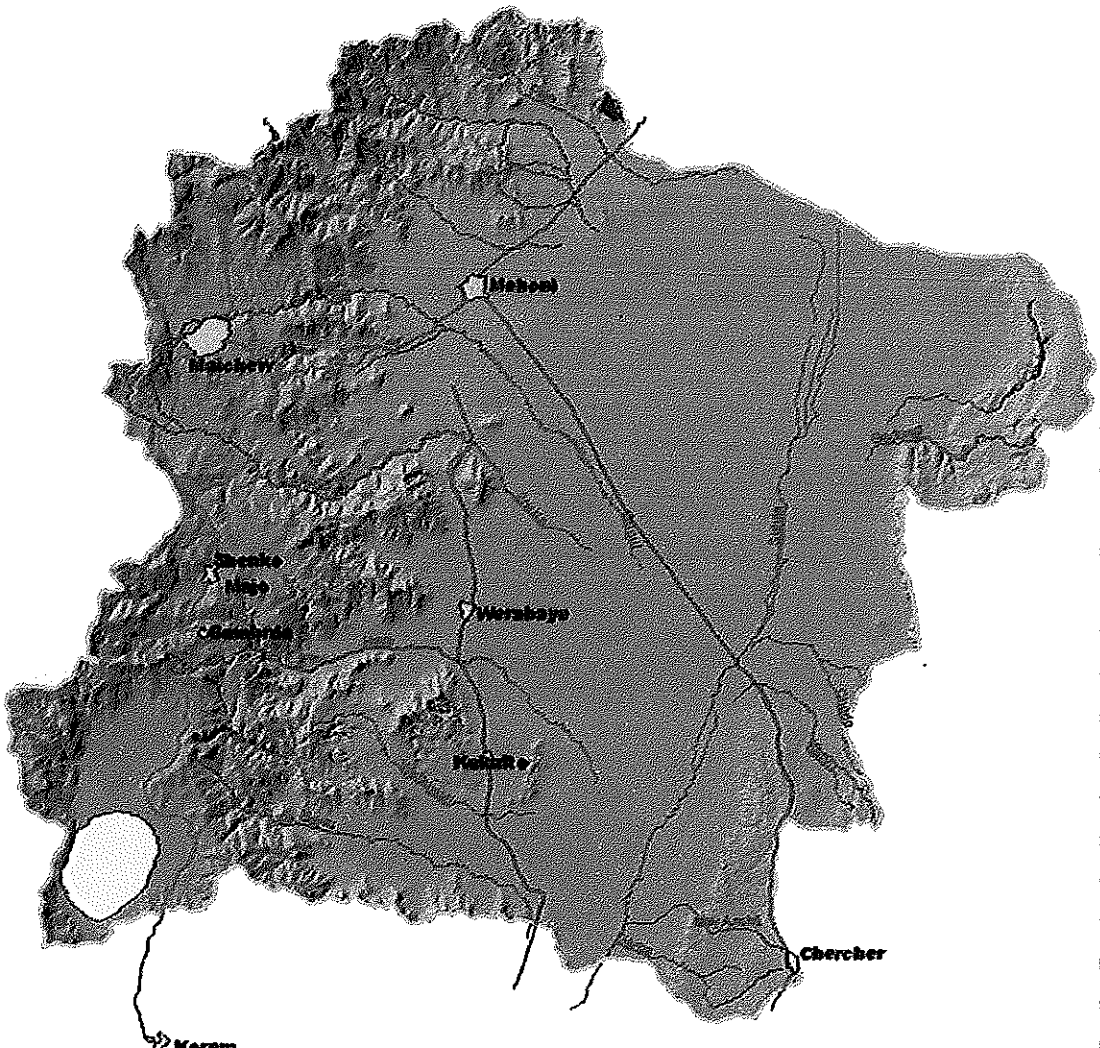
HYDROGEOLOGY OF MEHONI SUB-BASIN AND LAKE ASHANGE CATCHMENT IN THE RAYAVALLY (NORTHENR ETHIOPIA)



A THESIS
SUBMITED TO THE SCHOOL OF GRADUATE STUDIES
ADDIS ABABA UNIVERSITY

IN PARTIAL FULFILMENT OF THE REQUIREMENTS FOR THE DEGREE OF MASTER IN HYDROGEOLOGY

BY ERMIAS HAGOS
JUNE 2005

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DEPARTMENT OF EARTH SCIENCES

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ABSTRACT

The study area is found in northern Ethiopia at the northern portion of the Raya valley, which forms its own trough at the western shoulder of the Afar Depression. Its total aerial coverage is about 1302 km^2. It is largely covered by the Mehoni Sub-basin, which is surrounded by volcanic hills with sediment filled depression at the center. The Lake Ashange basin is also located on the south-western highlands of the study area. The main objective of this work was to study the overall hydrogeological and hydrochemical setup of the study area.

The overall water balance of the study area was computed with an aim of estimating the amount of annual recharge to the groundwater. Accordingly, the mean annual rainfall as the only water input into the study area was estimated to be 722.42 mm/year. On the out put side; the actual annual evapotranspiration was estimated to be 687.24mm/year, the annual surface water out flow was estimated to be 52.3 MCM, the annual lake surface evaporation was estimated to be 16.73 MCM and the annual bareland evaporation was assumed to be half the lake surface evaporation. Finally, a annual recharge to the groundwater of the study area is approximated to be 60.9 MCM.

The study area comprises two major hydrolithologic units. They are the alluvial aquifers and the basaltic aquifers where the alluvial aquifers especially on the western flank of the valley floor and along large rivers were found to be good groundwater potential zones and wells sunk into these aquifers have fairly good efficiencies and productivities. Generally, the groundwater has good chemical quality with the exception high EC (TDS) and NO_3 values in some areas.

The Lake Ashange water does not have similar chemistry to that of the springs that emanate from the western highlands and the groundwater in the valley floor of the Mehoni sub-basin. Therefore, it is concluded that the lake may not have significant sub-surface outflow towards the sub basin.

Lowering of the static water level by about 2m within 2 years time was detected in a monitoring well located at the groundwater irrigated field around Werabaye. This could be

attributed to the large-scale pumping from the groundwater and intensive diversion activities in the upstream areas.

Generally, two ground water flow systems were detected. One is in the Mehoni sub-basin where the groundwater flows from all directions towards the central valley floor with probable groundwater out flow at the southeastern part of the study area. And the other is in the lake Ashange basin where groundwater flows from all dirrections towards the lake.

CHAPTER ONE

1. Introduction

1.1. General Information

Ethiopia is one of the most populous countries in sub-Sahara Africa with a wide range of physiographic regions and climatic variability. More than 80% of its people live in rural areas and they are dependent on subsistence agriculture. In the peripheral regions of the country the most livelihood practice is livestock breading. Ethiopia contains the largest livestock population in Africa. But these two agricultural practices in the country are traditional and backward. They are easily disturbed by the natural hazards such as drought. Due to these reasons most of the rural community is poor and usually subjected to famine and migration. Tigray region is one of the recurrently drought and war affected regions. The study area is also found in the southern part of this regional state. The annual amount of rainfall in most parts of the region is so small and unevenly distributed in time that crop failure and early drying of surface waters is common. The domestic water supply of the rural people is from the intermittent streams, low yield springs, shallow wells and some deep boreholes. Therefore, shortage of water supply due to shortage of enough recharged from rainfall to the stated shallow ground water and surface water sources is also one serious problem in the region. Even some deep boreholes serving urban areas are suffering from shortage of enough recharge from the rainfall whereas the ever-increasing population especially in the urban areas of the region may result to a continually increasing demand of domestic water supply. Some of the shallow and deep well sites in the region are selected with little localised studies or with out any investigation. This may affect their sustainability. Water quality problems are also observed in some urban and rural areas.

The lake Ashange and couple of intermitent and perenial rivers and springs are the surface water sources in the area. The springs and rivers that flow from the western and eastern highlands towards the valley floor become influent as soon as they encounter the valley floor and dry out not far from the foot of the mountains.

The Raya valley is dominantly composed of loosely compacted basin fill deposits and it is usually considered as a groundwater potential area for agricultural development.

1.2. Previous Works

Hunting Technical services (HTS) during the period of 1974 to 1975 has covered 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) is undertaken in the area between Kobo and Alamata towns covering part of the Alamata sub basin. This study 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.

The most recent hydrogeological study in the Raya valley was by 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. Accordingly, 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 m^2/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 – HCO_3 or Mg – Ca – HCO_3 type groundwater is common in the Raya valley. In different to the 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.

1.3. Purpose of the study

Now a days, the project area is characterised by low yield or total failure of seasonal crops and shortage of enough drinking water and grass for the livestock. Especially in the occasional dry periods, people and livestock become starved and migrated from their place to other areas of the country. This usually resulted in large-scale poverty and other socio-economic problems. Such problems in the study area can be avoided by having a proper study, development and management of both the surface and ground water resources. Therefore, this project deals with the assessment of water resources in the study area with special emphasis to the groundwater resources.

The specific objectives of this research work are:-

- Estimating the hydrologic balance of the study area.
- Determining the ground water flow directions.
- Assessing different hydrochemical properties of natural waters in the study area.
- Assessing the groundwater potential especially for large-scale irrigation development.
- Producing detailed hydrogeological map of the area.

1.4. Location and Accessibility

The study area is located in the northern eastern part of Ethiopia between $12^{0}32'$: $12^{0}54'$ north latitude and $39^{0}28'$: $39^{0}52'$ east longitude and specifically in the south-eastern part of Tigray region (Figure 1.1). It constitutes about 60 % of the Raya Valley, which is considered as one of the interconnected Valleys of the Ethiopian rift system (RVADP, 1998). Its areal coverage is about 1302 km^2 and accessed by the Maichew-Alamata, Mehoni-Maichew, Mehoni-Alamata and Mehoni-Chercher-Alamata all weather roads and other branching dry weather roads. The dry weather roads are usually dissected by traditional spate irrigation canals, which result in difficult driving. 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.

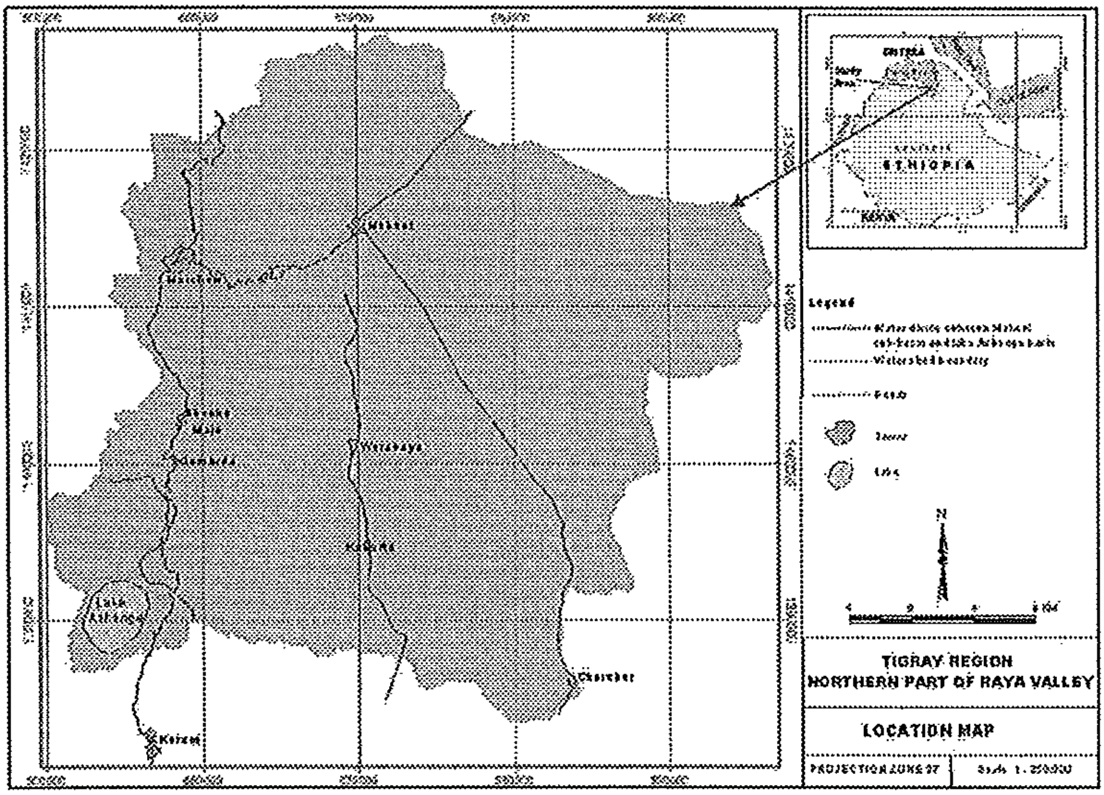


Figure1.1 Location map of the study area.

1.5. Population

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).

1.6. Climate and Vegetation

Due to wide altitude variations the study area is characterized by wide range of climatic elements. It is characterised 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 °c and 13.93 °c respectively for the western high lands and 23.44 °c and 19.64 °c respectively for the valley floor.

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 of cultivation land. Cactus tree also invades considerable land in the study area.

1.7. Geomorphology and Drainage

1.7.1. Geomorphology

Geomorphologically, the study area is characterised by an asymmetric graben and horst like structure bordered by fault scarps (Figure 1.2). The graben is filled with Quaternary sediments that are mainly derived from weathering and erosion of the Tertiary volcanics in the uplifted landforms. The landform of the project area can be a result of either one particular geologic process or sets of processes (such as; volcanism, tectonism, sturucrual deformation and further modifications by weathering and erosion processes). The study area can be classified into five physiographic units (Figure 1.3).

1.7.1.1. 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 faulting, fracturing, tilting and sliding. These disturbances resulted either to deep fault scarps that fall in to the valley floor or step like landform. The eastern plateau is characterised by rolling to dissected low hills. The plateau zone forms about 16.22% of the study area.

1.7.1.2. 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 not part of the large graben in the study area. One particular case of such basins is the lake Ashange basin. In general, this physiographic unit forms about 4.25% of the study area.

1.7.1.3. The Fault Scarp or Free Face Zone

This zone constitutes the steeply dipping escarpments or free faces. At places this zone rises as high as 2000m from the level of the valley floor. Some times this zone is characterised by detachment of boulders and rock falls that are induced by gravity and rainwater. The fault scarp zone forms about 21.39% of the sub-basin.

1.7.1.4. The Pediment or Alluvial Fan Zone

This zone is found between the pediment zone and the flood plain. It is characterised 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 14.71% of the sub-basin.

1.7.1.5. The Flood Plain

The flood plain constitutes the central part of the valley floor. It has more or less flat topograph. 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 favoured. This zone covers an area of about 570 km^2 and it is about 43.77% of the sub-basin.

1.7.2. 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 in Dessie Nadew, 2003) 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, Habro, Burka, Faje, Garba, Mai-Anishti, Karyo, Bulah and Gadwo (Figure 1.4). 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. All rivers in the lake Ashange basin drain towards the lake. Most of them are orientated in the north-south direction and they are also sub-parallel to each other indicating tectonically originated basin.

1.8. Methods and Materials

The methods employed in this work include literature review, field investigation and chemical laboratory analysis. The base maps used were topographic maps of scales 1:50,000 and 1:250,000, aerial photographs of 1:50,000 scale, and different geological and hydrogeological maps. Lithologic logs of 52 wells, pumping test data of 20 wells and important meteorological data were also compiled. The drainage map and 3-D model (TIN) are produced based on the 1:50,000 scale topographic maps by using the Arcview GIS software.

The fieldwork was aimed to acquire on spot observations and measurements and to collect water samples for chemical analysis. Geological Hammer, compass, stereoscope, deep meter, TDS meter, PH meter and thermometer were some of the field materials used. A total of 44 water samples were collected from the field. The water chemistry analysis was undertaken in the water laboratories of the Ethiopian Geological Survay (EGS) and Water Works Design and Supervission Enterprise (WWDSE). Pumping test and hydrochemical data were analysed by using softwares of the Waterloo-Hydrologic named as AquiferTest v3.5 and Aquachem 4.0 respectively. And then, different maps are produced by using Arcview GIS.

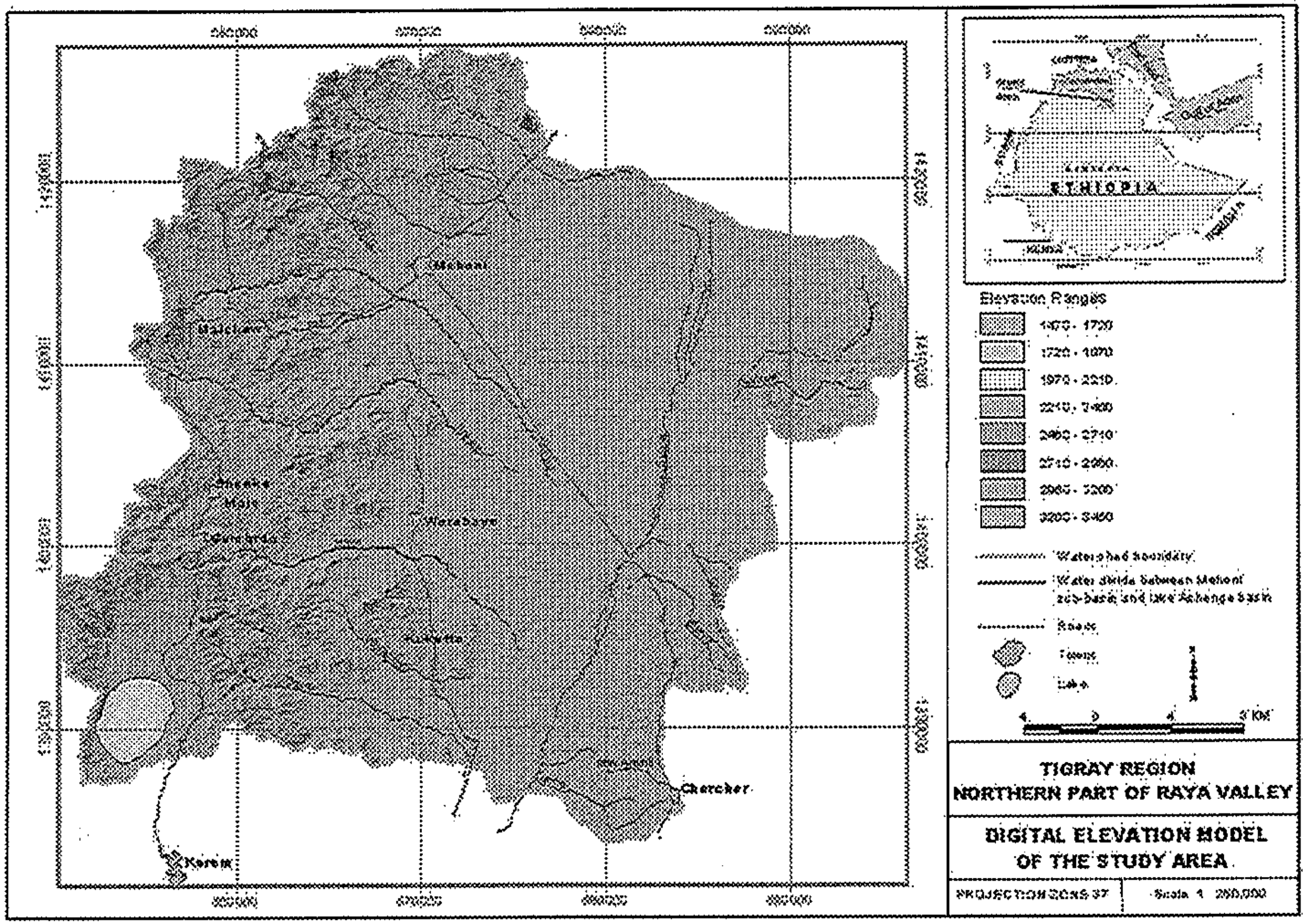


Figure 1.2. Digital Elevation Model of the study area.

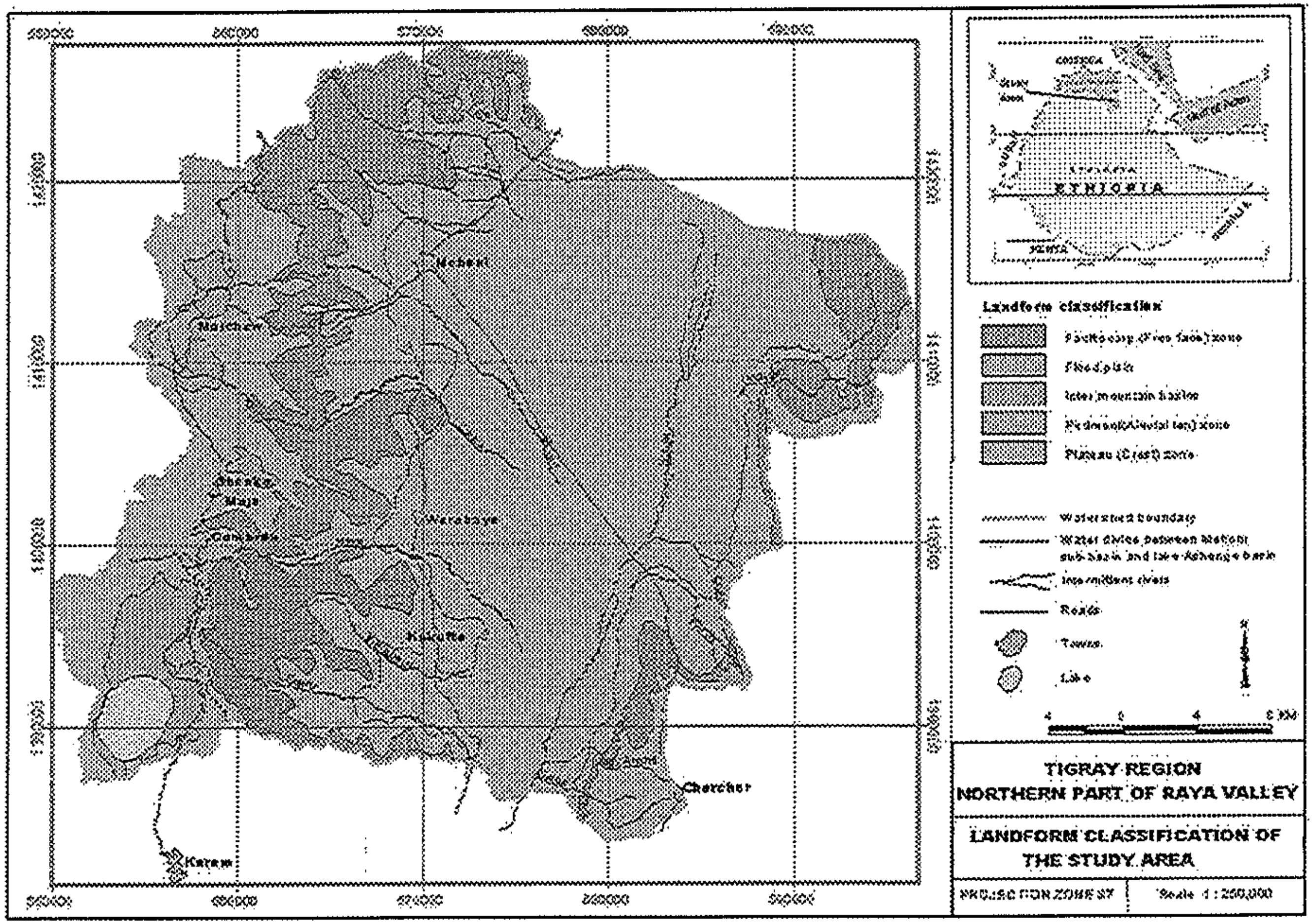


Figure 1.3. Landform clasification of the study area.

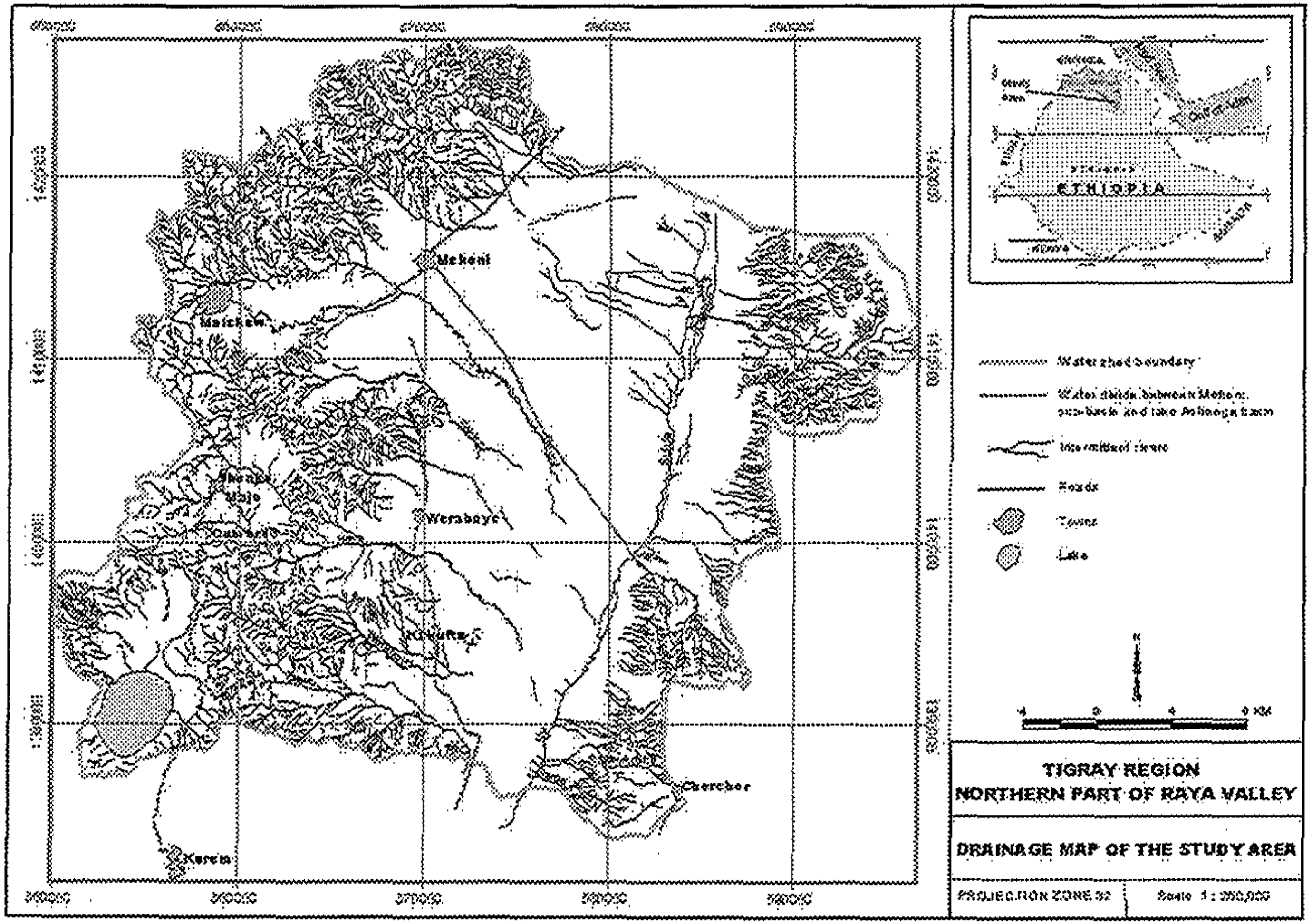


Figure 1.4. Drainage map of the study area.

