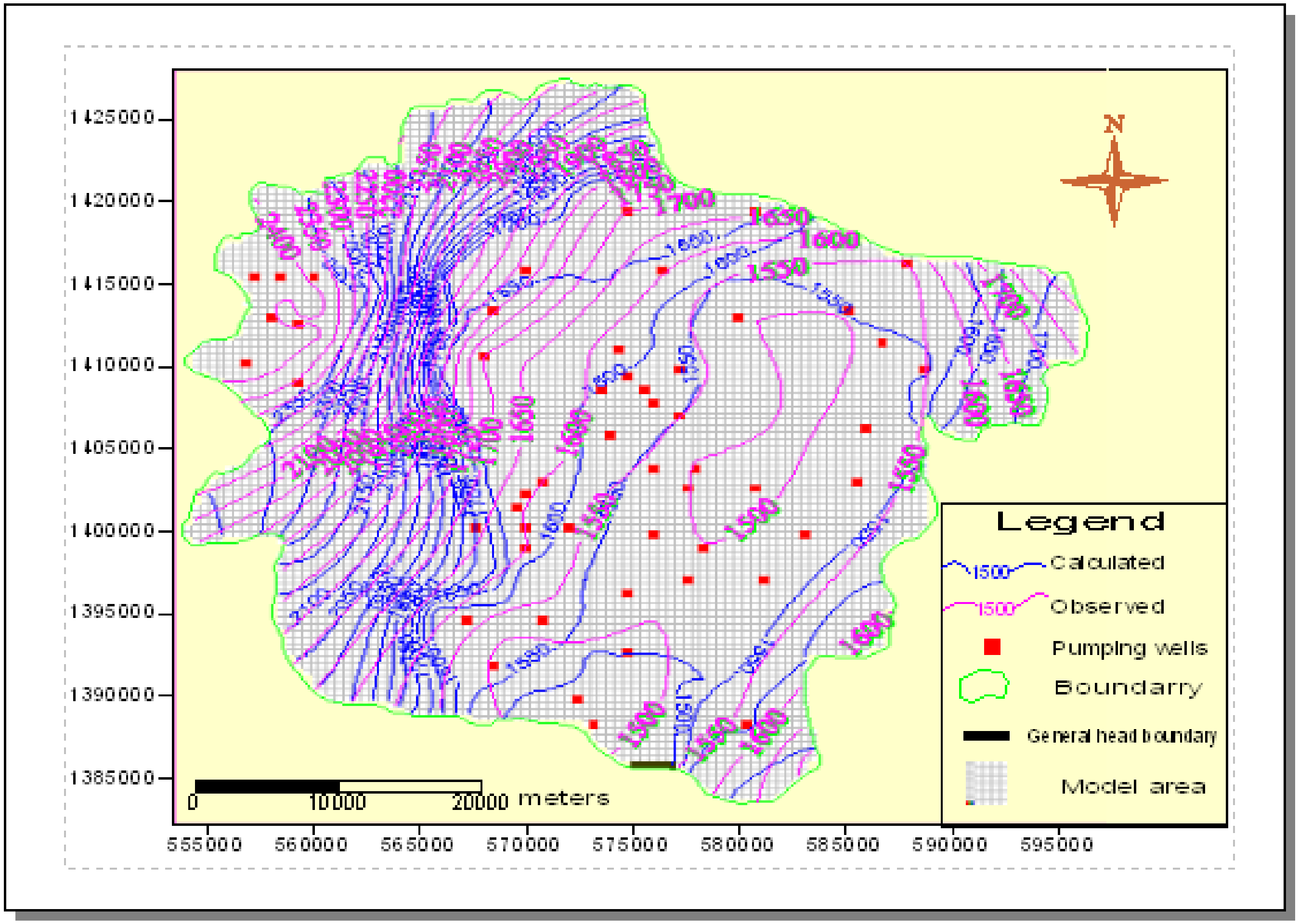

Fig6.2 A Comparison Of Observed and Calculated Hydraulic Head

A second calibration criterion for steady state simulation includes matching 95% of all wells to within ±15m of the observed hydraulic heads. This second criterion was considered adequate because 15m is less than 15% of the difference between the maximum and minimum hydraulic head of the estimated average of wells in the model area.

Simulated hydraulic heads matched observed values within ±15 m difference for 97.5% of the 40 observed wells and 100% of the observed wells is matched to the simulated heads within ±20m difference . A histogram shows the distribution of difference between observed and simulated hydraulic heads (residuals) (fig 6.3)

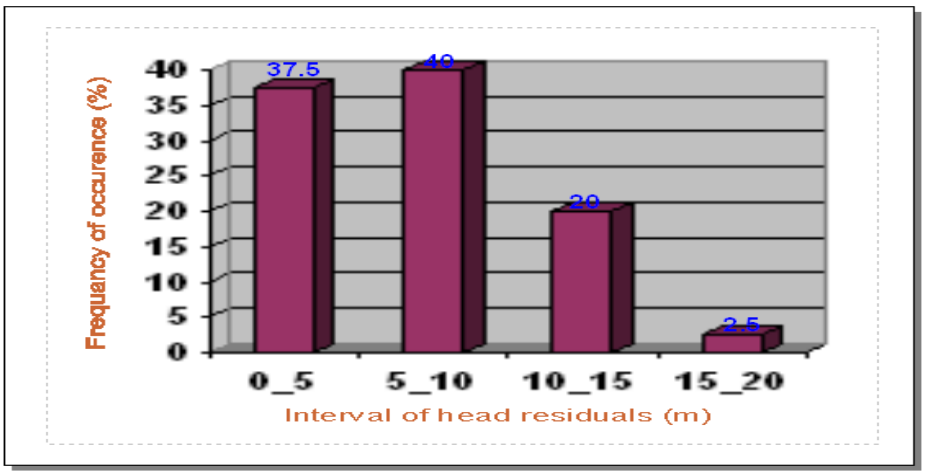

Fig 6.3 Histogram Showing the Level of Calibration for the Selected Head Observation Wells

