ADDIS ABABA UNIVERSITY

SCHOOL OF GRADUATE STUDIES DEPARTMENT

OF CIVIL ENGINEERING

NUMERICAL GROUNDWATER FLOW MODELING OF THE GERADO RIVER CATCHMENT

Thesis Submitted to Addis Ababa Institute of Technology, School of Post Graduate Studies in partial fulfillment of the requirements for the Degree of Masters of Science in Civil and Environmental Engineering, Hydraulic Engineering stream.

By:

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ACKNOWLEDGMENTS

I am highly indebted to the Ethiopian road authority (ERA) and Kombolcha institution of technology for sponsoring me to attend my postgraduate study.

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ABSTRACT

In this study, a one dimensional groundwater flow model under steady state condition was constructed as a tool to understand the aquifer system of Gerado River catchment using numerical groundwater flow modeling. The study was ambitious in the sense that the modeling was conducted with limited data, particularly for the highland tertiary volcanic part of the basin due to absence of wells in the highland. The groundwater storage is in the alluvium and in the

fractured volcanics. The aquifers in the rocks and alluvial sediments are assumed to be hydraulically linked or interconnected.

The groundwater flow system in the two aquifers of the basin was modeled using MODFLOW (McDonald and Harbaugh, 1988). The model was run for steady-state conditions in for unconfined aquifer. The grid cell size of the model was taken 250 x 250m. Model area and the layer top elevation were delineated by the DEM processing and use of topographic maps. The hydraulic conductivity values were found from pumping test data analysis and literature review for the alluvial sediment aquifer and the volcanic aquifer respectively.

Recharge for the area was estimated using the water balance approach and gave 148.95mm/year or 13% of the rainfall. The model Calibration was using trial and error approach results were evaluated using statistical lumped sum description of the average differences of residuals between simulated and measured heads (mean error, mean absolute error, root mean square error & R_2) were calculated for the residuals and found to be 1.5m,5.29m,6m and 0.88m respectively.

The average recharge obtained by the model is about 151mm/year which closer to the manually calculated value of 148.95mm/year. The total groundwater in Gerado catchment is 39.3 MM^3/year. This volume of water compared with the total abstraction in the basin (about 2.6 MM^3/year) is good storage and can sustain well abstraction at the existing rate or more than the existing in the future.

Simulated heads were sensitive to recharge and hydraulic conductivity in similar fashion. As observed from the values, the model was most sensitive to both decrease in the recharge and hydraulic conductivity values, mainly beyond 50 percent and less sensitive to both recharge and hydraulic conductivity increasing of the value.

The responses of the aquifer to different scenarios were investigated with increased withdrawals and decreased recharges. The effects of increased in groundwater withdrawal rate by 50% cause decline in groundwater level with a maximum of the water level by 1.08 m and a minimum of 0.19m,decreas groundwater discharge through general head boundary and Base flow to river. The second scenario is effect of decrease in groundwater recharge by 25 % due to less than normal precipitation cause maximum decline of the water level by 2 of the water level by 23.09m and a minimum of 2.98m , decrease of groundwater discharge through the general head boundary & Base flow.

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List of abbreviations

GHB	General Head boundary
FAO	Food and Agricultural Organization of the United Nation
GIS	Geographic Information Systems
m.a.s.l	Meter above sea lvel
m/s	Meters per second
m^2/d	Meter square per day
m^3/d	Meter cubic per day
UTM	Universal Transvers Mercator
DEM	Digital Elevation Model
PW	Pumping well
FAO	Food and Agricultural Organization
K	Hydraulic conductivity
IFD	Integrated finite difference
d	Day
ETo	Potential Evapotranspiration
DWL	Dynamic Water Level
SWL	Static Water Level
MAE	Mean absolute error
ME	Mean error
M^3m	Million Cubic Meters
RMS	Root mean square error

ABSTRACT

In this study, a one dimensional groundwater flow model under steady state condition was constructed as a tool to understand the aquifer system of Gerado River catchment using numerical groundwater flow modeling. The study was ambitious in the sense that the modeling was conducted with limited data, particularly for the highland tertiary volcanic part of the basin due to absence of wells in the highland. The groundwater storage is in the alluvium and in the fractured volcanics. The aquifers in the rocks and alluvial sediments are assumed to be hydraulically linked or interconnected.

The groundwater flow system in the two aquifers of the basin was modeled using MODFLOW (McDonald and Harbaugh, 1988). The model was run for steady-state conditions in for unconfined aquifer. The grid cell size of the model was taken 250 x 250m. Model area and the layer top elevation were delineated by the DEM processing and use of topographic maps. The hydraulic conductivity values were found from pumping test data analysis and literature review for the alluvial sediment aquifer and the volcanic aquifer respectively.

Recharge for the area was estimated using the water balance approach and gave 148.95mm/year or 13% of the rainfall. The model Calibration was using trial and error approach results were evaluated using statistical lumped sum description of the average differences of residuals between simulated and measured heads (mean error, mean absolute error, root mean square error & R2) were calculated for the residuals and found to be 1.5m,5.29m,6m and 0.88m respectively.

The average recharge obtained by the model is about 157.2 mm/year which closer to the manually calculated value of 148.95mm/year. The total groundwater in Gerado catchment is 39.3 MM^3/year. This volume of water compared with the total abstraction in the basin (about 2.6 MM ^{3}/year) is good storage and can sustain well abstraction at the existing rate or more than the existing in the future.

Simulated heads were sensitive to recharge and hydraulic conductivity in similar fashion. As observed from the values, the model was most sensitive to both decrease in the recharge and

hydraulic conductivity values, mainly beyond 50 percent and less sensitive to both recharge and hydraulic conductivity increasing of the value.

The responses of the aquifer to different scenarios were investigated with increased withdrawals and decreased recharges. The effects of increased in groundwater withdrawal rate by 50% cause decline in groundwater level with a maximum of the water level by 0.866 m and a minimum of 0.131,decreas groundwater discharge through general head boundary and base flow to river. The second scenario is effect of decrease in groundwater recharge by 25 % due to less than normal precipitation cause maximum decline of the water level by 10.623 and a minimum of 0.583m , decrease of groundwater discharge through the general head boundary & base flow.

CHAPTER ONE

1.1Background

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

Groundwater is the sub-surface water that occurs beneath the water table in soils and geologic formations that are fully saturated (Freeze and Cherry, 1979). It is one of the most valuable natural resources, which supports human health, economic development and ecological diversity. Because of its several inherent qualities (e.g. consistent temperature, continuous availability, excellent natural quality, reasonable development cost, drought reliability relatively less risk to pollution, etc.), it has become an important and dependable source of water supplies in all climatic regions including both urban and rural areas of developed and developing countries (Todd, 2005).

Numerical models using computer programs that are used for analyzing flow in groundwater system played an increasing important role in the evaluation of alternative approaches to groundwater development and management.

Because of extended drought condition, increase in population size and advancement in civilizations, and there by change in land use and land cover, variation in recharge amount, groundwater storage is decreasing and water demand is increasing at an alarming rate. This generalization is true for Gerado River catchment too.

This research work is to model aquifer system of Gerado River catchment at steady state by assuming one layer unconfined aquifer system, to calibrate the model based on observed head, and to analyze model sensitivity to different model parameters like hydraulic conductivity and recharge.

In addition, this thesis work gives an understanding about the response of Gerado River catchment regional groundwater flow system to different possibly occurring scenarios like decreasing in recharge and increasing in well withdrawal. So this model may be used as a tool for water resource managers to assess the regional effects of change in hydrogeological stresses to the steady state system. Moreover, it improves understanding of the groundwater system and the regional effects of various groundwater use alternatives on the water resource of the area.

1.2 Problem Statement

Dessie and its surrounding areas are currently getting their water supply from the boreholes and springs which are found in Hote, Dawido, Erobit, Borumeda and Gerado. But the Gerado catchment covers 70% of water supply for the people in the Dessie town and surrounding areas. There are seven boreholes that have different depths and yields are found drilled in different parts of the catchment.

Currently the need for additional water sources for Dessie town and its surrounding areas is highly increasing due to the increasing number of population and the increasing number of industries. To satisfy this need, additional boreholes are going to be drilled in the Gerado River catchment. This intensive development plan of groundwater in the catchment prior to detail investigation will expose the groundwater resource of the catchment to risk, and ultimately the problem threatening the water supply become in evitable. Groundwater modeling is one of the essential tool to evaluate the groundwater flow and for quantifying its potential. Hence, this thesis can be an input to groundwater developers is the catchment.

1.3 Objectives of the research

1.3.1General objective

The general objective of this research work was to develop a better understanding of the ground flow system of the Gerado River catchment.

1.3.2 Specific objectives

- ❖ Determine areal recharge amount of the catchment.
- ❖ Conceptualization the general groundwater flow system.

- To build a numerical model which can be used to simulate the groundwater flow pattern under steady state condition
- To analyze the sensitivity of the model to which parameter or stress the flow condition is sensitive.
- To evaluate the behavior of the groundwater system under possible future utilization scenarios.

1.4 Research questions

To address the above objectives, the following research questions are posed:

- Does the conceptual model be transformed to a numeric model?
- Does the model provide reliable results?
- Do the simulated hydraulic heads match the observed heads?
- Can a steady state groundwater flow model of the area improve the understanding of flow patterns and predict the effect of future abstractions?

1.5 Structure of the tissues

The content of the thesis is briefly outlined as follows;

- Chapter one: introduction part, as it was discussed above impotency of groundwater and cause of groundwater storage decrease. It also contains the objectives and statement of the problem for this thesis.
- Chapter two: It deals with review of the previous work in the study area and relevant citation of literatures that are related to the topic of this research paper. The portion is covered by types of model, how to building the conceptual module, assigning parameter values, calibration process and calibration techniques.
- Chapter three: it gave a brief describes the study area location, Physiography and Drainage system, climate condition and Land use and land cover of the study area. It also contains geology and the slop of the study area.
- Chapter four: integrates the methods used to carry on the investigations and data collected. It is also discussed about meteorological data analysis; it can be used for determination of recharge. This analysis includes determination of areal depth of rainfall and evapotranspiration. It also contains developing conceptual model and numerical model.

- Chapter five: This chapter consists of the result and discussion of the findings made by the analysis part. The result includes determination of areal rainfall, potential evapotranspiration actual evapotranspiration and surface runoff. And also estimate recharge of the study area which one of the input of the model.
- Chapter sex: the final chapter is the conclusion part. On this part some conclusion are reached and stated. Summary of contributions and recommendation are also covered in this chapter.

CHAPTER TWO

2. Literature Review

2.1 Review of previous study

Several Geological and Hydrogeological studies have been conducted around Dessie area was conducted during the year 1992 by Shawel. The major objective of the research was for the identification of well sites for Dessie town water supply. To accomplish the work different methodologies methodologies or multidisciplinary approacheslike geological, hydrogeological and geophysical investigation were deployed. From the geological study, Dessie area is composed of hard rocks and unconsolidated deposits. According to Shawel, 1992 the higher elevation is composed of hard rock mainly basalt and the depression and valleys are filled by unconsolidated deposits. The grabens around Dessie (Gerado and Boru Meda) are filled by unconsolidated sediments of alluvial and lacustrine formations. Accordingly this study, the main well fields (Gerado, Boru Meda and Hote) were evaluated for the possibility of potential water source for Dessie Water Supply and finally Gerado Well field was selected.

Hydrogeological studies have been conducted in connection with the water supply of the town Dessie (ADSWE, March 2002). Based on this study, average depth of alluvial aquifer is 200m. The groundwater recharge is mainly from the south direction and the hydraulic conductivity of the river bed is estimated to be about 0.01m/day.

Dereje Gidafie (2012) carried out Groundwater potential and water quality investigation of Gerado River catchment, in his Msc thesis. In his research he had carried out conventional hydrogeological investigation in order to define aquifer characteristics, aquifer Productivity, and Recharge and Discharge Zones of the study area. In his research, he determent Groundwater Recharge, Runoff, Evapotranspiration and water quality of the study area.

2.2 Groundwater Modeling

A groundwater model may be defined as a simplified version of the real groundwater system that approximately simulates the excitation- response relations of the groundwater system. The real system is very complicated and difficult to use it directly for the purpose of planning and making management decisions. The simplification is introduced in the form of a set of assumptions that express our understanding of the nature of the system and its behavior. These assumptions will tend to smooth out the effect of various heterogeneities. Because the model is a simplified version of the real system, there exists no unique model for a given groundwater system (Bear, J. & Verruijt, A., 1994).

A computer program or code solves a set of algebraic equations generated by approximating the partial differential equations (governing equation, boundary conditions, and initial conditions) that form the mathematical model (Anderson & Woessner, 1992).

With the introduction of computers and their application in the solution of numerical models, physical models and analogs have become laid off as tools for predicting future groundwater regimes. The selection of the appropriate model to be used in any particular case depends on the objective or objectives of the investigation and the available resources. The later include time, budget, skilled manpower, high capacity computers and codes (Bear, J. & Verruijt, A., 1994).

Groundwater models are an attempt to represent the essential features of the actual groundwater system by means of a mathematical counterpart (Todd 1976). These models have a capacity to test and quantify the consequences of various errors and related model-based forecasts.

Groundwater models according to Todd are physically based mathematical models derived from Darcy's law and the law of conservation of mass. The following five numerical methods are used in groundwater modeling: (Anderson & Woessner, 1992).

1. Finite differences
2. Finite elements
3. Integrated finite differences
4. The boundary integral equation method, and
5. Analytic elements.

The boundary integral equation method (Liggett and Liu, 1983; Liggett, 1987) and analytic elements (Strack, 1987, 1988) are relatively new techniques and are not yet widely used. Integrated finite difference (IFD) techniques are closely related to the finite element method. Finite differences and finite elements are more commonly used to solve flow problems.

The choice between a finite difference and a finite element model depends on the problem to be solved and on the preference of the user. The accuracy of the solutions (model predictions) is dependent upon the reliability of the estimated model parameters and the accuracy of the prescribed boundary conditions.

Finite differences are easy to understand and program. In general, fewer input data are needed to construct a finite difference grid.

Finite elements are better able to approximate irregularly shaped boundaries than standard finite differences. (Integrated finite differences, however, can handle irregular boundaries as well as finite elements.) It is easier to adjust the size of individual elements as well as the location of boundaries with the finite element method, making it much easier to test the effect of nodal spacing on the solution.

There is a fundamental difference in philosophy, however, between the two methods. Finite difference methods compute a value for the head at the node which also is the average head for the cell that surrounds the node. No assumption is made about the form of the variation of head from one node to the next. Finite elements, on the other hand, precisely define the variation of head within an element by means of interpolation (basis) functions. Heads are calculated at the nodes for convenience, but head is defined everywhere by means of basic functions.

The general form of the finite difference expression for Eqn. 2.1is written for the computational molecule shown in Fig. 2.1a as follows:

$$\frac{d}{dx}\left(Kxx,\frac{dh}{dx}\right)+\frac{d}{dy}\left(Kyy,\frac{dh}{dy}\right)+\frac{d}{dz}\left(Kzz\frac{dh}{dz}\right)-W=Ss\frac{dh}{dt} \qquad \text{Eq2.1}$$

$$Bh_{i-1,j,k} + {}_{Chi,j+1,k} + Dh_{i+1,j,k} + Eh_{i,j-1,k} + Fh_{i,j,k+1} + Gh_{i,j,k-1} + Hh_{i,j,k} = RHS_{i,j,k} \qquad \text{Eq2.2}$$

The equation for the head at the node i, j, k ($h_{i,j,k}$) involves the head at the node itself as well as heads at the six surrounding nodes. Each head is multiplied by a coefficient (B, C, D, E, F, G, or H) that is a function of the hydraulic conductivity between the nodes. Coefficient H is also a function of the storage term. The term $RHS_{i,j,k}$ includes storage and recharge terms on the right-hand side of the equation. A form of Eqn. 2.2 is used in MODFLOW. The two-dimensional finite difference equation, is written for the computational molecule.

$$Bh_{i-1,j} + Ch_{i,j+1} + Dh_{i+1,j} + Eh_{i,j-1} + Hh_{i,j} = RHS_{i,j} \qquad \text{Eq2.3}$$

Where again B, C, D, E, and H are coefficients. Both Eqns. 2.2 and 2.3 can be written as matrix equations of the form [A]{h} = {f}, where [A] is the coefficient matrix, {h} is the array of unknown heads, and {f} is the array of terms on the right-hand side (RHS) of the equation.