CHAPTER TWO

2. Geology

2.1. Regional Geological Setting

The African and Arabian plates are now separated by the Red Sea and Gulf of Aden Spreading Centres. Extension 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 (R. 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 and Mohr & Zanettin, 1988 in Dessie Nadew, 2003).

New structural and geochronological data (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 (Welde Gabriel et al., 1990; Ebinger et al., 2000). Thus, the MER, which is the northern 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 lithospheric 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, In press).

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, In press). There is a rift ward migration of the locus of magmatism and faulting through time.

The felsic volcanic strata, capping the flood basalt sequences, are preferentially localised on or near the border fault, indicating that the age of the rhylites 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 al., 2005, In press).

The mineralogical and chemical composition of the flood basalts is relatively uniform. Most are aphyric to sparsely phyric, and contain phenocryst of plagioclase and clinopyroxene with or without olivine. Most have tholeitic to transitional compostions (Mohr, 1983a; Mohr & 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 Huchon (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 un deformed, flat lying tholeiitic basalts. Instead, it consists of a number of volcanic centers with different magmatic character and with a large range of ages.

2.2. Local Geology

The continental flood basalts of Ethiopia are commonly termed as the trap series to distinguish them from the post rift Aden 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: Ashage basalt; and Oligocen – Miocene stage: Aiba basalt, Alaji fissural basalts and rhyolites and Termaber basalts (B. Zenetin et al., 1980, Berhe et al., 1987 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 (R. Pik et al., 1998 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 Alaji 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 (B.Zebetin et al., 1980 in Dessie Nadew, 2003) (Figure 2.1)

2.2.1. Eastern highlands

The basalts that makeup the eastern highland is vesicular, friable and scoracious with variegated colours. 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 consists mudstone at the top, sand stone 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 50^0W and N20^0E). These orienatations are also common in fracture system of the basalt.

2.2.2. Western highland

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 ruminants 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.

2.3. Rock Description

The study area is basically composed of the Tertiary volcanics as a frame and unconsolidated sediments in the valley floor and intermountain depressions. Detailed geochemical and pertrological investigation was not done because it was not the scope of this work. But in the previous study of RVADP (Geo, 1996) a sort of rock classification was made and from the field verification in this work it was possible to acceptable this classification as a starting point to understand the rock formations in the study area.

2.3.1. Volcanic rocks

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.

2.3.1.1. Amygdaloidal basalt

This 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.

2.3.1.2. Alkali olivine basalt (A.O.B)

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.

2.3.1.3. Aphanitic (fine grained) basalt

This 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 intrussions were acting as feeders for the thick flood basalt accumulations (P.Mohr, 1988 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.

2.3.1.4. 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.

2.3.1.5. Volcano–Sedimentary unit

The volcano–sediments, which are encountered by the RVADP (1996) in the study area on the way from Maichew to Mekele, 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.

2.4. 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 26^0 towards NE and strike $N10^0W$. 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.

2.5. 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 (Isaak Gershanovich, 2000). Therefore the study area is one of the graben and horst lineaments with in the rift system. Where the graben is filled with sediments that 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 in RVADP, 1998).

These Quaternary deposits can be considered as alluvial fan deposits that interwave each other (Figure 2.3). And the gentle slopes of the valley floor might promote changes of stream coursesfrom time to time. Therefore, buried streams of predominantly alluvial sediments can be expected in the valley (Isaak 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, Shenko Majo and in the Lake Ashange closed basin. Lacustrine deposits such as diatomite and organic rich silty clay soils are also found only in the Lake Ashange basin.

2.6. Geological Structures

The study area and its surroundings are structurally similar to the step fault scheme of western Afar margin (Zanetin et al, 1978) (Fig 2.2). Such structural set up was clearly visible in the aerial photographs and 3D model of the study area (Fig 1.2). 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 throws 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. Lake Ashange is also a fault controlled basinal water fill.

Ingeneral, 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).

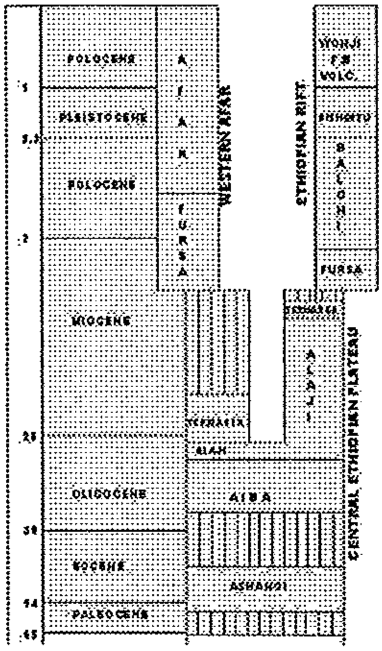


Fig 2.1. Schematic stratigraphic section of the Tertiary volcanics in Ethiopia (G. Zanettin et al in RVADP 1993)

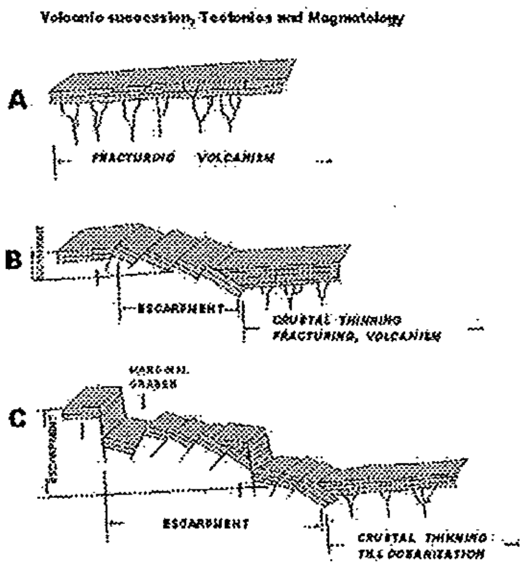


Fig 2.2 Tectonic and volcanic evolution of the western margin of the Afar, from A to C; 1) the fracturing, magmatic injection and volcanism are restricted to increasingly narrower belts; 2) the width and height of the escarpment increase while the rift narrows; 3. the narrowing of the rift is expected by thinning of the crust (G. Zanettin et al in RVADP 1993)

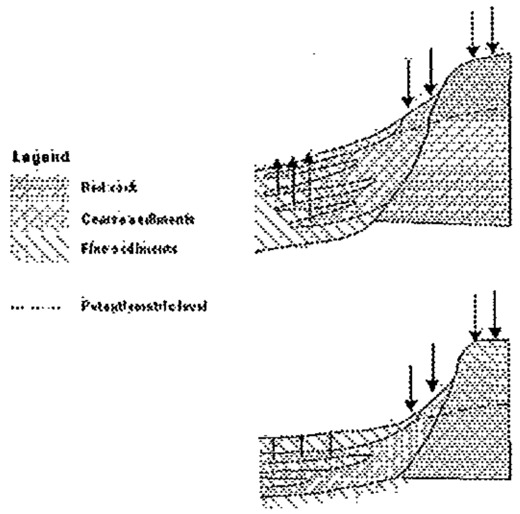


Fig. 2.3. CLASSICAL PATTERNS OF FAN-LIKE AQUIFERS
(after Bindeman N.N. - Estimation of Groundwater xploitation Reserve,1982)

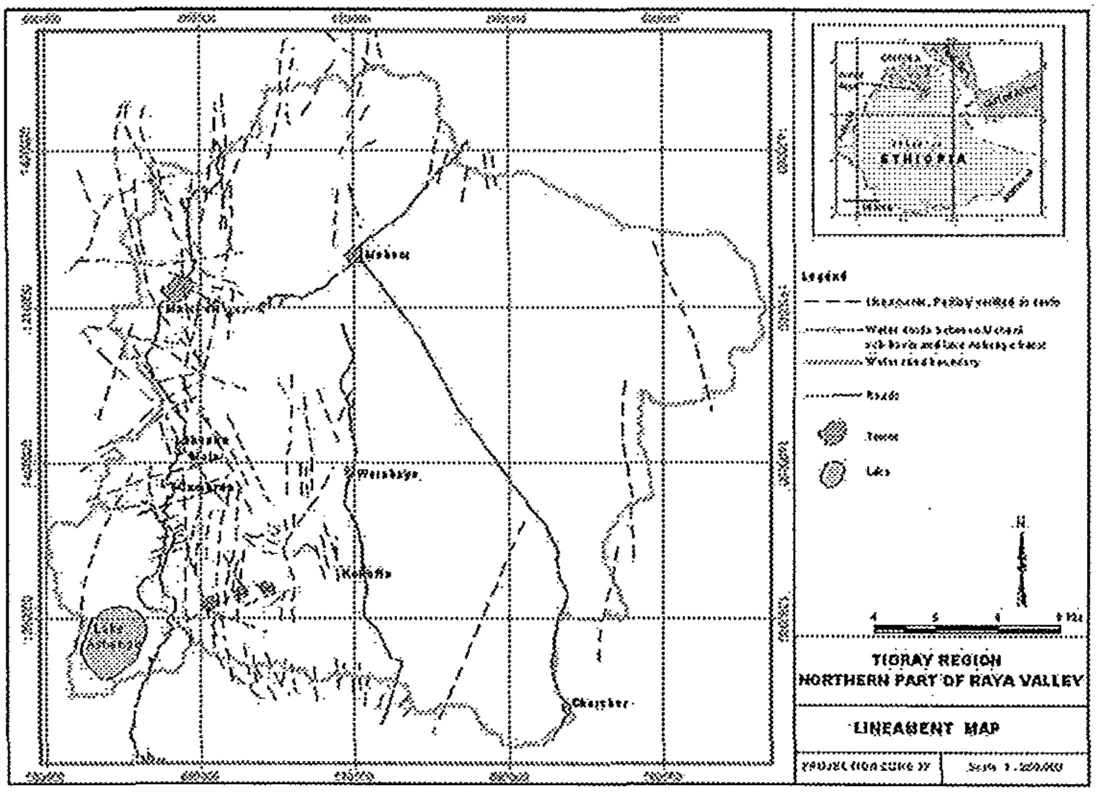


Figure 2.4. Lineament map of the study area

CHAPTER THREE

3. Hydrometeorology and Water Balance

3.1. Introduction

The available meteorological data are collected from six stations located in the towns Chercher (Weira wuha), Mehoni, Maichew, Meswaet, Korem and Alamata. The first five stations are located with in the range of study area while the Alamata station is south of the study area. With the exception of the stations at Mehoni and Meswaet, which are controlled by the respective Wereda Agricultural Bureaus, the rest are controlled by the National Meteorological Agency (NMA). Readings of wind speed, relative humidity, and solar radiation are available for the Maichew station only whereas rainfall and temperature were read in all the stations.

Table 3.1. The number of years with readings of 7 months and above for each parameter in the stations

Station	Precipitation	Temperature	Relative Humidity	Wind Speed	Solar Radiation	Remark
Chercher	9 years	5 years	-	-	-	
Mehoni	9 years	6 years	-	-	-	
Maichew	38 years	25 years	only mean monthly	9 years	9 years	
Meswaet	only mean monthly	-	-	-	-	
Korem	30 years	10 years	-	-	-	
Alamata	18 years	17 years	-	-	-	

3.2. Rainfall and Air Temperature

The spatial and temporal variations of rainfall in Ethiopia are strongly controlled by the inter-annual movement of the position of the Inter Tropical Convergence Zone (ITCZ). The ITCZ represents a low-pressure area of convergence between Tropical Easterlies and Equatorial

Westerlies along which equatorial wave disturbances take place. During its movement to the north and south of the equator, the ITCZ passes over Ethiopia twice a year and this migration causes the on set and withdrawal of winds from north and south. (Tenalem Ayenew and Tamiru Alemayehu, 2001)

The big summer (Keremt) rains, which occur when the ITCZ is found north of Ethiopia, are results of the Equatorial Westerlies from South Atlantic Ocean and Southerly winds from the Indian Ocean. The small rains of spring (Bulg), which occur when the ITCZ lies in the southern part, are results of winds from the Gulf of Aden and the Indian Ocean (anticyclone).

The study area, as most parts of Ethiopia receives a bi-modal type of rainfall. The Keremt rains occur during July to September and the Belg rains during March to May (Figure 3.1).

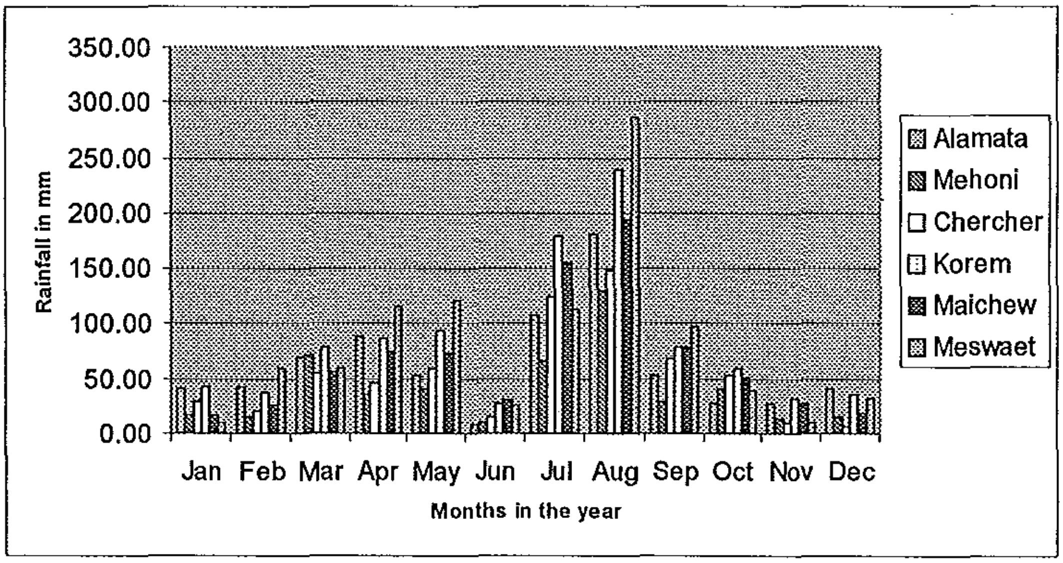


Figure 3.1. Monthly distribution of rainfall in the Mehoni sub-basin

3.2.1. Relationship of altitude to rainfall and air temperature

In the study area three stations are located on the western highlands, one station on the eastern highland and one station in the northwestern part of valley floor. The mean annual rainfalls of these stations have shown clear variations with altitude. That is, increasingly higher rainfalls are recorded with increasing altitude. The correlation coefficient between the mean annual rainfall and altitude is found to be 0.92. On average there is about 51mm annual rainfall increase for every 100m change in altitude. Unlike precipitation, air temperature is inversely related to the altitude.

Table 3.2. Mean annual rainfall and temperature in each station in the study area

Station	UTM location			Mean annual rainfall (mm)	Mean annual air temp. (^{0}c)
	X	Y	Z		
Alamata	561800	1373150	1560	995.11	15.97
Chercher	583294	1387187	1825	740.84	22.69
Korem	556135	1383742	2500	803.19	16.12
Maichew	557882	1415078	2400	637.14	21.47
Mehoni	570544	1415107	1650	485.43	20.89
Meswaet	557905	1404020	2400	971.6	-

3.2.2. Estimation of aerial depth of precipitation

The aerial depth of precipitation is estimated using isohyetal method. In the study area this method is preferable than the Theison Polygon method because there is uneven rainfall distribution due to orographic effects. The isohyetal method accounts for these effects (Tenalem Ayenew and Tamiru Alemayehu, 2001).

On the basis of the isohyetal map shown in figure 3.4 and data presented on table 3.3, the mean annual precipitation of the study area is estimated to be 723.57mm.

3.3. Evaporation from Lake Ashange

Lake Ashange is one of the highland lakes in Ethiopia but with small size. In this work the Penman nomogram method is used to estimate the free water surface evaporation from the lake. For the application of this method the mean daily temperature of the Korem station; and mean daily relative humidity, mean daily wind speed, and mean daily sunshine hours of the Maichew station are used. The mean daily solar radiations in a horizontal plane for the lake latitude are taken from the calculated values in the work of Dessie Nadew (2003) and the mean daily maximum possible sunshine hours at 13^{0}00' latitude.

Table 3.3 Inter-isohyetal areas and average annual precipitation

Isohyets in mm (A)	Estimated EUD in mm (B)	Inter isohyetal area in m^2 (C)	Percent of total area (D)	Weighed precipitation in mm (B*D)
>950	980	36831664.21	0.028	27.785
900-950	925	135114509.05	0.104	96.21
850-900	875	75237490.08	0.058	50.676
800-850	825	95407210.96	0.073	60.59
750-800	775	106119566.94	0.817	63.31
700-750	725	121669061.38	0.094	67.90
650-700	675	401807860.93	0.309	208.78
600-650	625	169389710	0.13	81.49
550-600	575	92107049.22	0.071	40.78
500-550	525	53351065.35	0.041	21.561
<500	485	12057210.62	0.009	4.5
	Total area	**1299092399**	Net EUD(mm)	**723.57**

EUD : Equivalent Uniform Depth

Table 3.4. Mean monthly rainfall as calculated from isohyetal method

Station	Jan.	Feb.	Mar.	Apr.	May	Jun.	Jul.	Aug.	Sept.	Oct	Nov.	Dec.
Alamata	40.73	42.44	70.73	88.64	53.21	7.71	107.49	181.30	52.54	27.84	27.45	40.77
Chercher	17.92	15.61	71.28	35.85	40.52	9.62	66.7	129.84	28.84	40.5	12.82	15.93
Korem	28.48	21.07	54.88	45.7	59.8	15.71	124.69	148.81	68.14	52.96	9.91	7.00
Maichew	42.71	37.82	79.31	87.7	94.59	27.83	179.96	238.4	78.74	59.45	33.28	35.34
Mehoni	16.44	24.92	56.8	74.57	73.18	30.79	155.77	194.63	77.91	50.55	28.05	19.58
Meswaet	10.4	60.6	60.1	115.5	120.4	25.3	111.9	286.4	98.1	39.4	10.4	33.1

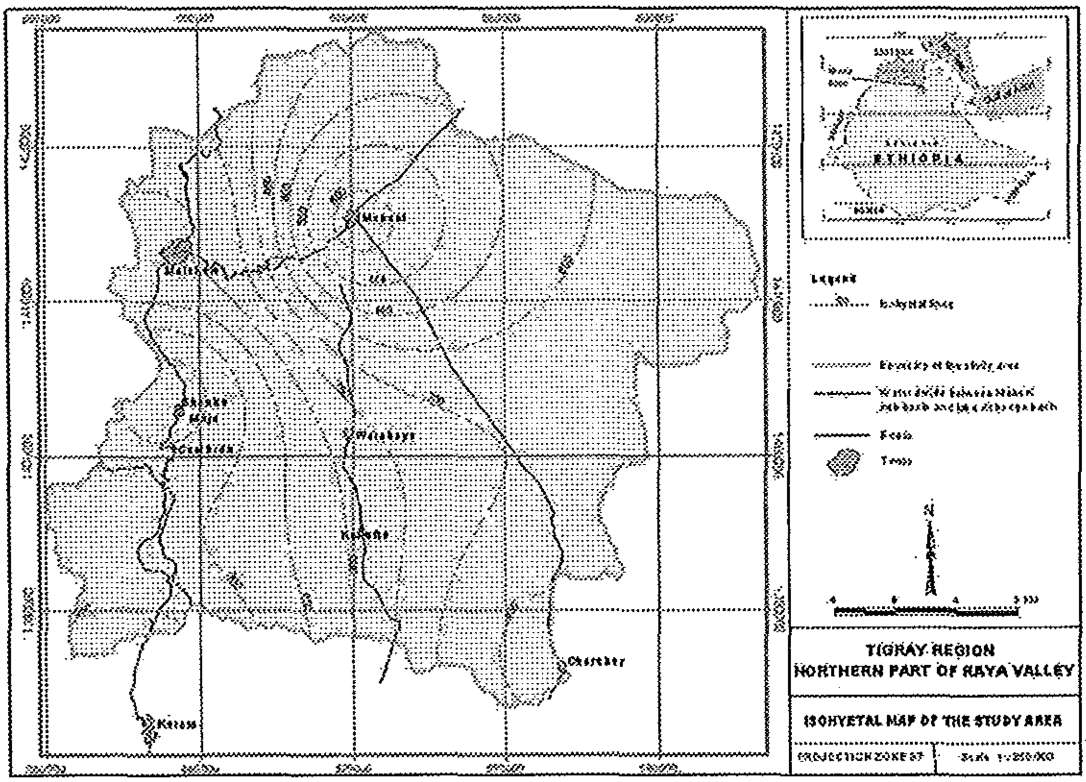


Figure 3.2. Isohyetal map of the study area.

Table 3.5. Data used for the estimation of the lake evaporation

Months	Jan.	Feb.	Mar.	Apr.	May	Jun.	Jul.	Aug.	Sep.	Oct.	Nov.	Dec.
Mean monthly air temp. (^{0}c)	14.78	14.4	15.56	16.57	16.81	17.54	17.74	17.21	16.68	15.37	14.68	14.27
Mean monthly RH at 0600(%)	78	68	76	74	73	56	70	0.74	78	85	0.82	78
Mean monthly wind speed at 2 meters a.g.l	1.17	1.2	1.34	1.34	1.41	2.33	3.41	2.53	1.28	1.15	1.14	1.12
n/N	0.72	0.78	0.69	0.66	0.76	0.47	0.42	0.44	0.52	0.67	0.73	0.71
Solar radiation (Laglays/day)	455	467	523	575	573	521	424	426	507	492	461	449

RH : relative humidity

n : number of sunshine hours per day

N : maximum possible number of sunshine hours per day

Finally by plotting these data in Penman nomogram, the lake evaporation is estimated to be 722.42 mm/year. But plotting of the data and estimating the evaporation values in this method is highly dependent on individual judgment that may lead to unpredictable errors.

In the previous studies evaporation from this lake is estimated by using different methods. In the works of Dessie Nadew (2003) and RVADP (1998) the evaporation from the lake is estimated to be 1204 mm/year and 1169mm/year respectively. For the water balance calculation in this work the estimate of Dessie Nadew (2003) that apply the U.S. Weather Bureau nomograph (Kohler, Nordeson and Fox, 1955) is used.

3.4. Evapotranspiration

Evapotranspiration can be defined as a natural process that transfers water vapour to the atmosphere from a vegetation covered area. It incorporates both the evaporation from the land surfaces and transpiration from the vegetation surface. There are two types of evapotranspiration. If sufficient moisture is always available to completely meet the needs of vegetation that fully cover an area, the resulting evapotranspiration is called potential evapotranspiration (PET). This time PET no longer critically depends on the soil and plant factors but depends essentially on climatic factors. The real evapotranspiration occurring in a specific climatic and soil moisture conditions is called actual evapotranspiration (AET).

3.4.1. Estimation of PET and AET in the study area

In this work the method employed to estimate potential evapotranspiration is the Thornth weite method. The formula of Thornthweite was developed from data of eastern USA and uses only the mean monthly temperature together with an adjustment for day-lengths.

The method can be expressed by

$$PET = 16\, L_a \frac{(10T_n)^a}{(I)^a}$$

Where

PET : Monthly gross potential evapotranspiration in mm

L_a : Adjustment for the number hours of daylight and number of days in the month

T_n : Mean monthly air temperature in ^{0}c

I : Annual heat index obtained by summing the monthly heat indices (I_n)

$$I_n = \frac{(T_n)^{1.514}}{(5)^{1.514}}$$

n : 1, 2, 3, ..., 12

a = 0.000000675 I^3 - 0.0000771 I^2 + 0.01792 I + 0.49239

Therefore, PET of the study area as estimated by the Thornthweite formula is equal to 1124.3 mm/year. As estimated from the PET and precipitation values for each month (table 3.8) and assuming soil moisture accretion and utilization to be 100 mm/year, the annual AET of the area is approximated as 725 mm/year. This value is almost equal to the mean annual precipitation of the study area.

Table 3.6. Mean monthly air temperature (^{0}c) for different stations in the study area.

Station	Type	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec	Mean
Alamata	max T	26.8	27.7	29.1	29.7	31.7	33.9	31.8	30.5	30.8	30.2	28.7	27.8	29.9
,,	min T	12.9	13.7	15	16.3	17.3	18	17.7	16.8	15.8	15.3	14	13.1	15.5
	mean	19.9	20.7	22.1	23	24.5	26	24.8	23.6	23.3	22.7	21.3	20.4	**22.7**
Chercher	max T	25.1	25.4	25.8	26.9	29.1	33.3	29.9	28.8	27.3	27.1	26.5	25.3	27.5
,,	min T	13.3	13.6	15	16.1	16.9	19	16.7	16.2	16.5	15.1	13.8	12.6	15.4
	mean	19.2	19.5	20.4	21.5	23	26.1	23.3	22.5	21.9	21.1	20.2	18.9	**21.5**
Korem	max T	20.6	21.1	22.3	22.8	23.4	24	23.1	22.6	23.2	21.9	21.4	20.8	22.3
,,	min T	8.92	7.71	8.82	10.4	10.2	11.1	12.3	11.8	10.2	8.83	7.98	7.77	9.67
	mean	14.8	14.4	15.6	16.6	16.8	17.5	17.7	17.2	16.7	15.4	14.7	14.3	**16**
Maichew	max T	22.1	22.6	23.6	23.9	25.7	26.4	23.7	23.6	24	22.8	22.4	21.5	23.5
,,	min T	6.43	7	8.76	9.76	10.1	11.6	12.1	11.4	9.2	6.82	5.9	5.7	8.72
	mean	14.2	14.8	16.2	16.8	17.9	19	17.9	17.5	16.6	14.8	14.2	13.6	**16.1**
Mehoni	max T	26.8	26.8	27.6	28.3	29	29.8	28.3	26.6	27.4	27.3	27	26.6	27.6
,,	min T	12.9	13.3	13.1	13.3	14.7	15.3	16.2	15.3	14.7	14.5	13.8	12.9	14.2
	mean	19.9	20.1	20.4	20.8	21.8	22.5	22.3	20.9	21	20.9	20.4	19.7	**20.9**

Table 3.7. Computation of PET and AET by the Thornthweite method.