Table6.1 List of Head Observation Wells Used For Calibration Process

Observation wells	Easting	Northing	Observed head value
BH01	558641	1413019	2371.4
BH02	559360	1412779	2390.54
BH03	559938	1412807	2391.1
BH04	557946	1412790	2414.48
BH06	559903	1412019	2422.9
BH09	558702	1409774	2385.72
BH10	558853	1409498	2377.55
BH11	556671	1414869	2432.87
BH12	556892	1409954	2405
BH13	569309	1394926	1589
BH14	571531	1394926	1542.4
BH17	580500	1388250	1564.65
BH20	569675	1401249	1603.28
BH21	569881	1415748	1575.5
BH23	574204	1410936	1555.6
BH24	576688	1406735	1498.7
BH25	577755	1402683	1531.5
BH26	587750	1416000	1503
BH27	580035	1413220	1702.6
BH29	574336	1419297	1609.5
BH30	570637	1403161	1609.5
BH33	586046	1406271	1515
BH34	569929	1399050	1602
BH36	567959	1410599	1693
BH38	574668	1396412	1507
BH40	574600	1409500	1594
BH41	575889	1407832	1588.84
BH42	573600	1408650	1594
BH43	574000	1406000	1583
BH45	574932	1408935	1595.1
BH46	575500	1408500	1590.8
BH51	577250	1409991	1582.2
BH52	571089	1402927	1591.1
BH54	567162	1394661	1579.52
BH56	571951	1400364	1556
BH58	577527	1396864	1534.75
BH61	574151	1405561	1605
BH68	568477	1392062	1585
BH70	586059	1403104	1506
BH73	560052	1415237	2396.5

Statistical comparison between the simulated and measured values of hydraulic heads (table 6.2) was done to quantitatively asses the calibration match. The mean error (ME), root mean square error (RMSE) and the mean absolute error (MAE) provide ways to determine the overall goodness-of-fit between the simulated and the measured hydraulic heads.

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

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

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

$$MAE = \frac{1}{n}\sum_{i=1}^{n}\left|(h_m - h_s)_i\right| \text{ ----------------------------------2}$$

3. The root mean squared error (RMSE) or the standard deviation is the average of the squared differences in measured and simulated heads.

$$RMSE = \left[\frac{1}{n}\sum_{i=1}^{n}(h_m - h_s)_i^2\right]^{0.5} \text{ ------------------------------3}$$

(Anderson and Woessner, 1992).

The summary of statistics, after calibration for residual heads between simulated and observed values was calculated, for 40 water levels measured during the study time and drilling water level reports. Based on this, the mean error (ME), the mean average error (MAE) and the root mean square error (RMSE) were computed for the number of wells (Table 6.2). The mean error of the calibration results for Mehoni sub-basin groundwater flow model is -0.27. This indicates that, in the overall calibration of head levels, calculated heads were greater than observed heads by about 0.27m. The root mean square error shows whether the calibration criteria set prior to or during calibration has been met or not. The mean absolute error and the root mean squared error for the calibration are 8.12 and 8.52 m respectively. The statistical computation for residual errors is summarized in tables 6.2.

Table6.2 Simulated Head and Measured Heads in Observation Wells

BHID	Simulated value (h_s)	Measured value (h_m)	$(h_m - h_s)$	$(h_m - h_s)^2$	$\lvert h_m - h_s \rvert$
BH01	2383.093	2371.4	-11.693	136.72625	11.693
BH02	2385.711	2390.54	4.829	23.319241	4.829
BH03	2385.735	2391.1	5.365	28.783225	5.365
BH04	2424.004	2414.48	-9.524	90.706576	9.524
BH06	2433.003	2422.9	-10.103	102.07061	10.103
BH09	2384.336	2385.72	1.384	1.915456	1.384
BH10	2384.417	2377.55	-6.867	47.155689	6.867
BH11	2433.19	2432.87	-0.32	0.1024	0.32
BH12	2423.786	2405	-18.786	352.9138	18.786
BH13	1481.002	1472	-9.002	81.036004	9.002
BH14	1534.432	1542.4	7.968	63.489024	7.968
BH17	1565.389	1564.65	-0.739	0.546121	0.739
BH20	1597.439	1603.28	5.841	34.117281	5.841
BH23	1588.162	1575.5	-12.662	160.32624	12.662
BH24	1555.958	1555.6	-0.358	0.128164	0.358
BH25	1514.187	1498.7	-15.487	239.84717	15.487
BH26	1525.224	1531.5	6.276	39.388176	6.276
BH27	1512.051	1503	-9.051	81.920601	9.051
BH29	1713.438	1702.6	-10.838	117.46224	10.838
BH30	1601.408	1609.5	8.092	65.480464	8.092
BH33	1527.716	1515	-12.716	161.69666	12.716
BH34	1594.6	1602	7.4	54.76	7.4
BH36	1673.77	1693	19.23	369.7929	19.23
BH38	1518.36	1507	-11.36	129.0496	11.36
BH40	1595.651	1594	-1.651	2.725801	1.651
BH41	1576.418	1588.84	12.422	154.30608	12.422
BH42	1594.165	1594	-0.165	0.027225	0.165
BH43	1573.808	1583	9.192	84.492864	9.192
BH45	1590.795	1595.1	4.305	18.533025	4.305
BH46	1586.041	1590.8	4.759	22.648081	4.759
BH51	1571.272	1582.2	10.928	119.42118	10.928
BH52	1600.316	1591.1	-9.216	84.934656	9.216
BH53	1592.788	1594.52	1.732	2.999824	1.732
BH54	1571.486	1579.52	8.034	64.545156	8.034
BH56	1561.36	1556	-5.36	28.7296	5.36
BH58	1524.851	1534.75	9.899	97.990201	15.899
BH61	1592.915	1605	12.085	146.04723	12.085
BH68	1580.628	1585	4.372	19.114384	4.372
BH70	1515.245	1506	-9.245	85.470025	9.245
BH73	2386.418	2396.5	10.082	101.64672	10.082
			ME=0.27	RMS=8.52	MAE=8.12