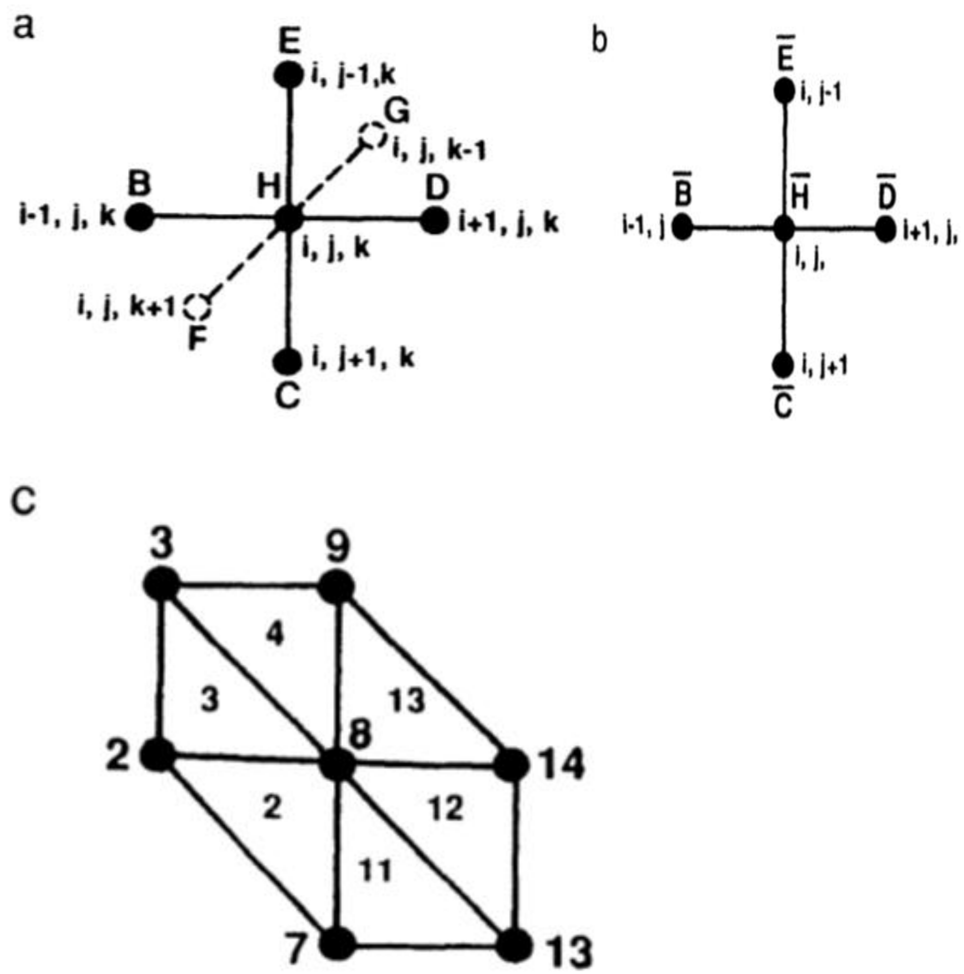

Figure 1:-Computational molecules

a) Three-dimensional finite difference molecule.
b) Two-dimensional finite difference molecule.
c) Patch of six finite elements around node 8. Both node numbers (large type) and element numbers (small type) are shown.

AQUIFEM-1 uses a two-dimensional finite element approximation based on the Galerkin finite element method that yields a matrix equation of the form [G]{h} = {f}, where [G] is the coefficient matrix, also known as the conductance matrix. The matrix equation is assembled by calculating contributions to each term in the coefficient matrix from a patch of elements like the one shown in Fig. 1 (Anderson and Woessner, 1992).

2.2.1Types of Models

There are several ways to classify groundwater flow models. Models can be either transient or steady state, confined or unconfined, and consider one, two, or three spatial dimensions. (Anderson and Woessner,1992). We can classify models in terms of spatial dimension as

- Two-dimensional areal,
- Two dimensional profile,
- Quasi three-dimensional, and
- Full three-dimensional.

Two-dimensional areal and quasi three-dimensional models assume the aquifer viewpoint, while two-dimensional profile and full three-dimensional models use the flow system viewpoint.

I. Two –Dimensional areal models

Two-dimensional areal simulations may consider four different types of aquifers. These are

- Confined aquifers
- Leaky confined aquifers
- Unconfined aquifers, and
- Mixed aquifers.

Confined Aquifers

When simulating confined aquifers, transmissivity and storage coefficient are specified for each node, cell, or element. Variation in transmissivity may represent changes in either hydraulic conductivity or aquifer thickness (Figures. 2). In a two-dimensional areal model, anisotropy in transmissivity is represented by the difference between transmissivity in the x and y directions (Tx and Ty). Input to the model may consist of the two transmissivity arrays or the Tx array and an anisotropy factor to compute the Ty array from the Tx array. Transmissivity can be estimated from literature values of hydraulic conductivity and estimates of aquifer thickness. Values for transmissivity and storage coefficient are generally obtained from pumping test results.

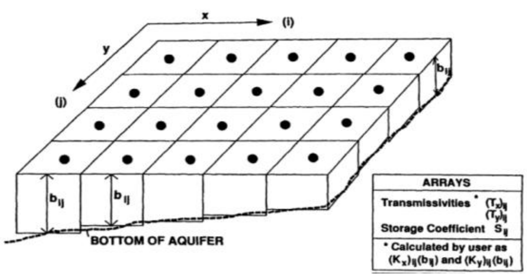

Figure 2:-Block -centered grid for a two-dimensional areal model of a confined aquifer

Leaky Confined Aquifers

In a leaky confined system, the confining bed and adjacent aquifer that supplies leakage to the confined aquifer are not explicitly represented in the model but are simulated by means of a leakage term (see Eqn. 2.1). The leakage term is a function of the leakance, which is the ratio of the vertical hydraulic conductivity (Kz) of the confining bed to the thickness of the confining bed (b')

Leakance = K'_z/b'

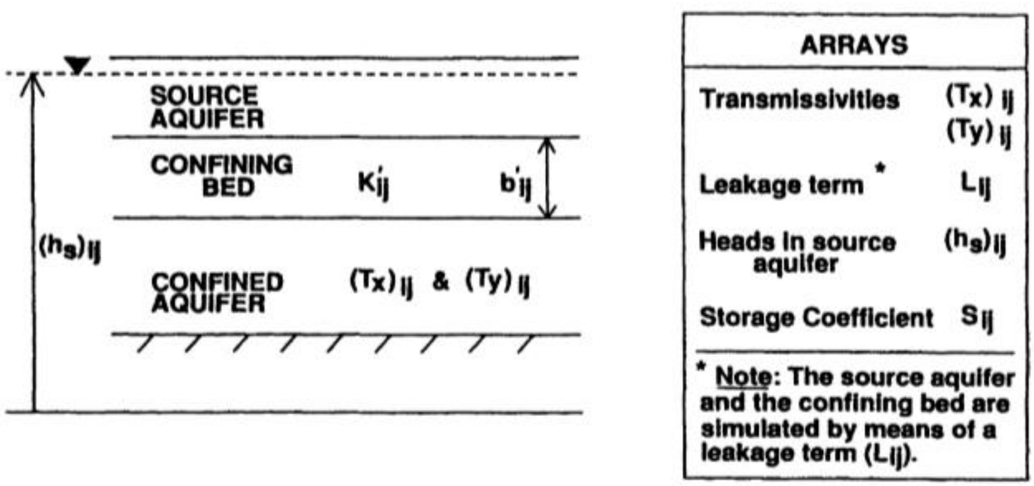

Figure 3:-Cross section of a two-dimensional areal model of a leaky confined aquifer.

Unconfined Aquifers

Most modeling applications involving unconfined aquifers use the Dupuit assumptions, which ensure horizontal flow by requiring that there is no change in head with depth. Use of the Dupuit assumptions in effect turns a three-dimensional problem into a two-dimensional areal problem and a two dimensional profile problem into a one-dimensional problem. The model calculates the elevation of the water table for each node.

Simulations involving an unconfined aquifer require arrays specifying hydraulic conductivity, specific yield, and the elevation of the datum (Fig 4). Because the simulation is usually done in two-dimensional areal view, hydraulic conductivity values should be vertically averaged when using point data or obtained from pumping tests.

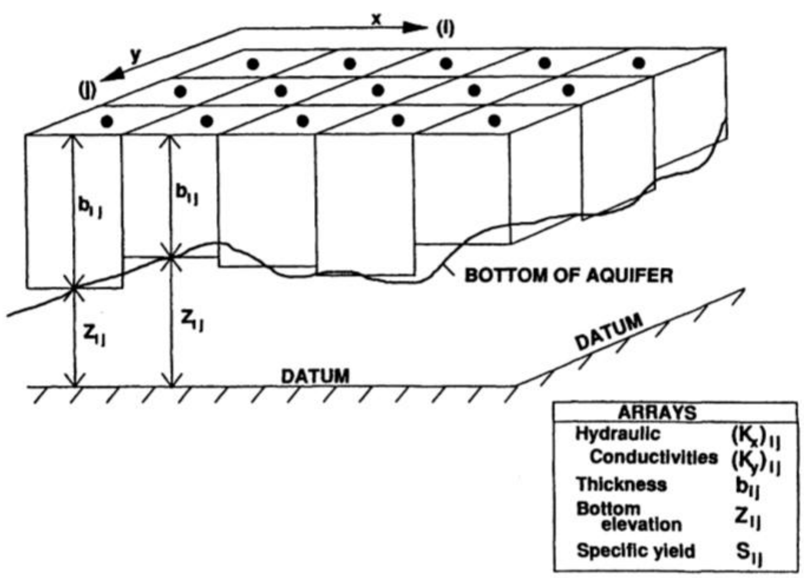

Figure 4:-Block-centered grid for a two-dimensional areal model of an unconfined aquifer

Mixed Aquifer Systems

A mixed aquifer system consists of some combination of the above three aquifer types. An aquifer may vary spatially from unconfined to confined conditions. Or an aquifer may undergo conversion from confined to unconfined conditions as a result of pumping.

II. Quasi three dimensional models

A quasi three-dimensional model simulates a sequence of aquifers with intervening (between) confining layers .Like two-dimensional areal models of leaky confined aquifers confining layers are not explicitly represented in a quasi three-dimensional model, nor are heads in the confining beds calculated. The effect of a confining bed is simulated by means of a leakage term ($L_{i,j}$) representing vertical flow between two aquifers. The leakage term is a function of the leakance and the head difference across the confining bed (Eqn. 2.1). Release of water from storage within the confining bed typically is not considered in this approach. In a quasi three-dimensional model the head in the unit overlying the top confining bed, usually an unconfined aquifer can be calculated directly by the model.

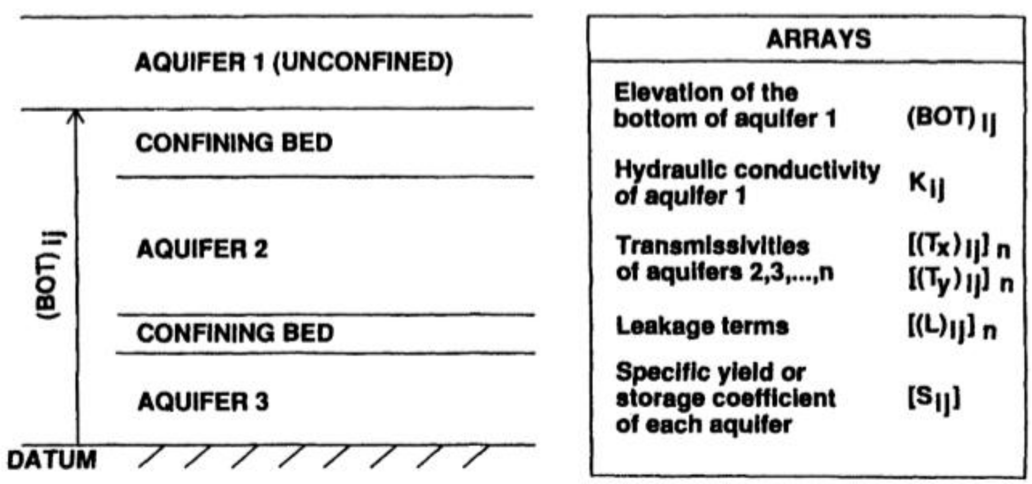

Figure 5:- Schematic view of a quasi three-dimensional model

III. Profile and full three dimensional

Two-dimensional profile models and full three-dimensional models assume the flow system viewpoint. Full three-dimensional models have essentially the same array requirements as two-dimensional areal models except that parameter arrays must be specified for each layer of the model (Fig. 6).

Profile or full three-dimensional models are used to simulate unconfined aquifers when vertical head gradients are important. In these models, the water table (and seepage face, if present) forms part of the boundary. Both finite difference and finite element models are able to simulate aquifers in profile, but movement of the water table and seepage face is more easily handled with a finite element model.

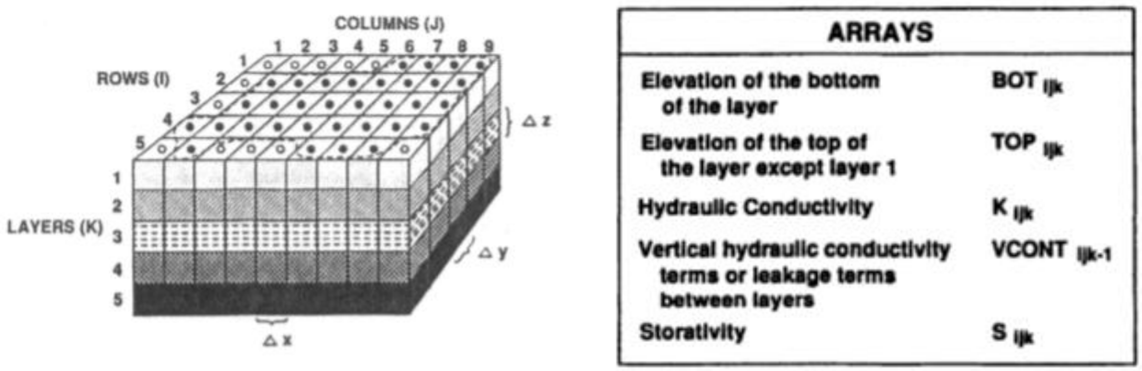

Figure 6:-Schematic diagram of a full three-dimensional model

2.2.2Building the Conceptual Model

A conceptual model is a pictorial representation of the groundwater flow system, frequently in the form of a block diagram or a cross section. The purpose of building a conceptual model is to simplify the field problem and organize the associated field data so that the system can be analyzed more readily. Simplification is necessary because a complete reconstruction of the field system is not feasible. The key data requirements in the process of conceptualization include data about hydro-stratigraphic units, surface water bodies, physical and hydraulic boundaries, recharge and discharge zones.

In theory, the closer the conceptual model approximates the field situation, the more accurate is the numerical model. However, in practice it is desirable to strive for parsimony, by which it is implied that the conceptual model has been simplified as much as possible yet retains enough complexity so that it adequately reproduces system behavior. It is critical that the conceptual model be a valid representation of the important hydrogeologic conditions; failure of numerical models to make accurate predictions can often be attributed to errors in the conceptual model.

The first step in formulating the conceptual model is to define the area of interest, i.e., to identify the boundaries of the model. Numerical models require boundary conditions, such that the head or flux must be specified along the boundaries of the system

The finite difference method, as applied in the computer code MODFLOW, was used in this study. The code is based on the physical theory of groundwater movement Darcy's law and the continuity equation. The program supports seven additional packages, which are integrated with the original MODFLOW (Chiang and Kinzelbach, 2001).

2.2.3 Assigning Parameter Values

DATA NEEDS

Table 1:-Data needed for groundwater flow models

Data Requirements for a Groundwater Flow Model

A. Physical framework 1.
   1. Geologic map and cross sections showing the areal and vertical extent and boundaries of the system.
   2. Topographic map showing surface water bodies and divides.
   3. Contour maps showing the elevation of the base of the aquifers and confining beds.
   4. Isopach maps showing the thickness of aquifers and confining beds.
   5. Maps showing the extent and thickness of stream and lake sediments
B. Hydrogeologie framework
   1. Water table and potentiometric maps for all aquifers.
   2. Hydrographe of groundwater head and surface water levels and discharge rates.
   3. Maps and cross sections showing the hydraulic conductivity and/or transmissivity distribution.
   4. Maps and cross sections showing the storage properties of the aquifers and confining beds.
   5. Hydraulic conductivity values and their distribution for stream and lake sediments.
   6. Spatial and temporal distribution of rates of évapotranspiration, groundwater recharge; surface water-groundwater interaction, groundwater pumping, and natural groundwater discharge.