Months		Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec	Total
Mean daily air temp.(^{0}c)		17.6	17.9	18.9	19.8	20.8	22.2	21.2	20.4	19.9	19	18.2	17.4	
Monthly heat index "i_n"		3.5	3.6	3.8	4	4.2	4.5	4.2	4.1	4	3.8	3.6	3.5	
Standardised PET (mm)		81.8	83.5	89.4	94.28	101	109	103	97.9	95.2	89.8	85	80.6	
Lat. Correction factor (L_a)		0.97	0.98	1	1.03	1.05	1.06	1.05	1.04	1.02	0.99	0.97	0.96	
Corrected PET (mm)		79.3	81.9	89.4	97.1	106	116	108	101.8	97.1	88.9	82.4	77.4	1124.3
Mean monthly rainfall "P" (mm)		13.1	28.4	56.6	69.8	69.9	25.1	124	186.7	70.1	44.8	19.8	18.2	726.2
PET minus P	Negative	-66.2	-53	-32.8	-27.3	-36	-91			-27	-44	-62.7	-59	-498.5
	Positive							15.8	84.93					100.71
Uncorrected AET		13.1	28.4	56.6	69.8	69.9	25.1	108	101.8	70.1	44.8	19.8	18.2	625.5

Annual AET = 725.56 mm (assuming soil moisture utilization to be 100 mm

The actual evapotranspiration is also calculated by another method using the formula of Turc, Langbein and Wundit as follows:

$$\overline{E} = \frac{\overline{P}}{\sqrt{0.9 + \left(\frac{\overline{P}}{J}\right)^2}}$$

Where:

$J = 300 + 25T + 0.05\,T^3$

$\overline{P}$: Mean annual precipitation in millimetre

$\overline{E}$: Mean annual actual evapotranspiration in millimetre

T : Mean annual air temperature in ^{0}C

In case of this work $\overline{P} = 723.57$mm ; T = 19.43 ^{0}C ; J = 1152. 52 and $\overline{E} = 636$ mm/year

For the water balance calculations in this work, the AET is taken as 687.24 mm/year that is the average of the estimates by the two methods.

3.5. Run off

The runoff from the western highland and escarpment areas is approximated by using the annual flow generated by the SCS model (Soil Conservation Service in US America) for dam sites in the study area and its surroundings (RVADP, 1998). Accordingly, by taking the average of these flows a value of 176mm of annual runoff is approximated to flow from the western escarpment to the valley floor. Taking the 75% dependable flow a value of 105.8mm is found (Dessie Nadew, 2003). As it can be seen from the drainage map and by asking local people of the study area, it is found that almost all the flow from the western, northern and northeastern parts of the study area percolate into the unconsolidated sediments of the valley floor. From the above figure the approximate runoff coefficient of the mountainous area could range from 0.22 to 0.13 (Dessie Nadew, 2003). In general, the runoff that could possibly leave the study area during rainy season is assumed to be derived from the eastern and southwestern parts. Therefore, the water that leaves the Mehoni sub-basin in a form of runoff is estomated to be 52.3 MCM by taking the area converge of the eastern and south western drainage systems (i.e. 32849469 m^2), the runoff coefficient as estimated by the RVADP (1998) (i.e. 0.22) and the average annual rain fall of the study area from isohyetal method.

3.6. Water Balance

The water balance represents the hydrological gains and loses of a given system (reservoir, column of soil, aquifer, river basin, etc) over a specific period. The water budget for the recharge area of an aquifer is a very useful means of determining groundwater recharge.

The general form of water balance equation is as follows:

Inflow - Outflow = $\pm$ change in storage

In case of calculation for ground water recharge, the above equation can be rewritten as follows:

Ground water recharge = (Precipitation + Surface water inflow + Imported water + Ground water inflow) – (Evapotranspiration + Reservoir evaporation + Surface water outflow + Exported water + Ground water outflow) $\pm$ change in storage

(Fetter, 1994).

The annual recharge of the study area, which is assumed as a closed groundwater basin, can be therefore estimated by the following equation.

Recharge = Precipitation – (Actual Evapotranspiration + Lake evaporation + Surface water outflow).

The mean annual precipitation of the Mehoni sub-basin as estimated from the isohyetal method is about 723.57 mm and the total area of the study area is estimated to be 1302.3km^2. Based on this, the volume of input water in a form of precipitation is estimated to be 942.3 million cubic meters.

The annual actual evapotranspiration for the vegetation covered area, which accounts for about 83.93% of the sub-basin as computed from the Turc method, is estimated to be 695.2 million cubic meters.

The annual reservoir evaporation from lake Ashange, which covers about 1.067% of the sub-basin as, calculated by the United States Water Bureau nomogram method (Dessie Nadew, 2003), is estimated to be 16.73 million cubic meter.

The annual evaporation from the bare land, which accounts about 15% of the sub-basin approximated as half the free water surface evaporation, is found to be 117.6 million cubic meters.

The annual surface water outflow is estimated to be 52.3 million cubic meters.

Then by substituting these values into the above equation a value of 60.49 million cubic meters will be the estimate for the annual recharge to the groundwater in the study area. Therefore, this amount of water can be abstracted from the groundwater in the whole study area with out affecting the water table. But this will be true only if the assumptions set for the water balance calculation are found to be true.

CHAPTER FOUR

4. Hydrogeophysics

4.1. Introduction

To determine the properties of the sub-surface formations it would have been good to make observation or measurement of the material directly. However, for many reasons this may not be always possible. Therefore, the effective and widely used approach is to sample the sub-surface materials at selected locations by using geophysical methods, and then extrapolate and/or interpolate the properties between the sample locations. Of course, the number of sample locations and the spacing between them limit the success of this approach. Large number of systematically located samples leads to more reliable predictions. In practice the effective approach, to characterize a target, is to combine the geophysical methods with observations from several sample locations to do a guided interpolation between the sample points.

Geophysics is the science of measuring variation of physical properties of the subsurface material and interpreting it in terms of earth sciences (geology, hydrogeology, and/or others). It is used to detect contrast of physical properties that enable us to differentiate one hydrolithologic unit from the other. Its suitability to study large area with a modest cost per unit area and with in short period of time is appreciable.

Some of the objectives of hydrogeophysics are:

- to acquire qualitative knowledge about geological features, including water-bearing formations, and perform quantitative analysis (i.e., estimate their thickness, lateral dimensions, and physical properties like electrical resistivity/ conductivity, magnetic susceptibility, etc).
- to identify/delineate structurally weak zones like faults, fracture zones and fissures that usually serve as conduits for the ground water to flow into or out of the aquifers.
- to map the depth and morphology of impermeable bed rocks and outline their regional trends.

- to delineate the interfaces between fresh and saline water and also approximately evaluate degree of mineralization of underground water.
- to study the ground water dynamic (velocity and direction of ground water flow/movement).

The geophysical methods that are mostly applied in hydrogeology are the electrical resistivity method (in profiling & sounding modes), electromagnetic method, induced polarization (IP) method, self-potential (SP) method, magnetic method, seismic refraction method and other geophysical well logging methods.

4.2. Hydrogeophysical survey in the study area

The vertical electrical sounding (VES) with the shlumberger array was domonantly applied for ground water exploration and well sitting purposes in the study area. During VES survay the depth of penetration increases with increments of the separation between the current electrodes while the center of the electrode configuration is kept fixed. The depth wise resistivity variations are measured directly below this central point. In the previous works the sounding data were initially interpreted using curve-matching techniquè, and then the first model parameters were improved through an iterative process of curve generation and fitted to model curves using the computer program **"Resist"** (1982).

The depth/thickness and resistiity parameters acquired by quantitative interpretation of sounding curves were used to construct geoelectric sections so as to show the distribution of different lithologic units both vertically and laterally (EIGS, 1996).

Almost all the electrical resistivity surveys in the study area were undertaken on the valley fill unconsolidated sedimentary deposits with the aims of determining loose sediment thickness, depth to the saturated media and its thickness and depth to the bedrock. In addition some prediction of the lateral lithological variations and identification of fault zones in the bedrock were made.

The geophysical surveys in the Mehoni sub-basin and the Raya valley as a whole are conducted by the German consult (1978), Ethiopian Geological survey (1996), Bureau of Water Resources (1999, 2003), and Freelance consult (2004).

The VES survey of the German consult is undertaken in the Alamata sub-basin in the Raya valley. They conduct their survey along different traverses with the vertical electrical sounding. Their maximum electrode spacing (AB/2) was 350m and the maximum thickness of the loose sediment deposit is found to be greater than 250m. It has also been observed that the total thickness of the sediment and the aquifer thickness increase from west to east whereas transmisivity decreases in the same direction. (RVADP, 1998).

Seismic refraction method was also carried out to supplement resistivity method in the Alamata sub-basin by the German consult (1978), with 350m maximum depth of penetration. From the seismic refraction interpretation made by the German consult, two layers were identified. The upper layer comprised of dry unconsolidated sediments at the top underlain by water-saturated sandy material with seismic velocities of 500 to 800 m/s and 1200-1800m/s respectively. Below these sediments unaltered basaltic rock was assumed, since a sudden change of velocity up to 2500-3500m/s was manifested and this velocity value usually corresponds to velocity of compact rocks. But in the saturated layer some minor discontinuities were also observed from the seismogram due to the inter fingering of lenticular aquifers and equators with differing contents of ground water and air in the pores.

The study area in particular was largely covered by the vertical electrical soundings of the EIGS (1996). The VES points were aligned along four traverse lines: Kukufto-Mehoni, Mehoni-Ardabasa, Mehoni-Chercher and Chercher-Kukufto (Figure 4.1). In this survey the VES measurements were made with maximum spread length (AB/2) of up to 500m and 750 in some places. A summary of these VES traverses is presented in table 4.1 (after RVADP, 1998). Three geo-electric sections along Kukufto-Mehoni, Mehoni-Ardabasa and Mehoni-Chercher are also included in this work (Figure 4.2, 4.3 and 4.4).

From the VES survey of EIGS the following conclusions were drawn (after RVADP, 1998).

- The thickness of the unconsolidated material (basin fill deposits) increases from west to east and attains its maximum (>250m) in the eastern part.
- The material is coarser in the west (coarse sand and gravel) and gets finer and finer towards the east.
- Depth to bedrock is relatively shallow in the west (in most cases less than 100m) getting deeper and deeper in the central and eastern part (greater than 250m).

Table 4.1. Summery of VES traverses in the study area (After RVADP, 1998)

No	Traverse line	Resistivity (ohm-meter)	Thickness (meter)	Lithology
1	Kukufto-Mehoni (VES 8, 9, 10, 11)	12 - 67	5 - 20	Sandy soils with rock fragments
		10 - 16	40 - 120	Thick saturated sand
		60 - 93	infinity	Bed rock
2	Mehoni-Chercher (VES 16, 17, 18, 19, 20, 21)	10 - 36	1 - 3	Sandy soils with rock fragments
		23 - 36	1-21	Dry coarse sand with boulders
		7 - 17	37-95	Saturated coarse sand
		8 - 23		
3	Mehoni-Ardabasa (VES 12, 13, 14, 15)	8 - 34	2 - 10	Silt, clay and sandy soil
		11 - 68	6 - 10	Coarse sand and boulders
		13 - 37	50 - 100	Fine sand (possibly saturated)
4	Kukufto-Chercher (VES 22, 23)	12 - 28	1-5	Clayey sand
		21 - 43	20 - 50	Sand
		>80.9		Bed rock

Generally, the sediments in the valley floor were classified in to the following three lithologic units.

i. The top layer, which is characterised by thickness range of 1 to 20m and relatively high apparent resisitivity. These are attributed to dry sand, silt and sandy clay.

ii. The middle layer, which is composed of fine to medium gravel and coarse sand in the west and coarse to medium sand with intercalations of fine sand and clay in the central and eastern parts. This layer has relatively low apparent resistivity. This layer is predicted as the aquifer zone of the valley floor loose sediments and has a thickness range of 20-60m in the west and 50-160m or larger in the central and eastern parts.

iii. The bottom layer, which is characterised by very low resistivity and relatively higher thickness of sediment. It is found in the central and eastern parts underlying the middle layer. It is composed of finer materials, possibly lacustrine deposits and has unknown thickness.

An exception to the above generalization was the geo-electric section along the Mehoni-Ardabasa where the middle layer is composed of finer materials with a thickness of about 60m and the bottom layer is composed of coarse sediments (Figure 4.3).

More detailed VES survey was undertaken by the Tigray Bureau of Water Resources (TBoWR, 1999) in the western part of the valley floor around Werabaye village (WF9 and WF10). It incorporated 30 VES (V5 toV16 and V17 to V34 in figure 4.5) and 172 points of resistivity profiling within an area of about $3km^2$. The VES and resistivity profiling data collection was conducted in three parallel lines each separated with 500 meters and the distance between VES points being 400m. The electrode spacing (AB/2) in most of the VES points ranges between 400m and 500m. And the profiling was conducted with $\overline{AB}$ of 150m, 300m, 600m and 900m with total 172 readings in order to see the lateral variations, which were not clearly identified by the Pseudo Sections constructed from the VES data. The quantitative interpretation of the VES data was done by using a software called Resist version 87 and the out put model parameters have been used to construct three geo-electric sections. But the profiling data was interpreted only qualitatively.

In general, from the quantitative VES interpretations of TBoWR (1999) the composition of the sub-surface materials has been summerised in table 4.2.

With in the area coverage of this survey the estimated depth to the bedrock ranges from 75m to 180m, with an average of 130m. Depth to the water table also ranges from 40m to 60m.

Pseudo-sections that were constructed from the VES data of TBoWR (1999) has shown that there are some lateral variations in the resistivity of the sub surface materials due to variability in the dominance of either coarse or fine sediments and due to the lateral variations of the depth to the bed rock.

Similar results were found from three VES points of TBoWR (2003) in the South Western part of the sub-basin (Figure 4.5) because of similarity in the alluvial fan deposits in both areas.

Table 4.2. Summery of the VES survey around Werabaye

Unit	Resistivity range	Thickness range	Lithology
Upper unit	Dominantly $10\Omega m$ - $40\Omega m$	20m – 50m	Dry clay, silt and sand soils
Middle unit	$5\Omega m$ – $16\ \Omega m$	60m–150m dominantly around 100m	Medium sand to fine gravel sometimes with intercalations of clay silt layers and fine sand
Lower unit	Around $34\Omega m$ for the weathered saturated basalt and $60\Omega m$-$67\Omega m$ sometimes upto $150\Omega m$ for the dry massive bedrock.	4m–10m weathered & saturated basalt underlain by dry massive basaltic bedrock.	Basaltic bed rock

Five VES points (V40 to V46 in figure 4.5) undertaken by the Freelance consult (2004) in the area north of Mehoni town also revealed that the depth to the bed rock is relatively shallower (i.e., less than 70m) and the anticipated aquifer material in that area is composed of sand and coarser materials. The information from the wereda Mehoni TBoWR also indicated that most of the boreholes that are drilled in the northern extreme areas of the Mehoni sub basin were either with low yields or dry. This can be because of extremely high permeability and shallow thickness of the valley floor sediments and absence of enough recharge.

Another five VES points (V35 to V39 in figure 4.5) that were done by the Freelance consult (2004) in the central part of the sub-basin around Ganda-Chafe (figure 4.5) has indicated that the depth to the bed rock in that area can be greater than 300m and the water bearing zone has resistivities that range between $6.5\Omega m$ and $8\Omega m$. Silts and fine sand dominate the sediments in this area.

Table 4.3. Summery of the VES survey around Ganda-Chafe

Depth	Resistivity	Lithology
0 – 35m	12Ωm - 42Ωm	Dry silty clay soil
35m – 300m	6.5Ωm - 8Ωm	Silt dominated intercalations of clay, silt and sand (saturated)
Below 300m (?)	> 49Ωm	Basaltic bed rock (weathered / fractured)

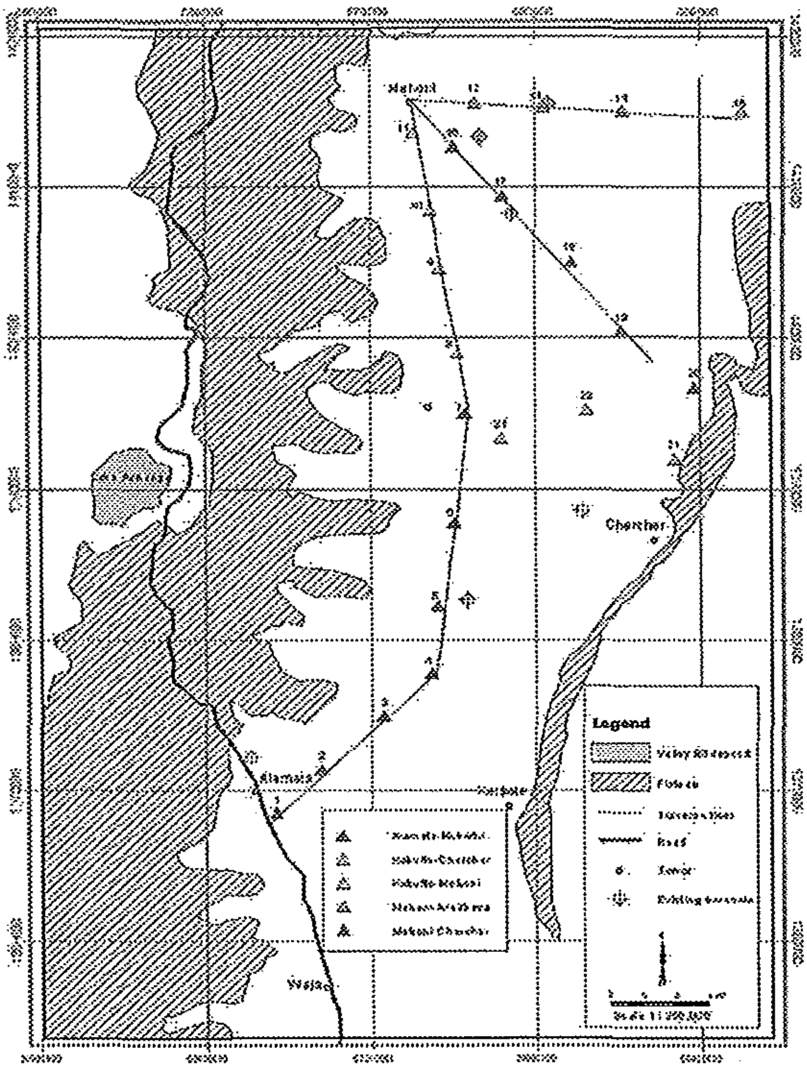


Figure 4.1. Location of EIGS geophysical survay sites with in Raya Valley (modified from RVDP (1998)

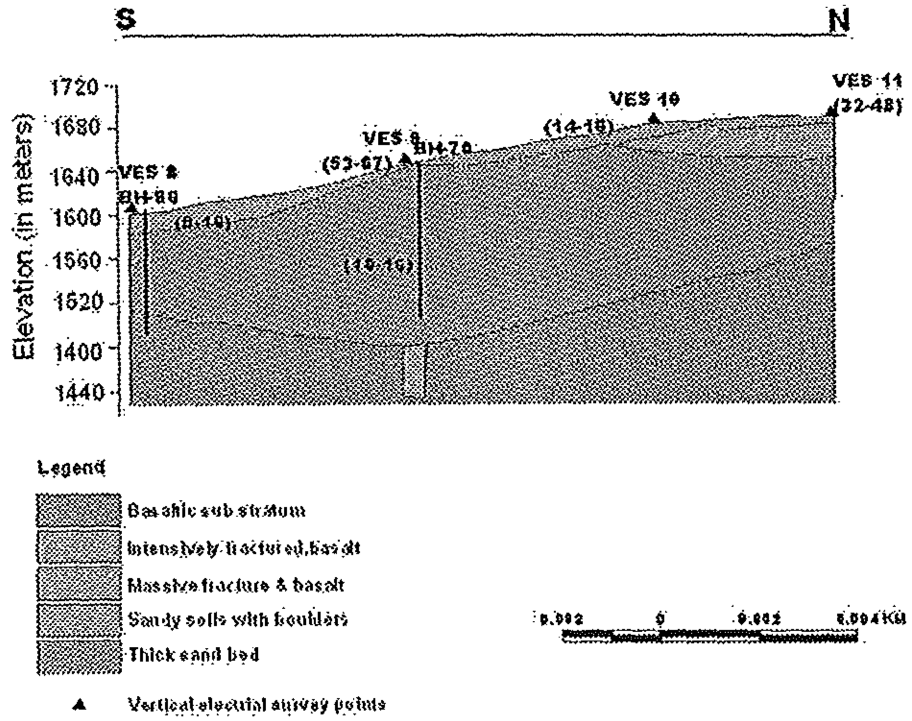


Figure 4.2. Geoelectric section along Kukufto-Mehoni regional traverse line (refer figure 4.1)

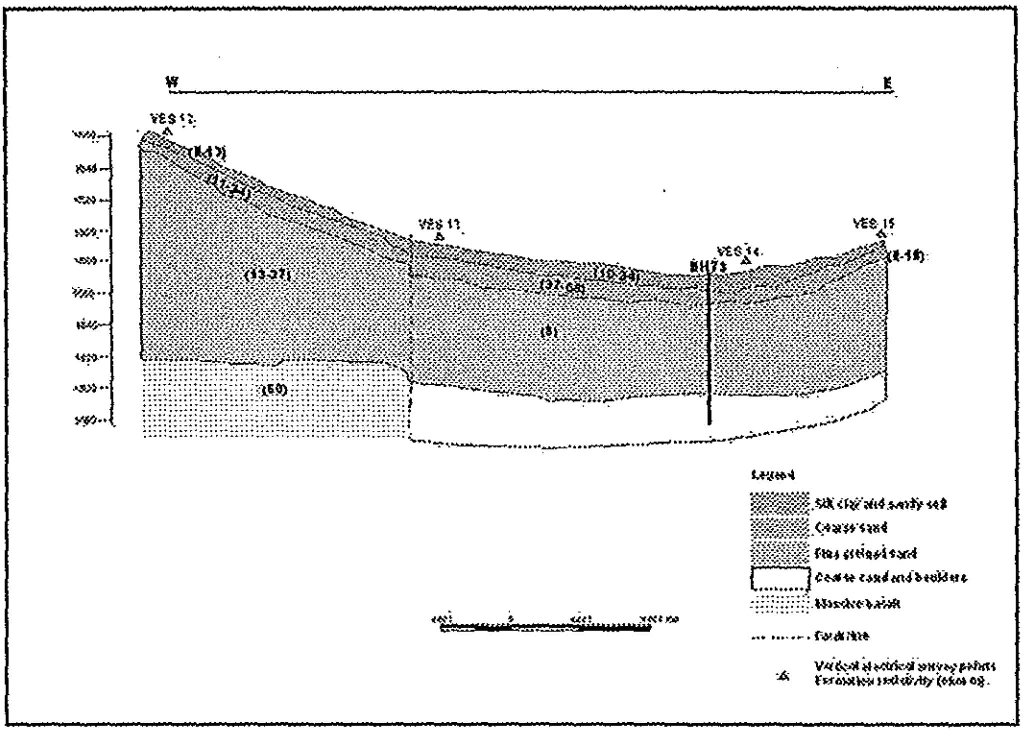


Figure 4.3. Geoelectric section along Kukufto-Ardabasa regional traverse line (refer figure 4.1)

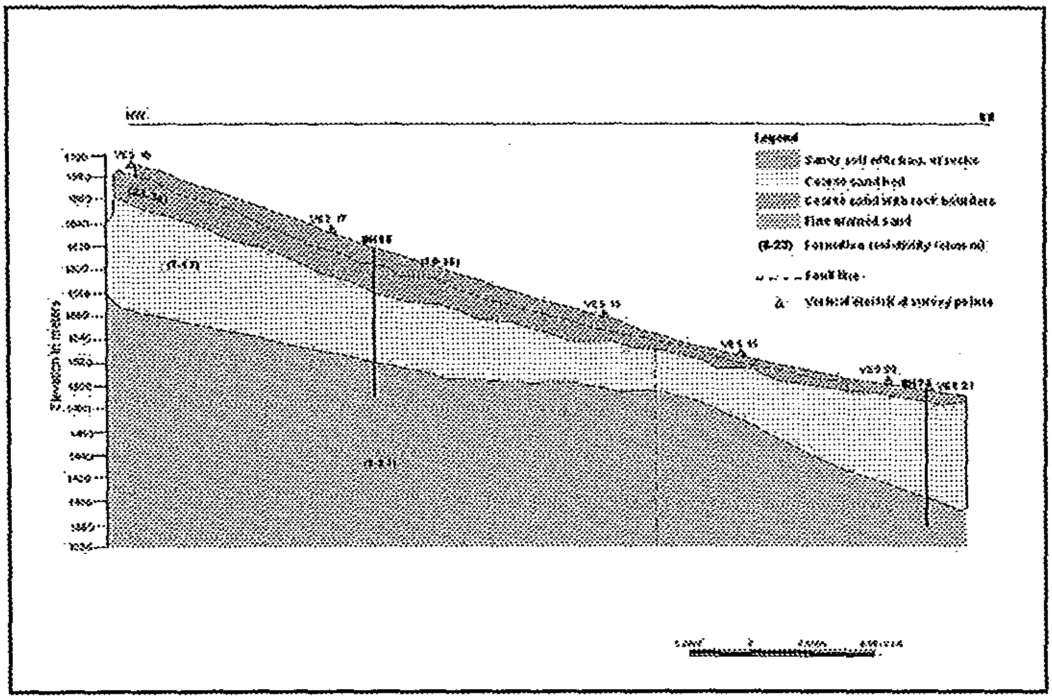


Figure 4.4. Geoelectric section along Mehoni-Chercher regional traverse line (refer figure 4.1)

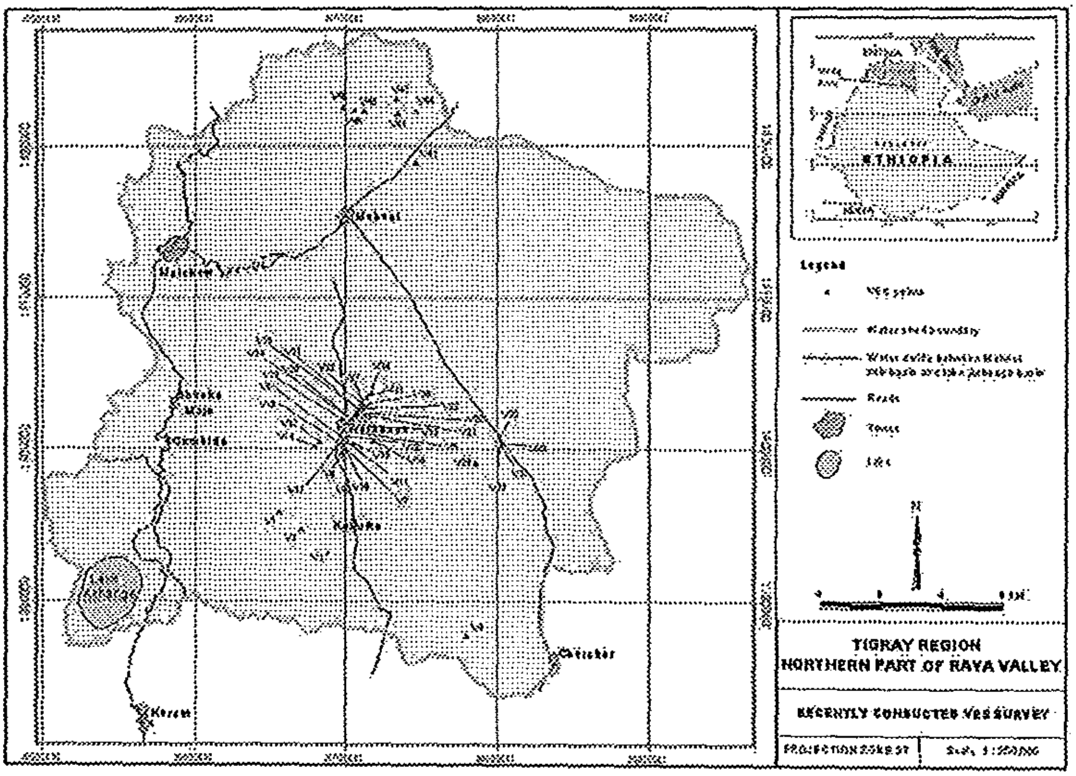


Figure 4.5. Location of recently conducted VES surveys

CHAPTER FIVE

5. Hydrolithologic units and groundwater conditions

The hydrolithologic units in the study area can be broadly classified as alluvial aquifers and volcanic aquifers. The alluvial aquifers are composed of poorly compacted Quaternary deposits, which fill the valley bottom and intermountain troughs. Whereas, the volcanic aquifers are that found within the fractured and weathered parts of the Tertiary volcanics in the highlands and escarpments of the study area.