In addition to contours, scatter plot of simulated heads against observed heads was used to show calibration fit (fig6.5). Observation points that lie on the straight line show exact fit between head points, observation points that are above the straight line show higher calculated heads and those points below the straight line show higher observed heads. Overall, it shows that the head differences are normally distributed and, high and low simulated values are evenly distributed through most parts of the area. Location of observation points is shown in figure 6.4.

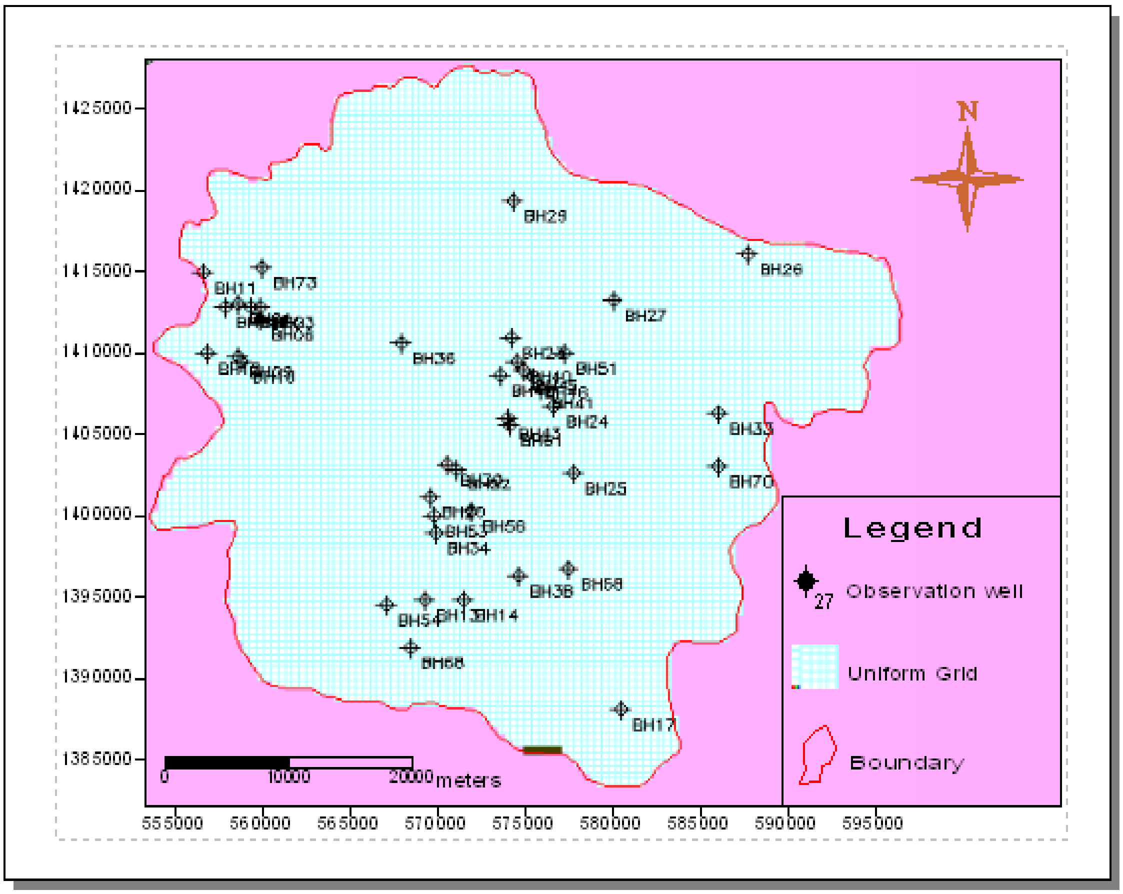

Fig 6.4 Head Observation Distribution

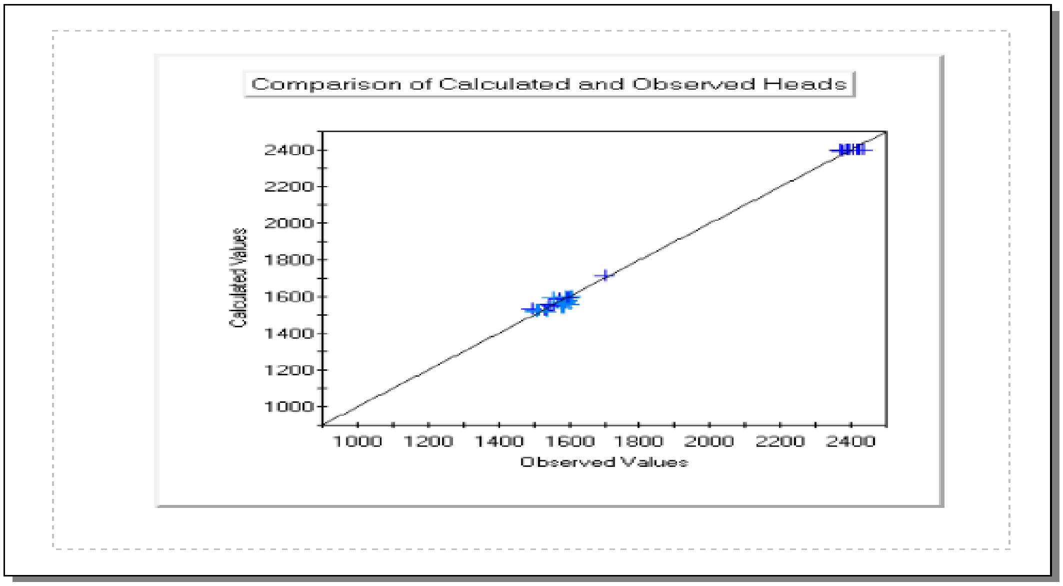

Fig 6.5 Scatter Plot of the Simulated Heads versus Measured Heads

6.2 Model-Derived Groundwater Budget

The calibrated model can be used to derive components of the groundwater budget or estimate the response of the regional system to new stresses such as alternative groundwater management strategies. Water resource managers can use that information to make informed decisions to plan for future groundwater development.

A groundwater budget for the steady state simulation in the model area is expressed by the following equation:

$$GW_{in} + R = GW_{out} + D + \Delta S \quad \text{-----}4$$

Where

GW_{in} is groundwater inflow to the model area,

GW_{out} is groundwater outflow from the model area,

R is recharge ,

D is discharge, and

ΔS is change in groundwater storage.

Recharge to the groundwater system occurs primarily as precipitation. Discharge from the system occurs as evaporation of groundwater from soils and transpiration from plants, as groundwater outflow, and as pumping from wells. A more detailed representation of the groundwater budget of Mehoni sub-basin model is provided by the equation:

$$GW_{in} + R_{ppt} = GW_{out} + D_{et} + D_{ppg} \quad \text{---5}$$

Where

R_{ppt} is recharge from precipitation,

D_{et} is groundwater discharge by evapotranspiration

D_{ppg} is pumping from wells

All water-budget components can be quantified on the basis of the calibrated model except discharge by evapotranspiration and change in groundwater storage. Evapotranspiration from the groundwater system is unknown but was largely accounted for in the determination of recharge and assumed to be relatively insignificant (D_{et}=0).the Mehoni sub-basin model is a steady state representation of the system ,that is, inflow to the system is assumed to be equal to outflow from the system ,resulting in no change in the volume of water stored within the system(ΔS=0).the conceptual model for the Mehoni sub-basin assumes no groundwater inflow from outside the sub-basin.