Adapted from Moore, 1979

Data under category A, the physical framework, define the geometry of the system including the thickness and areal extent of each hydrostratigraphic unit. Hydrogeologic data include information on heads and fluxes (items B.l and B.2 in Table 1), which are needed to formulate the conceptual model and check model calibration. Hydrogeologie data also define aquifer properties and hydrologie stresses (items B.3-B.6 in Table 1).

Transferring field data to the grid

The first consideration in translating field data to the grid is to match parameter values to the scale of the model. For example, profile and full three-dimensional models require point measurements of hydraulic conductivity. Ideally, these data are obtained from point measurements of hydraulic conductivity in the field. Two-dimensional areal models and quasi three-dimensional models require vertically averaged values that may be obtained indirectly by averaging point measurements or directly from pumping tests in wells that fully penetrate the aquifer.

When the field data are determined to be compatible with the scale of the model, aquifer properties may be assigned to each hydrostratigraphic unit identified in the conceptual model. The grid is divided into zones so that certain sets of nodes have similar aquifer properties based on the areal extent of the hydrostratigraphic units. Thickness of each hydrostratigraphic unit is also assigned to each node. When hydrostratigraphic units are defined at a local scale, interfingering of two or more types of material may occur within a single cell or element. In this case, average properties for the cell or element are computed.

KRIGING

Assigning parameter values to the grid is difficult because the model requires values for each node, cell, or element and field data are typically sparse. Interpolation of measured data points can help in defining the spatial variability over the problem domain.

Kriging is the interpolation method used most frequently for this purpose. However, other interpolation methods may also be used (e.g., Williams and Williamson, 1989) or parameter values may be assigned to nodes using hydrogeologic judgment. Kriging is a statistical interpolation method that chooses the best linear unbiased estimate (BLUE) for the variable in question.

Kriging differs from other interpolation methods because it considers the spatial structure of the variable and provides an estimate of the interpolation error in the form of the standard deviation of the kriged values.

Once the conceptual model is translated into a numerical model in the form of governing equations, with associated boundary and initial conditions, a solution can be obtained by transferring it into a numerical model and writing a computer program (code) for solving it. This includes, design of grid, setting boundary and initial conditions and preliminary selection of values for aquifer parameters. The input parameters include model grid size, layer elevations, boundary conditions, hydraulic conductivity, recharge, and additional model input for steady state condition. Model calibration consists of changing values of model input parameters in an attempt to match field conditions within some acceptable criteria (Anderson and Woessner, 1992).

2.2.4 Calibration Process

Calibration of a flow model refers to a demonstration that the model is capable of producing field-measured heads and flows which are the calibration values (Fig.7). Calibration is accomplished by finding a set of parameters, boundary conditions, and stresses that produce simulated heads and fluxes that match field-measured values within a preestablished range of error (Fig.7). Finding this set of values amounts to solving what is known as the inverse problem. In an inverse problem the objective is to determine values of the parameters and hydrologie stresses from information about heads, whereas in the forward problem system parameters such as hydraulic conductivity, specific storage, and hydrologie stresses such as recharge rate are specified and the model calculates heads. Most classroom problems are formulated as forward problems but most field problems require solving an inverse problem.

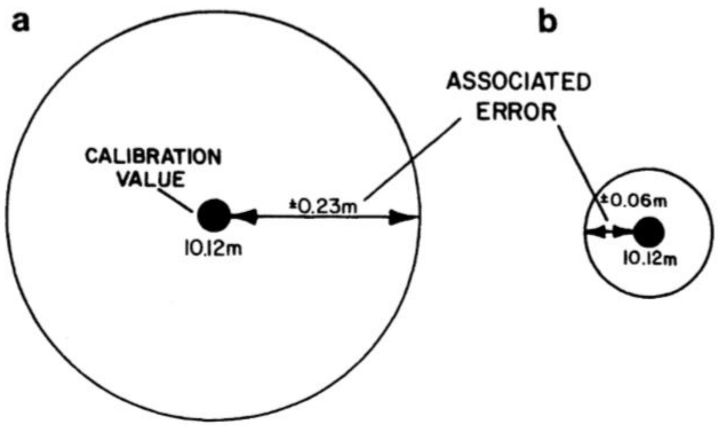

Figure 7:-A calibration targets is defined as a calibration value and its associated error.

2.2.6 CALIBRATION TECHNIQUES

There are basically two ways of finding model parameters to achieve calibration, i.e., of solving the inverse problem:

1. Trial-and-Error Calibration
2. Automated Calibration

Trial-and-Error Calibration

Manual trial-and error calibration was the first technique to be used and is still the technique preferred by most practitioners. In trial-and-error calibration, parameter values are initially assigned to each node or element in the grid. During calibration, parameter values are adjusted in sequential model runs to match simulated heads and flows to the calibration targets.

Prior to calibration, the range of uncertainty in each parameter value is quantified. Some parameters may be known with a high degree of certainty and therefore should be modified only slightly or not at all during calibration. The results of each model execution are compared to the calibration targets; adjustments are made to all or selected parameters and/or boundary conditions, and another trial calibration are initiated. Tens to hundreds of model runs are typically needed to achieve calibration. Trial-and-error calibration may produce non unique solutions when different combinations of parameters yield essentially the same head distribution.

Automated Calibration

Automated inverse modeling is performed using specially developed codes that use either a direct or indirect approach to solve the inverse problem.

Direct solution

In a direct solution, the unknown parameters are treated as dependent variables in the governing equation and heads are treated as independent variables. This means that values for head must be input for all nodes. Heads are known only at points where there are observation wells, making it necessary to estimate heads elsewhere in the grid, usually by kriging. The solution minimizes the nodal mass balance errors caused by using these heads and the computed parameter values. Direct solutions are prone to instability. Furthermore, they do not recognize measurement errors. According to Carrera (1988), it may be this failure to provide a statistical framework in which to view field data that has caused direct methods to be practically abandoned.

Indirect approach

The indirect approach is similar to performing trial-and-error calibrations in that the forward problem is solved repeatedly. However, an inverse code automatically checks the head solution and adjusts parameters in a systematic way in order to minimize an objective function, an example of which would be to minimize the sum of the squared residuals, i.e., differences between simulated and observed heads.

Although more stable than direct solutions of the inverse problem, indirect solutions may be unstable and give unreasonable solutions involving negative parameter values. It may be possible to control instability by proper zonation of aquifer parameters.

Indirect solutions are formulated in a statistical framework in which errors in heads and parameters are quantified. In a weighted least squares statistical framework, head measurements and prior information on parameter values are weighted to indicate the relative confidence in the measurement. In this way it is possible to place greater emphasis on measurements that are thought to be of higher reliability. The objective function becomes a weighted sum of the squared differences between observed and simulated heads and between initial and current parameter estimates. In this statistical framework, errors are assumed to be normally distributed and have zero mean.

To date, automated inverse models have had limited application. They are criticized because of problems with non uniqueness and instability, but according to Sampler et al. (1990) non uniqueness and instability "depend on the problem and the way it is posed, not on the approach used for calibration." Non uniqueness is especially a problem in the absence of prior information on transmissivities (Neuman et al.; 1980). Sampler et al. (1990) believe that the calibrated model produced with automated techniques is not necessarily superior to one produced using manual trial-and-error calibration. Rather, an advantage to using automated calibration codes is that they speed the modeler through the most time-consuming and frustrating part of the modeling process.

After model calibration, Sensitivity analysis is useful in determining which parameter or parameters most influence the model results. These parameters will be emphasized in the future data collection attempting to improve model accuracy.

CAPTER THREE

3 DESCRIPTION OF THE STUDY AREA

3.1Location

The study area is located in Amhara Regional State, South Wollo administrative zone; around Dessie town . Dessie is located 480km from Bahirdar and 401 km from Addis Ababa. Gerado catchment is situated in the south western part at 5 km from Dessie town. It is located in zone 37 UTM grids bounded between 552427 to 572575m E and 1210450m to 1243824m N.

The catchment is delineated based on surface water divide using topographic maps of scale 1:50,000. The catchment covers a total area of 250 km^2. There is one major asphalt road which runs from Addis Ababa to Mekelle and access to the Gerado catchment is possible through a gravel road constructed from Dessie to Mekaneselam.

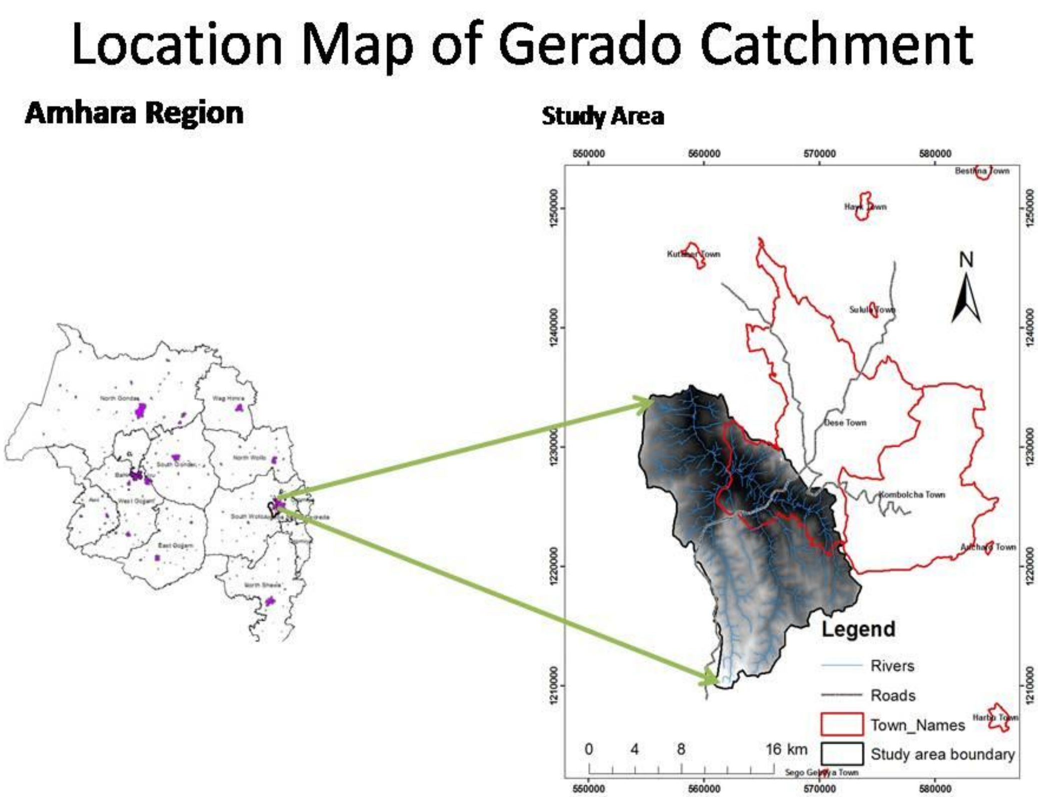

Figure 8:-Location Map

3.2 Physiography and Drainage

Gerado River catchment is situated in the Ethiopian highland plateau adjoining the western escarpment of the Rift valley and found in the Abay River basin. The general topography of the catchment is undulating hills, flat and valley. It gradually decreases in elevation to the north and north-west. Numerous narrow and shallow river valleys originated from mountain ranges and merge subsequently and form Gerado Perennial River. Intermittent rivers like Kelina, Yito and Negeweli are among the main tributaries of Gerado River. Gerado River flows to Beshlo River which is one of the major tributaries of Blue Nile River.

Plate 1:- Gerado flat land near Dessie-Mekaneselam road

Small valleys originated from ridges and hills form dense drainage pattern at the upstream and sides of the catchment where the slope are relatively steeper and become less dense at the flat land and gentle to flat land of the study area (Figure 9:- drainage map). These small valleys and streams are controlled by main faults, joints, fractures or a combination of these.

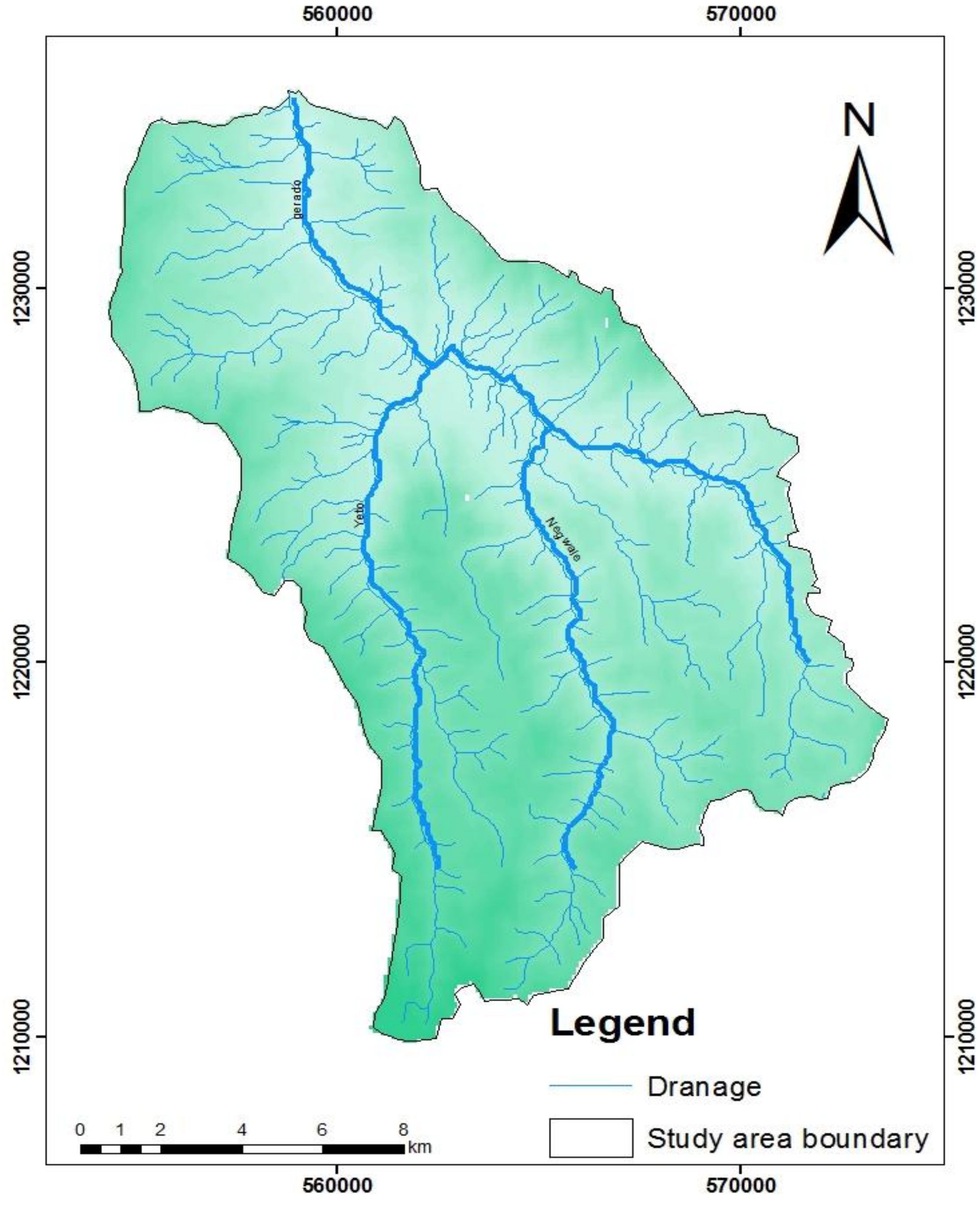

Figure 9:-Drainage map

Plate2:- The North West Physiography of Gerado river catchment

3.4 Climate

Based on Ethiopian climate classification (Danieal Gemechu, 1997), the study area is classified in the Woina Dega to Dega climatic zone having an altitude that ranges from 2100 to 3510m 3510m.a.s.l. with small rains in November to February and heavy rains in July to September. The maximum amount of mean annual rainfall goes up to 1297.1mm at Guguftu, and Kombolcha has the minimum 981.9mm. The study area has 15.32 ^{0}C mean annual temperature, May to July have the highest and November and December have the lowest. The mean monthly wind speed at Kombolcha station varies from 0.8-1.3 m/s as measured 2m above ground surface, June and July have the peak wind speed whereas October and November have the least.