5.1. Aquifers

In previous works the alluvial aquifers in the Raya valley were further classified into three sub-units (RVADP,1998 and Dessie Nadew,2003). But in this work after closer study on systematically sampled well lithological logs and pump test analysis results the alluvial aquifers in the Mehoni sub-basin are classified only into two sub-units. They are the Quaternary colluvial and alluvial deposits (Q_{col} and Q_{al}) and the Quaternary interfluvial, fan foot plain and valley bottom deposits.

5.1.1. Quaternary colluvial and alluvial deposits (Q_{col} and Q_{al})

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

The Quaternary alluvial deposits (Q_{al}) 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 Q_{col} and Q_{al} aquifers. Also water supply for people around Maichew town and in the lake Ashange basin is pumped from wells drilled into Q_{col} and Q_{al} aquifers.

5.1.2. Interfluvial, Fan foot plains and Valley bottom Quaternary deposits (Q_{smc})

Larger part of the valley bottom unconsolidated deposit is comprised of the interfluvial, fan foot plains and valley bottom Quaternary deposits (Q_{smc}). 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 Q_{smc} 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.

5.1.3. Volcanic aquifers

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. This can be approved by the perennial rivers (Guguf and Haya) and perennial springs (CS8, CS10, CS6, CS1 and HS1) that emanate from the western highlands. There are also 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.

5.2. Groundwater level fluctuation

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 and in the lake Ashange basin wells close to the lake have no significant fluctuation but those far from the lake have shown fluctuations that reach up to 5m. In the valley floor at BH24 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 (Figure 5.5b). In this period the water level had shown a declining trend with temporary rises in the rainy months of the year.

One of the expected reasons for this decline in water level was the change in climate in the recharge area, which may decrease the amount of ground water recharge to the aquifers in the study area. To study this possibility the daily air temperature and rainfall data of the Maichew station for the same period as that of the water level reading period were plotted on a linear graph and compare with that of the water level fluctuation versus time graph (Figures 5.1a, 5.1b, 5.1c). But as it was observed from these graphs the diurnal variation of the maximum air temperature and rainfall before and after the start of pumping was more or less similar with slightly drier Bulg in the year 2004. This can indicate that the climate has only slight effect on the observed water level fluctuation. Other factors, which are considered as more important, are the large-scale pumping for the irrigation practices around Kara Adishawo and relatively more diversion activities in the up stream areas for spate irrigation and for recharging old and newly constructed communal ponds. The pumping may discharge considerable amount of groundwater from the sub-surface while the diversion significantly decreases the amount of water that could have been infiltrated in to the sub-surface.

Figure 5.1a Daily Max. Temperature of Maichew station

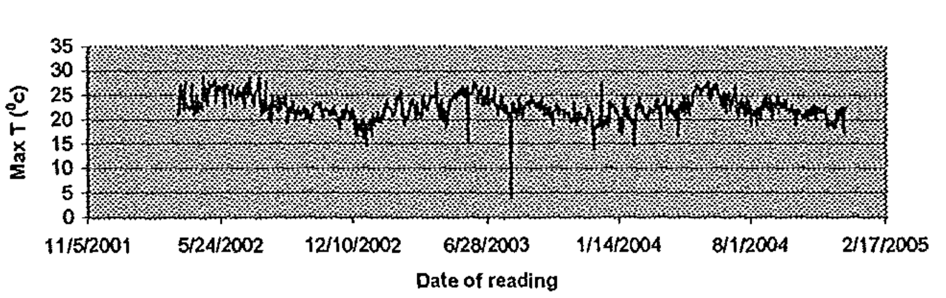


Figure 5.1b Groundwater level fluctuation measured from monitoring well (MW) in Kara in the study area (at Kara Adishawo)

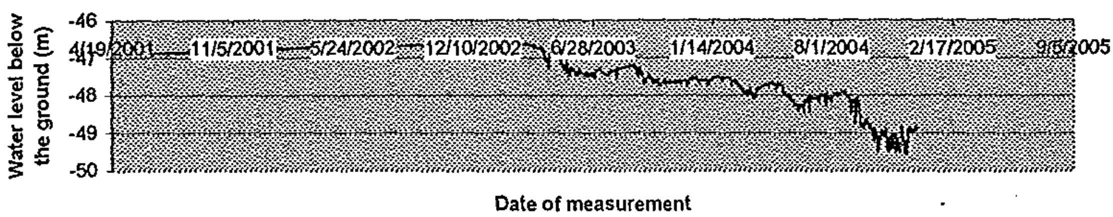
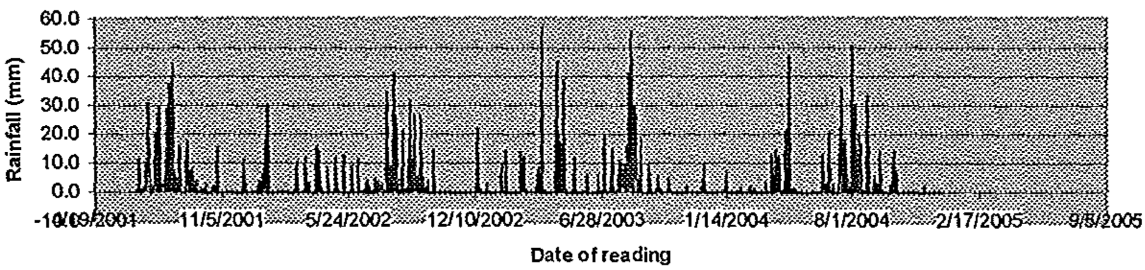


Figure 5.1.c Daily rainfall reading of Maichew station



5.3. Groundwater flow systems

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 pumptest 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 BH24 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. Elevation data of lake level, 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 Arcview GIS environment and contour map was prepared (Figure 5.2) 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 5.2). 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. Such trend was also observed from the spatial distribution of different hydrochemical parameters in the study area (chapter 7). In addition, the bed rock elevation map in Dessie Nadew(2003) also shows a depression in the central part of the valley floor.

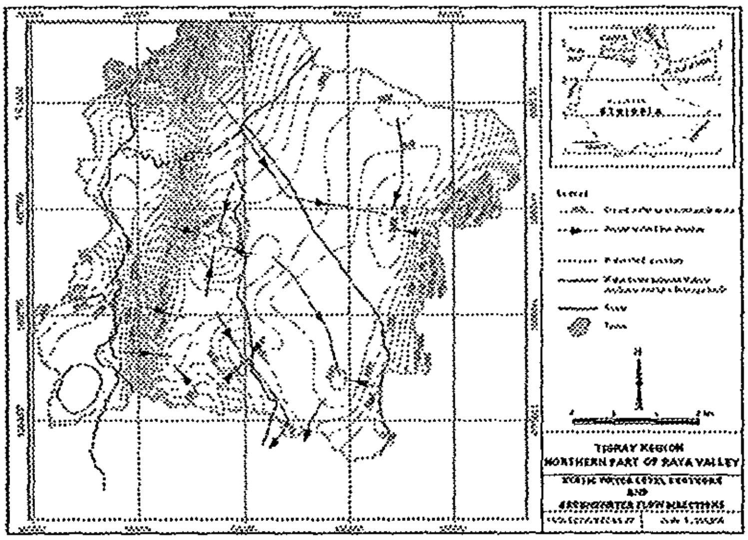


Figure 5.2. Groundwater level contour map and groundwater flow directions in the study area.

CHAPTER SIX

6. Pumping test and Aquifer characterisation

6.1. Introduction

The objective of unsteady state flow tests is to determine aquifer parameter estimations from time-drawdown field measurements in the main and/or observation wells by processing them through physically based and simplified mathematical models. These models help to infer aquifer, well or pump properties and finally to make the future predictions of discharge, drawdown, or distance between wells for the groundwater resources planning, design, operation and management.
Some of the information that can be found from properly completed aquifer tests are

- About the general aquifer parameters such as transmisivity, storativity, hydraulic conductivity (if aquifer thickness is known), leakage factor, drainage factor and hydraulic resistance;
- About the geometry of the aquifer including distance, direction and nature of impermeable barriers and recharge boundaries;
- About the lateral variations in the aquifer vicinity.

(Zekai Sen,1995)

In this work step drawdown test data of 23 wells and continuous pumping test data of about 24 wells were collected and analysed. Pre-existing transmisivity and specific capacity data of 3 wells were also used.

All the pumping tests undertaken in the study area were single well tests where drawdawn measurements were taken only from the pumping well. The duration of each step in the step tests range from 60 minutes to 120 minutes and the number of steps range from 3 steps in 17 wells to 4 steps in 5 wells. And the duration of the continuous tests range from 270 minutes to 4320 minutes. The depths of wells range from 80m to 164m and their screened sections range from 17.1m to 72m. Well diameters range from 8.5 inch to 17 inch and casing diameters ranges from 6 inch to 12.25 inch. With the exception of BH10 and BH11 all the other wells were considered as fully penetrating.

The aquifer type in the study area is mostly unconfined. But semi-confined conditions may be encountered in the central and eastern parts of the valley floor. This was observed from well lithologic logs. But for most pumping test analysis exercises in this work unconfined porous media aquifer condition was assumed. During analysis many physical parameters of the well and the hydrogeological conditions of the aquifer around the well were taken in to consideration.

6.1.1. Methods employed for the analysis

The collected pump test data was analysed by using different methods such as Cooper- Jacob time-drawdown method, Theis recovery model, Theis method, Neuman method, Cooper-Jacob step test method, Theis step test method, specific capacity estimation method, Hantush-Bierschenk well loss estimation method, and Moench fracture flow method. With the exeption of the Neuman method others were developed for confined groundwater flow conditions. But the flow in unconfined aquifers can be corrected by the Jacob correction for confined conditions. His formula was as follows:

$$s_{cor} = s - \left(\frac{s^2}{2D}\right)$$

where S_{cor} : the corrected drawdown

S : measured drawdown

D : original saturated aquifer thickness

(Zekai Sen, 1995)

This formula accounts for the change of aquifer thickness with drawdown in the well that penetrates into unconfined aquifer. Therefore, the models that were developed for the confined aquifer conditions can also be used with reasonable accuracy for the unconfined case by correcting the drawdowns with the above formula.

The following assumptions were also considered during the analysis:

- aquifers are homogeneous, isotropic and of uniform thickness over the area influenced by pumping,
- hydraulic gradient prior to the start of pumping was negligible,
- aquifers have "apparent" infinite extent,
- well diameters are so small that well storage is negligible.

Based on the type of data and type of aquifers in the study area, the aquifer parameter estimates of Cooper-Jacob time-drawdown and Theis recovery method were considered in this study. The results of the Theis recovery method are taken as final estimates. Theis recovery method is applicable for single well recovery tests conducted in confined, leaky or unconfined aquifers (Kruseman and Deroder, 1990).

For well characterization, the values of well specific capacity and well loss coefficients were used. In general, the software program named Aquifer Test Version 3.5 was used to analyze the raw pump test data collected in this work. This software is developed by Waterloo Hydrogeologic, Inc.

6.2. Aquifer parameters

6.2.1. Wells drilled into Q_{al} and Q_{col} aquifers in the valley floor

There are 12 wells (BH1, BH2, BH4, BH5, BH6, BH20, BH35, BH37, BH38, BH40, BH43 and BH66) in the study area that tap groundwater from the alluvial and colluvial aquifers. These aquifers are mainly found in the western portion of the valley floor. Total depths of these wells range between 84m in BH5 and 151m in BH1 where all the wells except BH4 and BH40 were drilled up to the bedrock. Depth to static water level ranges between 19.06m in BH2 and 77.48m in BH4 and their yields range between 3 l/s in BH43 and 44l/s in BH1. From the Theis recovery analysis the transmisivity values found from 7 of the wells were greater than 900 m^2/day that shows highly potential aquifers and in 5 wells indicate moderately potential aquifers with transmisivities ranging between 216 m^2/day and 475 m^2/day (Zekai Sen, 1995).

According to Zekai Sen (1995) aquifer potentiality is classified as in table 6.1.

Table 6.1. 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

The wide range of transmisivity values in the near by located wells like BH35, BH37, BH38, BH40, and BH66 (216 m^2/day in BH38 to 967 m^2/day in BH66) can be attributed to heterogeneity of the aquifer materials or different level of properness in the design and construction of the wells. The same is true for BH20 as compared to BH2, BH5 and BH6 (table 6.2).

6.2.2. Wells drilled into the Q_{smc} aquifers

There are 7 wells drilled into the aquifers of the interfluvial, fan foot plains and valley bottom Quaternary deposits (Q_{smc}). Out of these wells BH7, BH8, BH10, and BH11 have raw pumping test data that are analysed in this work. Only transmisivity data as analysed by Cooper-Jacob and Theis methods are found for the wells BH15, BH16 and BH27 from RVADP (1998) (table 6.2). The transmisivitie computed by Theis recovery analysis for BH15, BH16 and BH27 are 121 m^2/day, 1126.5 m^2/day and 8.52 m^2/day respectively.

Total depths of BH7, BH8, BH10 and BH11 are 164m, 147m, 120m and 100m respectively. Static water levels in these 4 wells range between 36.25m in BH10 and 97m in BH8 and their yields range between 2 l/s in BH8 and 7 l/s in BH11. BH7 is fully penetrating. BH8, which is near to BH9 is also considered as fully penetrating. Whereas, BH10 and BH11 are also assumed so, just to have an idea about the aquifer parameters. Having this in mind, the transmisivity values computed from the continuous pumping tests in the wells that tap groundwater from the Q_{smc} aquifers by the Theis recovery model range from 8.52 m^2/day in BH27 to 436 m^2/day in BH10 and exceptionally high value of 1155.7 m^2/day in BH16. In general, their transmissivity values found from BH7, BH10, BH11 and BH15 indicate moderately potential aquifers while BH27 indicates low aquifer potentiality (tables 6.1 and 6.2).

6.2.3. Wells drilled into the volcanic aquifers

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, BH42, BH46, BH72, BH84, BH86, BH87, BH88, BH89, BH90, BH91, Menora and Burka. BH25 was the only well with full pump test data among those found in the eastern basaltic areas. Menora and Burka are located in the north-western escarpment areas. And the remaining 10 wells are located on the western highlands mainly around Maichew town. BH42 is located near Korem town. Their total depth ranges between 43.25m in BH91 and 100m in BH25. Their static water levels also range between 1.7m in Burka and 48.55m in BH86 and their yields range between 0.7 l/s in BH84 and 5.62 l/s in BH42.

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 in BH90 and 81.25 m^2/day in BH91. The two extreme values were found from wells close to each other which may indicate a short distance variablity 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 Zekai Sen's classification (table 6.1) the transmisivities found from BH25, BH89 and BH91 indicated moderately potential aquifers; BH42, BH72 and BH87 indicated low potential aquifers and BH84, BH86, BH88, BH90, Menora and Burka indicated weak potential aquifers (tables 6.1 and 6.2).

6.2.4. Wells drilled into Q_{al} and Q_{col} aquifers in the western highlands

These aquifers are treated separately from similar aquifers in the valley floor due to their presence in a different geomorphological and climatic environment and because they are not as extensive and heterogeneous as that of the valley floor. Such aquifers are found in the lake Ashange basin and around Maichew town. The wells that penetrate into these aquifers are BH70, BH 71, BH 73, BH 74, BH 75, BH 76, BH 77, BH 78, BH 79, BH 80, BH 81, BH 83, BH92 and BH93. The first 12 wells are located in the lake Ashange basin and the remaining 2 are located south of Maichew town. Total depths of these wells range from 24.8m in BH81 to 86.15m in BH78 and their litologic logs indicated that the_Q_{al} and Q_{col} aquifers in the western high lands are mainly composed of sandy gravel or gravely sand sediments with thin clay interbeds. The static water level in these wells range between 4.7m in BH73 and 53.2m in BH83 and their yields range between 1.5 l/s in BH93 and 3.45 l/s in BH79. All the wells have pre-existing data of transmisivity values as computed by the Cooper-Jacob time-drawdown model (table 6.2). Accordingly, transmisivities of the Q_{al} and Q_{col} aquifers in the western highlands range between 6.5 m^2/day in BH72 and 606 m^2/day in BH79. That of BH74 and BH79 indicate highly potential aquifers; BH71, BH77, BH81, and BH83 indicate moderately potential aquifers and BH70, BH 73, BH 75, BH 76, BH 78, BH 92 and BH 93 indicate low potential aquifers. The variation in transmisivity of the Q_{al} and Q_{col} aquifers in the western highlands can be attributed to the variable proportions of the sand, gravel and clay in the aquifers at different places even within the lake Ashange basin.

Table 6.2. List of transmisivity values found from pumping tests in different parts of the study area.

Well No	SWL (m)	DWL (m)	Constant discharge pumping rate (l/s)	Duration of constant pumping (minutes)	Transmissivity values computed by		Remark
					Cooper-Jacob straight line model (m^2/day)	Theis recovery model (m^2/day)	
BH1	28.9	34	44	2880	3330	4200	Computed
BH2	19.06	30.02	35	3120	1220	1210	Computed
BH4	77.48	80.42	5.7	1920	1070	1240	Computed
BH5	36.7	37.8	7.1	2880	1250	-	Computed
BH6	29	29.76	7.1	1920	1090	965	Computed
BH7	64.2	66.71	6.8	1440	366	419	Computed
BH8	97	104.65	2	2700	33.2	-	Computed
BH10	36.25	38.91	6.2	1920	568	436	Computed
BH11	38.92	44.14	7	1920	120	98.9	Computed
BH19	36.34	39.34	22	600	1150	1110	Computed
BH20	21.73	33.13	30	3120	391	344	Computed
BH25	43.23	57.63	2.6	1920	20.8	94.2	Computed
BH30	41.8	60.02	3.45	2880	28.3	42	Computed
BH35	35.07	38.35	22	420	1300	910	Computed
BH37	33.18	44.7	22	600	476	294	Computed
BH38	52.9	72.65	18	1680	389	212	Computed

BH39	40.35	77.35	0.91	270	3.61	-	Computed
BH40	40.48	41.26	3.6	1440	474	474	Computed
BH42	1.98	29.8	5.62	2700	10.8	15	Computed
BH43	58.42	60.83	3	2520	184	286	Computed
BH44	39.12	59.52	2	2700	5.51	14.1	Computed
BH66	46.74	47.95	20.3	4320	3270	967	Computed
Menora	39.8	63.3	1.5	2880	-	-	3.67m^2/day (Moench fracture flow model)
Burka	16.8	56.7	1.8	2880	-	-	1.7m^2/day (Moench fracture flow model)
BH70	23.15	32.62	1.82	360	> 9.5 (incomplete test)	-	ADCS
BH71	15.93	23.12	3.4	360	59.73	-	ADCS
BH72	41.04	52.44	2.64	360	6.52	-	ADCS
BH73	4.7	9	3.45	360	45.46	-	ADCS
BH74	13.22	14.07	3.28	360	545.9	-	ADCS
BH75	15.1	26.2	3.4	360	29.22	-	ADCS
BH76	22.2	32.75	3.24	360	25.49	-	ADCS
BH77	32.79	35.34	2.98	360	130.88	-	ADCS
BH78	46.5	-	2.46	360	10.51	-	ADCS
BH79	17.65	18.5	3.45	360	606.1	-	ADCS
BH80	20.22	-	10	-	High	-	ADCS
BH81	9.22	11.43	3.4	360	66.37	-	ADCS
BH82	34.9	-	1.1	-	-	-	ADCS
BH83	53.2	53.81	2.58	360	331.65	-	ADCS
BH84	37.29	60.24	0.7	120	4.1 upto 4 minutes and 0.96 after 4 minutes may be due to impermeable boundary.	-	ADCS
BH85	20.95	-	3.5	360	35.93	-	ADCS
BH86	48.55	74.97	1	40	0.86	-	ADCS
BH87	17.1	26.73	3	360	28.92	-	ADCS
BH88	21.2	72.22	1.5	360	2	-	ADCS
BH89	41	45.52	2.6	360	87.46	-	ADCS
BH90	24.28	51.36	0.78	50	0.66	-	ADCS
BH91	12.45	16.33	3.7	360	81.25	-	ADCS
BH92	12.13	18.2	1.8	360	19.49	-	ADCS
BH93	39.1	43.75	1.5	360	22.16	-	ADCS

ADCS : Adigrat Catholic Church

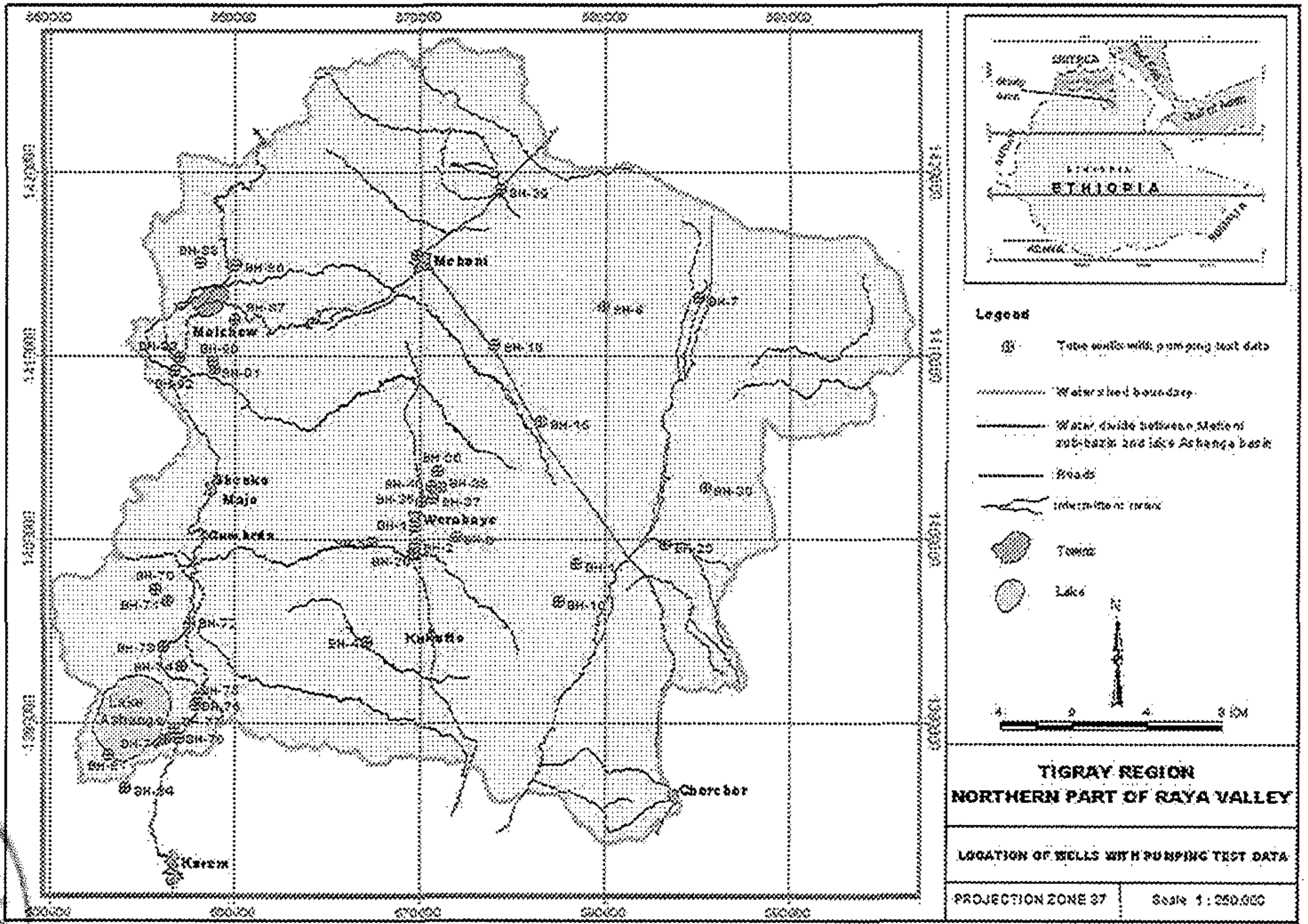


Figure 6.1. Location of wells with pumping test data.