Substituting the calibrated-model values and above assumption into equation 6 yields the following (all terms are as defined above). From conceptual and simulated groundwater budgets shown in table 6.3 one can clearly observe the steady state model simulated more inflows and outflows compared to the estimated one. The only source of groundwater inflow considered in the conceptual budget was areal recharge from precipitation and was approximated 23,061m^3/d. There is 0.49% discrepancy between the calibrated groundwater model budgets. This is due to low estimates in recharge amounts or exaggerated estimates in one of the outflow components.

Table 6.3 Simulation Result Water Budget

CUMULATIVE VOLUMES L**3	RATES FOR THIS TIME STEP L**3/T
------------------	------------------------
IN:	IN:
---	---
CONSTANT HEAD = 0.0000	CONSTANT HEAD = 0.0000
WELLS = 0.0000	WELLS =0.0000
HEAD DEP BOUNDS = 0.0000	HEAD DEP BOUNDS = 0.0000
RECHARGE = 23061.3594	RECHARGE =23061.3594
TOTAL IN = 23061.3594	TOTAL IN =23061.3594
OUT:	OUT:
----	----
CONSTANT HEAD = 0.0000	CONSTANT HEAD = 0.0000
WELLS = 15141.2607	WELLS =15141.2607
HEAD DEP BOUNDS = 7806.7710	HEAD DEP BOUNDS = 7806.7710
RECHARGE = 0.0000	RECHARGE = 0.0000
TOTAL OUT = 22948.0313	TOTAL OUT = 22948.0313
IN - OUT = 113.3281	IN - OUT = 113.3281
PERCENT DISCREPANCY = 0.49	PERCENT DISCREPANCY = 0.49

Given the uncertainties in the estimation of recharges and discharges, the overall observed water balance can be taken as a sufficient approximation of the actual groundwater flow system. As a whole, the steady state numerical model has reasonably represented the behavior of groundwater system under consideration.

6.3 Model Sensitivity Analysis

The purpose of sensitivity analysis is to quantify the uncertainties in the calibrated model caused by uncertainty in the estimates of aquifer parameters, stresses and boundary conditions (Anderson and Woessner, 1992). Different methods are available to conduct such analysis, but there is no one method to conclusively determine model sensitivity. In this study, a traditional approach was used by adjusting the most important parameters by selected percentages and documenting the resulting change in simulated water levels and groundwater fluxes in different part of the model area.

The model is considered sensitive to parameter when a change of parameter value changes the distribution of the simulated hydraulic head. When the model is sensitive to an input parameter, the value of that parameter within the model is more accurately determined during model calibration because small changes to the parameter value cause large change in hydraulic head. If the change of parameter value does not change the simulated hydraulic head, the model is considered insensitive to that parameter.