Table 2:-Classification of Ethiopian climate

Altitude(masl)	Temperature (^{0}c)	Classification	Local Name
Below 500	25 and above	Hot	Bereha
500-1500	20-25	Hot Temperate	Kola
1500-2300	15-20	Temperate	Woina Dega
2300-3300	10-15	Cool temperate	Dega
3300 and above	10 or less	Cool	Kur

Source: Daniel Gemechu, 1997.

3.5 Land use and land cover

Land use is important in the hydrological and groundwater studies, because it is a prominent factor affecting the recharge. Based on the study conducted by Amhara National Regional State, Finance and Economic Development Bureau (FEDB, 2000) , Ministry of water resources study for the preparation of Abay basin master plan and the land use classification made using Land sate , the dominant land use and land cover of the study area are classified in to four:

Cultivation land: Cultivation land is highly concentrated on the flat, gentle to flat and gentle part of the study area. But the intensity of cultivation is decreasing down ward and along the mountains, where the soil type and the slope of the area are gradually changing in to the rocky type and steep slope respectively. The degree of cultivation especially on the flat and gentle to flat land of the study area is highly intensified, due to the presence of good groundwater potential, as a result the communities are producing different cereal crops throughout the year, which is the main supply for Dessie town and surrounding areas. But there are also some cultivation activities around the steeper parts which is not as much intensive as that of the flat land.

Grazing land: The coverage of grazing land is highly pronounced at the flat land of the study area, where it is covered by grass. But there is some grazing land covered by grass at the gentle to flat and gentle part of the study area.

Shrub: is a plant commonly characterized by vegetation dominated by shrubs, often also including grasses, herbs, and woody plants. Most of the shrubs in the study area are currently becoming deforested through the activities of the surrounding community for different purposes. Shrub lands in the study area are usually fairly open so grasses and other short plants grow between the shrubs.

Forest: Most of the forest found in the study area is eucalyptus and Habesha Tid**.**

Based on FAO land use land cover classification, the dominant land use land cover in the study area is classified in to four major groups called Cultivation, Grazing land, Shrub and plantation. In this specific research, for the determination of different hydrological parameters like actual evapotranspiration and surface runoff of the catchment, the FAO land use land cover classification is adopted.

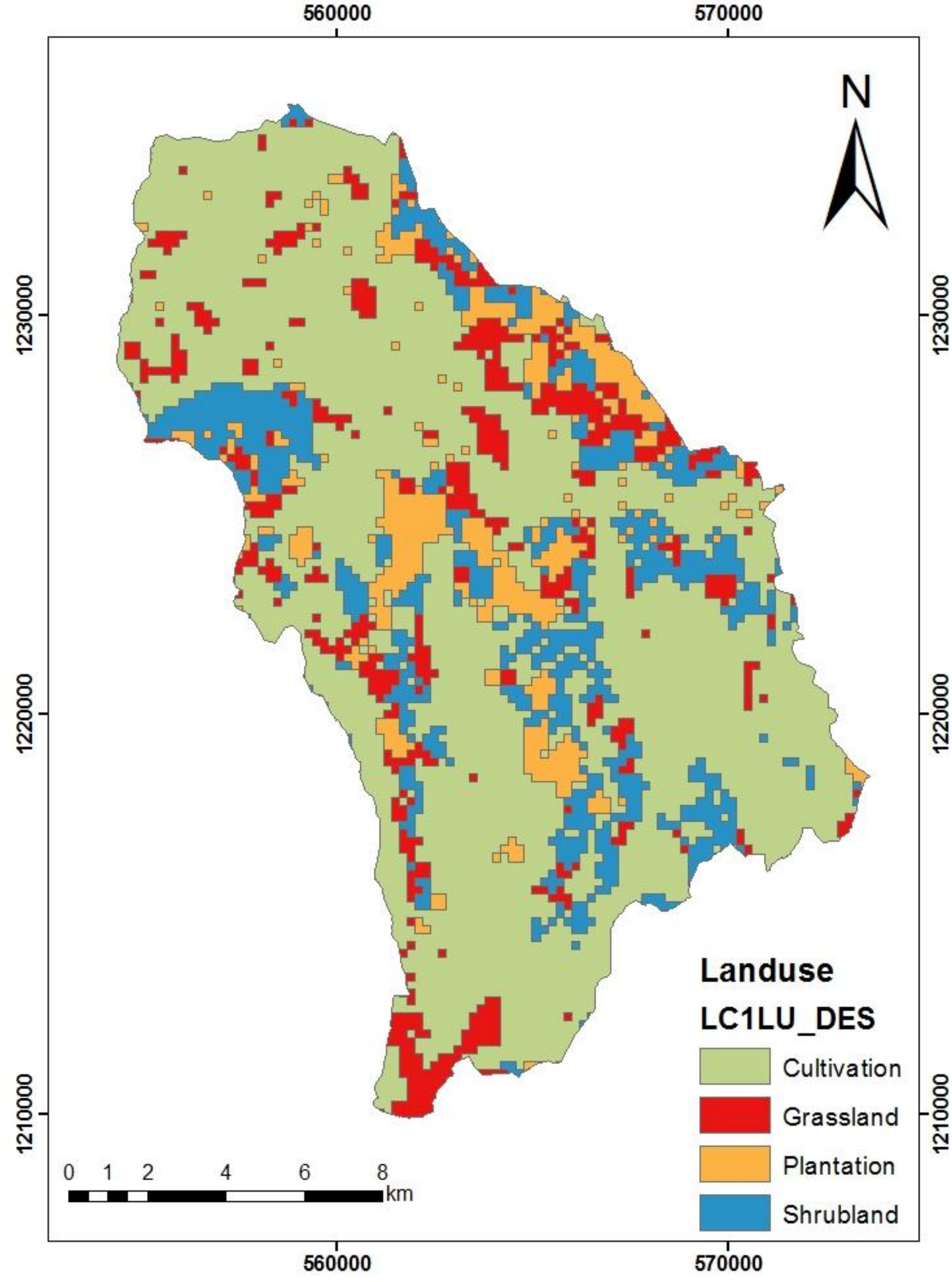

Figure 10:-land use land cover

3.6 Soil

Based on the actual field soil classification made using physical observations and finger testing method, the study conducted by Amhara National Regional state, Finance and Economic Development Bureau (FEDB, 2000) and the soil classification done by Ministry of water resource for the preparation of Abay basin master plan, texturally the dominant soil types of the study area are classified in to clay, sandy cay and rocky. But there are some none mappable soil types like boulders, cobbles, pebbles and gravel which are highly concentrated along the course of Yito, Negeweli and main Gerado (Dereje Gidafie, 2012).

Clay soil: The soil is typically black in colour but at the contact between other soil types the colour of the soil is becoming blackish brown. The soil is highly exposed on the left side of the river along flat, gentle to flat and gentle land of the study area. Such types of soil are highly covered with cultivation where the main cultivation activity is cereal crops, at the gentle to gentle flat land and grazing at the flat land of the study area.

Sandy clay*:* The soil is having brownish to black colour. The soil is highly exposed on the right side of the river along flat, gentle to flat and gentle land of the study area. Such types of soil are highly used for cultivation at the gentle, gentle to flat land and flat land of the study area.

Rocky: Such soil coverage is highly exposed along the steeper part of the catchment. The rocky soil, especially on the steeper part of the catchment is covered with shrubs and plantation like eucalyptus and cereal crops at the thin to very thin soil cover.

For the determination of different hydrological parameters like potential and actual evapotranspiration of the catchment of the study area FAO soil classification was used. Based on FAO soil map the dominant soil types in the study area are classified in to three major groups called Cambisols, lithosols and rock surface.

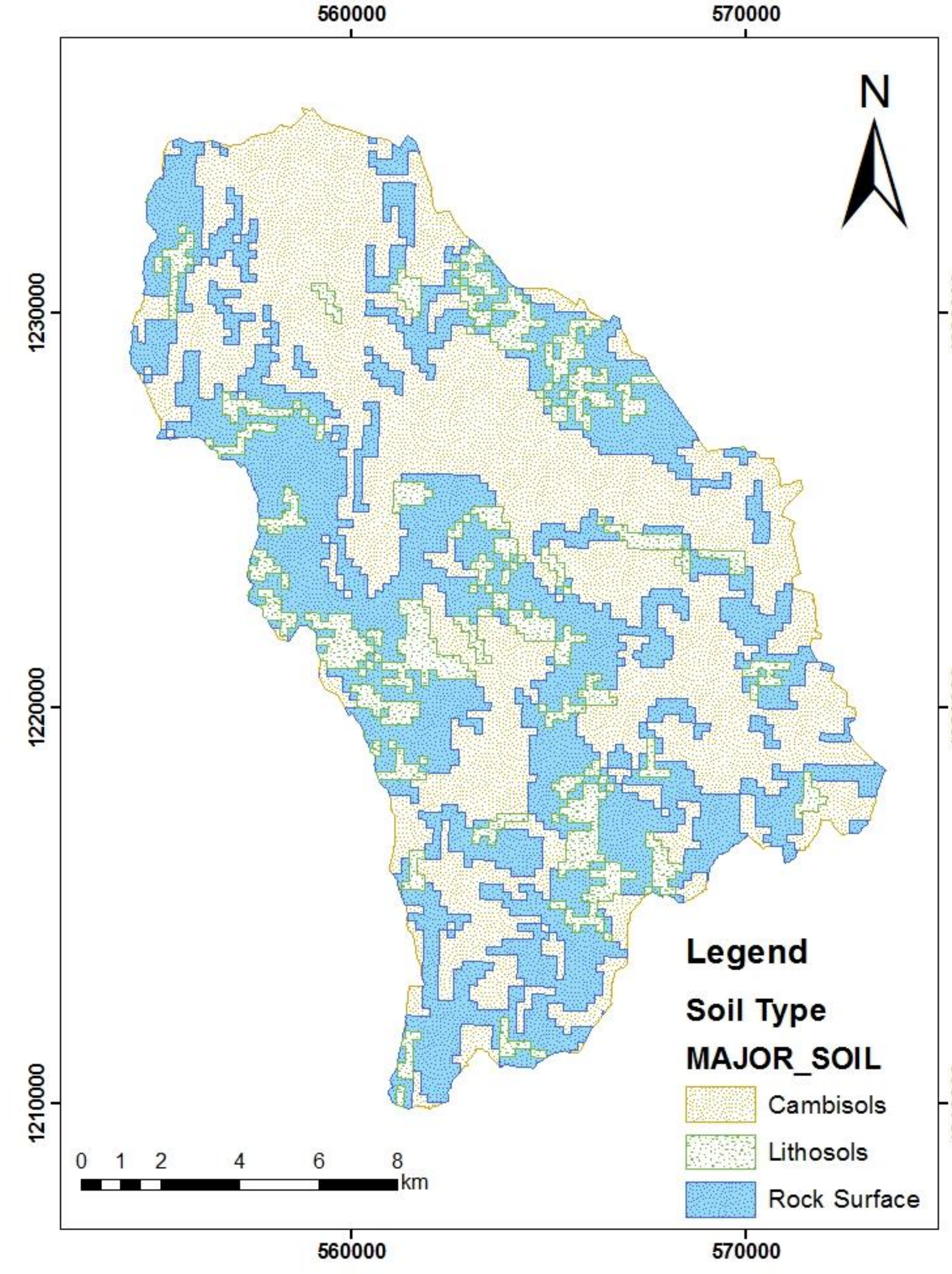

Figure 11:-Soil map of the study area

3.7 Geology of the study area

The previous hydro geologist researcher Dereje Gidafie (2012) identified the major litho logical units in the study area, volcanic rocks and alluvial deposits. The volcanic units cover 123.4km^2 of the catchment which is 49.36% of the total study area. The alluvial deposits account 126.6km^2 of the catchment which is 50.64% of the total study area.

3.7.1Volcanic Units

A. Lower Basalt

At the most north western part of the study area, the basalt is found at the river bed overlaid by ignimbrite. The basalt is black in color, fine grained, moderately weathered, highly jointed and vertically fractured, where the dominant orientations of the fractures are N-S and NNE-SSW. The thickness of the basalt ranges from 20-25 m.

Ignimbrite and Tuff

The ignimbrite is grayish pink in color, slightly weathered, jointed, vertically and irregularly fractured. The dominant orientations of the fractures are N-S and NNW-SSE. It is a coarse grained rock that contains phenocrysts of quartz and feldspars and the thickness of this lithology ranges from 4-5m. The ignimbrite is found overlain by tuff and other lithological units and is highly exposed on the northwest and at the outlet of the Gerado river catchment.

The tuff is whitish, highly weathered, horizontally, vertically and irregularly fractured. Its thickness ranges from 3-4 m. It is found exposed in the north western parts of the studied area.

Upper Basalt

The third main lithological unit overlying the ignimbrite and tuff is the upper basalt having more than two lava flows, where the first one is highly weathered, grayish black in color, moderately to course grained, highly fractured and overlain by unweathered (fresh) and highly fractured basalt.

The second lava unit overlying the highly weathered and fractured basalt is unweathered (fresh), blackish in color, and highly fractured. The fractures found in these lithological units are vertical and irregular in nature. This geological formation exposed continuously from Dessie southwards to Tossa Felana and the road from Dessie to Kombolcha.

3.7.2 Alluvial Deposit

This unit is exposed on the flat, and flat to gentle part of the study area (Plate 1).From field observation and lithological log of the boreholes found in the flat land, the alluvial deposit is composed of black cotton clay, silt, sand, rounded gravel and boulders. The thickness of the alluvium can reach up to 200m.

The alluvial deposits are becoming fine grained at the center of the catchment whereas at the base of the surrounding mountains and along the river courses, the deposit becomes coarse grained and sub-rounded and angular in shape. The alluvium covers 126.6 km^2 of the catchment which is 50.64 % of the total study area.

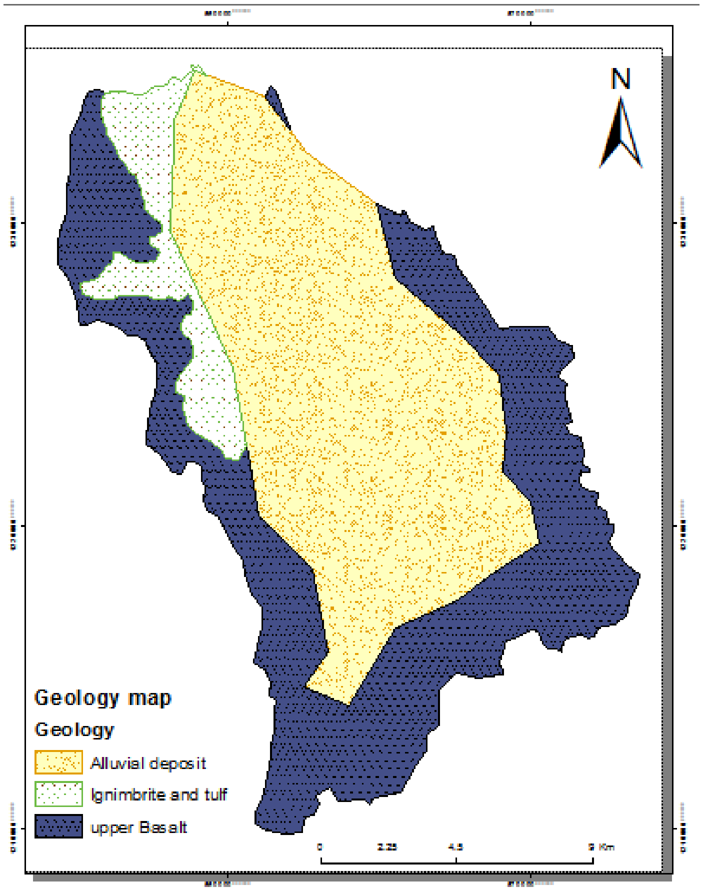

Figure 12:- Geological map (modified after Dereje Gidafie2012)

CHAPTER 4

4 Methodology

In order to achieve the objectives of the research project, the followings methods and Techniques were employed:-

4.1 Data Collection

Data collection is an important aspect of any type of research study. Inaccurate data collection can impact the results of a study and ultimately lead to invalid results. Basically data is classified as primary and secondary data. Even though, for research purpose the collection of primary data is thought to give the research paper a more reliable outcome. With this intention river width and height at different part of the river were measured to determine the physical boundary, borehole information collected and current conditions of wells seen for each well during the field trip in the study area.