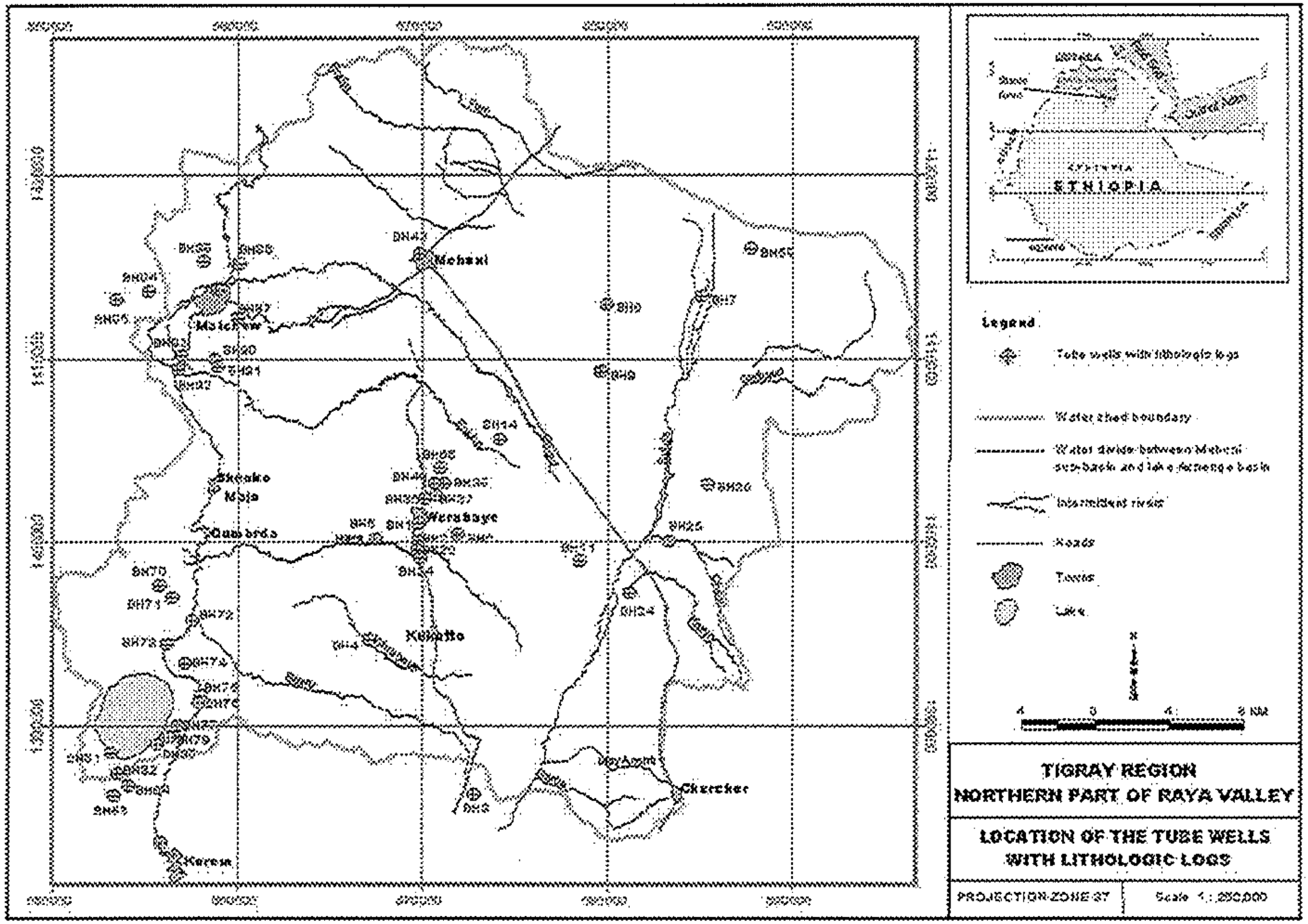


Figure 6.2. Location of wells with lithologic log data.

6.3. Well characterization

To determine the characteristic behavior of the relationship between specific drawdown and discharge, a set of pumping tests each with different discharge must be conducted in the field. This is usually done by continuous pumping consisting of discharge increments for specific time durations with no termination of pumping between consecutive steps. Therefore, the measured drawdowns during each step contain the residual drawdowns of the preceding step. It is recommended by Jacob(1946) that at least 3 epochs should be included in such step pumping tests (Zekai Sen, 1995).

In this work, step drawdown test data of 23 wells were collected and used to determine the linear and non-linear well loss coefficients and specific capacities of those wells. The durations for each step in the step tests range between 60 minutes and 120 minutes. In 17 wells the step tests comprise 3 steps while in 5 wells they comprise 4 steps. The rates of discharge and their corresponding drawdowns of the step tests and the computed well loss coefficients and specific capacities are presented in table 6.3.

6.3.1. Well loss coefficients

The Hantush-Bierschenk method requires fitting a straight line to the specific drawdown (s_w/Q) versus discharge scatter plot on arithmetic graph. The intercept of this straight line on the s_w/Q axis (ordinate) is equal to the linear well loss (aquifer loss) coefficient "B" and the slope equals to the non-linear well loss coefficient "C" (Zekai Sen, 1995).

The relationship of the drawdown in the pumping well and the discharge from it can be expressed as follows:

$$s_w = BQ + CQ^2$$

where s_w : drawdown in the pumping well

B : linear well loss (aquifer loss) coefficient

C : non-linear well loss coefficient

Q : discharge

The well loss coefficient is used to estimate the drawdown that is only due to the effect of pumping. The non-linear well loss that causes additional drawdown in the well than in the adjacent aquifer is due to turbulence flow through the well screens and into the pump intake. Therefore, the measured drawdown in the pumping well can be corrected for the well losses by the following equation:

$$s_{wc} = s_w - CQ^2 = BQ^2$$

where s_{wc} : corrected drawdown

Table 6.3. List of computed well loss coefficients and specific capacities of wells in the study area

Well No	Step	Discharge rate (l/s)	Duration of pumping (minutes)	Draw-down (m)	Aquifer loss coeff. (B) (s/m^2)	Well loss coeff. (C) (s^2/m^5)	Specific capacity for each step [(l/s)/m]	Specific capacity [(l/s)/m]
BH1	1	15	90	1.38	79	892	10.87	9.81
	2	20	90	2			10	
	3	25	90	2.7			9.26	
	4	30	90	3.9			9.12	
BH2	1	25	90	9.06	381	1500	2.76	2.92
	2	30	90	10.44			2.87	
	3	35	90	11.78			2.97	
	4	44	90	14.31			3.08	
BH4	1	3	120	1.52	449	5530	1.97	2
	2	4.1	120	2.05			2	
	3	5.6	120	2.74			2.04	
BH5	1	2.5	120	0.47	211	12500	5.32	6.66
	2	5	120	0.72			6.94	
	3	7.1	120	0.92			7.72	
BH6	1	3	120	0.35	139	8300	8.57	10.27
	2	5	120	0.48			10.42	
	3	7.1	120	0.6			11.87	
BH7	1	2	120	0.54	241	9600	3.7	3.37
	2	4.5	120	1.39			3.24	
	3	6.8	120	2.15			3.16	
BH8	1	0.79	60	3.45	5270	1210000	0.23	0.27
	2	1	60	4.2			0.24	
	3	2	60	6.16			0.33	
BH10	1	3	120	1.27	333	13400	2.36	2.49
	2	5	120	1.89			2.65	
	3	6.2	120	2.53			2.45	
BH11	1	4	120	3	324	54500	1.33	1.47
	2	6.5	120	3.96			1.64	

	3	7.1	120	4.92			1.44	
BH15	1	1.05		1.73	-	186624	0.61	0.8
	2	1.9		2.03			0.94	
	3	3.6		4.24			0.85	
BH16	1	1.88		0.22	-	3732.48	8.55	6.4
	2	2.88		0.51			5.65	
	3	4.44		0.75			5.92	
	4	5.2		0.95			5.47	
BH20	1	20	90	6.97	284	1800	2.87	2.92
	2	25	90	8.26			3.03	
	3	30	90	10.4			2.89	
	4	35	90	12.15			2.88	
BH25	1	1	120	4.41	2200	1270000	0.23	0.2
	2	2.5	120	10.57			0.24	
	3	3.3	120	23.12			0.14	
BH27	1	2		8.4	-	1866240	0.24	0.17
	2	3		25.12			0.12	
	3	3.6		26.73			0.14	
BH30	1	1.5	120	7.6	4220	412000	0.2	0.18
	2	2.2	120	13.1			0.17	
	3	3.5	120	20.1			0.17	
BH37	1	15	60	7.19	289	11500	2.09	1.95
	2	18	60	10.12			1.78	
	3	22	60	11.17			1.97	
BH38	1	14	120	13.11	749	9630	1.07	0.98
	2	15	120	18.71			0.8	
	3	20.5	120	19.23			1.07	
BH40	1	1	120	0.17	132	10600	5.88	6.07
	2	2	120	0.3			6.67	
	3	3	120	0.53			5.66	
BH42	1	3	60	2.29	-	439000	1.31	0.88
	2	4	60	5.62			0.71	
	3	5.62	60	9.05			0.62	
BH43	1	1	120	0.3	143	113000	3.33	2.72
	2	2	120	0.68			2.94	
	3	3	120	1.58			1.9	
BH44	1	1	60	4.2	3320	695000	0.24	0.21
	2	2	60	9.05			0.22	
	3	3	60	18.47			0.16	
Menora	1	1	60	5.35	2220	2710000	0.19	0.15
	2	1.3	60	8.15			0.16	
	3	1.7	60	11.87			0.14	
	4	2.5	60	25.4			0.10	
Burka	1	0.7	60	1.98	-	7480000	0.35	0.19
	2	1.2	60	6.59			0.18	
	3	1.8	60	11.82			0.15	
	4	2.5	60	45			0.06	

Zekai Sen (1995) has classified wells based on their non-linear well loss coefficients (table 6.4). Accordingly, among the wells drilled into the Q_{al} and Q_{col} aquifers BH1 is properly designed and developed; BH2 and BH20 are mildly deteriorated or clogged; BH4, BH6, BH37 BH38 and BH40 are severely deteriorated or clogged; BH5, BH43 and BH44 are difficult to restore to their original capacity. Among the wells drilled into the Q_{smc} aquifers BH7 and BH16 are severely deteriorated or clogged; BH8, BH10, BH11, BH16, BH27 and BH30 are difficult to restore to their original capacity. All the wells in the volcanic aquifers for which the well loss coefficients are determined (i.e. BH25, BH42, Menora and Burka) have very high well loss coefficients may be due to low aquifer transmissivities.

Table 6.4. Classification of wells based on their well loss coefficients "C"(Zekai Sen, 1995)

"C" in s^2/m^5	Well condition
C < 1490	Properly designed and developed
1490 < C < 2980	Mild deterioration or clogging
2980 < C < 11920	Severe deterioration or clogging
C > 11920	Difficult and sometimes impossible to restore the original capacity

6.3.2 specific capacity

The specific capacity of a well is defined as the ratio of discharge "Q" to the drawdown "s_w" in the pumping well (Q/s_w). It gives some information about the behaviors of the well and its productivity. Wells are said to be productive if small drawdowns are coupled with high discharges. The larger the specific capacity of a well the better is the well.

The specific capacity of a well is not constant because it is dependent on time. A high specific capacity indicates an efficient well design and construction. A decline in its value may indicate failure of well screen due to clogging or decline in the storativity and transmisivity of the aquifer. Comparison of two or more wells is possible provided that the same time duration is considered and the well diameters are equal. Better estimations of specific capacity can be achieved if the steady state or quisi-steady state drawdowns are considered. Wells can also be classified based on their average specific capacities (table 6.5).

Table 6.5. Classification of wells based on their specific capacities (Zekai Sen, 1995).

Specific capacity (Q/s_w) in [(l/s)/m]	Well productivity
$Q/s_w > 5$	Highly productive
$0.5 < Q/s_w < 5$	Moderately productive
$0.05 < Q/s_w < 0.5$	Low productive
$0.005 < Q/s_w < 0.05$	Very low productive
$Q/s_w < 0.005$	Negligibly productive

According to table 6.3 and table 6.5 among the wells drilled into the Q_{al} and Q_{col} aquifers BH1, BH5, BH16 and BH40 are highly productive wells; BH2, BH4, BH20, BH37, BH38 and BH43are moderately productive and BH44 is low productive. Among the wells drilled into the Q_{smc} aquifers BH6 and BH16 are highly productive; BH7, BH10, BH11 and BH27 are moderately productive and BH8 and BH15 are low productive. And among the wells that tap groundwater from the basaltic aquifers only BH42 is moderately productive while BH25, Minora and Burka are low productive.

In general, the relatively lower productivities and poor efficiencies of the wells in the Q_{smc} aquifers can be attributed to presence of considerable amount of fine sediments like fine sand, silt and clay in these aquifers. The flow of these fine sediments towards the well with the moving groundwater can result to reduction of permeability of the aquifer materials adjacent to the well and they also clog and deteriorate the well screens and the gravel pack. In addition, poor well completions and low transmisivities of aquifers may result to such low well efficiencies and productivities. But most of the wells in the Q_{al} and Q_{col} aquifers even with similar well completions are moderately to highly productive and they exhibit fairly good efficiencies. This can be attributed to the fewer occurrences of fine sediments in these aquifers and their higher transmissivities.

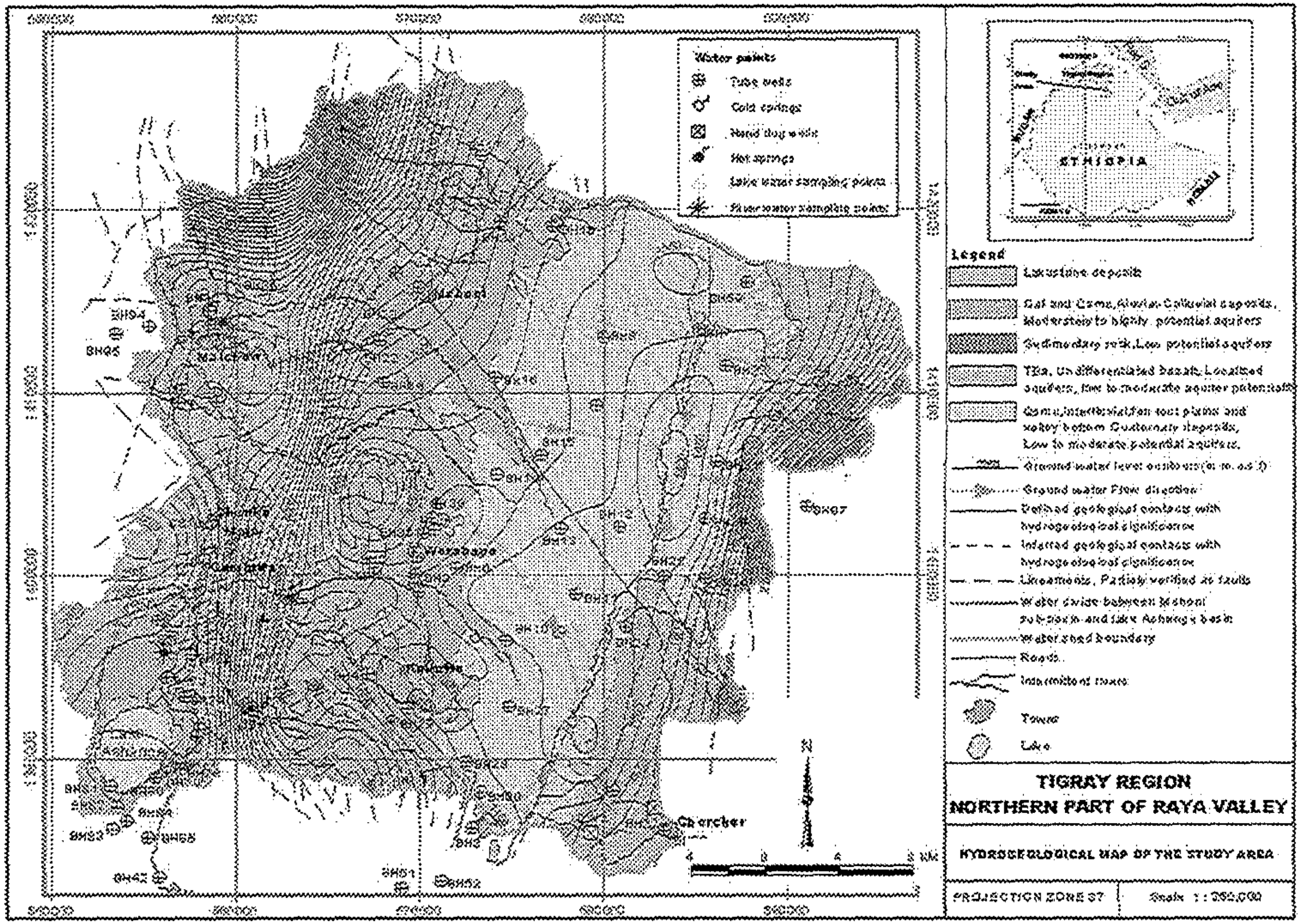


Figure 6.3. Hydrogeological map of the study area

CHAPTER SEVEN

7. Hydrochemistry and Water Quality

7.1. Introduction

The relationship between the pattern of flow and hydrochemistry is valuable both in the study of the genesis of hydrological and hydrogeological phenomena and in recognizing the distribution and movement of sub-surface waters. The concentrations, relative proportions and rates of transport of dissolved substances in natural waters reflect their sources, paths and interactions with different sub-surface materials (Tenalem Ayenew, 1998).

In this study the hydrochemical data is used to have an idea of the ground water flow systems in the study area and to study the distribution of different hydrochmeical parameters from water quality point of view.

7.2. Water Sample Collection and analysis

First it was tried to collect as many exisiting data as possible from different sources. Accordingly, some sort of hydrochemical data of 12 well samples and one lake sample were acquired from RVADP (1998) and Dessie Nadew (2003). Then in order to fill some gaps in the previous data and to get additional data for the purpose of this thesis work, a total of 44 new water samples were collected and analysed. 31 samples were collected from tube wells, 2 samples from hand dug wells, 7 samples from springs, 3 samples from rivers and 1 sample from lake Ashange. One tube well was sampled twice in order to study weather there is any major seasonal variation in the hydrochmeisry of the ground water in the valley floor or not. The locations of the sampling points data are shown in figure 7.1.

All the water samples from wells were collected directly from the pumped water at the inlet to the reservoir. When the pump was not functional for many days, sample was taken after pumping for some time so that the stagnant water in the well would be discharged first. Spring and river water samples were collected from points of strong flow and no interference from humans or animals. Lake samples were collected from the top few centimetres of the surface

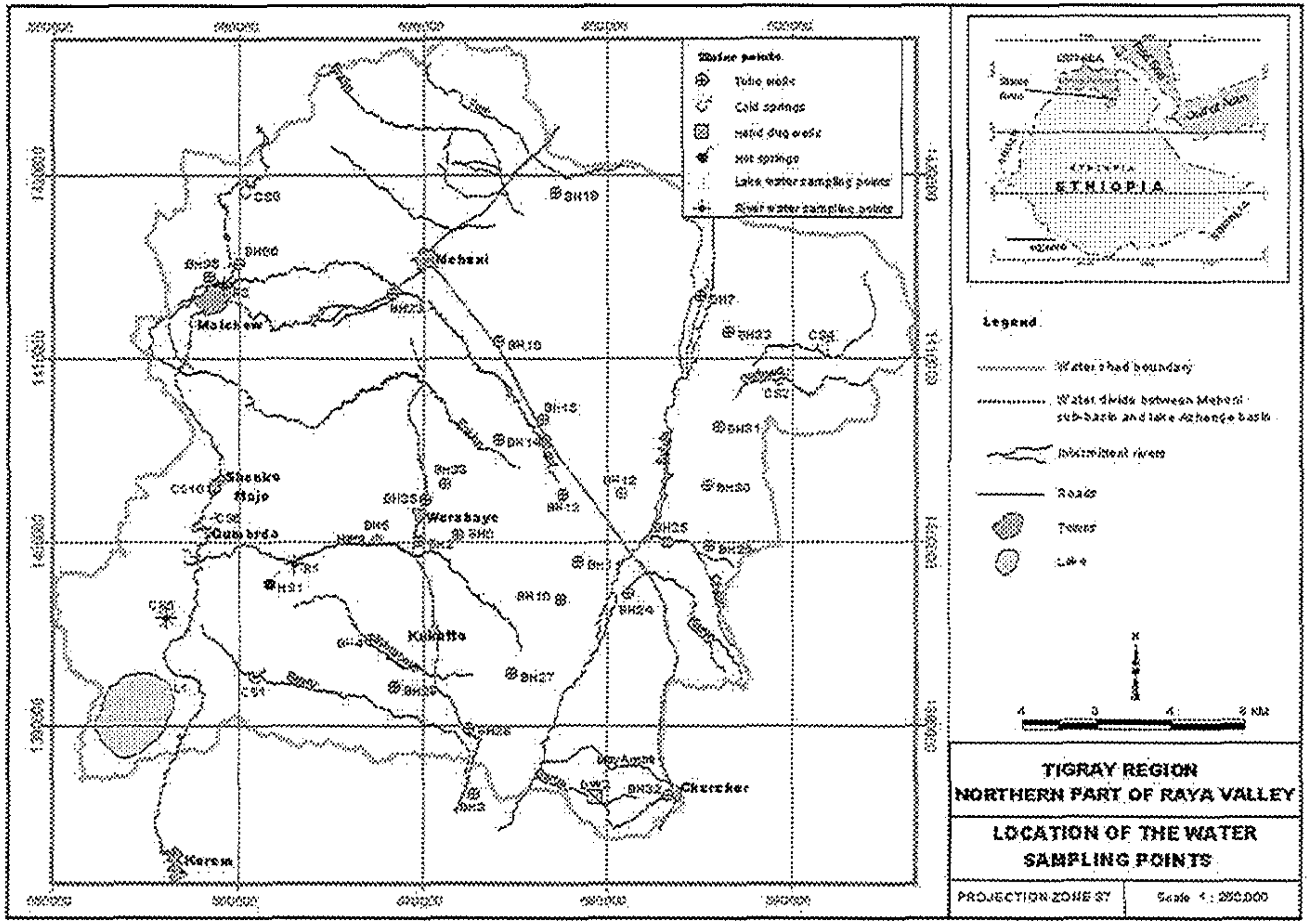


Figure 7.1. Location of water sampling points in the study area

and at a distance of about 10 meters from the shore. The rivers and the lake were sampled facing opposite the flow and wave movement directions respectively and away from samplers feet. In general, care was taken during sampling in order to obtain reliable hydrochemical data. And any other contamination was avoided during and after sampling.

The type of sampling undertaken at a particular place and time is called batch sampling and the samples that are collected in this study were such type. Clean inert plastic bottles of one litre volume with rubber stoppers were used to carry the water samples to the laboratory. The sampling bottles were repeatedly rinsed with the water to be sampled ant then completely filled with it leaving no space for air. The water samples were kept in a cool place until they are analysed and their laboratory analysis was under taken with in a maximum of two month after sampling.

Collection and analysis of 31 samples was undertaken in the period between 6 / 9 / 2004 and 24 / 11 / 2004 and that of 13 samples in the period between 22 / 2 / 2005 and 24 / 2 / 2005. The analysis of the first 31 samples was undertaken in the Ethiopian Geological survey (EGS) water laboratory and the remaining 13 samples were analysed in the Water Works Design and Supervision Enterprise (WWDSE) water laboratory. These laboratories are found in Addis Ababa, Ethiopia. All the samples were analysed for the major cations and anions, NO_3, F, total ionic species (EC) and PH. In addition SiO_2, CO_2, and Fe were tested for 33 samples that are locate in the valley floor. Total hardness ($CaCo_3$) and alkalinity ($CaCo_3$) of 11 water samples and iodine in 3 samples were also measured.

Certain properties of water such as temperature and pH are closely related to the environment at the sampling site and sampling instant. These properties can be altered after sampling due to the physical and chemical changes of water and storage conditions. Therefore the parameters temperature, pH, TDS, EC and turbidity of many samples were measured in situ.

Table 7.1. Methods applied for the chemical analysis of the water samples

Parameter	Analysis method
PH, HCO_3^-, CO_3^{2-} & CO_2	Electrometric titration method measured by PH meter
F^-	Ion selective electrode method
NO_3	Ultraviolet spectrometric method
Sio_2	Molybdosilicate method (yellow method) measured by spectrophotometery
Cl^-	Argentometri method (Titration)
SO_4	Turbidimetrtic method measured by spectrophotometric
Na^+, K^+, Ca^{2+}, Mg^{2+}, Fe	Atomic absorption method

Accuracy of the chemical analysis results was evaluated by using the electro neutrality and comparison of calculated conductivity with measured conductivity. Electro neutrality is a measure of the balance between cations and anions. It is calculated by the following formula:

$$\text{Reaction error} = \frac{(\sum \text{cations} - \sum \text{anions})}{(\sum \text{cations} + \sum \text{anions})} \times 100$$

where concentration of cations and anions is expressed in meq/l.