Ground-water modeling results are affected by various model parameters and assumptions, including the 1) geometry of the hydrogeologic units 2) Vertical and horizontal spacing of the model grid, 3) types and locations of model boundaries, 4) magnitudes and areal distribution stresses such as ground-water recharge and pumpage, 5) conductance of stream, drain and general head boundary cells, and 6) horizontal and vertical hydraulic conductivities of the aquifers and confining units. In addition, transient mode modeling results also are affected by the length and number of stress periods and the storativities of aquifers and confining units.

The sensitivity of the model in steady-state mode for the study area was determined by systematical increasing or decreasing the model parameters and stresses from the calibrated values. The adjusted parameters and stresses include; horizontal hydraulic conductivity and recharge. The adjustment was done by increasing or decreasing the values by 25, 45 and 55 percents. Effects of the adjustments on the simulated water level and groundwater flux were calculated (table 6.4 and fig 6.6). As observed from the values, the model was most sensitive to horizontal hydraulic conductivity and less sensitive to recharge for both decreasing and increasing of the value. In general, it revealed that, the model is more sensitive for decreasing the scenarios than increasing for the hydraulic conductivity while small changes appear when the value of recharge decreases. Therefore, for further study of the sub-basin care has to be taken during hydraulic conductivity zonation.

Table 6.4 Sensitivity Analysis of the Effect of the Of Hydraulic Conductivity and Recharge on the RMS Value of Water Level

Percent change in HHC & R from calibrated value (%)	RMS value for the selected parameters	
	HHC	Recharge(R)
-55	27.521	9.553
-45	19.293	9.254
-25	12.948	8.854
0	0.000	0.000
25	8.499	8.290
45	11.860	8.385
55	13.356	8.571

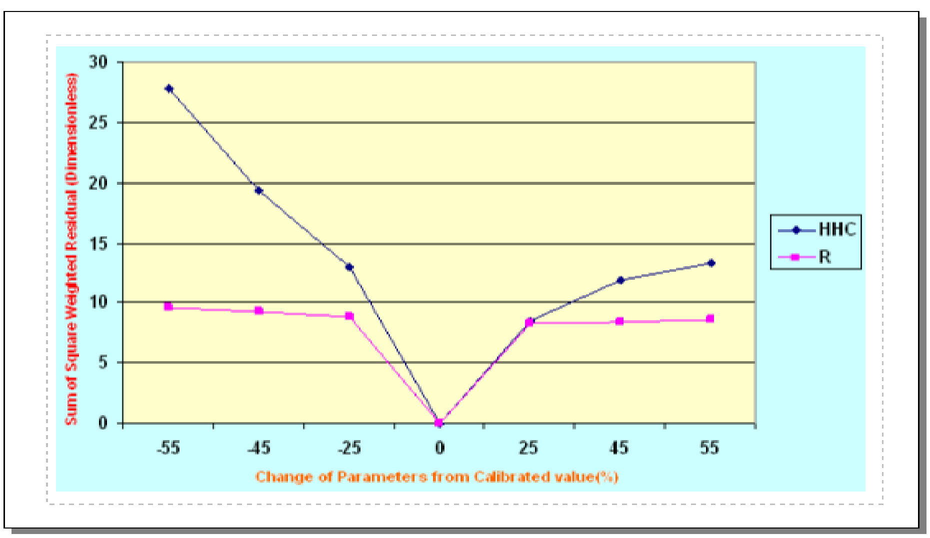

Fig 6.6 Sensitivity Analysis of the Effect of the Hydraulic Conductivity and Recharge on the RMS Value of Water Level

6.4 Scenario Analysis

It is possible to use the model to simulate the resulting change in water levels and fluxes due to the new proposed scenarios and project the likely effect, if having a calibrated model that was tested for sensitivity as capable of simulating water level elevations or fluxes and given its associated limitations.

One of the objectives of numerical flow model simulated in this study was intended to test the response of the hydrologic system to different scenarios. So, alternative scenarios were developed to test the response of the hydrologic system to changes in water used or hydrologic stress under steady state condition. System response was evaluated by using fluxes and heads of the calibrated model as a baseline changes in water table elevation and changes in groundwater outflow in the new scenario simulation.

Changes in water levels and fluxes caused by groundwater withdrawal in the whole sub-basin and the effect of local increase in groundwater withdrawal of irrigation activity takes place were simulated using the model. The effect of changes in water levels and fluxes caused by decreased recharge due to less than normal precipitation in the high lands that may result from weather modification was also simulates.

It should be noted that the results of the scenarios depend on future land use, population growth, weather conditions, hydrologic stress etc, and may not be used as predictive tool to generate absolute amounts in the future, but used primarily to test the response of the system. In general, the results of the scenarios or their accuracy depend on the validity of the assumptions behind the scenarios. Moreover, errors introduced due to limitations associate with the model also affect the results of the scenarios and should be taken into consideration during interpretation and application result.

The first scenario, which is increment of withdrawal by three fold of the existing average withdrawals for the whole sub-basin, was conducted to represent possible future changes in water use in the sub-basin. The heads calculated for this scenario show a maximum decline of the water level by 2.33 m and a minimum of 0.56 m.

In the second scenario, increases of withdrawal at Maichew and its surrounding; and at the alluvial aquifer (valley fill basin) separately, all the other parameters was constant. In this case the two aquifer type is not affected each other. When withdrawal at Maichew and its surrounding increases by three fold of the average existing withdrawal the maximum decline was 0.856m and minimum 0.725m no change has taken place in the valley fill aquifer. On the other hand, when withdrawal increases by three fold of the existing average withdrawal at Kara-Adishawo and Wrabaye the maximum and the minimum decline was detected 1.288 and 1.271m; and 1.338 and 1.266m respectively.