The secondary data such as metrology data, borehole logs, pumping test data, and topographic map, geological and hydrogeological map of the study area and pertinent documents like, different studied documents around the study area, well completion report were collected from different organizations (water works and supervision enterprise, Ethiopian mapping agency, Dessie Water Supply Project Office, Kombolcha metrological agency and Ethiopian Geological Survey) giving particular attention on the quality of documents.

Delineating of the study area was first done by extracting the catchment from topographic maps with a scale of 1:50,000 and Digital Elevation Model (DEM 90).

4.2 Data Analysis

In order to determine the recharge of the study area, identification and classification of the major land use, soil and slope of the catchment were done and important meteorological data were collected from Kombolcha meteorological station. Finally various thematic maps like drainage, land use, soil, and slope were produced and recharge was determined.

Areal depth of rainfall of the study area was determine using three methods called Arithmetic, Thiessen polygon and Isohyetal methods:

Arithmetic method:

$$P = \frac{P1+P2+P3+\cdots Pn}{n} \qquad \text{Eq 4.1}$$

Where P is the average depth of precipitation of the area, P1, P2, P3 and Pn are the rainfall records at the stations 1, 2, 3 etc and n is the number of meteorological stations.

Thiessen Polygon Method:

$$P = \frac{P1A1+P2A2+P3A3\ldots\ldots\ldots PnAn}{At} \qquad \text{Eq 4.2}$$

Where P is areal average precipitation, P1, P2 ... Pn are mean annual rainfall recorded at each rainfall stations and A1, A2...An are polygonal areas around each gauging stations enclosed in the catchment and At is the total area of the catchment.

Isohyetal Method:

$$P = \frac{P1A1+P2A2+P3A3\ldots\ldots\ldots PnAn}{At} \qquad \text{Eq4.3}$$

Where P is areal average precipitation, P1, P2 ... Pn are mean annual rainfall calculated between consecutive isohyets and A1, A2...An etc. is enclosed area found between Isohyets and At is the total area of the catchment.

Thornthwaite method (Thornthwaite and Mather, 1957) was used to estimate the potential evapotranspiration of the catchment. This method uses air temperature as an index of the energy available for evapotranspiration, assuming that air temperature is correlated with the integrated effects of net radiation and other controls of evapotranspiration, and that the available energy is shared in fixed proportion between heating the atmosphere and evapotranspiration (Dunne and Leopold, 1978).

The Thornthwaite empirical equation is given by (Shaw, 1988):

$$PETm = 16\ Nm\ (10Tm/I)^{a}\ mm \qquad \text{Eq 4.4}$$

Where

- PET is Potential evapotranspiration in mm/month
- m is the months 1, 2, 3……12
- Nm is the monthly adjustment factor related to hours of daylight. It is found from standard table by dividing possible sunshine hours for the appropriate latitude (10^0 N) of the study area (Gerado river catchment) by twelve,
- Tm is the monthly mean temperature in ^{0}C,
- I is the heat index for the year, given by:

  $I = \Sigma im = \Sigma(Tm/5)^{1.5}$ for m=1... 12 and Eq 4.5

  And a is given by

  $a=6.7x10^7I^3 - 7.7\ x10^{-5}\ I^2 + 1.8x10^{-2}I+0.49$ Eq 4.6

For the determination of the actual evapotranspiration the Thornthwaite soil water balance model (Dunne and Leopold, 1978) was utilized. The required parameters to determine actual evapotranspiration using this model are mean monthly precipitation, mean monthly potential evapotranspiration, water holding capacity of the dominant soil type and monthly soil moisture storage.

To evaluate and check for the shortage or addition of moisture to the soil, the difference between precipitation and potential evapotranspiration is determined. As a result, positive values are indicatives of the addition of moisture to the soil while negative values are showing the monthly demand of moisture by the vegetation which is not satisfied by the monthly rainfall.

Accumulated potential water loss, which is obtained by cumulating the negative values of the differences between monthly precipitation and evapotranspiration, is used to calculate soil moisture for the dry months.

$Sm=W*\exp(-APWL/W)$ Eq 4.7

Where Sm is the soil moisture during the month m (mm), APWL is the accumulated potential water loss and W is the available water capacity of the root zone (mm).

The actual evapotranspiration,

$AET = PET$ Eq 4.8

If the mean Precipitation of the month is greater than the respective Potential evapotranspiration, other wise

$$AETm = Pm + Sm - S\,m - 1 \qquad \text{Eq 4.9}$$

Where m stands for month, Sm and Sm-1 and are soil moisture during the month m (current month) and m-1 (earlier month) respectively.

The soil moisture deficit

$$SMD = PET - AET \qquad \text{Eq 4.10}$$

The soil moisture surplus which is in excess of soil moisture values(Sm) especially in wet season and it is given by

$$Sm = P - AET + \Delta Sm \qquad \text{Eq 4.11}$$

The actual evapotranspiration, AET, for the dominant soil types and the respective land use in the area was weighted according to the proportion of the area it represents, and the weighted AET was calculated as:

$$AETw = \sum \frac{(AETi)ai}{At} \qquad \text{Eq 4.12}$$

Where

- AETw is weighted actual evapotranspiration:
- AETi is actual evapotranspiration of the dominant soil types:
- ai is area of each soil coverage: and,
- A is total area of soil coverage

The volume of runoff from the catchment was also computed by using the runoff coefficient method, which employed the following formula (Garg, 1987),

$$Q = K.P.A \qquad \text{Eq 4.13}$$

Where

- Q is runoff, m^3:

- K is a constant also called runoff coefficient depends upon the imperviousness of the drainage area:
- P is precipitation (mm): and,
- A is area of the catchment (m^2).

For the estimation of groundwater recharge of the studied area, water balance method was utilized. For the estimation of groundwater recharge of the studied area, water balance method was utilized.

Inflow =Outflow ± Change in storage Eq 4.14

$$P + Gi = AET + SRO + R + Go \pm \Delta S \qquad \text{Eq 4.15}$$

Where,

- P = Annual precipitation
- Gi =Groundwater inflow
- AET = Actual evapotranspiration
- SRO = Annual surface run-off
- R = Groundwater recharge
- Go = Groundwater out flow and
- ΔS = Change in water storage

Various assumptions have been made to derive the water balance equation for the studied area and these are summarized below:

1. Surface and Subsurface water exchange with neighboring catchment is assumed to be zero.
2. Since the computation is made on annual basis, net change of soil moisture and groundwater storage is assumed to be zero.
3. Assuming no artificial diversion from other catchments.

Based on the assumptions made above, the general water balance equation is reduced to

$$P - AET - SRO - R = 0 \qquad \text{Eq4.16}$$

Where

- AET = Evapotranspiration
- R = Recharge
- P = Precipitation and
- SRO = River discharge from the catchment (Surface Runoff).

4.3 Conceptual Model Development

Developing the appropriate conceptual model for a given problem is one of the most imperative steps in the modeling process. Over simplification may lead to a model that lacks the required information, while under simplification may result in the lack of data required for model calibration. A conceptual model describes how water enters an aquifer system, flows through the aquifer system and leaves the aquifer system. Briefly, it describes the hydrologic system with respect to aquifer properties, flow characteristics and boundary conditions. According to Anderson and Woessner (1992) there are three steps in building a conceptual model: defining hydrostratigraphic units, preparing a water budget and defining the flow system.

Although, there was very limited data particularly for the highland part, a simplified conceptual model was developed for the groundwater flow system in Gerado River catchment. To develop the conceptual model, some simplifying assumptions were made. The assumptions include: the model consists of a single layer, the model is two dimensional, and the aquifer is assumed to be unconfined as the main flow to the alluvial aquifers is from adjacent volcanic aquifers with some direct recharge in the alluvium. There are few clay layers in the alluvial aquifers which confine the water below the clay layers in around the wells. However, this clay bed is not continuous and doen't disrupt flow from volcanic aquifers to the alluvial aquifers and even vertically in the alluvial aquifers as the clay is silty to loamy clay in nature which allows water to pass somehow. Moreover, pump positions are below the confining layers and depth to water table in some wells can drop below the confining layer in which case the storage is assumed to be unconfined.

Simplification is important because complete reconstruction of the filed system is not feasible. The conceptual model should be simplified as much as possible while it is still remains complex enough to represent the system behavior (Anderson and Woessner, 1992). In this study, to simplify the complex nature of the Gerado river catchment, a simplified conceptual hydrogeological model of the groundwater system was developed based on information about geology, hydrogeology and hydrology.

4.3 Numerical Groundwater flow modeling

General Concept and Modeling Approach

Numerical ground water flow modeling helps to have a good understanding of the current or to predict the long term tendencies of a hydrogeological system and it allows an analysis of the movement of water through hydrogeologic unit that constitute the groundwater flow system.

It is mandatory to have good initial data on boundary conditions, fluxes and aquifer hydraulic parameters for a model to give simulation output that approaches the real situation. In other words, models can only be good if the input data is good enough. Especially, input parameters that have the most control on the model output have to be carefully investigated and correctly estimated.

Numeric models describe the entire flow field of interest at the same time providing solutions for as many data points as specified by the user. The area of interest is subdivided into many small areas (usually referred to as cells or elements) and a basic ground water flow equation is selected to solve for each cell. The solution of a numeric model is the distribution of hydraulic heads at points representing individual cells.

The Gerado River catchment ground water flow model developed in this thesis was simulated to study the response of the system to different hypothetical scenarios of withdrawal and recharge.

The approach followed to develop this numerical model includes definition of system boundaries, estimation of well withdrawal, hydraulic conductivities, collection of water level data, selection of an appropriate computer code/governing equation for simulation, calibration of calculated heads/fluxes to field observed heads/fluxes and simulation under different scenarios to understand the response of the system.

Governing Equation and Model Code

Based on Darcy’s law and the conservation of mass (McDonald and Harbaugh, 1988), the movement of groundwater through porous media is described by the following partial differential equation:

$$\frac{d}{dx}\left(Kxx,\frac{dh}{dx}\right)+\frac{d}{dy}\left(Kyy,\frac{dh}{dy}\right)+\frac{d}{dz}\left(Kzz\frac{dh}{dz}\right)-W=Ss\frac{dh}{dt} \qquad \text{Eq4.17}$$

Where:

- *Kxx*, *Kyy*, and *Kzz* :-are values of hydraulic conductivity in the x, y and z directions along Cartesian coordinate axes, which are assumed to align with principal directions of hydraulic conductivity
- *h* is hydraulic head
- R is a general sink/source term that is defined to be intrinsically positive to represent recharge and negative for withdrawals of groundwater
- *Ss* is the specific storage of the porous material and
- *t* is time

Equation (4.17) describes the distribution of hydraulic head and flow throughout a continuous region. Derivations of equation (4.17) can be found in Freeze and Cherry (1979) and Anderson and Woessner (1992).

In order to simulate the unconfined aquifer of upper Gerado River catchment equation (4.18) is used which is the Boussinesq Equation.

$$\frac{d}{dx}\left(Kx\mathrm{h}\frac{dh}{dx}\right)+\frac{d}{dy}\left(Ky\mathrm{h}\frac{dh}{dy}\right)=\mathrm{Sy}\frac{dh}{dt}-W \qquad \text{Eq4.18}$$

It is assumed that $T_x = K_x h$ and $T_y = K_y h$, where h is the saturated thickness, and S_y is the specific yield. Available data are limited to horizontal properties in aquifers and no relation can be established regarding the anisotropy of units. Thus, for this study, a hydraulic property within the layer is assumed isotropic. Consequently K_x and K_y are considered to be equal at any given location and K_x and K_y are replaced in this discussion by the single term K to describe horizontal hydraulic conductivity. Since the study is on a Steady State condition there is no change in head with time, therefore, this part of the right hand side of equation (4.18) becomes zero and it can be re-written as:

$$\frac{d}{dx}\left(Kx\mathrm{h}\frac{dh}{dx}\right)+\frac{d}{dy}\left(Ky\mathrm{h}\frac{dh}{dy}\right)-W=0 \qquad \text{Eq4.19}$$

This equation describes the distribution of hydraulic head and flow throughout a continuous region. It is continuous in space and time, and generally cannot be solved analytically for practical applications involving complex system (Anderson and Woessner, 1992). Practically, the continuous system described in the above equation is replaced by a set of spatially and temporally discrete points using numerical methods, which form a set of simultaneous algebraic equations that describe the distribution of hydraulic head at each point and flow through the system in response to this head distribution. These simultaneous equations are set up in a matrix form and then solved.

A computer code or program solves a set of algebraic equations generated by approximating the partial differential equations that form the mathematical model. In this study, ground water flow simulation was done by using the computer code, MODFLOW (McDonald and Harbaugh, 1988). As applied in this study, the program uses a block centered, finite difference method to simulate groundwater flow through porous media in a horizontal grid.

A MODFLOW consists primarily of a set of input files that contain information on the physical properties of the modeled system such as the geometry, boundary conditions, internal properties (like distribution of hydraulic conductivity) and sources/sinks such as ground water recharge, Rivers and pumping wells (Anderson and Woessner, 1992). Once these files are created, the model program is run to solve a set of equations that describe the distribution of head at discrete points within the system and flow in response to that head distribution. This numerical method requires that the modeled domain be divided into discrete volumes, known as cells and the properties of material in each cell is assumed to be homogeneous.

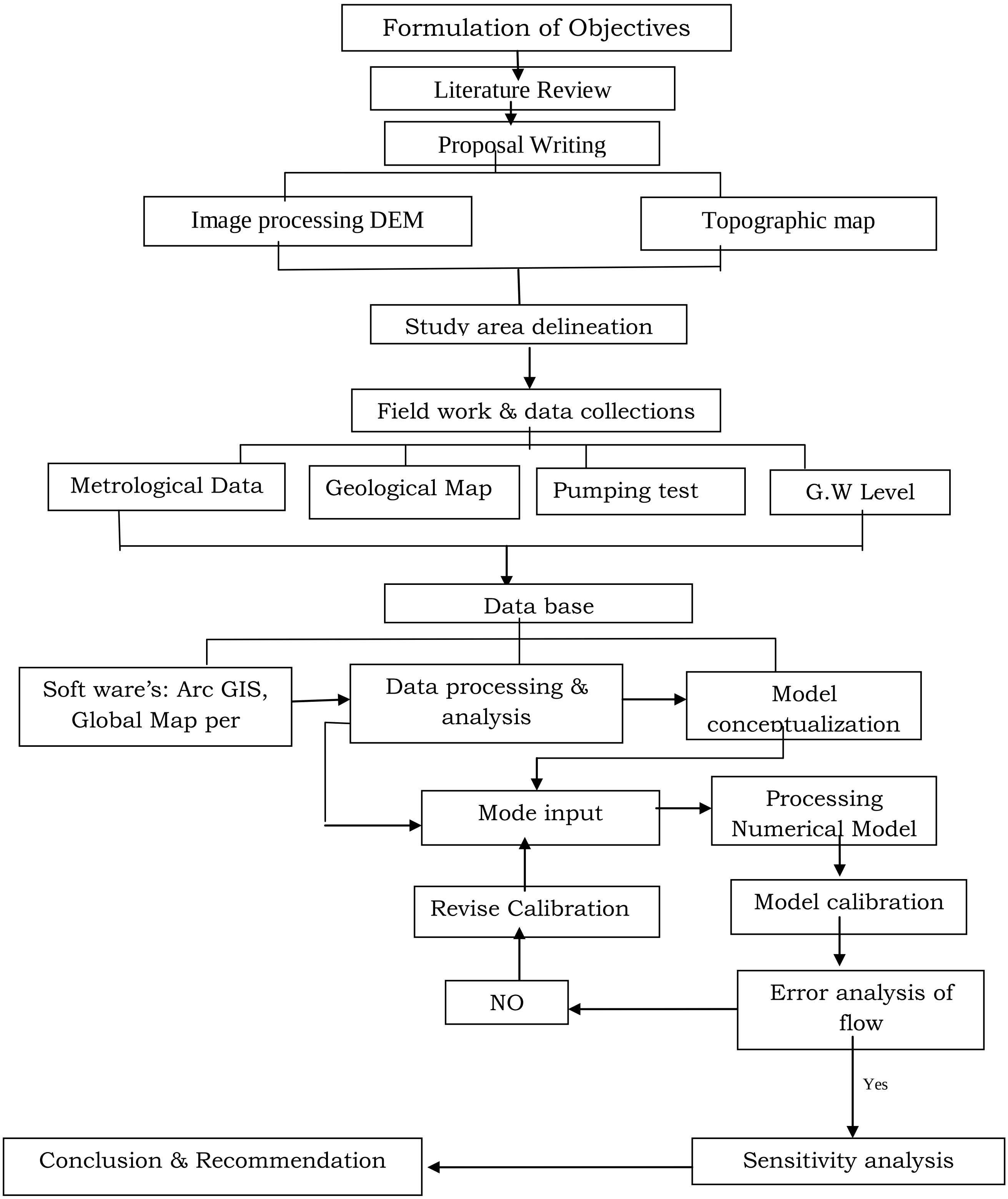

Figure 13:- Flow chart of the methodology

CHAPTER FIVE

5 RESULT AND DISCUSSION

5.1 Hydrometeorology

Long term measurements of meteorological data such as precipitation, temperature, wind speed, relative humidity and sunshine hours are essential to determine the water balance of a certain catchment.