The second method is also defined by the following relationship (Appelo et.al, 1996):

$\sum$ anions $=$ $\sum$ cations $=$ EC/100

where concentrations of cations and anions are expressed in meq/l and EC in µs/cm.

As it can be seen from table 7.2, only 7 analysis results has reaction error values greater than 5% or less than -5% which may indicate that either there is some problem in the chemical analysis or one or more ionic species that have not been identified are present in significant amounts. But generally most of the analysis results have fairly good accuracy.

Table 7.2. **Field pH, field total dissolved solids(TDS), field electrical conductivity (EC), sum of cations, sum of anions, EC/100 and reaction error of recent water samples.**

S.no	Sample Id	Fieild PH	Field TDS (ppm)	Lab. EC (µs/cm)	Temp. (^{0}c)	Σcations (meq/l)	Σanions (meq/l)	EC/100 (µs/cm)	Reaction error (%)
1	BH2	-		549		5.86	6.16	5.49	-2.52
2	BH3	-	323	673	30.5	7.23	7.64	6.73	-2.8
3	BH4	-	355	750	25	8.01	8.39	7.5	-2.29
4	BH5	-	273	585	23	6.38	6.66	5.85	-2.17
5	BH6	7.6	273	572	22.5	6.23	6.6	5.72	-2.87
6	BH7	8.1	211	446	28	4.59	4.77	4.46	-1.88
7	BH10	-	230	479	30	5.15	5.31	4.79	-1.51
8	BH11	7.8	335	694	27	7.56	7.56	7.87	-2.0
9	BH12	7.8	242	503	28	5.57	5.51	5.03	0.53
10	BH13	7.5	802	1772	26	19.3	26.51	17.72	-15.72
11	BH14	7.8	268	570	27	6.2	6.51	5.7	-2.45
12	BH15	7.2	268	569	26	6.19	6.5	5.69	-2.46
13	BH16	7.8	296	641	24.5	6.67	7.24	6.41	-4.1
14	BH18	7.2	430	915	26.5	9.29	10.74	9.15	-7.22
15	BH23	7.4	322	668	23.5	7.04	8.0	6.68	-6.39
16	BH24	7.6	982	2702	30	26.67	31.12	27.02	-7.71
17	BH25	7.8	227	453	31	4.85	5.04	4.53	-1.9
18	BH26	8.1	194	405	30	4.23	4.3	4.05	-0.82
19	BH27	-	-	488	-	5.34	5.44	4.88	-0.95
20	BH28	7.3	319	653	28.5	7.23	7.42	6.53	-1.27
21	BH29	-	-	603	-	6.56	6.87	6.03	-2.3
22	BH30	8.2	195	444	26	4.37	4.81	4.44	-4.7
23	BH31	7.8	245	483	39	4.98	5.09	4.83	-1.03
24	BH32	7.4	644	1150	26	10.78	13.26	11.5	-10.32
25	BH33	7.8	229	486	26	5.25	5.45	4.86	-1.83
26	BH35	-	271	575	24	6.06	6.38	5.75	-2.58
27	BH38	7.3	348	714	28	7.64	8.25	7.14	-3.84
28	BH86	>8.2	159.2	253	22.9	2.96	3.08	2.53	-1.96
29	BH96	>8.2	152	239	25.8	3.1	3.35	2.39	-3.83
30	R1	-	207	444	29	4.9	5.16	4.44	-2.59
31	R2	8.2	268	443	27.3	5.95	5.76	4.43	1.61
32	R3	8.2	252	415	21.7	5.91	5.83	4.15	0.74
33	CS1	-	-	481	-	5.47	5.47	4.81	-5.5
34	CS2	-	-	768	-	8.88	8.83	7.68	-4.33
35	CS4	-	-	1102	-	13.29	13.29	11.02	-4.74
36	CS6	-	205	340	19.7	4.53	4.48	3.4	0.6
37	CS8	7.6	214	351	17.1	4.73	4.73	3.51	-0.6
38	CS10	7.8	228	377	21.6	4.93	4.94	3.77	-0.16
39	HS1	>8.2	140	263	35	2.52	2.88	2.63	-6.64
40	DW2	-		325		3.29	3.55	3.25	-3.76
41	L1	>8.2	1065	1776	21.3	27.62	30.04	17.76	-4.2

All recent and pre-existing chemical analysis results are included in appendix 2.

7.3. Evaluation of Hydrochemicial Parameters

7.3.1. Temperature, pH, TDS and EC

The groundwater temperatures as measured in situ range between 17.1^{0}C and 39^{0}C. It seems that groundwater temperature is slightly higher in the valley floor than in the highlands (table 7.2), which coincides with the variation of the air temperature with altitude (table 3.2). The temperatures observed in HS1 and BH31 were exceptionally high (i.e. 35^{0}C and 39^{0}C respectively). These high groundwater temperatures may indicate some sort of geothermal activities that can be associated with deep faults. Similar PH values were found from the field and laboratory measurements. But the field measurements were taken by titration method, which needs visual judgement of resulting colours. It also could not read pH values greater than 8.2. Therefore, the laboratory PH is taken for comparison between samples. Based on this, the PH of natural waters in the study area range between 7.2 and 9.64 showing alkaline nature. The total dissolved solids (TDS) and electrical conductivity (EC) were also measured in the field and in the laboratory respectively. The values of these two parameters were linearly correlated. Both TDS and EC of the ground water samples show an increasing trend with decrease in altitude because the highland groundwaters are more flashed than those in the valley floor and the residence time of the groundwater increases towards the lowland areas. The highest TDS and EC values were found in the lake Ashange, 1065 mg/l and 1776 µs/cm respectively. This can be attributed to the high rate of evaporation from the lake. When we compare the EC of the lake water with that of the groundwaters in it's surrounding that range between 215 $\mu s / cm$ and 715 $\mu s / cm$, it seems that the groundwater flows from all directions towards the lake. Similarly, there is an increase of EC and TDS values from all directions towards the centre of the Mehoni sub-basin (Figure 7.2). This may also indicate that the groundwater in the sub-basin flows from all directions toward the area where EC and TDS were high (i.e. EC & TDS increase with increase of residence time of the groundwater). But there are some localities with exceptionally low or high EC and TDS. This can be due to local flow systems or due to highly soluble rock intercalations (Figure 7.2).

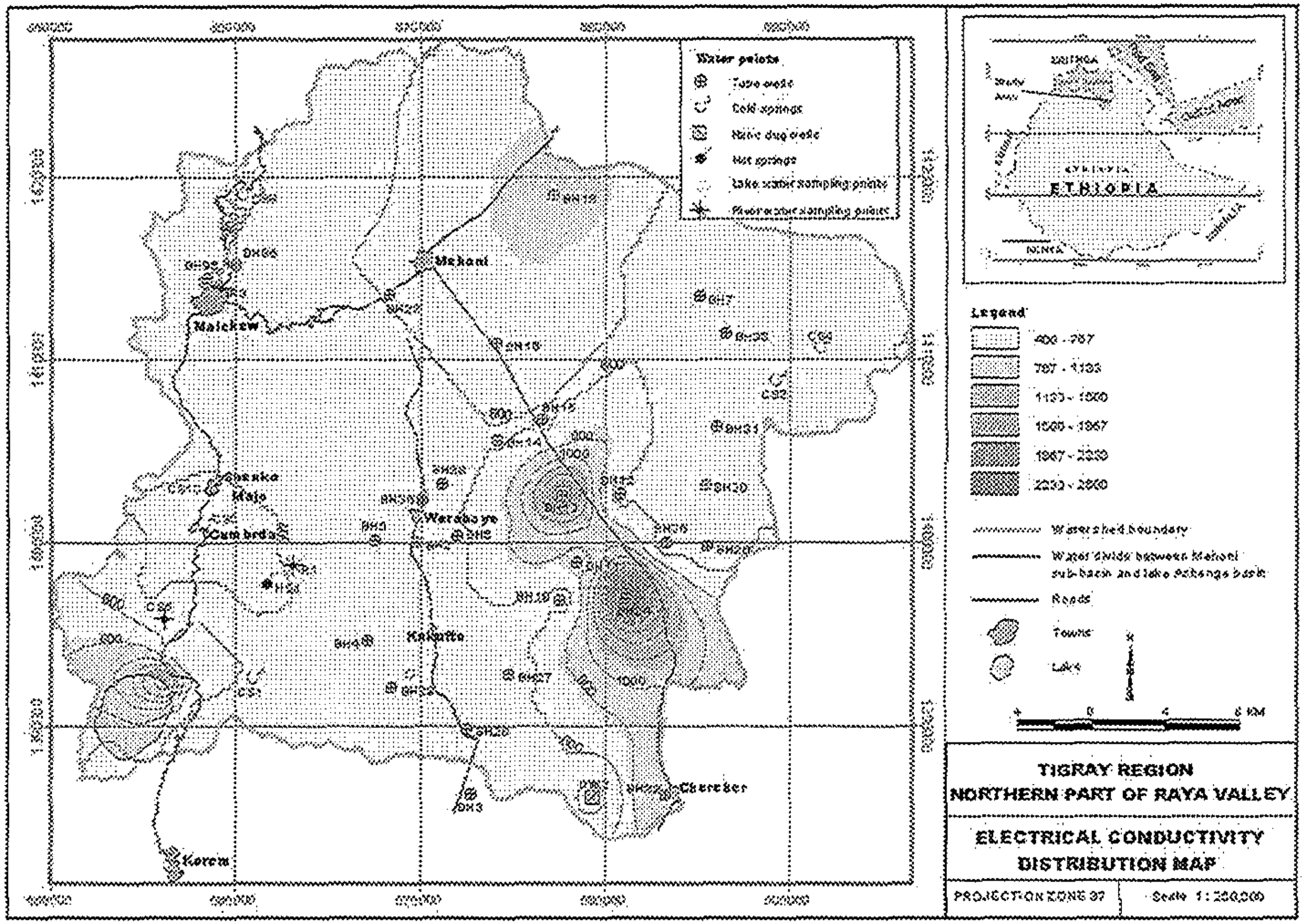


Figure 7.2. Distribution of EC in the study area.

7.3.2. Major ions

7.3.2.1. HCO_3^-, Ca^{2+} and Mg^{2+}

The dominant anion in the study area is HCO_3^-. Over most of the normal PH range of groundwater (i.e. 6 to 9), HCO_3^- is the dominant carbonate species (Freeze & Cherry, 1979). The main source of this anion in the study area is the weathering of silicate minerals that make up the basalts and detrital materials derived from the basalts. The dissolution of albite, which is the common feldspar in basaltic rocks, can be the main source of the HCO_3, Na and SiO_2. The possible reaction for this dissociation process is as follows:

$$\underbrace{NaAlSi_3O_8\,(s)}_{\text{Albite}} + H_2CO_3 + (9/2)H_2O \leftrightarrow$$

$$Na^+ + HCO^-_3 + 2H_4SiO_4 + \underbrace{(½)Al_2Si_2O_5(OH)_4(S)}_{\text{Kaolinite}}$$

The dissolution of albite increases with increasing partial pressure of CO_2. In BH86, HS1 and L1 which have PH values greater than 9, the CO_3^{-2} is also one of the major anions. Another source of HCO_3^- in the study area, especially in the northeastern part of the study area is the sedimentary succession comprising of mudstone, sandstone and limestone. In particular limestone can introduce considerable amount of HCO_3^- and Ca into the ground water from the dissolution of calcite by the following reaction:

$$CaCO_3 + H_2CO_3 \leftrightarrow Ca^{2+} + 2HCO_3^-$$

The highest concentration of HCO_3^- is found in the lake with a value of 885.5ppm, which show slight depletion as compared to previous data (917ppm) in Dessie Nadew (2003). In the well samples the highest concentration was found in BH13 (i.e. 534ppm) and the lowest in BH86 (i.e. 22.7ppm). But most of the samples from wells, springs and rivers have HCO_3^- concentrations that range between 203ppm and 350ppm.

The dominant Cation in the study area is calcium followed by sodium and magnesium respectively. This makes the natural waters of the study area to be hydrochemically different from that of the Main Ethiopian Rift, which are exclusively Na-HCO_3 type (Dessie Nadew, 2003). In the analysed water samples, Ca ranges between 1.1ppm in HS1 and 210ppm in BH24 with a mean concentration of 53.95ppm while Mg ranges between <0.1 in HS-1 and 150.1ppm in L1. In the study area calcium is mainly derived from weathering of calcic plagioclase (anorthite) and calcium rich pyroxene (diopside). And magnesium is derived from weathering of magnesium rich olivine and pyroxene, like diopside and hypersthene. The chemical analysis of rocks from the study area and its surroundings indicates that these minerals are abundantly present in the lithologic formations of the area (R. Pik et.al, 1998 in Dessie Nadew; 2003). Additional source for the calcium, particularly in the northeastern part of the area, can be from weathering of calcite in the limestone and sandstone.

7.3.2.2. Na^+ and K^+

As described in the previous section sodium is the second abundant cation in the study area. Its concentration in the analysed water samples ranges between 12ppm in DW2 to 125ppm in BH24. And there is exceptionally high value of 325ppm in the sample taken from lake Ashange. Its concentration increases towards the center and central east of the valley floor. This can be an important indication of the ground water flow direction because the calcium ions may continually be replaced by the sodium as the groundwater percolates through the unconsolidated sediments in the valley floor while calcium precipitate in the sediments due to saturation of the groundwater by calcite (Ca^{2+}). As described previously the main source of sodium in the study area is the sodium plagioclase (albite), which is found with in the mineral assembledge of the basaltic rocks (Dessie Nadew; 2003).

Potassium is the least abundant of the major cations in the study area. Its concentration in the analysed water samples ranges between <0.01ppm in two samples (CS1 and HS1) and 6.4ppm in CS4. Exceptionally high K^+ concentration of 23ppm is detected in L1. Only in wells its concentration ranges between 0.4ppm in two samples (BH5 and BH35) and 2.6ppm in BH26. In both the springs and the wells the concentration of K^+ tends to increase from west towards the east. Generally, the low concentrations of K^+ in the natural waters of the study area could be attributed to its less abundance in the basaltic rocks and its low solubility (Dessie Nadew 2003). Potassium feldspar, which is a minor constituent in the basaltic rocks, can be the main source of K^+ in the natural waters of the study area.

Here it should be noted that the large contrast in the concentration of the major anions and cations between the lake Ashange and the groundwaters in the Mehoni sub-basin could be one important indicator for the absence of large amount of leakage from the lake towards the sub-basin.

Well water samples in the lake Ashange catchment have higher Ca^{2+}concentrations than in the lake samples. Conversely, in the lake water samples Na^{+} over dominates Ca^{2+}. The Ca^{2+} ions in the lake may vigorously react with the highly concentrated HCO_3^- and CO_3^+ ions in the lake to form $CaCO_3$, which later precipitated leaving higher concentrations of Na in the solution. On the other hand this contrasting concentrations of Na and Ca in the water samples taken from the lake and surrounding groundwater may indicate that they are two separate hydrologic systems.

7.3.2.3. Cl^- and SO_4

Chloride is the second abundant anion in the study area with concentrations that range between 5.96ppm in CS6 and 298ppm in BH24. Next to chloride the third abundant anion in the study area is sulphate. Its concentration ranges between 1ppm in two samples (BH33 and CS10) and 370ppm in BH24. In a similar way to the other major ions large concentrations of these two anions are detected in the tube wells of the central part of the valley floor and in the springs located in the eastern highlands and escarpments.

Generally, the most probable source for the chloride ions in the valley floor is the dry fall outs that are brought from the Afar area by wind. But other sources such as dissolutions of apatite in the basaltic rocks and liquid inclusions can also contribut to some extent (Devis et.al-1996 in Dessie N. 2003). And the sulphate ions may come from sources such as oxidation of pyrite, fertilizers (Di Ammonium Phosphate & Urea) and atmospheric fallouts.

7.3.2.4. NO_3

The nitrate concentration in the study is highly variable from place to place. It ranges between <0.04ppm in two samples (BH26 and HS1) and 398.7ppm in BH13. But most of the samples show concentrations between 3ppm and 20ppm. In the study area, high nitrate concentrations are mostly associated with highly populated rural areas and small towns.

Therefore, the main source of the nitrate can be human and animal excreta with some contribution from fertilizers (mainly urea). Like the other major ions maximum nitrate concentration is detected in the central and central east of the valley floor.

7.3.2.5. SiO_2

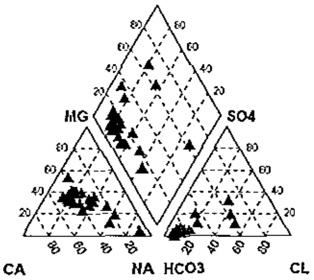
Silica is the most common non-charged constituent in natural waters. In the study area it is the second most abundant constituent next to HCO_3, especially in the groundwaters. Its concentration in the analysed water samples ranges between 26ppm in DW-2 and 76ppm in BH33. There are peculiarly high concentrations of silica in the north central and south central parts of the valley floor. Such high silica concentrations in these areas could be attributed to the weathering of the volcanic rocks that surround the valley floor, which are also sources for the unconsolidated sediments in the valley (Davis et. al, 1966 in Dessie N., 2003). Silica is derived from dissociation of the common silicate minerals in the basaltic terrain like pyroxenes and plagioclase feldspar (Dessie N., 2003).

7.4. Water Quality

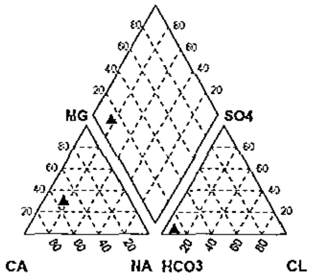
Water quality is closely linked to water use and to the state of economic development. The three main classes of use are domestic, agricultural and industrial. Scientists develop water quality criteria. They provide basic scientific information about the effects of water pollutants on a specific water use and also describe water quality requirements for protecting and maintaining an individual use. Water quality criteria are based on variables that characterize the quality of water and/or the quality of the suspended particulate matter, the bottom sediment and the biota. Many water quality criteria set a maximum level of the concentration of a substance in a particular medium (i.e., water, sediment or biota), which will not be harmful when the specific medium is used continuously for a specific purpose. In this work, the chemical quality of water will be studied giving particular emphasis to its use for irrigation.

7.4.1. Drinking water quality

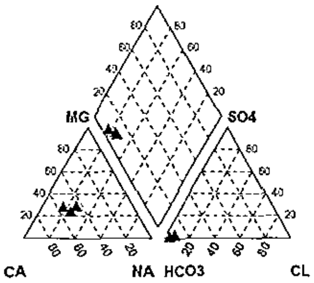
Water has a profound influence on human health. The quantity and quality of the water supplied are important in determining the health of individuals and whole community. The first priority must be to provide access for the whole population to some form of improved water supply. However, access may be restricted by low coverage, poor continuity, insufficient quantity, poor quality and excessive cost relative to the ability and willingness to pay. In the study area the sources of water for domestic use are the springs, hand dug wells, ponds, and tube wells. The first three sources are highly susceptible to microbiological contaminations because they are easily exposed to the contaminant sources and in most cases both the villagers and their animals use from one source with little or no safety. But the tube wells that tap groundwater from relatively deeper groundwater sources are less probable to be microbiologically contaminated. But it would be better if we have clear understanding of the hydrogeological systems so that care will be taken to protect sensitive parts from any contaminantion. Periodical chlorination will also improve microbiological quality of water source.



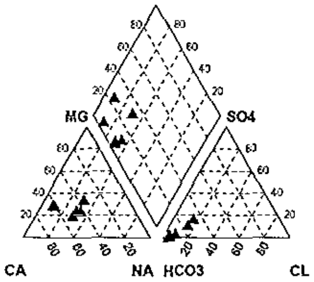
A. Piper plot of borehole samples



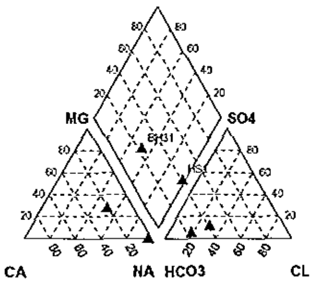
B. Piper plot of a hand dug well sample



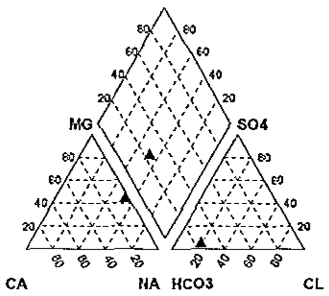
C. Piper plot of river samples



D. Piper plot of cold spring samples



E. Piper plot of hot spring and geothermal well samples



E. Piper plot of a lake samples

Figure 7.3. Piper plots of the natural waters in the study area grouped by source.

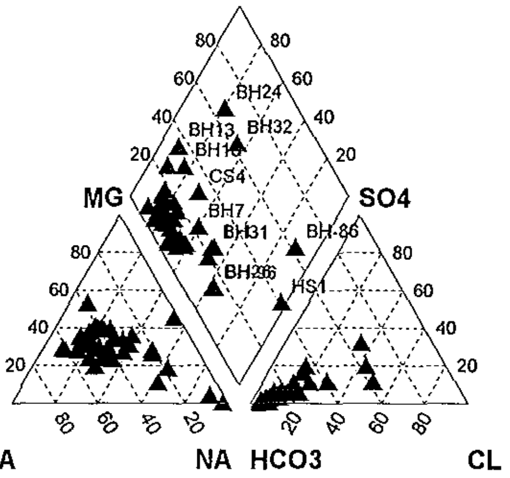


Figure 7.4. Piper plot of all the water samples collected from the study area.

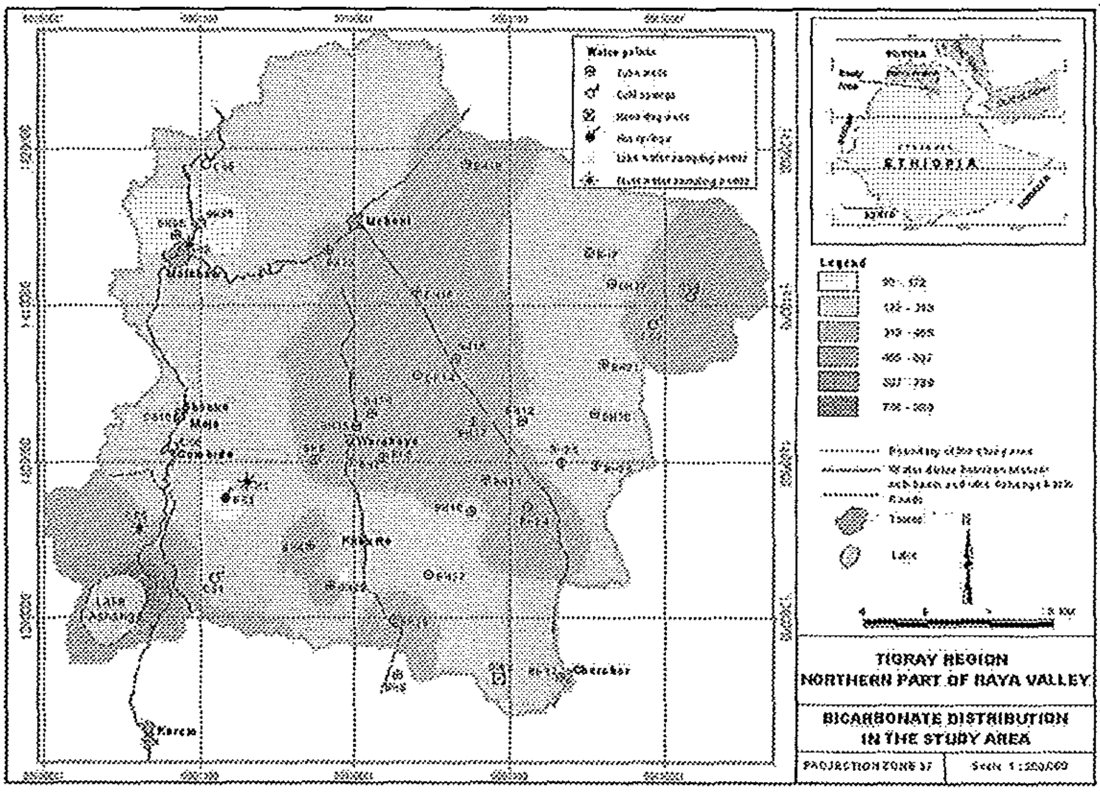


Figure 7.5. Spatial distribution of HCO_3^- in the study area.

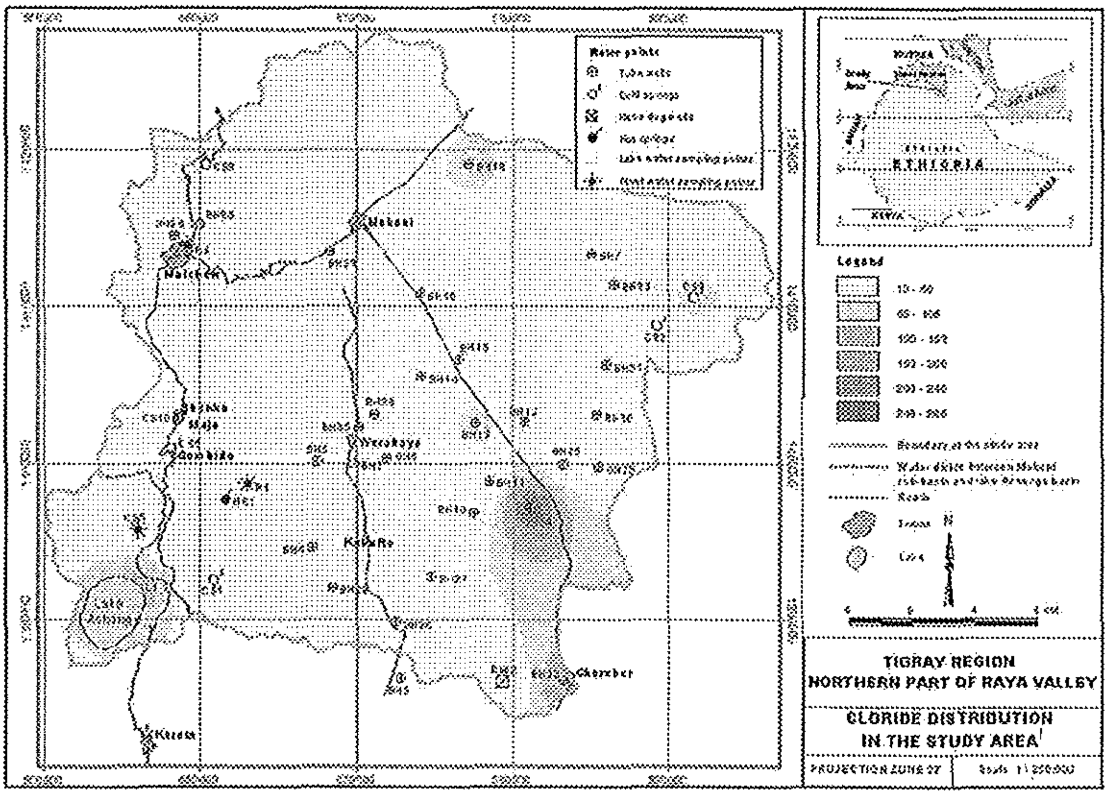


Figure 7.6. Spatial distribution of Cl^- in the study area.