In the third scenarios, 25 wells with daily discharge of 1000m^3/d were selected and were distributed over the whole area the maximum drawdown is appeared in quaternary alluvial and colluvial aquifer and its value was 3.25m. The minimum decline was in head on the interfluvial, fan foot plain and valley fill deposit aquifer and this was 0.143m.

The fourth scenario simulated by the model is a decrease in recharge by 50 percent which could be the case if mild drought conditions were imposed on the aquifer system while water extraction is maintained at current rates. simulated drought condition showed a reduction of water level up to 3.117meters at BH-36and a minimum of 0.056meters at BH-25 large difference in water level occurs in the alluvial alluvial-colluvial deposits(Q_{al} & Q_{col}) aquifer and the smaller differences result from well found in the high lands and interfluvial, fan foot plain &valley bottom quaternary deposits(Q_{smc}).

6.5 Model Limitations

A groundwater flow model represents a complex, natural system with a set of mathematical equations that describe the system. Intrinsic to the model is the error and uncertainty associated with the approximations, assumption and simplifications that must be made. Hydrologic modeling errors typically are the consequence of a combination of: input data, representation of the physical processes by algorithms of the model and parameter estimation during the calibration procedure.

Data on types and thickness of hydrogeologic units, water levels, and hydraulic properties were taken from different phase studies of RVDP (1996, 1998), Desie Nadow (2003) and Ermias Hagos (2005) and represent only an approximations of actual values especially the hydraulic conductivity and recharge estimation. Those data having full pumping test data were concentrated in small portion of the study area. Broad ranges of hydraulic property parameter values are possible. No short or longer term data exist that provide information about how the system responds to changing condition, and this wide time range data is used for the model construction as well as calibration with the assumption of no significant change occurred. However, with no log-term data, it is unknown if the simulated steady state response is truly representative of the real condition.

A wide range of parameter values such as hydraulic conductivity and recharges were used. The previously worked researches have not zoned these parameter. So, numbers of different trial and error model runs were made with various combinations of parameter values during calibration process to arrive at the calibration target; and various combinations can result in low residual error yet improperly represent the actual system. Acceptable degree of agreement between simulated and measured values does not guarantee that estimated model parameter values reasonably represent the actual parameter values.

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CHAPTER SEVEN
CONCULUSION AND RECOMMENDATION

7.1 CONCULUSION

Mehoni sub-basin which is found in the Raya Valley is located in the southern part of Tigray Regional State of the Northern Ethiopia.the area covers 60% of the Raya valley with area coverage of 1183km^2.

The study area is important water source for irrigation, and domestic water supplies. Development of water project in the area increases withdrawal of water from the aquifer. This would lead to the decline in water level in the aquifer. Water resource tools are needed to evaluate the management and environmental issue associated with the aquifer, such as planning for source water protection and estimating sustainable aquifer withdrawals. This report describes a conceptual model of groundwater flow in the aquifer and documents the development and calibration of a numerical model to simulate groundwater flow.

In this study, a homogeneous, an isotropic and single layer aquifer is considered. Geological, hydrogeological and other parameters were collected from different bureaus, institutions and previous works. From these collected data, 49 wells with static water level measurements were selected from 81 wells. Map of equipotential line and groundwater flow direction was developed and the flow direction was determined from West-East, East-West and finally North –South direction.

Groundwater recharge is only from precipitation for the highlands and precipitation and runoff for the valley fill. The out flow from the ground system includes well withdrawal and surface outflow.

Numerical groundwater flow was used to study the response of Mehoni sub-basin groundwater flow system to different scenarios. As numerical groundwater flow models represent the simplification of complex natural systems, different parameters were assigned into the conceptual model to represent the system in a simplified form. The numerical groundwater flow model was simulated using MODFLOW, 1996(McDonald and Harbaugh, 1998). The area was represented by a uniform grid cells arranged in 116 rows and 111 columns each cell with sides of 400m by 400m.a two dimensional profile model under steady state condition is developed to study the response of the system to different scenarios.

Estimated horizontal hydraulic conductivity used for the numerical model input ranges from 0.0022 to 79.096 m/d.

The calibration was calibrated in steady state condition using trial and error calibration techniques. This method involves determination the set of optimal parameter values that best fit the observed hydraulic heads and flows by using number of trial and error of input parameters such as recharge and horizontal hydraulic conductivity and was adjusted and refined using calibration.

The result of calibration was evaluated using steady state graphical model fit approach by comparing the pattern and arrangement of contour map ,and hydraulic gradient from simulated and measured heads and by statistical model fit analysis approach by calculating the average difference between simulated and measured heads using groundwater level data at 40 head observation wells. Though the three statistical measuring parameters were calculated, the standard deviation (RMS) value was used to evaluate the calibration result. In the case of contour matching the general shape and slope of the simulated water table and the overall hydraulic, head gradient matched the geometry and hydraulic heads of observed data. The calculated standard deviation of hydraulic head for the calibration result was 8.52m. This is considered good, since it is within the set of calibration target. The difference especially lower observed inflow was due to under estimation of recharge or exaggerated estimates in one of the outflow.

Model simulated heads was found to be more sensitive to horizontal hydraulic conductivity than recharge. In both cases, value decreased the sensitivity increased. This is done by changing of the parameters by ± 25, ± 45 and ± 55.