To determine the various water balance components of the study area, important long term meteorological data have been taken from Kombolcha Meteorological Service Agency.

The meteorological station within the study area and the vicinity is displayed in the following table and currently all the stations are working properly.

Table 3:-Meteorological stations within and around Gerado river catchments

Stations	Location (UTM)/ADINDAN		
	Easting	Northing	Elevation
Dessie	569342	1231348	2400
Guguftu	553039	1203569	3450
Kutaber	558212	1245405	2600
Kombolcha	576448	1227015	1825
Tita	573693	1233807	2480

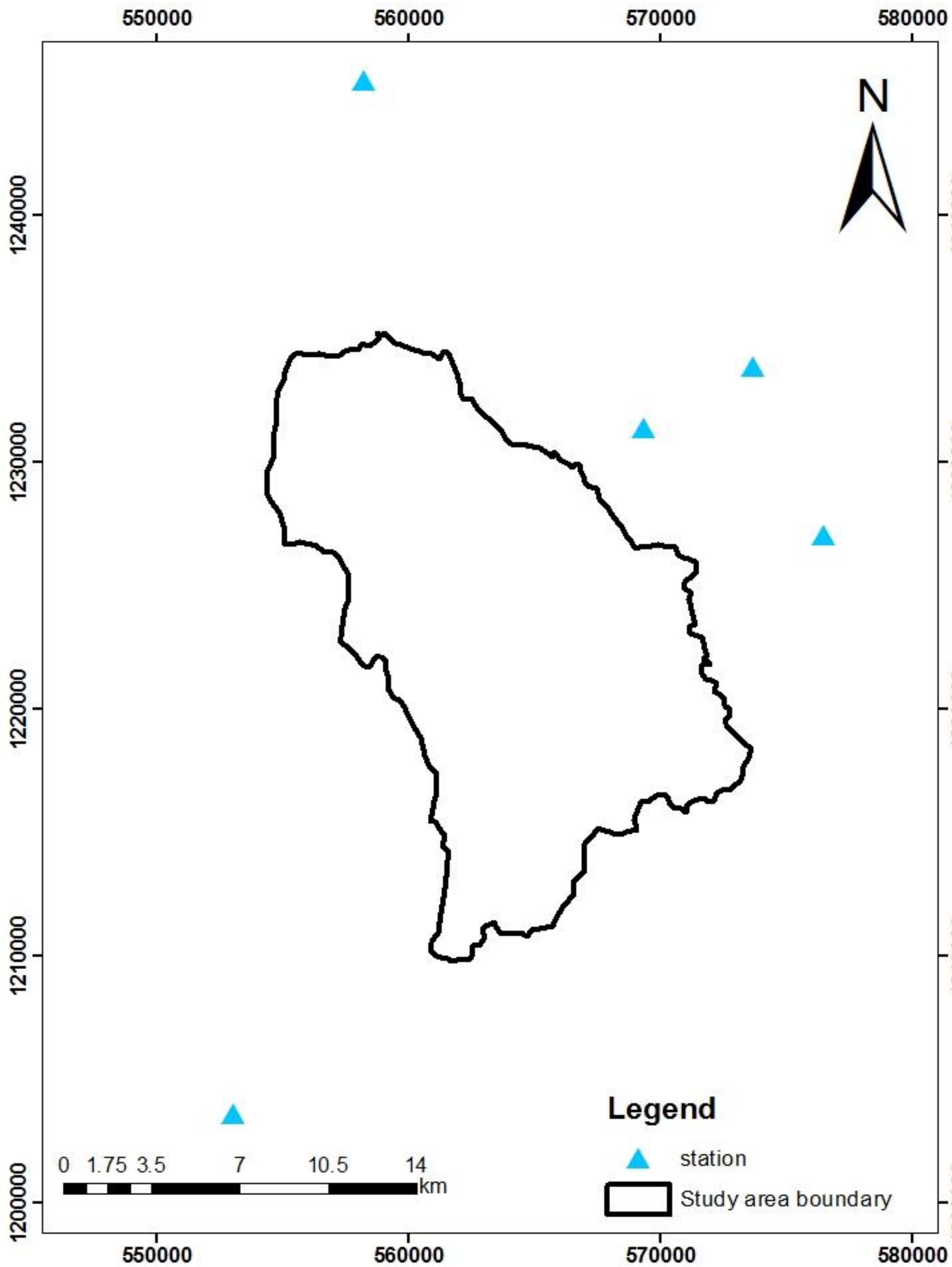

Figure 14:- Meteorological stations found within and around the study area

Kombolcha, Guguftu, Dessie and Tita meteorological stations record rain fall and temperature, whereas Kutaber meteorological station record only rainfall.

5.1.1Rainfall

5.1.1.1Determination of areal depth of rainfall

Rainfall measurement is a point observation and may not be used as a representative value for the area under consideration. Therefore, it is necessary to obtain effective uniform depth of rainfall of the catchment to get a more reliable and representative results. Areal depth of rainfall in the catchment is estimated by using simple arithmetic mean, Isohyetal and Thiessen polygon methods.

Arithmetic mean

The arithmetic mean method is the simplest method of determining areal average rainfall. It involves averaging the rainfall depths recorded at a number of gauges. For the determination of mean monthly rainfall of the study area, five meteorological stations i.e. Dessie, Guguftu, Kombolcha, Tita and Kutaber meteorological stations have been used. Based on equation 4.1, the mean annual rainfall of the Gerado river catchment is 1114.2mm (Table 4).

Table 4:-Arithmetic method to calculate the annual rainfall of Gerado river catchment

Stations	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec	annual
Dessie	18.4	22.9	50.5	84.3	60.7	35.5	303.6	324.1	93.1	52.1	18.6	20.0	1083.7
Guguftu	22.9	21.2	68.8	74.5	65.6	67.0	428.3	391.0	93.8	34.2	13.8	15.9	1297.1
Kutaber	11.7	11.2	52.6	52.7	52.7	43.7	354.8	341.8	122.7	42.8	19.0	5.7	1111.6
Kombolcha	19.2	10.4	59.4	80.5	42.2	25.9	304.7	289.7	79.4	37.9	17.7	14.9	981.9
Tita	24.4	27.7	74.1	97.5	66.8	34.7	307.5	309.2	79.3	42.3	16.5	16.9	1096.8
mean	19.3	18.7	61.1	77.9	57.6	41.4	339.8	331.2	93.7	41.9	17.1	14.7	**1114.2**

Thiessen Polygon Method

The Thiessen Polygon method of determining areal depth of precipitation assumes that at any point in the area under consideration, the rainfall is the same as that as the nearest gauge so the depth recorded at a given gauge is applied out to a distance halfway to the next station in any direction. The relative weights for each gauge are determined from the corresponding areas created in a Thiessen polygon network; the boundaries of the Polygons are formed by the perpendicular bisectors of the lines joining adjacent gauges (Chow, 1988). This method provides for none uniform distribution of rain gauge by determining a weighted factor for each gauge and it is generally more accurate than the arithmetic mean method but it is not particularly good for mountainous areas, since altitudinal effects are not allowed for by the areal coefficients, nor is it useful for deriving areal rainfall from intense local storms.

Based on equation 4.2, the annual mean rainfall of Gerado river catchment computed by Thissen polygon method is 1087.3 mm (Table 5 and Figure 15).

Table 5:-Thiessen polygon method to calculate the annual rainfall of Gerado River catchment

Stations	mean rain fall	Area of influence (km^2)	Area Wtd	Weighted Area (%)	Weighted rainfall (mm)
Dessie	1083.7	169.0	0.676	67.6	732.58
Guguftu	1297.1	26.2	0.105	10.5	136.2
Kutaber	1111.6	6.8	0.027	2.7	30.01
Kombolcha	981.9	48	0.192	19.2	188.53
		Total=250		100	1087.3

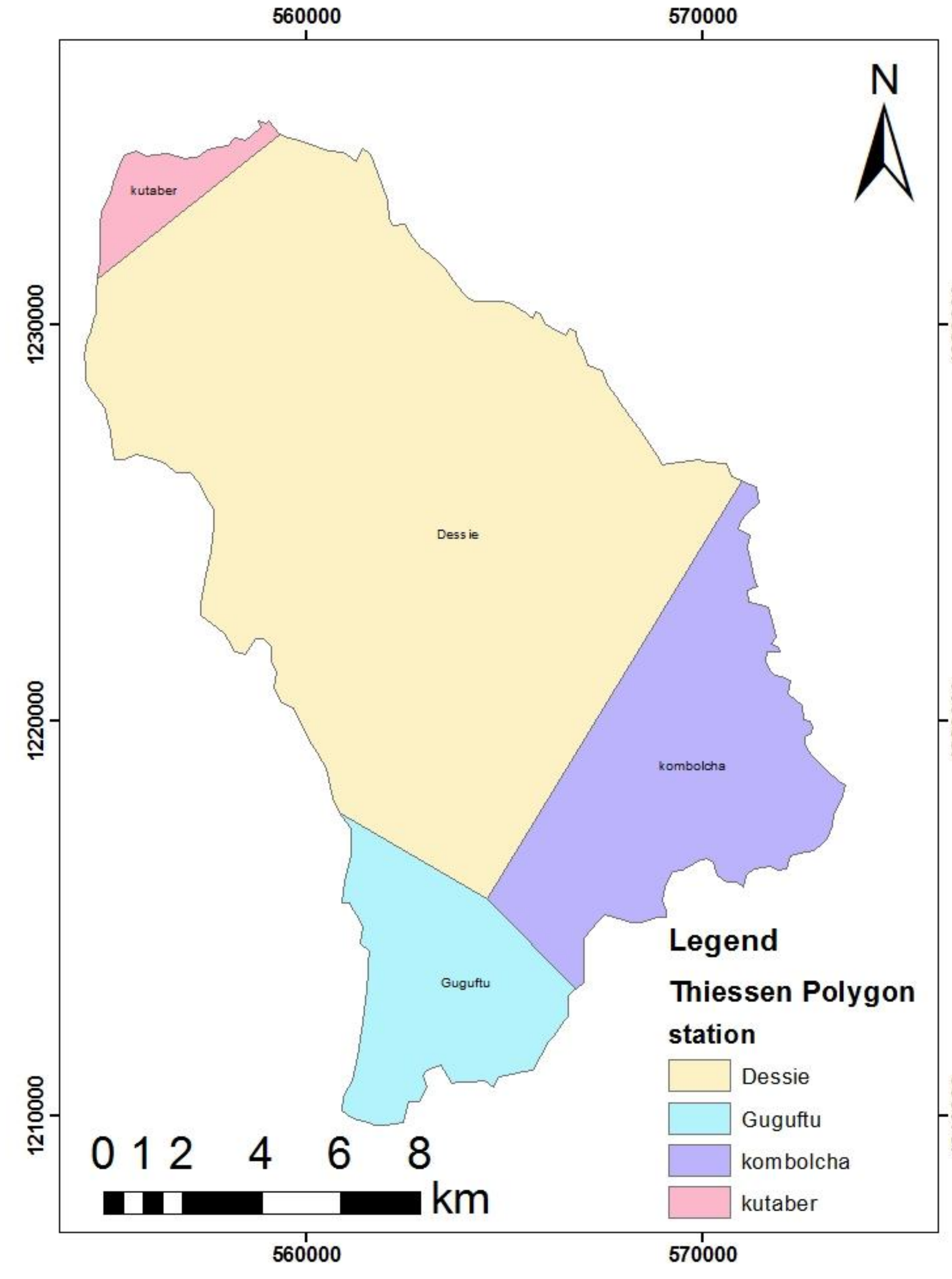

Figure 15:-Thiessen Polygon map of the study area

Isohyetal Method

This method takes in to account the influence of physiographic parameters which includes elevation, slope and distance from the coast and exposure to rain bearing winds (Shaw, 1988). This method is possibly the best of the three and has the advantage that the isohyets may be drawn to take account of local effects like prevailing wind and uneven topography (Wilson, 1990). It is employed by drawing contours of equal areal depth of precipitation and then measures their inter-Isohyetal area.

Based on equation 4.3, the annual mean rainfall of Gerado river catchment computed is 1136.18 mm (Table 6 and Figure 16).

Table 6:- Isohyetal method of calculating annual rainfall of Gerado river catchment

No	Isohyetalrainge	Average Isohyetal	Area (km ^{2})	Weight Area	Weight area %	Weighted rainfall (mm)
1.00	<1060	1050.00	0.50	0.00	0.20	2.10
2.00	1060-1080	1070.00	7.20	0.03	2.88	30.82
3.00	1080-1100	1090.00	27.26	0.11	10.90	118.85
4.00	1100-1120	1110.00	70.95	0.28	28.38	315.02
5.00	1120-1140	1130.00	58.80	0.24	23.52	265.78
6.00	1140-1160	1150.00	30.45	0.12	12.18	140.07
7.00	1160-1180	1170.00	17.72	0.07	7.09	82.93
8.00	1180-1200	1190.00	12.54	0.05	5.02	59.69
9.00	1200-1220	1210.00	9.15	0.04	3.66	44.29
10.00	1220-1240	1230.00	7.76	0.03	3.10	38.18
11.00	1240-1260	1250.00	6.28	0.03	2.51	31.40
12.00	1260-1280	1270.00	1.39	0.01	0.56	7.06
		Total	250.00		100.00	1136.18

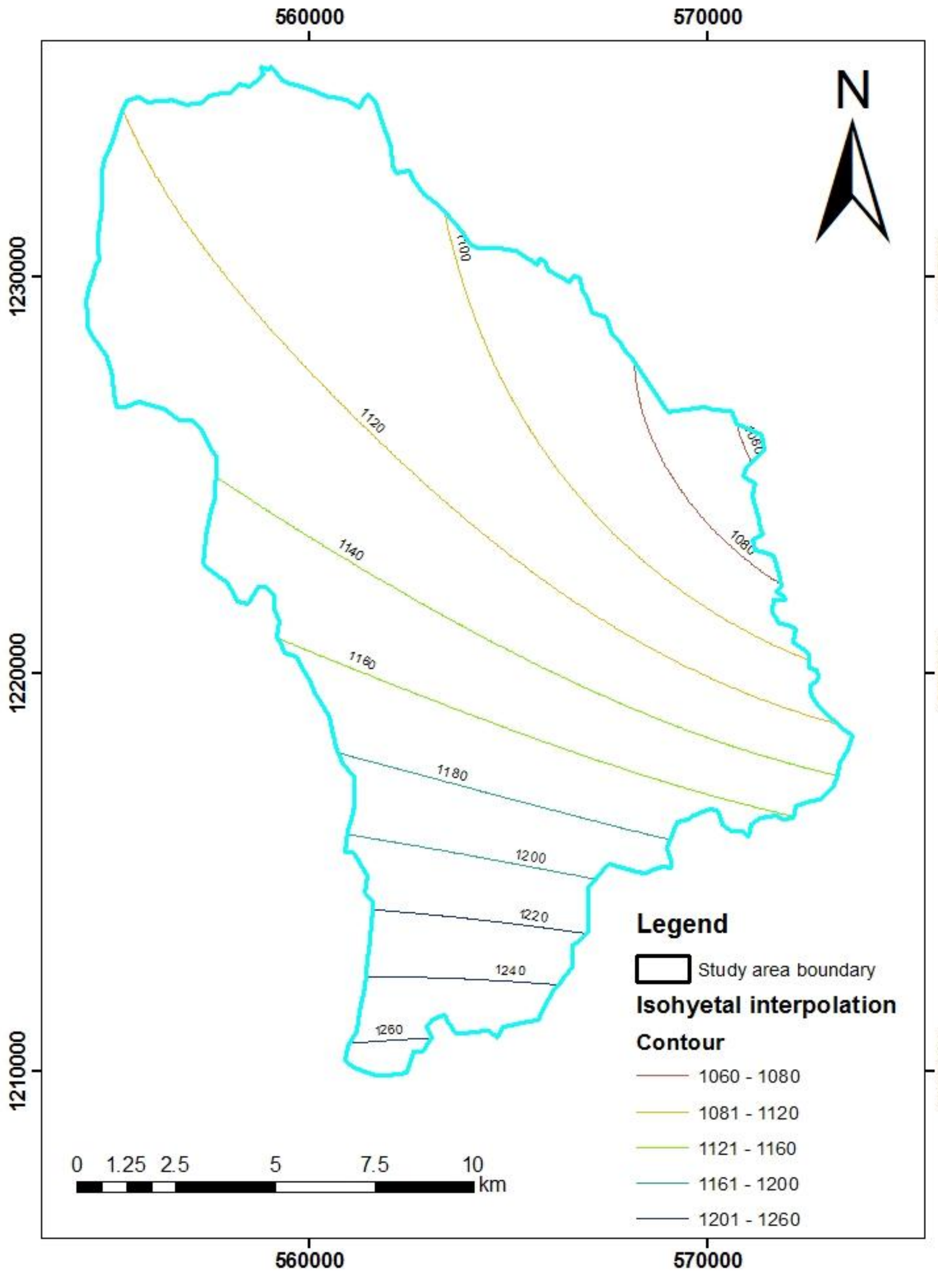

Figure 16:- Isohatal map of the study area

From three methods, which are used for the determination of mean annual rain fall of the study area, the value which is determined by Isohyetal method is used for the determination of actual evapotranspiration, as it considers different physiography effects.