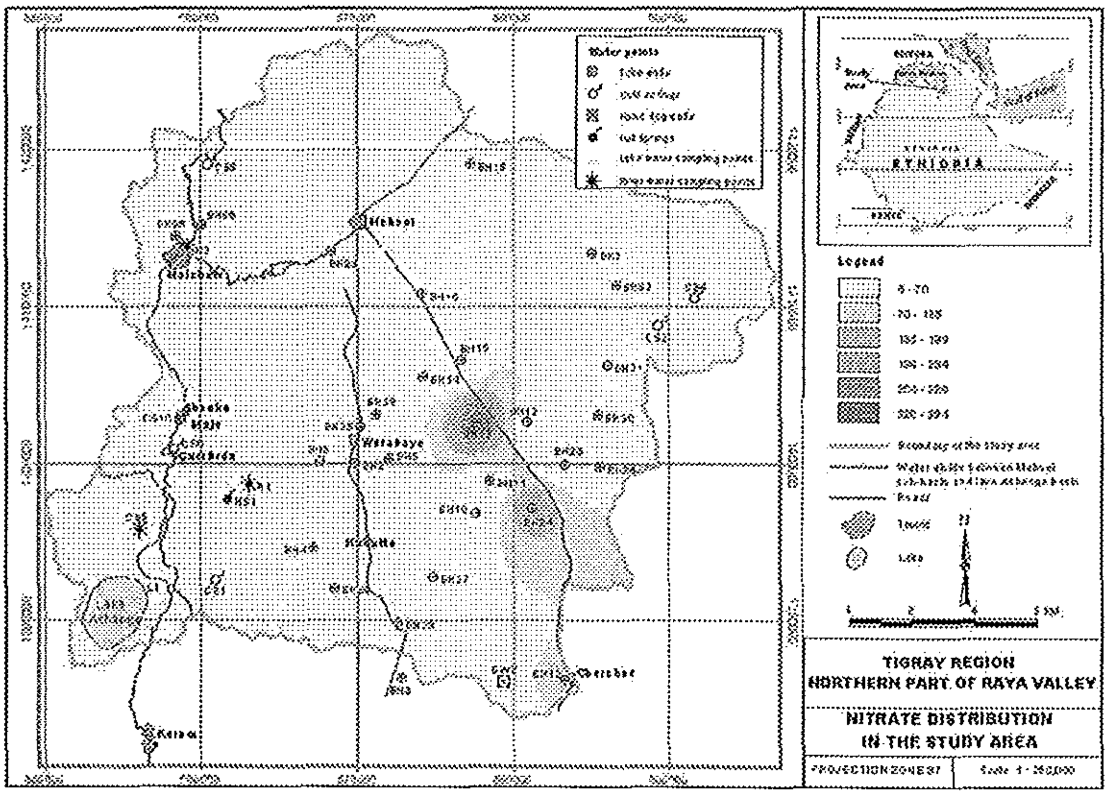


Figure 7.7. Spatial distribution of NO_3^- in the study area.

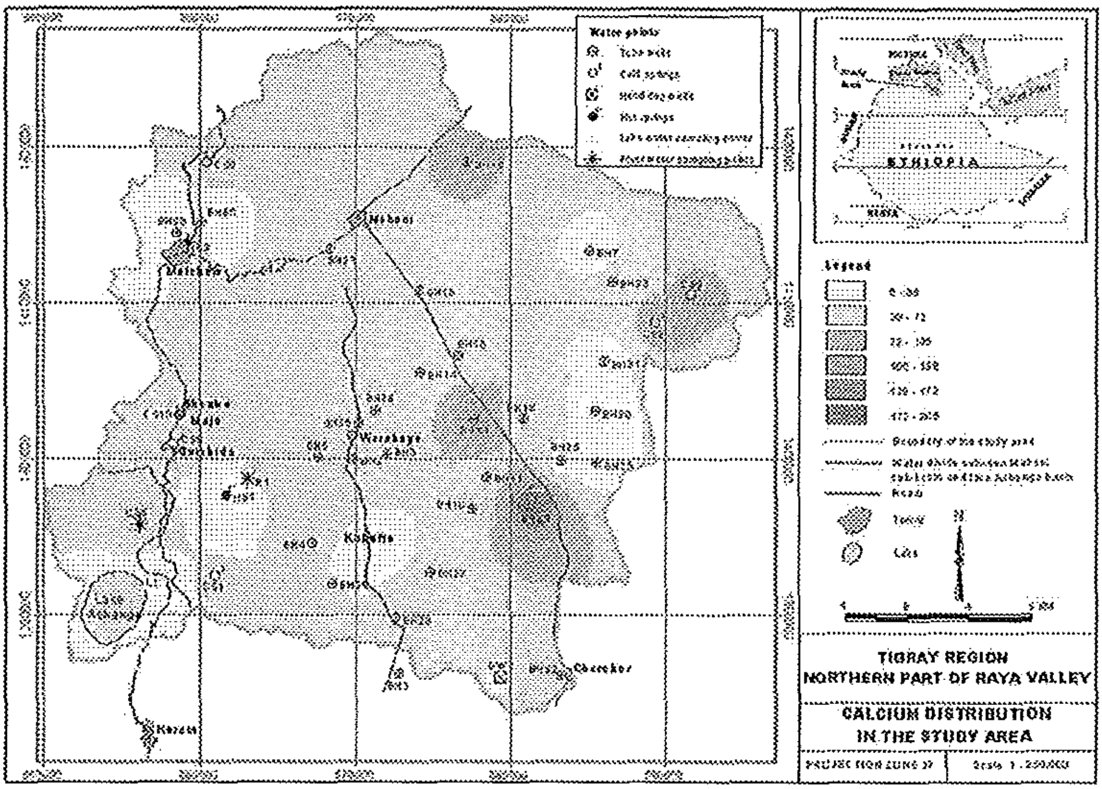


Figure 7.8. Spatial distribution of Ca^{2+} in the study area.

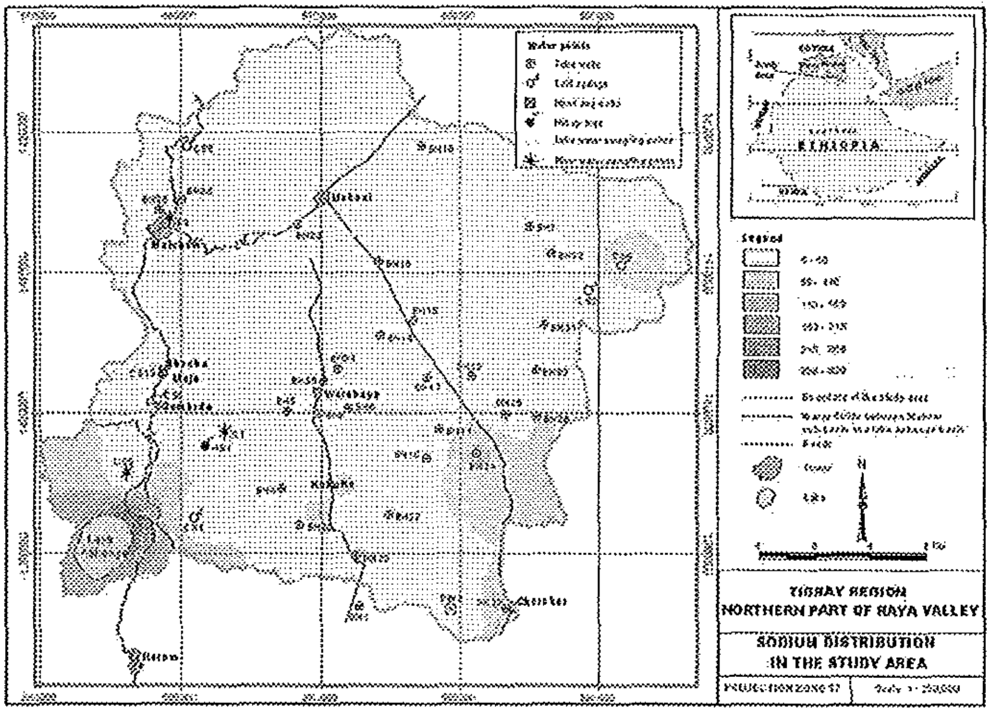


Figure 7.9. Spatial distribution of Na^{+} in the study area.

Chemical quality of the drinking water in the study area is treated in the following sections with respect to some measured parameters.

7.4.1.1. NO_3^-

Excess nitrate in drinking water has been linked to methaemaglobinaminia (blue-baby syndrome) in infants and the old. Here nitrate leads to the oxidation of normal haemoglobin to methaemoglobin which is unable to transport oxygen to the tissues. In addition, wide spread chronic effects are likely to be noted including a possible greater likelihood of gastric cancer. Nitrate is a conservative ion in natural groundwaters. Therefore, once large-scale nitrate contamination has occurred, it will take a considerable period of time before it is naturally attenuated through denitrification or dilution.

The Ethiopian standard for drinking water published in Negarit Gazeta No 12/1990 has declared the maximum permissible nitrate to be 45mg/l (RVADP, 1998). Accordingly the wells BH13, BH24 & BH32 have nitrate concentrations that are by far greater than the maximum permissible value (i.e. 398.7mg/l, 221.5mg/l and 110.75mg/l respectively) and that of BH18 (66mg/l) is slightly greater than the Ethiopian maximum permissible limit for nitrate. Therefore, the sub-surface waters from which these samples were collected seem to be exposed to some form of higher nitrate pollution. The most probable sources of nitrate in the study area are the human and animal excreta and nitrogen fertilizers such as urea ($CO(NH_3)_2$). But, detailed study may be required in order to know its exact sources and to minimize additional inputs to the groundwater.

7.4.1.3. Drinking water standards of other measured parameters

In table7.3, the concentrations of F^-, Cl^-, SO_4^{2-}, Ca^{2+} and Mg^{2+} in the water samples from the study area and their TDS and PH values are compared with the Ethiopian maximum permissible and highest desirable levels in drinking water. Most of the tested water samples full fill the maximum permissible levels with exception of BH24 and BH13, which contain excess levels in some of the parameters. But all the tested samples fulfill the Ethiopian highest desirable levels for the parameters under consideration.

Table 7.3. Comparison of some measured chemical parameters in natural waters of the study area with the Ethiopian standards for drinking water

Measured Parameters	Ethiopian standard for drinking water		Dominant range of values (ppm)	Exceptions
	Highest desirable level (ppm)	Maximum permissible level (ppm)		
F^-	1.0	1.5	0.2 – 0.96	-
Cl^-	200	600	5.96 – 149	298ppm in BH24
SO^{2-}_4	200	400	1 – 130	370ppm in BH-24
Ca^{2+}	75	200	1.1 – 71	145ppm in BH13, 83ppm in BH15, 210ppm in BH24,
Mg^{2+}	50	150	4.3 – 150.1	< 0.1 in HS1
TDS	500	1500	152 – 430	802ppm in BH13, 982ppm in BH24, 644ppm in BH32, 1065ppm in L1
pH	7.0 – 8.5	6.5 – 9.2	7.33 – 8.36	9.17ppm in L1, 9.63ppm in HS1, 8.76ppm in BH96, 9.64ppm in BH86,
Fe	0.1	1.0	< 0.1 – 0.1	-

7.4.2. Water Quality for Irrigation Purposes

Water quality, soil types and cropping practices all play a role in successful irrigation. Good quality of water permits maximum yields consistent with proper soil and water management. Study of soil types is used to determine the infiltration rates and leaching of mineral salts. Leaching is essential to the reduction of salinity in the topsoil. Equally important is to have an understanding about the tolerance of specific crops to certain elements.

7.4.2.1. Salinity and Sodium Adsorption Ratio

Excessive salinity occurs when there is an accumulation of salts in the topsoil. Salts in soils or water can reduce water availability to the crop to such an extent that yield is affected. Salts are added to soils with each irrigation. These salts will reduce crop yield if they accumulate in the rooting depth to damaging concentrations. The crop removes much of the applied water from the soil to meet its evapotranspiration demand but leaves most of the salt behind and become concentrated in the shrinking volume of soil water. A portion of the added salt must therefore be leached from the root zone before the concentration affects crop yield. Leaching is done by applying sufficient water so that some of it will percolate through and below the entire root zone carrying a portion of the accumulated slats. Salinity in water is usually measured by the specific electrical conductance (EC) or TDS of the water.

Sodium ion also has far-reaching effects on soils. When sodium-rich water is applied to soil, some of the sodium is taken up by clay and the clay gives up calcium and magnesium in exchange. This reaction is called base-exchange reaction and it is particulerly important montmorillonitic clays (smectite) because it causes large changes in permeability. Clays of montmorillonite group can expand and contract in response to changes in the composition of the adsorbed cation between the cay platelets. The hydrated radii of Na+ and Ca^{2+} are such that two hydrated ions of Na^{+} require more space than one Ca^{2+} ion. Hence, replacement of Ca^{2+} by Na^{+} on the exchange sites causes an increase in the dimension of the crystal lattice. This results in a decrease in permeability of the soils. Clays that take up sodium also become sticky and slick when wet or form hard clods when dry. These effects may result to degradation of soil agricultural productivity and difficult cultivativation.

The United States salinity laboratory proposed the following relation to calculate the sodium hazard:

$$\mathrm{SAR} = \frac{Na}{\sqrt{\frac{Ca + Mg}{2}}}$$

Where SAR is defined as sodium adsorption ratio and the concentration of the ions is expressed in milliequivalents per liters.

The United States Development of Agriculture has developed a widely accepted classification of irrigation waters based on the sodium adsorption ratio as an index for sodium hazard and specific electrical conductance (EC) as an index of salinity hazard (table7.4).

Table 7.4. Suitability of natural waters for irrigation based on their EC and SAR

Water class	**EC in $\mu s / cm$**	**Alkali hazard (SAR)**
Excellent	Less than 250	Up to 10
Good	250 – 750	10 – 18
Medium	750 – 2250	18 – 26
Bad	2250 – 4000	Grater than 26
Very bad	Greater than 4000	-

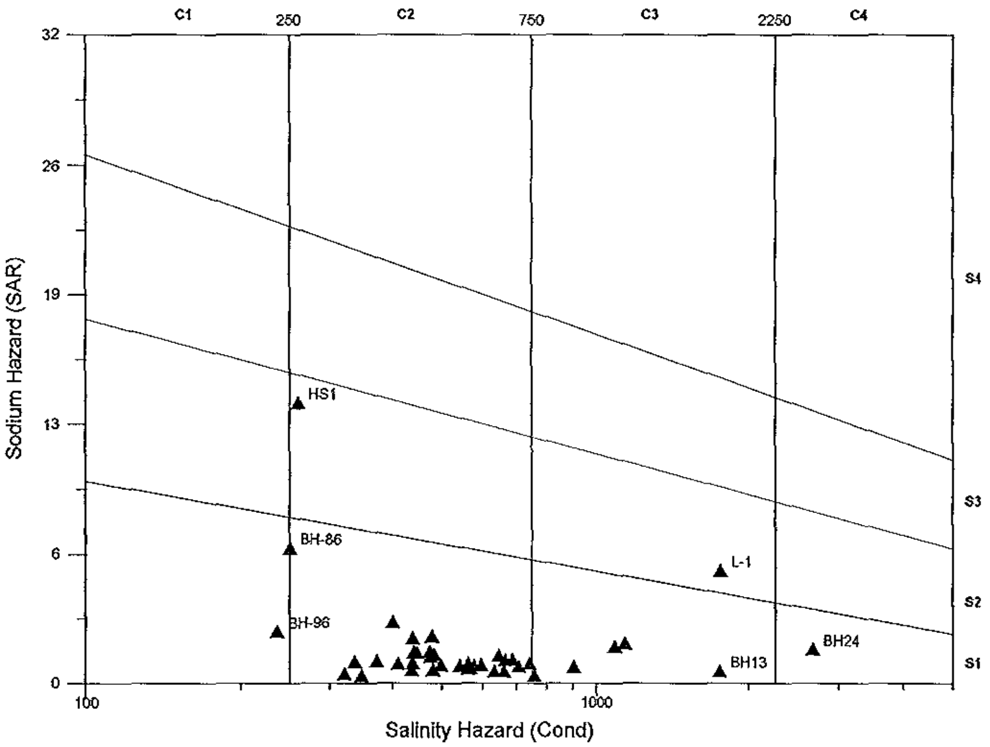


Figure 7.3. suitability of ground water of the study area for irrigation purpose (Wilcox plot).

Therefore, development of excessive sodium in soils will result from irrigation water that has high SAR values (18 or above). Values below 10 indicate little danger of sodium hazard.

The SAR of all water samples in the study area (except HS1) varies from 0.37 to 6.67. Therefore, almost all the natural waters in the study area are excellently suitable for irrigation from the point of their SAR values. The hot spring (HS1) has SAR value of 13.83, which is in the good class (table 7.4) and it may also need cooling before use.

Also based on EC as an index of salinity hazard (Table 7.4) the water samples of the Mehoni sub-basin can be classified as follows:

Table 7.5. Suitability of the natural waters in the study area for irrigation based on their EC.

Water Samples	Range of EC in $\mu s/cm$	Water class
BH – 96	239	Excellent
BH10, BH11, BH12, BH14, BH15, BH16, BH2, BH23, BH25, BH26, BH27, BH28, BH29, BH3, BH31, BH33, BH35, BH38, BH4, BH5, BH6, BH7, BH86, CS1, CS10, CS6, CS8, DW2, HS1, R1, R2, R3	253 – 750	Good
BH13, BH18, BH32, CS2	750 - 2250	Medium
BH24	2702	Bad

Where :-

C_1 : Low salinity hazard S_1 : Low sodium hazard

C_2 : Medium salinity hazard S_2 : Medium sodium hazard

C_3 : High salinity hazard S_3 : High sodium hazard

C_4 : Very high salinity hazard S_4 : Very high sodium hazard

N.B. only the extremes are labelled in the Wilcox plot.

As it was also observed from the Wilcox plot (U.S. Department of Agriculture diagram) on figure 7.3, almost all the tested water samples lie into the zones of low sodium hazard and medium to high salinity hazard. Exceptions are HS1 and L1 with medium sodium hazards and BH24 with very high salinity hazard. In general, high salinity hazard is detected in the

groundwater of the northern and central parts of the valley floor and eastern basaltic areas (refer EC distribution map in figure 7.2). Therefore, care must be taken while using the highly saline waters for irrigation.

Medium salinity waters ($250\,\mu s/cm$-$750\,\mu s/cm$) can be used where moderate amount of leaching occurs and plants with moderate salinity tolerance can be grown in most cases without special practices for salinity. But high salinity waters ($750\,\mu s/cm$-$2250\,\mu s/cm$) can be used on soils that have adequate drainage and special salinity management is required.

It is also equally important that any migration of saline waters from their current locations towards fresh water zones due to excessive pumping of fresh ground waters should be monitored.

7.4.2.2. Specific ion toxicity

Certain ions from soil or water accumulate in sensitive crops to concentrations high enough to cause crop damage and reduced yields. Toxicity problem is different from salinity problem in that it occurs within the plant itself and is not caused by a water shortage. Toxicity normally results when the toxic ions are taken up with the soil-water and accumulate in the leaves during water transpiration to an extent that can damage the plant. The usual toxic ions in irrigation water are chloride, sodium and boron. They cause crop damage either individually or in combination. Most annual crops are not sensitive at concentrations that may severely affect tree crops and woody perennial type plants.

The toxic ions sodium and chloride can also be absorbed directly into the plant through the leaves moistened during sprinkler irrigation. This occurs typically during periods of high temperature and low humidity. The leaf absorption speeds the rate of accumulation of a toxic ion and may be a primary source of the toxicity.

7.4.2.2.1. Chloride

In most of the tested water samples from the study area, the chloride concentration ranges from 0.168meq/l in BH8 to 2.17meq/l in BH13. The water samples with in this range can be readily used for irrigation purpose with no risk of chloride toxicity. But the samples L1, BH32 and BH24 with chloride concentrations of 3.81meq/l, 4.2meq/l and 8.41meq/l respectively can only be used with slight to moderate restriction on use (table 7.6).

7.4.2.2.2. Sodium

The sodium concentration in most of the tested water samples in the study area ranges from 0.52meq/l in DW2 to 2.63meq/l in BH26 and BH86. The water samples with in this range can be used for irrigation with out any restriction on use. Where as CS4, BH32, BH24 and L-1 have sodium concentrations ranging between 3.65meq/l and 14.14meq/l. Therefore, these samples should be used for irrigation with slight to moderate restrictions on use (table 7.6).

7.4.2.2.3. Boron Toxicity

Boron unlike sodium is an essential element for plant growth. (chlorine is also essential but in such small quantities that it is frequently classed as non-essential). Boron is needed only in small amounts, however, if present in amounts appreciably greater than needed, it becomes toxic. Surface water rarely contains enough boron to be toxic but water from well or springs occasionally contain toxic amounts, especially near geothermal areas and earthquake faults.

In this work boron could not be tested due to absence of laboratory chemicals for boron test. But in the previous work by Dessie Nadew (2003) it is indicated that for the water samples collected from the study area and its surroundings boron ranges from less than 0.16ppm to 2.53ppm in lake Ashange and 90% of these samples have a value less than 1ppm. Therefore, almost no boron hazard is detected in the study area from previous this work.

Table 7.6. Guidelines of specific ion toxicity and miscellaneous effects for interpretations of water quality for irrigation

Potential irrigation problem		Unit	Degree of restriction on use		
			None	Slight to moderate	Severe
Specific ion toxicity (affects sensitive crops)					
	Sodium (Na^+)				
	Surface irrigation	**SAR**	<3	3-9	>9
	Sprinkler irrigation	meq/l	<3	>3	
	Chloride (Cl^-)				
	Surface irrigation	meq/l	<4	4-10	>10
	Sprinkler irrigation	meq/l	<3	>3	
	Boron (B)	mg/l	<0.7	0.7-3	>3
Miscellaneous effects (affects susceptible crop)					
	Nitrogen (NO_3 - N)	mg/l	<5	5-30	>30
	Bicarbonate (HCO_3) (over head sprinkling only)	mg/l	<1.5	1.5-8.5	>8.5
	pH		Normal range 6.5-8.4		

7.4.2.3. Miscellaneous effects

7.4.2.3.1. Excess nitrate (NO_3^-)

Nitrogen is a plant nutrient and stimulates growth. Natural soil nitrogen or added fertilizers are the usual sources, but nitrogen in the irrigation water has much the same effect as soil-applied fertilizer nitrogen and excess of it will cause over stimulation of growth, delayed maturity or poor quality of several commonly grown crops. These effects in turn result to low productivity and poor marketability of the grown crops.

In the study area six of the analysed water samples have nitrate concentrations less than 5ppm, twenty-eight samples range from 5ppm to 30ppm, and six samples greater than 30ppm (appendix 1). Therefore, the first group can be used for irrigation with no restriction on use, the second group with slight to moderate restriction and the third group with severe restriction (table 7.6).

7.4.2.3.2. Scale Deposits

Irrigation water that contains high proportion of slightly soluble salts such as calcium, bicarbonate and sulphate may result to the problem of white scale formation on leaves or fruits when sprinklers are used. Although there is no toxicity involved, the deposits often build up on the leaves and fruits. This problem is given special concern when flowers, vegetables or fruits are grown for the fresh market. These deposits reduce the marketability of fruits and foliage.

Out of the tested water samples from the study area BH86 and HS1 had HCO_3^- concentrations of 0.37meq/l and 0.84meq/l respectively; in other thirty-seven samples it ranges from 2.14meq/l to 8.15meq/l; its concentrations in BH13 and L1 were 8.75meq/l and 14.51meq/l respectively. Therefore, the first group can be used for irrigation with no restriction on use, the second group with slight to moderate restriction and the third group with severe restriction (table 7.6).

7.4.2.2.3. Abnormal PH

PH is an indicator of the acidity or alkalinity of water, but is seldom a problem by itself. The main use of PH in a water analysis is for detecting abnormal water. The normal PH range for irrigation water is from 6.5 to 8.4. Abnormal PH warns that the water needs further evaluation. Irrigation water with a PH outside the normal range may cause a nutritional imbalance or may contain a toxic ion. The greatest direct hazard of an abnormal PH in water is the impact on irrigation equipment. Especially PH lower than 6.5 is very corrosive and may rapidly corrode pipelines, sprinklers and control equipment.

In the study area only four samples (L1, HS1, BH86 and BH96) has PH>8.5 (ranging 8.76 to 9.64) the remaining tested water samples has PH ranging from 7.33 to 8.36 which is with in the normal range for irrigation waters. Therefore, most of the natural waters in the study area are non-corrosive from PH point of view.

7.4.2.2.4. Magnesium Problem

In magnesium-dominated waters (Ca/Mg<1) the potential effect of sodium may be slightly increases. In other words, a given SAR value will show slightly more damage if the Ca/Mg ratio is less than 1. The lower the ratio, the more damaging is the SAR. But in the study area, except for L1 with Ca/Mg ratio of 0.07, the Ca/Mg ratio of the rest water samples is greater than 1. Therefore, there is no magnesium hazard in the study area except for the lake Ashange.