As the model is intended to study the response of the hydrologic system the following four scenarios was assigned: In the first and second scenarios which apply for the whole sub-basin and selected local areas, withdrawals were increased by three fold of the average existing withdrawals. In the third scenarios, 25 wells with daily discharge of $1000m^3/d$ were selected. The fourth scenario applies for the whole sub-basin by decreasing the recharge of 50%. The effects of the scenarios are evaluated with respect to change induced groundwater heads to the steady state simulated heads. Accordingly, an increase of withdrawals by three fold over the whole area resulted in an average decline of steady state water level was 2.331m. Locally increase of withdrawals by three fold at Maichew and its surrounding; and Kara-Adishwo and Wera-Abaye separately the other parameters were constant the two system is not affected each other. This indicates that the two aquifer system (volcanic aquifer and alluvial aquifer) are not interconnected. The maximum and minimum declines which were detected at Maichew and Kara-Adishwo were 0.856m and 0.725m; 1.288m and1.271m respectively. The 25 wells were distributed over the whole area the maximum drawdown is appeared in quaternary alluvial and colluvial aquifer and its value was3.25m. The minimum decline was in head on the interfluvial, fan foot plain and valley fill deposit aquifer and this was 0.143m. When recharge was decreased by 50% the whole sub basin, the maximum and minimum decline in head was 3.117m and 0.056m respectively.

7.2 RECOMMENDATION

- ☞ The concerned bodies should collect the data and put it in appropriate place and make it accessible for researchers who are interested to conduct further studies which are beneficial for both the researchers and organizations in charge of water affairs.
- ☞ The area should be zoned as per the value of hydraulic conductivity and recharge so that there would be a facilitated condition to estimate the groundwater potential of the area of the area.
- ☞ Due to variation of the thickness of the alluvial deposit, it is hard changes in geology and simulation of vertical head gradient. Thus, it is important to divide the aquifer system into different layers and estimate their respective hydraulic parameters. This is useful to carry out a three dimensional groundwater flow model simulation properly and efficiently in the sub-basin.
- ☞ Groundwater level monitoring wells should be placed in the whole sub-basin in order to understand the general fluctuations in groundwater levels. This helps to carry out transient groundwater flow modeling, so that system response can be predicted with accuracy.

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ANNEXIES

Annexis 1 Charcher station monthly rainfall

	Jan.	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec
1975	93.1	57.7	24.2	70.9	43.8	39.7	186.8	268.5	130	1.5	0	15.9
1976	0	8.6	45.5	107	81	11.6	98	124.3	28.5	36.5	36.8	4.2
1977	28.2	2.9	69.2	31	76.6	32.1	126.1	307.8	62.2	168.1	10	12.6
1978	4.1	21.3	37.1	40.1	47.8	2.3	139.1	58.9	25.9	0	12.7	9.3
1979	83.4	0	63.7	16.1	102	11.1	120.3	X	X	X	X	X
1980	2.7	54.6	0	12	11.9	28.9	79.3	114.1	93.4	76.8	0	0
1881	0	0.4	184.4	24.4	30.3	0	123.2	162.2	70.5	36.1	0	0
1982	16.3	23	14.9	64.1	85	X	X	X	X	X	X	X

X data not available

Annexis 2 **Chercher** station mean **Minimum Temperature**

	Jan.	Feb	Mrz	Apr	Mai	Jun	Jul	Aug	Sep	Okt	Nov	Dez
1976	X	X	X	X	X	X	X	X	X	X	13.8	12.
1977	13.5	12.2	14.3	16.2	16.5	18.3	16.9	15.6	16.3	15.3	13.8	12.
1978	12.8	13.5	15.2	16.8	17.7	19.3	16.5	16.5	16.9	15.5	13.7	12.
1979	13.4	14.4	15.9	16	17.1	18.3	17.7	X	X	X	X	X
1980	X	X	X	X	X	X	X	X	16.4	15.8	13.9	12.
1981	13.2	13.9	14.6	15.2	16	18.9	15.9	16.3	15.6	14	13.9	12.
1982	13.8	13.9	15.2	16	17.3	20.1	X	X	X	X	X	X

Annexis 3 **Chercher** station mean **maximum Temperature**

	Jan	Feb	Mrz	Apr	Mai	Jun	Jul	Aug	Sep	Okt	Nov	Dez
1976	X	X	X	X	X	X	X	X	X	28.9	25.3	25
1977	22.6	23.6	26.7	28	29.7	31.8	31	27.9	28.3	24.9	24.3	24
1978	25	24.8	26.7	29.4	31.7	32.8	28.9	28.7	29.8	28.4	26.2	24
1979	22.4	25.2	26.5	28.5	29.2	31.9	31.1	X	X	X	X	X
1980	X	X	X	X	X	X	X	X	28.1	28.5	27.2	25
1981	26.2	26.7	25.5	26.6	29.8	33.8	30	28.8	27.3	27.1	26.5	25

1982	24	23.5	26.2	27.3	28.4	32.8	X	X	X	X	X	X

Annexis 4 **Maichew Relative Humidity** at 0600

Year	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec.
1992	X	X	75	57	67	49	63	70	73	86	83	9
1993	86	80	88	84	84	61	75	72	76	81	77	7
1994	76	X	81	77	80	60	74	77	78	79	84	7
1995	69	75	77	86	80	57	75	75	81	78	79	8
1996	86	74	86	83	83	59	68	71	72	80	77	7
1997	73	69	77	76	74	68	67	71	79	X	X	X
1998	87	77	78	72	75	48	X	72	72	83	76	5

Annexis 5 **Maichew Relative Humidity** at 1200

Year	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec.
1992	X	X	57	40	37	26	46	60	51	54	56	7
1993	75	75	58	74	57	40	55	48	47	54	47	4
1994	40	X	58	48	39	36	61	67	49	41	50	4
1995	40	56	54	61	66	27	56	64	50	40	37	5
1996	62	44	56	53	54	45	56	52	41	38	45	4
1997	59	38	53	49	38	40	60	50	38	X	X	X
1998	68	63	57	43	54	37	X	67	58	49	36	3