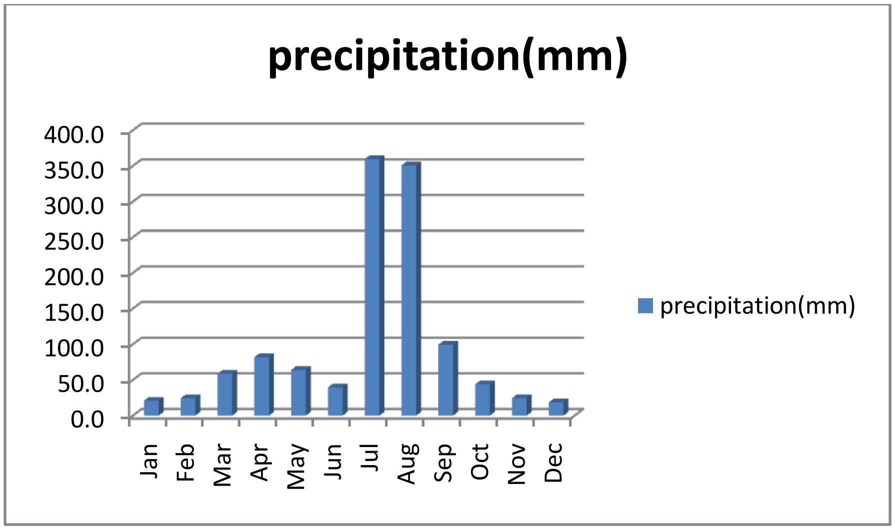

Figure 17:-Monthly rainfall distributions of Gerado River catchments

5.1.2. Temperature

Water temperature and air has straightforward effect on evaporation by developing the environment hot and allows the passage of liquid state of water to vapor condition. According to Shaw, 1988 the higher the air temperature, the greater extent of water vapor it can carry, and in the same appearance if the temperature of evaporating water is high, it can more readily vaporized.

From five meteorological stations used for this work but four stations (Kombolcha, Dessie Guguftu and Tita) have records of monthly maximum and minimum temperature. The mean annual temperature of the Gerado River catchment is 15.32 (See table 7).

Table 7:- Average temperate of Gerado River Catchment

station	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec	Annual
Tita	13.63	14.70	15.63	16.47	17.53	18.61	17.77	17.25	16.37	14.80	13.72	13.10	15.80
Guguftu	9.59	9.68	10.13	10.50	10.57	10.38	9.75	9.74	9.83	9.19	8.84	9.13	9.78
Kombolcha	17.5	18.6	19.8	20.5	21.5	22.8	21.7	20.9	20.1	18.3	17.0	16.8	19.63
Dessie	13.8	14.5	16.0	16.8	17.4	18.6	18.2	17.5	16.7	15.4	14.6	13.4	16.06
mean	13.61	14.37	15.39	16.08	16.75	17.61	16.84	16.34	15.76	14.42	13.54	13.11	15.32

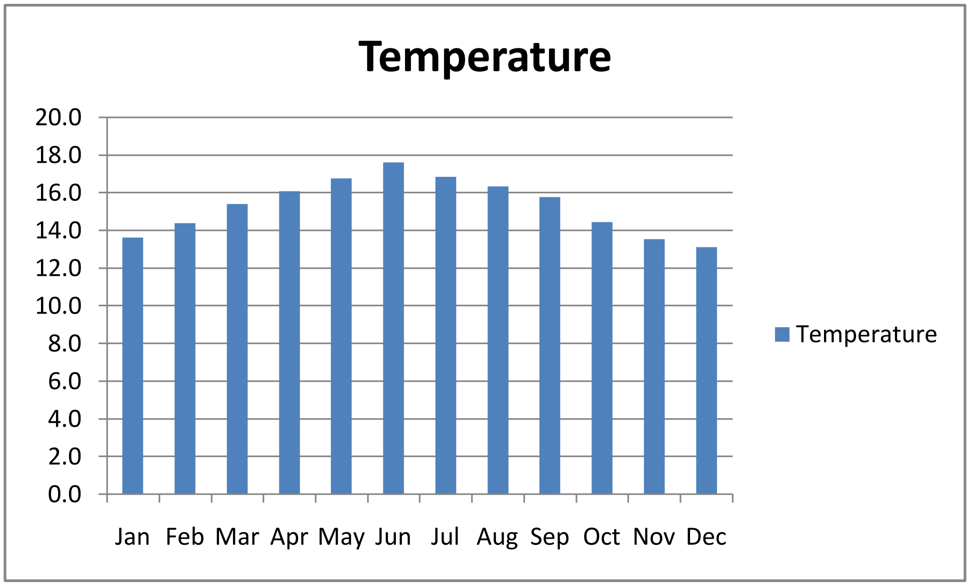

Figure 18:-Mean Monthly temperature distributions of Gerado River catchments

5.1.3Evapotranspiration

Evapotranspiration (ET) gives a detailed account of the discharge of water to the atmosphere from vegetations and land surface.

Estimation of potential evapotranspiration (PET)

Potential evapotranspiration is evapotranspiration from vegetation cover if sufficient water is supplied to obtain optimum growth or the maximum amount of vapor which might be transferred under existing metrological condition, water is not the limiting factor. Potential evapotranspiration can be calculated with various methods based on the available metrological data. In this specific study area the only available meteorological data has temperature and rain fall. So due to this reason, the method which is used for the determination of potential evapotranspiration is Thornthwaite method.

Based on equation 4.4, the potential evapotranspiration has been computed and the result is displayed in table 8. Accordingly, the potential evapotranspiration of the catchment is 692.2mm/year.

Table 8:-Estimation of potential evapotranspiration

month	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec
Tm	13.61	14.37	15.39	16.08	16.75	17.61	16.84	16.34	15.76	14.42	13.54	13.11
N(po)	11.60	11.80	12.00	12.30	12.60	12.70	12.60	12.40	12.10	9.00	9.70	9.70
Nm	0.97	0.98	1.00	1.03	1.05	1.06	1.05	1.03	1.01	0.75	0.81	0.81
i(m)	4.49	4.87	5.40	5.77	6.13	6.61	6.18	5.91	5.60	4.90	4.46	4.24
I	64.55											
A	1.51											
Em	49.36	53.63	59.46	63.54	67.56	72.87	68.16	65.08	61.64	53.90	49.00	46.66
PETm	47.73	52.75	59.48	65.14	70.96	77.15	71.58	67.26	62.17	40.44	39.62	37.72

Actual Evapotranspiration Estimation

Actual evapotranspiration is the amount of water that actually returns to the atmosphere depending on the availability of water. A value of actual evapotranspiration (AET) over a catchment is often obtained by first calculating the potential evapotranspiration (PET), and then modifying the value by accounting for the actual soil moisture content (Shaw, 1988).

To calculate the actual evapotranspiration (AET) of the area in this research, are identified soil type and vegetation cover of the catchment. Based on these categories and meteorological data the actual evapotranspiration of the specific land use and soil type the catchment is calculated and summarized in the tables 9 and 10.

The actual evapotranspiration for the dominant soil types and the respective land uses in the area was weighted according to the proportion of the area they represent, and the weighted actual evapotranspiration was calculated using equation 4.12 to determine the actual evapotranspiration of the catchment (Table 11). Accordingly, the annual actual evapotranspiration of the catchment is 666.52 mm.

Table 9 Calculated AET using soil water balance method for silty loam soil covered with cultivated land (cereal crops) having root depth of 0.8m and available water capacity of the root zone 200mm. Root depth of cereal crops was adopted from FAO papers (FAO irrigation and drainage paper, No. 56) All values in the table are in mm.

Table 9:-Calculated AET using soil water balance method for silty loam soil covered with grass land

Month	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec	annual
P	20.0	24.4	57.5	80.9	59.9	40.3	346.6	343.0	101.0	44.7	19.6	17.6	1155.4
PET	47.7	52.8	59.5	65.1	71.0	77.1	71.6	67.3	62.2	40.4	39.6	37.7	692.0
P-PET	-27.7	-28.3	-2.0	15.7	-11.1	-36.8	275.0	275.7	38.8	4.3	-20.1	-20.1	
APWL	-67.9	-96.2	-98.2		-11.1	-47.9					-20.1	-40.2	
Sm	142.4	123.6	122.4	200.0	189.2	157.4	200.0	200.0	200.0	200.0	180.9	163.6	
AET	41.2	43.2	58.7	65.1	70.7	72.1	71.6	67.3	62.2	40.4	38.6	34.9	**666.1**
ΔSm	-21.2	-18.8	-1.2	77.6	-10.8	-31.8	42.6	0.0	0.0	0.0	-19.1	-17.3	
SMD	6.5	9.5	0.7	0.0	0.3	5.0	0.0	0.0	0.0	0.0	1.0	2.8	
S	0.0	0.0	0.0	93.3	0.0	0.0	317.6	275.7	38.8	4.3	0.0	0.0	

Table 10:- Calculated AET using soil water balance method for silty loam soil covered with grass land

Month	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec	annual
P	20.0	24.4	57.5	80.9	59.9	40.3	346.6	343.0	101.0	44.7	19.6	17.6	1155.4
PET	47.7	52.8	59.5	65.1	71.0	77.1	71.6	67.3	62.2	40.4	39.6	37.7	692.0
P-PET	-27.7	-28.3	-2.0	15.7	-11.1	-36.8	275.0	275.7	38.8	4.3	-20.1	-20.1	
APWL	-67.9	-96.2	-98.2		-11.1	-47.9					-20.1	-40.2	
Sm	190.5	170.1	168.8	250.0	239.2	206.4	250.0	250.0	250.0	250.0	230.7	212.9	
AET	42.4	44.8	58.9	65.1	70.7	73.1	71.6	67.3	62.2	40.4	38.8	35.4	670.7
ΔSm	-22.4	-20.4	-1.3	81.2	-10.9	-32.7	43.6	0.0	0.0	0.0	-19.3	-17.9	
SMD	5.4	7.9	0.6	0.0	0.2	4.1	0.0	0.0	0.0	0.0	0.8	2.3	
S	0.0	0.0	0.0	97.0	0.0	0.0	318.6	275.7	38.8	4.3	0.0	0.0	

Table 11:-Weighted Actual Evapotranspiration

Soil texture	Land use/land cover	Root	w(mm)	coverage(km^2)	Wtd	%	AET	Wtd.AET
Silty loam	cultivated land	1	200	95.44	0.91	90.79	666.09	604.75
Silty loam	Deep rooted (grass)		250	9.68	0.09	9.21	670.70	61.76
Total				105.12				**666.52**

Where

- p = precipitation
- PET = Potential evapotranspiration
- AET = Actual evapotranspiration

- Sm = Monthly soil moisture
- ∆Sm = Monthly change of soil moisture
- SMD = Soil moisture deficit
- APWl = Accumulated potential water loss
- S = Surplus moisture
- W=Available water capacity of the root zone

5.1.4 Runoff Estimation

The water which is not infiltrate, forms flows as thin sheet of across the land surface, which is called over land flow or surface runoff (Tenalem Ayenew and Tamiru Alemayehu, 2001).

Since there was no river gauging station in the catchment, it was impossible to determine the runoff of the catchment from recorded data. As a result the volume of runoff from the Gerado River catchment was computed by using the runoff coefficient method (Garg, 1987).

Accordingly, the volume of water which leaves as surface runoff from the catchment is calculated to be 78046867 m^3 (320.71 mm).

Table 12:-Runoff calculation using the runoff coefficient method

Land use	Soil Type	Slope	Area(m^2)	ppt(m)	K	Runoff(m^3)
Cultivation	Clay	Flat	24500000	1.13618	0.2	5567282
Grazing	Clay	Flat	3060000	1.13618	0.16	556273.728
Cultivation	Clay	Gentle, Gentle to flat	82000000	1.1362	0.3	27950520
Cultivation	Sandy loam	Gentle, Gentle to flat	48000000	1.1362	0.25	13634400
Grazing	Sandy loam	Gentle, Gentle to flat	25600000	1.1362	0.22	6399078.4
Plantation	Rock	Steep	23500000	1.13618	0.35	9345080.5
Shrubs	Rock	Steep	36700000	1.13618	0.35	14594232.1
Total			243360000			78046866.73

5.1.5 Groundwater Recharge Estimation

Groundwater recharge is defined as the entry into the saturated zone of water made available at the water table surface together with the associated flow away from the water table within the saturated zone (Freeze & Cherry, 1979). Quantifying the rate of recharge to aquifer is the most difficult of all measures in the evaluation of groundwater resources. Estimation of groundwater recharge requires modeling of the interaction between all the important processes in the hydrological cycle such as precipitation, infiltration, surface runoff, evapotranspiration, soil moisture and groundwater level variations (Jyrkama and Sykes, 2007).

There are many sources of recharge to groundwater systems. These include recharge from precipitation, rivers, irrigation losses and inter-aquifer flows. In many cases, combination of the various types of recharge will occur (Simmers, 1997).

In this study, recharge was estimated using water balance method which is based on the law of conservation of energy. Based on equation 4.16 the volume of water that recharges the groundwater of the catchment per year is 148.95mm (13% the annual precipitation).

5.2 Conceptual Model Result

5.2.1 System Boundary Conceptualization

The initial step in any ground water flow modeling is the definition of the boundary of the study area. To have a good conceptualization of a hydrologic system, it is essential to identify and assign system boundaries appropriately.

System boundaries are classified in to two: physical boundaries and hydraulic boundaries (Anderson and Woessner, 1992). Physical boundaries of groundwater flow systems are formed by the physical presence of an impermeable body of rock or a large body of surface water. Hydraulic boundaries are a result of hydrologic conditions, are invisible and they may include groundwater divides and streamlines.

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

The Gerado river catchment, aquifer system boundary was carefully assigned based on field visits made and existing works. The geographic boundaries of Gerado river catchment groundwater flow model approximately correspond closely with natural hydrologic boundaries across which groundwater flow was assumed to be negligible. This is true for the western, southern and northwestern boundaries of the system. In the western, southern and northwestern boundaries of the catchment the model coincides with the Amera Genda, Indod Ber and Tossa Mountain respectively. On these sides, major topographic divides were assumed to coincide with groundwater divides. It should be remembered that groundwater divide is not really a boundary in nature, but as groundwater on either side of the divide flows away from the divide and not across it, the divide itself acts as a no flow boundary. The subsurface outflow at Northern part of the study area is simulated by General Head boundary.

In addition, the upper boundary of the modeled catchment was assumed to be the water table and recharge was used to represent the boundary flux. Although the lower boundary was considered as a no flow because the aquifer was assumed to be underlain with an impermeable rock body.

5.2.2 Hydraulic property of aquifer

Through proper pumping test data collection, analysis and interpretation adequate information about the groundwater condition would be generated. Since, direct observation of groundwater movement is impossible, mathematical analysis offers a convenient and reliable way to predict what happens to water in the ground. It is therefore, imperative to derive simple mathematical expressions for describing the flow regime of water in the subsurface.