CHAPTER EIGHT

8. Conclusion and recommendation

8.1. Conclusion

The Mehoni sub-basin as part of the Raya Valley is considered as an intermountain trough structure that originated from local tectonic development attributed to the regional East African rift system (Isaek Gershanovich, 2000). Therefore the study area is one of the graben and horst lineaments with in the rift system where the graben is filled with sediments that drift out of the volcanic rocks on the western and eastern highlands mainly due to weathering and erosion of the rocks. The major volcanic rocks in the study area are alkaline olivine basalts with minor rocks of tholeiitic affinity that seems to have come later, as dikes, sills and cap rocks.

The mean annual precipitation of the Mehoni sub-basin as estimated from the isohyetal method is about 723.57 mm. By using Penman nomogram method, the free water surface evaporation from lake Ashange was estimated to be 722.42 mm/year. Therefore a value of 1204 mm/year estimated by the United States Water Bureau (USWB) nomogram method (Dessie N., 2003) is considered in the water balance calculation. The PET of the study area as estimated by the Thornthweite formula is equal to 1124 mm/year. The AET for the study area was 687.24 mm/year, which is estimated by taking the average of the values, computed by the Thornthweit and Turc methods. The runoff coefficient as estimated from previous works was taken as 0.22. And the bare land evaporation was assumed to be half of the lake evaporation. Finally by assuming a closed groundwater basin a value of about 66.5 million cubic meters was found for the annual recharge to the groundwater of the study area.

From the geophysical studies and lithologic logs of wells in the valley floor, it was possible to conclude the following:

- Thickness of the unconsolidated sediments is maximum (>250m) in the central parts of the valley floor and decreases towards peripheral areas.
- 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.

- Normally, three layers were detected in the western part of the valley floor and other peripheral areas. On the top it is dry sandy soil (1m-2m thick), in the middle it is saturated sand with some coarser materials (20m-150m thick) and the bottom layer is the basaltic bedrock. In the central part of the valley floor, a thick layer of clay dominated sediments was detected between the saturated sandy layer and the bedrock making the number of layers to be four in this part of the valley floor.

The hydrolithologic units in the study area can be broadly classified as alluvial aquifers and volcanic aquifers. The alluvial aquifers are composed of poorly compacted Quaternary deposits, which fill the valley bottom and intermountain troughs. They are further sub divided into the Quaternary colluvial and alluvial deposits (Q_{col} and Q_{al}) and the Quaternary interfluvial, fan foot plain and valley bottom deposits (Q_{smc}). The Q_{col} and Q_{al} deposits are sand dominated and the Q_{smc} deposits contain significant amount of silt and clay. Whereas, the volcanic aquifers are that found within the fractured and weathered parts of the Tertiary volcanics in the highlands and escarpments of the study area.

Normally, the seasonal groundwater level fluctuation in the valley floor was insignificant (about 0.2m). But the data logger fitted in a monitoring well around Kara Adishawo, where large-scale irrigation using the ground water was being practiced, has registered lowering of the groundwater level by about 2m within less than 2 years of pumping. This lowering of the groundwater level was not due to meteorological effects but rather due to the large scale pumping and intensive diversion activities in the up stream areas.

Based on the groundwater level contour map, the hydrochemical evaluations and the bedrock level map in previous work; the regional groundwater flow directions in the Mehoni sub-basin are from all directions towards its central valley floor with probable groundwater out flow at the southeastern part of the study area. But there are also some localized flow systems in the study area like in the lake Ashange basin where ground water flows from all directions towards the lake. No indication of sub-surface out flow from the lake Ashange basin toward the Mehoni sub-basin is detected in this work.

The aquifer type in the study area is mostly unconfined. But semi-confined conditions may also be encountered in the central and eastern parts of the valley floor. Based on the type of available pumping test data and type of aquifers in the study area, the transmisivity values estimated by the Cooper-Jacob time-drawdown and Theis recovery models were considered as reliable. Accordingly, most of the Q_{col} and Q_{al} 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 Q_{smc} 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 Q_{col} and Q_{al} aquifers have fairly good efficiencies and they are moderately to highly productive. Wells in the Q_{smc} aquifers have lower efficiencies and moderate productivity and those in the volcanic aquifers have weak efficiencies and low productivity. This can be attributed to the higher amount of fines in the Q_{smc} and low transmisivities in the volcanic aquifers.

The spatial distribution of EC (TDS) and the major ions was analysed by producing contour maps for each parameter in the Arcview GIS environment. Accordingly, EC (TDS) and most major ions in the Mehoni sub-basin show an increasing trend from the western highlands towards the central and eastern parts of the study area. In the lake Ashange basin, exceptionally high values of these chemical parameters was detected in the lake Ashange it self. The presence of basalt as the main rock formation and the high concentration of silica in the water samples may indicate that silicate weathering is the main source of most the chemical constituents in the natural waters of the study area.

The bicarbonate type of water is dominant in the study area. As to the cations in the water type formulations Ca, Mg and Na contribute major roles respectively.

With the exception of high TDS and nitrate concentrations in some areas most of the natural waters in the study area are fit for domestic water utilities. The most probable sources for the high nitrate values in some areas were the human and animal excreta and nitrogen fertilizers such as urea.

Almost all the tested water samples are characterised by low sodium hazard and medium to high salinity hazard. Exceptions are HS1 and L1 with medium sodium hazards and BH24 with very high salinity hazard. In general, high salinity hazard is detected in the groundwater of the northern and central parts of the valley floor and eastern basaltic areas. Therefore, care must be taken while using these highly saline waters for irrigation. The medium salinity waters ($250\,\mu s/cm$-$750\,\mu s/cm$) can be used where moderate amount of leaching occurs and plants with moderate salinity tolerance can be grown in most cases without special practices for salinity. But high salinity waters ($750\,\mu s/cm$-$2250\,\mu s/cm$) can be used on soils that have adequate drainage and special salinity management is required.

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.

8.2. Recommendations

In this study much data was collected and analyzed using different techniques within the time limit allowed for the thesis work. Therefore many new hydrogeological information of the study area can be found from this material. But further hydrogeological study is still recommendable so that more detailed and comprehensive knowledge of the groundwater in the study area will be acquired and used for effective planning and sustainable development.

Some detailed and integrated geophysical work is required in order to delineate the different hydrolithologic units in the valley floor and to study the bedrock morphology and structural setup of the valley floor and the lake Ashange basin.

Additional meteorological stations of higher class in the valley floor and stream flow gauging stations at the mouth to the valley floor of large rivers in the study area are required in order to have more reliable estimations of the water balance of the study area.

In order to have tube wells with higher efficiencies, careful design and construction of the wells is necessary where group of hydrogeologists may participate during the decision-making.

Pumping test with observation wells and group pumping test should be undertaken in order to better estimations of the transmissivities and storativities of the aquifers and thereby have an idea about the well interference conditions especially in irrigation well fields and their surroundings.

Monitoring wells fitted with data loggers that can at least measure changes in EC and groundwater level and some test wells are required at selected sites in the western, central and eastern parts of the valley floor for detailed and long-term groundwater study in the area. This can also help to have reliable groundwater models of the study area.

Isotopic study technique can also helps in order to study the hydrological interrelations between the rainwater, surface waters and the ground water of the study area. In the study area this technique helps to detect whether there is any subsurface leakage from the lake to other areas or not; to study the recharge conditions from rainwater and surface water to the ground water and to study the ground water flow systems in the study area.

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APPENDICES

Appendix 1: Static water levels and their dates of measurement

Well No	First measurement of static water level		Second measurement of static water level	
	SWL (m)	Date of measurement	SWL (m)	Date of measurement
BH1	28.9	7 / 7 / 2004	-	-
BH2	19.06	26 / 9/ 2004	-	-
BH3	70.2	-	-	-
BH4	77.48	12 / 9 / 2002	-	-
BH5	36.7	20 / 5 / 2003	-	-
BH6	29	17 / 7 / 2003	-	-
BH7	64.2	22/ 4 / 2004	-	-
BH8	92.51	20 / 9 / 2004	-	-
BH10	35.86	19 / 9 / 2004	-	-
BH11	38.52	19 / 09 / 2004	-	-
BH13	56.3	27 / 9 / 1996	-	-
BH14	47	15 / 9 / 1997	-	-
BH15	44.4	27 / 9 / 1996	-	-
BH16	60.5	27 / 9 / 1996	-	-
BH17	21.4	18 / 9 / 2004	-	-
BH20	21.73	16 / 9 / 2004	-	-
BH22	18	15 / 9 / 1997	-	-
BH24	34.9	12 / 9 / 2004	35.1	23 / 2 / 2005
BH25	43.23	2 / 7 / 2002	-	-
BH29	38	15 / 10 / 1997	-	-
BH30	41.65	10 / 9 / 2004	-	-
BH31	43	20 / 4 / 2001	-	-
BH32	50	-	-	-
BH34	21	-	-	-
BH39	40.35	30 / 5 / 1996	-	-
BH40	40.5	8 / 5 / 1996	-	-
BH42	5.8	20 / 8 /2003	-	-

BH43	58.7	29 / 4 / 1996	-	-
BH44	39.1	29 / 4 / 1996	-	-
BH45	22.4	29 / 4 / 1996	-	-
BH48	8.7	29 / 4 /1996	-	-
BH51	44	-	-	-
BH52	30.9	29 / 4 / 1996	-	-
BH74	13.22	19 / 12 / 1997	18.45	14 / 4 / 1998
BH77	32.79	18 / 11 / 1997	32.79	10 / 4 / 1998
BH79	17.65	29 / 12 / 1997	19.55	26 / 3 / 1998
BH86	30.42	6 / 12 / 1998	48.55	14 / 5 / 1998
BH87	14.65	4 / 12 / 1998	17.1	20 / 5 / 1998
BH88	14.48	2 / 12 / 1998	21.2	28 / 5 / 1998
BH89	37.96	10 / 12 / 1998	41	6 / 6 / 1998
BH90	22.65	28 / 11 / 1998	24.28	11 / 6 / 1998
BH91	11.37	30 / 11 / 1998	12.45	16 / 6 / 1998
BH92	10.68	13 / 11 / 1998	12.13	24 / 6 / 1998
BH93	34.37	15 / 11 / 1998	39.1	7 / 7 / 1998
DW3	5.9	14 / 12 / 1998	7.3	6 / 7 / 1998
DW4	10.05	14 / 12 / 1998	14.9	15 / 7 / 1998

Appendix 2. Hydrochemical data

Appendix 2.1 Location, water type and other additional informations of the collected water samples in this work

Sample ID	X	Y	Z	Sampling date	Sampling time	GEOLOGY	REFERENCE	W/TIPE	PH LAB.	TDS	TURB.
BH10	577527	1396864	1517	1/9/1997	12:26:00 AM	Qsmc	Generated	Na-Ca-Mg-HCO3	7.78	230	< 5
BH11	578474	1398916	1518	1/9/1997	11:20:00 AM	Qsmc	Generated	Mg-Ca-Na-HCO3	7.7	335	< 5
BH12	580815	1402660	1528	1/9/1997	10:36:00 AM	Qsmc	Generated	Mg-Ca-Na-HCO3	7.69	242	< 5
BH13	577637	1402592	1552	1/9/1997	9:51:00 AM	Qsmc	Generated	Mg-Ca-NO3-HCO3	7.54	802	< 5
BH14	574151	1405561	1625	1/9/1997	8:51:00 AM		Generated	Mg-Ca-Na-HCO3	7.78	268	< 5
BH15	576588	1406650	1607	1/8/1997	3:29:00 PM	Qsmc	Generated	Ca-Mg-HCO3	7.57	268	< 5
BH16	574110	1410875	1672	1/8/1997	2:58:00 PM		Generated	Ca-Mg-HCO3	7.59	296	< 5
BH18	577214	1419055	1715	1/8/1997	10:54:00 AM	Q_{al} or Q_{col}	Generated	Ca-Mg-HCO3-NO3	7.33	430	< 5
BH2	569802	1400032	1623	1/13/1997		Q_{al} or Q_{col}	Generated	Ca-Mg-Na-HCO3	7.57		
BH23	568318	1413524	1786	1/7/1997	6:00:00 PM	Q_{al} or Q_{col}	Generated	Ca-Mg-HCO3	7.46	322	< 5
BH24	581179	1397168	1513	1/14/1997	10:29:00 AM	Basalt	Generated	Mg-Ca-Cl-HCO3-SO4-NO	7.44	982	6.5
BH25	583291	1399961	1536	1/11/1997	10:25:00 AM		Generated	Na-Mg-Ca-HCO3	7.72	227	< 5
BH26	585553	1399786	1555	1/11/1997	11:06:00 AM		Generated	Na-HCO3	8	194	< 5
BH27	574827	1392822	1525			Qsmc	Genetated	Ca-Na-Mg-HCO3	7.62		
BH28	572536	1389879	1540		11:54:00 AM	Qsmc	Generated	Ca-Na-Mg-HCO3	7.6	319	
BH29	568477	1392062	1623			Q_{al} or Q_{col}	Generated	Ca-Mg-Na-HCO3	7.58		
BH3	572795	1386301	1484	1/13/1997	1:20:00 PM	Qsmc	Generated	Ca-Mg-Na-HCO3	7.37	323	45
BH30	585460	1403104	1542		4:49:00 PM		Generated	Na-Mg-HCO3	8.14	195	15
BH31	586059	1406286	1549	1/10/1997	3:59:00 PM		Generated	Na-Mg-HCO3	7.9	245	< 5
BH32	583294	1386187	1759	1/11/1997	12:40:00 AM	Basalt	Generated	Na-Mg-Ca-Cl-NO3-HCO3	7.34	644	
BH33	586575	1411430	1583	12/28/1996	9:13:00 AM	Qsmc	Generated	Mg-Ca-HCO3	7.76	229	< 5
BH35	570147	1402326	1653	1/15/1997	10:18:00 AM	Q_{al} or Q_{col}	Generated	Ca-Mg-HCO3	7.61	271	< 5
BH38	571178	1403127	1634	1/15/1997	9:23:00 PM	Q_{al} or Q_{col}	Generated	Ca-Mg-HCO3	7.51	348	< 5
BH4	567162	1394661	1657	1/13/1997	9:04:00 AM	Q_{al} or Q_{col}	Generated	Mg-Ca-Na-HCO3	7.64	355	< 5
BH5	567508	1400144	1672	12/1/1997	2:36:00 PM	Q_{al} or Q_{col}	Generated	Ca-Mg-Na-HCO3	7.61	273	< 5
BH6	571951	1400364	1585	12/1/1997	10:15:00 AM	Q_{al} or Q_{col}	Generated	Ca-Mg-Na-HCO3	7.6	273	< 5
BH7	585118	1413417	1573	10/1/1997	11:13:00 AM	Qsmc	Generated	Na-Mg-Ca-HCO3	8.09	211	< 5
BH-86	560052	1415237	2445	2/24/2005	10:54:00 AM	Basalt	Generated	Na-CO3	9.64	159.2	
BH-96	558460	1414488	2404	2/24/2005	2:30:00 PM		Generated	Na-Ca-HCO3	8.76	152	
CS1	560934	1392555	1893	11:30:00 AM			Generated	Ca-Na-Mg-HCO3	7.36		
CS-10	558637	1402791	2445	2/24/2005	5:15:00 PM	Basalt	Generated	Ca-Na-Mg-HCO3	7.53	228	
CS2	589262	1408760	1640	12/30/1996	11:00:00 AM	Basalt	Generated	Ca-Mg-HCO3	7.7		
CS4	591635	1410569	1703	12/28/1996	10:30:00 AM	Basalt	Generated	Ca-Mg-Na-HCO3	7.49		
CS-6	558278	1400938	2488	2/22/2005	4:45:00 PM	Basalt	Generated	Ca-Na-HCO3	7.82	205	
CS-8	560397	1419350	2941	2/24/2005	9:48:00 AM	Basalt	Generated	Ca-Mg-HCO3	7.65	214	
DW2	579313	1386058	1627	12/29/1996	3:30:00 PM	Basalt	Generated	Ca-Mg-HCO3	7.67		
HS1	561675	1397715	1917	1/12/1997		Basalt	Generated	Na-CO3-HCO3	9.63	140	
L-1	556124	1392159	2461	2/22/2005	11:09:00 AM	Lake	Generated	Na-Mg-HCO3-CO3	9.17	1065	
R1	563056	1398697	1772	1/12/1997	1:10:00 PM	River	Generated	Ca-Mg-Na-HCO3	8.36	207	77
R-2	556114	1395754	2489	2/22/2005	12:42:00 PM	River	Generated	Ca-Mg-HCO3	7.61	268	
R-3	559252	1413842	2363	2/24/2005	11:12:00 AM	River	Generated	Ca-Na-Mg-HCO3	8.22	252	

Appendix 2.2 Chemical data that is generated in this work

Sample ID	SITE	LOCATION	GEOLOGY	Na	K	Ca	Mg	Fe	CO3	HCO3	Cl	SO4	F	NO3	SiO2	CO2	Field PH	TEMP.	COND.	SAR
BH10	Adi Golo	Kukufto	Qsmc	45	1.8	35	17	0.1	0	266	17	8	0.37	8.86	72	19	8	30	479	1.56
BH11	Ganda Chafe	Kara Adi Shawo	Qsmc	45	1.7	52	36	< .1	0	393	21	22	0.39	11.08	70	35	7.8	27	694	1.17
BH12	Adi Tedla	Hade Alga	Qsmc	30	1.1	42	26	< .1	0	282	16	9	0.43	7.09	70	26	7.8	28	503	0.9
BH13	Adi Kuayaru	Kara ADI Shawo	Qsmc	40	1.4	145	125	< .1	0	534	77	130	0.28	398.7	60	74	7.5	26	1772	0.59
BH14	Koban	Kara Adi Shawo		34	1.3	46	29	< .1	0	348	16	7	0.4	5.76	68	24	7.8	27	570	0.97
BH15	Chekoni	Tsigea Wargiba	Qsmc	26	1	53	29	< .1	0	350	16	8	0.38	3.99	61	36	7.2	26	569	0.71
BH16	Tahtay Wargiba	Tsigea Wargiba		22	1.1	66	29	< .1	0	345	28	10	0.26	17.72	55	37	7.8	24.5	641	0.57
BH18	Adi Kirishame	Hawelti	Qal or Qcol	36	1.6	83	43	< .1	0	350	66	47	0.3	66.45	61	59	7.2	26.5	915	0.8
BH2	Aworamarsa 1	Werabaye	Qal or Qcol	29	0.4	54	23	< .1	0	305	15	9	0.23	16.83	48	18			549	0.83
BH23	Mehoni town	Mehoni town	Qal or Qcol	23	1	71	30	< .1	0	328	25	23	0.25	44.3	49	42	7.4	23.5	668	0.58
BH24	Kario	Hade Alga	Basalt	125	2	210	130	< .1	0	479	298	370	0.34	221.5	63	89	7.66	30	2702	1.67
BH25	Chadu	Hade Alga		42	1.9	25	21	< .1	0	265	14	8	0.43	3.54	64	22	7.8	31	453	1.5
BH26	Adi Guben	Hade Alga		60.5	2.6	15	9.5	< .1	0	223	15	10	0.26	< .04	48	10	8.1	30	405	3.01
BH27	Tsaeda Meda	Kukufto	Qsmc	43	1.4	44	15	< .1	0	292	10	8	0.41	5.76	72	26			488	1.43
BH28	Delagalto	Kukufto	Qsmc	50	0.8	58	26	< .1	0	387	19	10	0.32	9.75	67	42	7.3	28.5	653	1.37
BH29	Kalina	Kukufto	Qal or Qcol	33	0.7	61	25	< .1	0	375	14	8	0.28	4.43	49	35			603	0.9
BH3	Kelkala	Teo	Qsmc	43	0.8	64	26	< .1	0	359	30	21	0.35	14.18	60	30		30.5	673	1.15
BH30	Adi Berhe Reda	Hade Alga		52	2.3	18	14	< .1	0	216	22	13	0.31	11.08	39		8.2	26	444	2.23
BH31	Adi Adey	Mechare		58	2.4	20	17	< .1	0	223	31	13	0.3	8.42	42	11	7.8	39	483	2.3
BH32	Chercher town	chercher town	Basalt	86	0.9	68	44	< .1	0	217	149	92	0.22	110.75	34	35	7.4	26	1150	2
BH33	Adi Bedera town	Mechare	Qsmc	22	2.3	42	26	< .1	0	284	15	1	0.43	10.19	76	19	7.8	26	486	0.66
BH35	Production well	Kara Adi Shawo	Qal or Qcol	27	0.4	58	24	< .1	0	332	16	8	0.35	9.3	56			24	575	0.75
BH38	Production well	Kara Adi Shawo	Qal or Qcol	33	0.9	71	32	< .1	0	383	28	18	0.35	24.37	64		7.3	28	714	0.82
BH4	Marsa	Danisa	Qal or Qcol	40	0.6	61	39	< .1	0	394	38	21	0.36	12.4	57	27		25	750	0.98
BH5	Ganda Ajo	Werabaye	Qal or Qcol	31	0.4	61	24	< .1	0	322	18	12	0.26	19.05	50	18		23	585	0.85
BH6	Direito	Werabaye	Qal or Qcol	32	0.5	57	24	0.1	0	355	15	8	0.29	5.32	51	34	7.5	22.5	572	0.9
BH7	Arada Basea	Mechare	Qsmc	41	2.2	27	17	< .1	0	203	27	12	0.23	12.85	39	7	8.1	28	446	1.52
BH86	T.Haimanot	H.T.Haimanot	Basalt	60.5	0.6	3.56	1.62		52.8	22.7	17.9	5.5	0.96	8.5			8.5	22.9	253	6.67
BH-96	Maichew town	Maichew		44.5	0.6	16.02	4.3		14.4	130.4	8.9	4.5	0.96	10.5			8.4	25.8	239	2.55
CS1	Adi Gereb	Beiru DAnisa		38	< .01	42	14	< .1	0	266	15	8	0.2	15.95	41	23			481	1.3
CS10	Chinkomajo	Chinkomajo	Basalt	34	0.4	44.8	14.6		0	264.6	7.94	1	0.4	10.5			7.8	21.6	377	1.13
CS2	Gadwo	Mechare	Basalt	17	6.4	98	28	< .1	0	384	45	44	0.22	10.63	31	18			768	0.39
CS4	Adi Dera	Mechare	Basalt	84	4.8	84	50	< .1	0	497	70	96	0.47	35.44	42	34			1102	1.79
CS6	Aba Gabir	Gumbrda	Basalt	30.5	0.3	46.3	10.8		0	225.8	6.95	2.4	0.15	16.25				19.7	340	1.05
CS8	Boleta	Emba Hasti	Basalt	12.2	0.9	56.96	16.2		0	250.9	5.96	4.5	0.4	10.5			7.6	17.1	351	0.37
DW2	Dadacha base flow	Erba Adi Mekoni	Basalt	12	3.1	34	12	0.1	0	173	9	8	0.88	7.53	26	7			325	0.45
HS1	Owo(Latte)	Werabaye	Basalt	56.5	< .01	1.1	< .1	< .1	43	51	15	8	0.45	< .04	44	0	> 8.2	35	263	13.83
L1	Lake Ashange	Lake Ashange	Lake	325	23.5	10.7	150.1		312	885.5	135	52.8	1.1	5			8.8	21.3	1776	5.57
R1	Gereb Haya	Werabaye(Latte)	River	30	1.8	43	17	< .1	2	262	12	7	0.26	9.3	36	0		29	444	0.98
R2	Quiha	N. Ashange	River	23.5	3.2	64.1	20		0	301.1	7.94	2.4	0.5	16.25			8.2	27.3	443	0.66
R3	T.Haimanot	Near Maichew	River	33	1.4	60.5	17.3		16	267.9	12.9	5.5	0.4	12.5			8.2	21.7	415	0.96

Appendix 2.3 Chemical data collected from pravious works

Sample ID	SITE	Na	K	Ca	Mg	Fe	CO3	HCO3	Cl	SO4	F	NO3	SiO2	CO2	B	PH	COND.	REFERENCE
MW	Kara Adishawo					0.03			5	1.2	0.44	1.12	1.2	10			572	WWDE
BH13 (SBH-56)	Adi Wario	42	1.3	130	110			514	73	126	0.24	327.8	72	10	< .16	7.67	1641	Dessie N., 2003
BH15(BH-55)	Chekoni			57	25.6				12	2		0.6	26			7		RVADP, 1998
BH16(BH-54)	Wargiba			74	26				21	12.2		1.9	24					RVADP, 1998
BH16(SBH54)	Wargiba	23	1.4	65	28			345	25	11	0.21	13.73	51	3	0.76	7.91	614	Dessie N., 2003
BH31 (SBH85)	Soloka	60	2.4	19	17			219	31	18	0.27	7.97	43	< .16	1	7.98	482	Dessie N., 2003
BH44(BH53)	Mehoni	41	2	82	35	0		366	64	41.2		1.2						RVADP, 1998
BH47 (BH-03)	Maichew	32	0.9	63	18			300	18	11		35	44					RVADP, 1998
BH47((SBH03)	Maichew	21	0.6	67	28			401	13	4	0.22	16.39	41	4	0.63	7.56	605	Dessie N., 2003
BH52 (BH-16)	Adi Dergi			49	16.8	0			18	12.5		31.9	32					RVADP, 1998
BH52(SBH16)	Adi Dergi	37	0.8	44	16			295	23	14	0.36	11.96	68	4	0.24	7.72	516	RVADP, 1998
BH53 (BH-17)	kukufto	63	0.2	15	1			171	24	12		0.42	44					RVADP, 1998
BH54 (BH-18)	Bagile Dolo	22.4	0.8	80	35			439.2	17	10.5								RVADP, 1998
BH54(SBH17)	Kukufto	72	0.3	5.2	0.7		6	141	27	19	0.38	< .03	46	0	< .16	8.68	362	RVADP, 1998
BH55(BH-37)	Adi Barie			38	26.8				11	17.4		5.3	48			7.9		RVADP, 1998
BH57(SBH51/2)	Werabaye	32	0.6	62	25			388	753	2	0.29		64	6	< .16	7.78	600	Dessie N., 2003
BH59 (SBH66/1)	Adi Bidera	33	2.5	37	25			293	13	4	0.47	4.43	73	3	0.32	8.04	489	Dessie N., 2003
L-1	Lake Ashenge	340	24	154	3.6		225	914	108	101	0.29	9.3	5		2.53	9.31	2148	Dessie N., 2003
SBH66/2	Adi Bedera	32	1.2	72	42			450	35	19	0.21	27.46	54	4	< .16	7.8	797	Dessie N., 2003