Annexis 6 **Maichew Relative Humidity** at 1800

Year	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec.
1992	X	X	60	45	43	28	49	64	62	59	66	8
1993	82	88	73	79	70	44	60	58	55	58	56	5
1994	53	X	73	52	48	47	71	75	60	49	60	5
1995	52	71	69	72	71	28	65	71	57	47	48	7
1996	76	53	69	62	58	49	59	68	53	45	57	5
1997	73	50	69	56	42	45	62	54	44	X	X	X
1998	85	77	69	47	57	32	X	74	67	58	45	4

X data not available

Annex 7 **Michew** station mean **daily Sun Shine hours**

Year	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec
1991	7.2	6.8	8.3	10	9.5	8.1	5.4	3.9	6.8	8.4	7.6	9.
1992	X	X	7.5	9.7	9.3	8.2	5	4.8	6.6	6.5	5.3	5.
1993	5.1	5.6	8.3	5.5	8	7.5	5.1		6.6	7.1	9	X
1994	8.9	X	6.4	10.3	X	X	X	X	8.3	8.7	7.6	X

1995	8.8	6.6	7	7.3	8.2	9.4	X	X	7.7	8.2	X	X
1997	X	X	X	X	X	6.2	5.5	6.2	6.7	6.2	6.6	8.
1998	x	6.3	7.4	8.7	8.7	X	X	4.3	7.3	X	9	9.
1999	7.4	X	X	10.2	X	7.3	3.9	4.9	7	6	8.9	7.
2000	9	9.8	9.7	7.3	9.2	X	X	X	5.9	X	X	X

X data not available

Annex 8 **Michew** station mean monthly **wind speed** at 2 m from the surface in m/s

Year	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec
1992	X	X	1.4	1.5	1.7	2.3	3.7	2.8	1.6	1.2	1.1	1.
1993	1.1	1.1	1.4	1.1	1.2	2.2	3.5	2.6	1.6	1.4	1.3	1.
1994	1.4	X	1.6	1.5	1.5	3.3	3.8	2.5	1.4	1.3	1.2	1.
1995	1.3	1.1	1.4	1.2	1.3	1.9	3	2.5	1.4	1.3	1.2	1.
1996	1.1	1.3	1.3	1.3	1.3	2.9	3.8	2.7	1.2	1.2	1.2	1.
1997	1.1	1.3	1.3	1.3	1.5	1.9	3.1	2	1.3	1.2	1	
1998	0.9	1	1.2	1.4	1.4	1.9	X	2.6	1.2	1.2	1.3	1.
1999	1.2	1.4	1.4	1.5	1.6	2	3.3	2.2	1.1	0.9	1.1	1.
2000	1.2	1.3	1.3	1.4	1.4	2.2	3.4	2.8	1.1	0.9	0.9	X

X data not available

Annex 9 well lithologic log data

Site Boyegerarsa

depth	lithologic description
0-6	Sandy clay
6_9	clay brown in clour
9_12	fine sand
12_18	Sandy clay
18_24	Sandy clay
24_33	medium and fine sand
33_45	Sandy clay
45_48	fine sand
48_54	Sandy clay
54_60	fine sandy clay
60_69	Sandy clay
69_75	medium to fine sand
75_78	Sandy clay
78_84	clay brown in color
84_90	Sandy clay
90_93	coarse and fine sand
93_114	Sandy clay

Annex 10 well lithologic log data

Site Dederba

0_24	clay
24_30	medium to fine sand
30_60	clay intercalated with sand

60_63	medium to coarse sand
63_66	medium to coarse grained sand
66_69	fine sand with silt
69_84	coarse gravel to fine gravely sand
84_93	fine gravel, poorly sorted
93_96	fine to coarse sand
96_126	coarse gravel with sand
126_147	angular sub angular gravel

Annex 11 well lithologic log data

Site Degaga

0_3	clay
3_6	sandy silty clay
6_9	clayey sand
9_15	coarse sand
15_41	sandy clay
41_72	coarse sand with rounded gravel
72_92	slightly fractured basalt

Annex 12 well lithologic log data

Site Kara adishawo

0_2	top soil (dark brown plastic clay)
2_6	coarse sand with clay
6_12	medium grained sand with silt
12_22	silt with fine sand
22_34	medium sand with silt
34_38	brownish clay
38_42	sandy silt
42_58	sandy clay
58_84	fine sand with silt
84_110	fine sand with clay
110_114	coarse sand
114_128	clay with sand
128_138	sand with silt
138_144	weathered basalt
144_147	fresh massive basalt

Annex 13 well lithologic log data

Site Hade Alga

0_6	silty sand
6_12	medium grained sand
12_24	silty sand
24_27	coarse grained sand
27_30	highly weathered basalt
30_36	moderately to slightly weathered basalt
36_57	slightly weathered basalt
57_69	fresh basalt
69_75	moderately weathered basalt
75_90	basalt

Annex 14 well lithologic log data

Site Maichew

0_12	latertic clay
12_21	highly weathered basalt
21_33	dark fresh aphinitic basalt
33_39	scoraceous and fractured basalt
39_73	fresh and fractured aphinitic basalt

Annex 16 well lithologic log data

Site Maichew

0_3	reddish brown clay
3_6	weathered basalt
6_18	red and pale brown clay
18_24	black basaltic gravel
24_39	aphinitic basalt
39_57	fresh aphanitic fractured basalt
57_66	scoraceouse basalt
66_87	basalt