In this study, the analysis results of the pumping test data during drilling of each well was adopted to determine the aquifer hydraulic parameters such as transmissivity and hydraulic conductivity referred and hydraulic conductivity result was later used as one of the model input parameters.

The aquifers in the study area are mainly alluvial deposits, highly fractured and weathered basalts and tuff and ignimbrite. Totally 7 boreholes were inventoried in this study and all were located in the alluvial soil covered areas. For the purpose of this study, pumping test data of 5 large diameter (16’’Ø) Inches boreholes were referred. All these boreholes are pumping well. The boreholes have relatively complete pumping test data, deeper in depth but well not distributed. The drilling was mainly in the alluvium and pump testing results are assumed to represent storage in the alluvial aquifers.

There are no boreholes that tap the volcanic aquifer within the studied area. Therefore, it was found difficult to calculate the hydraulic parameters for the volcanic aquifer. Because of such limitation, the hydraulic parameter for the volcanic aquifer was adopted from literature given by Freeze and Cherry (1979).

Table 13:-Hydraulic conductivity of wells in the study area (adopted Dereje Gidafie, 2012 and water table data measured during field stay after pumping)

Well Id	Location		Depth (m)	SWL (m)	DWL (m)	Q (l/s)	DD(m) (m)	T (m^2d^{-1})	K (md^{-1})
	X	Y							
PW1	565035	1226086	63	1	19.9	35	19.9	246	12.3
PW2	566864	1226390	123	1.2	23.4	22.5	23.4	56.4	1.3
PW3	564424	1226780	116	1.7	18.4	15	18.4	25.8	0.7
PW4	565035	1225946	120	1	16.2	39.9	16.2	118	2.4
PW5	564073	1227366	81	1.6	27.9	16.5	27.9	21.1	0.4

5.2.3 Fluxes and Stresses

In Gerado River catchment groundwater entered to aquifer system through precipitation as a recharge and removed from aquifers in form of withdrawal through wells and spring for human consumption, subsurface outflow, and base flow to rivers. The following sub sections deal with each recharge and discharge types.

Ground Water Recharge

Groundwater recharge is a key component in any model of groundwater flow or contaminant transport and its accurate quantification is crucial to proper management and protection of groundwater resources.

Groundwater recharge estimated results are discussed in section 5.1.5. Generally the highland area of the catchment gets better direct recharge due to open, penetrative and interlinked fractures (faults) than the valley alluvial covered areas where there are semi-confining clay beds in the alluvium other than local morphologic effect minor rainfall variations. The elevated mountain ridges receive likely moderate to high rainfall amount due to orographic effect too which increases rainfall intensity and direct recharge ignoring the local and limited interflow in some portions of sloppy sides. Although, relatively short distances from the drainage divide, the river is perennial (mainly base flow, dry period groundwater contribution) and therefore the recharge has to be substantial. Major part of the area is covered by basalt and some parts with ignimbrite and tuff with penetrative fractures in some parts. The basalts also show weathering and fracturing that resulted in the formation of relatively thin soil cover over the slopes which contributed for direct percolation of water.

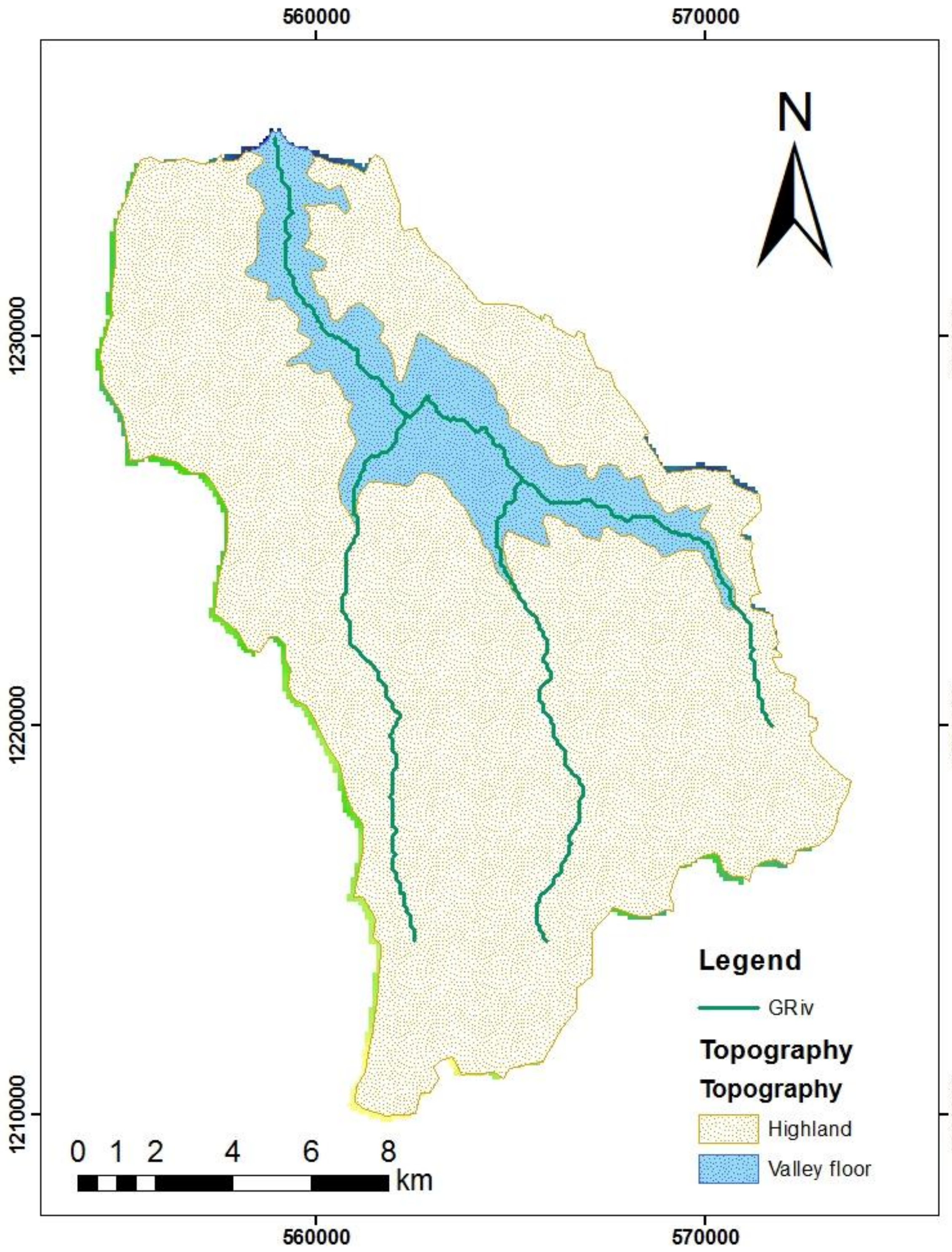

Figure 19:-Simple classification of the catchment for the purpose of recharge zonation.

In valley floor, recharge in this area is generally assumed to very minimum from direct precipitation because this lowland part of the study area has relatively lower annual precipitation and high annual actual evapotranspiration compared to the highland due to over 1000m elevation varitations. However, the valley floor gets regional flow of groundwater (indirect recharge) from the highlands (subsurface flow from high gradient to low gradient). The runoff from the highland area flow out to the valley floor along three main stream channels, Kelina, Negwaly and yito stream which in their way will contribute to recharge while crossing fracture zones and along loose and permeable river sediments.

Well withdrawal

The boreholes found in the study area are used for domestic purpose and are currently serving for Dessie water supply. The rate of abstraction varies from well to well, being highest in wells pw1, pw4 and harow and lowest in wells pw3 and pw6. These wells are 7 and 2 spring and the abstraction rate from these wells and two perennial spring is listed in table 14 based on 10 hours average pumping.

Table 14:-The estimated amount water abstracted from boreholes per annual

WELL ID	UTM		Elevation	Pumpage(m^3/h)	Average working time	Pumpage(m^3/d)
	Easting	Northing				
pw1	565163	1226291	2240	126	10	1260
pw2	564956	1226595	2236	81	10	810
pw3	564514	1226987	2231	54	10	540
pw4	565130	1226150	2242	143.64	10	1436.4
pw5	564124	1227809	2198	59.4	10	594
pw6	566737	1225871	2270	36	10	360
Harow	569303	1224795	2257	126	10	1260
Spring38	567887	1226708	2326	14.4	24	345.6
Spring43	561261	1224589	2355	21.6	24	518.4

Base Flow of rivers

The ways of groundwater discharge from the aquifer system is mainly by discharge to streams. The main stream (Kelina, Negwaly and Yito tributary) of Gerdao river that drains the southern and south Eastern**,** highlands gets its base flow from the volcanic aquifer and losses some amount when it reaches the valley floor (the alluvial deposits).

Therefore, the main mechanism of groundwater discharge from the catchment is in the form of base flow along Gerado River. Since, this river is not gauged and possesses both processes (gaining and losing) along its course within the basin.

In this study, base flow to rivers was calculated by using runoff coefficient method discuses in section 5.1.4. Based on this, base flow approximated to Gerado River as about 205,027.2m^3/day (320.71mm/y).

5.3 Numerical Model Results

5.3.1Grid Design/Spatial Discretization

In a numerical model, the continuous problem domain is replaced by a discretized domain consisting of an array of nodes and associated finite difference blocks (cells). The nodal grid forms the framework of the numerical model (Anderson and Woessner, 1992). A critical step in grid design is the selection of the size of the nodal spacing as the horizontal dimension. Factors that influence the size of nodal spacing are the availability of data, variability of aquifer properties and the size of the area to be modeled.

The numerical ground water flow modeling of Gerado river catchment has an aerial extent of the north-south and east-west is 30,000and 24,000m, respectively(shown in figure 19). The model uses a uniform grid size of 250 m by 250 m and contains 1 layer, 96 columns and 120 rows. The finite-difference model grid must be regular spacing, in order to facilitate data input from DEM files; certain cells may represent area outside the modeled area (which can have any shape). Such cells are considered inactive and are identified by assigning a value of “0 “entries of I-BOUND array in boundary condition and groundwater flow equation are not solved for such cells.

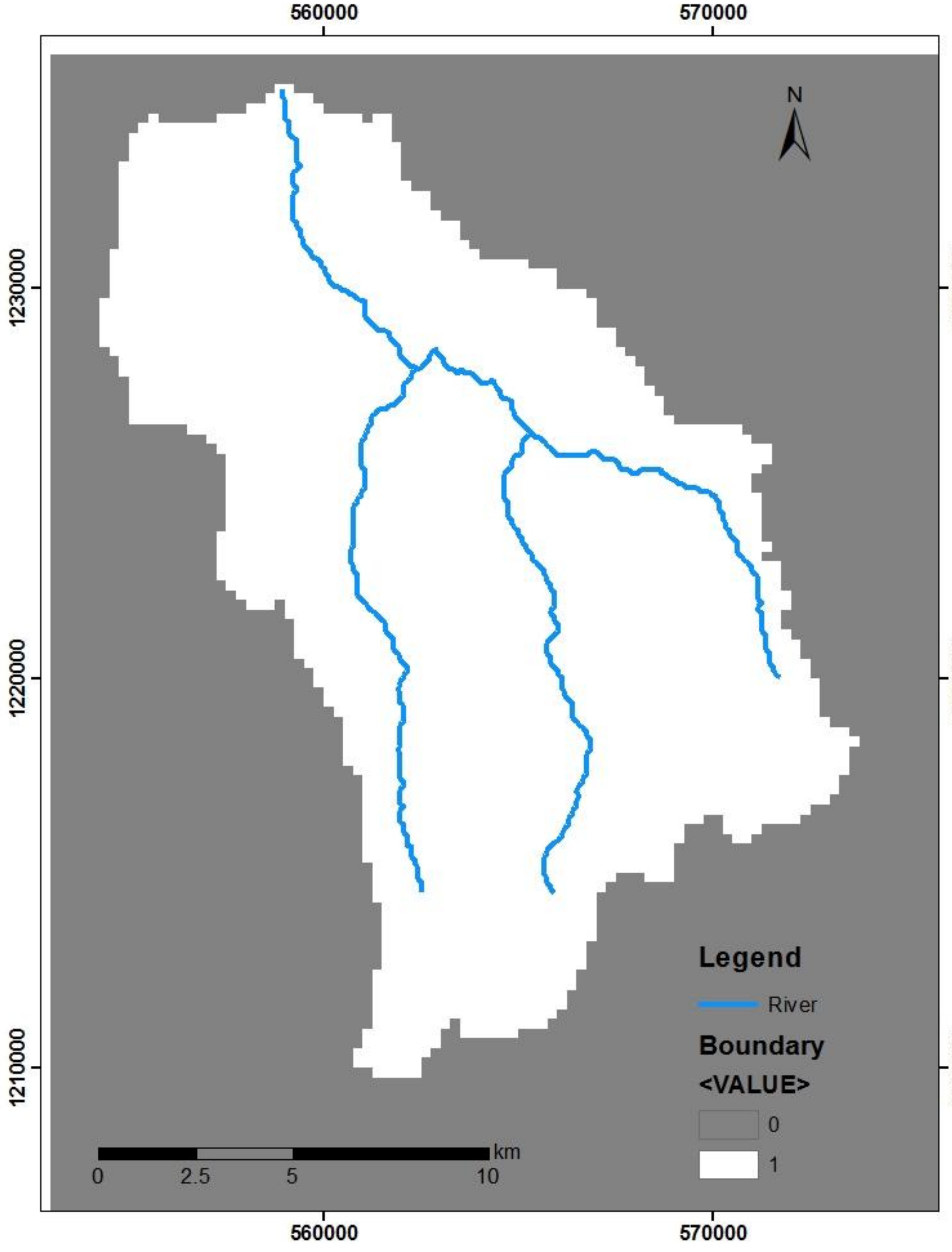

Figure 19:-Model Grid Design

5.3.2 Model Input Parameters

5.3.2.1Top and Bottom of layer

The top layer elevation was considered to be the elevation of ground surface and in this case nodal values of ground surface elevation were interpolated from DEM of 90m spatial resolution. The interpolation was done at a resolution of 250m by 250m and then loaded into MODFLOW top elevation array.

Layer bottom elevations were obtained by subtracting aquifer thickness from layer top elevation. Aquifer thickness of 250m is considered, except along the boundaries where ridges with high elevation are found. Elevated zones were simulated by giving relatively higher thicknesses at the cells in order to avoid drying of cells during simulations.

5.3.2.2 Initial & Prescribed Hydraulic Head

It is the initial stage at which groundwater level stood in aquifer system. Processing MODFLOW pro needs this initial heads to start simulation. For this simulation, it uses initial and prescribed hydraulic head by subtracting 10 m from the layer top elevation. This figures consider water table in highlands with volcanic cover and over 3000m.s.l. (low evapotranspiration areas) and alluvial covered lowlands which get lateral flow from adjacent area and recharge from streams.

5.3.3 Boundary Conditions

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

A thorough understanding of boundary conditions and the different ways to simulate them is required to select the best mathematical representation in a groundwater flow model because many physical features that are hydrologic boundaries can be mathematically represented in more than one way.

Boundary conditions may be three types: no-flow boundaries, specified-flux boundaries, and head-dependent flux boundaries. The contact between the permeable ground-water flow system and nearly impermeable bedrock is an example of a no-flow boundary. Known or estimated hydrologic fluxes, such as recharge and well discharge, are represented using specified-flux boundaries. A head-dependent flux boundary is one across which ground water moves at a rate proportional to the hydraulic-head gradient between the boundary and the ground-water system. Streams are usually represented as head dependent flux boundaries because the movement of ground water to or from a stream is proportional to the difference between the head in the aquifer and the stage of the stream.

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

The lateral boundaries of the study area flow model generally represented as no flow boundaries, with the exception of a narrow north surface outlet ,where it was represented as head dependent flux boundaries is assumed at the locality of topographically low surface water divides and where simulated with the general head- boundary (GHB) module of MODFLOW. Cells assigned as GHB are chosen based on the groundwater flow direction (See figure 20).

The upper boundary of the system was simulated as specified flux as recharge was applied to the water table. The lower boundary was assumed to be a no flow boundary. As the remaining boundaries of the catchment were not defined as unique boundary conditions, MODFLOW automatically assumes such boundaries as inactive. The value '0' was assigned to cells at the external of such boundaries to make them inactive and no flow occurs across such boundaries. Head values are not calculated for cells represented as inactive cells.

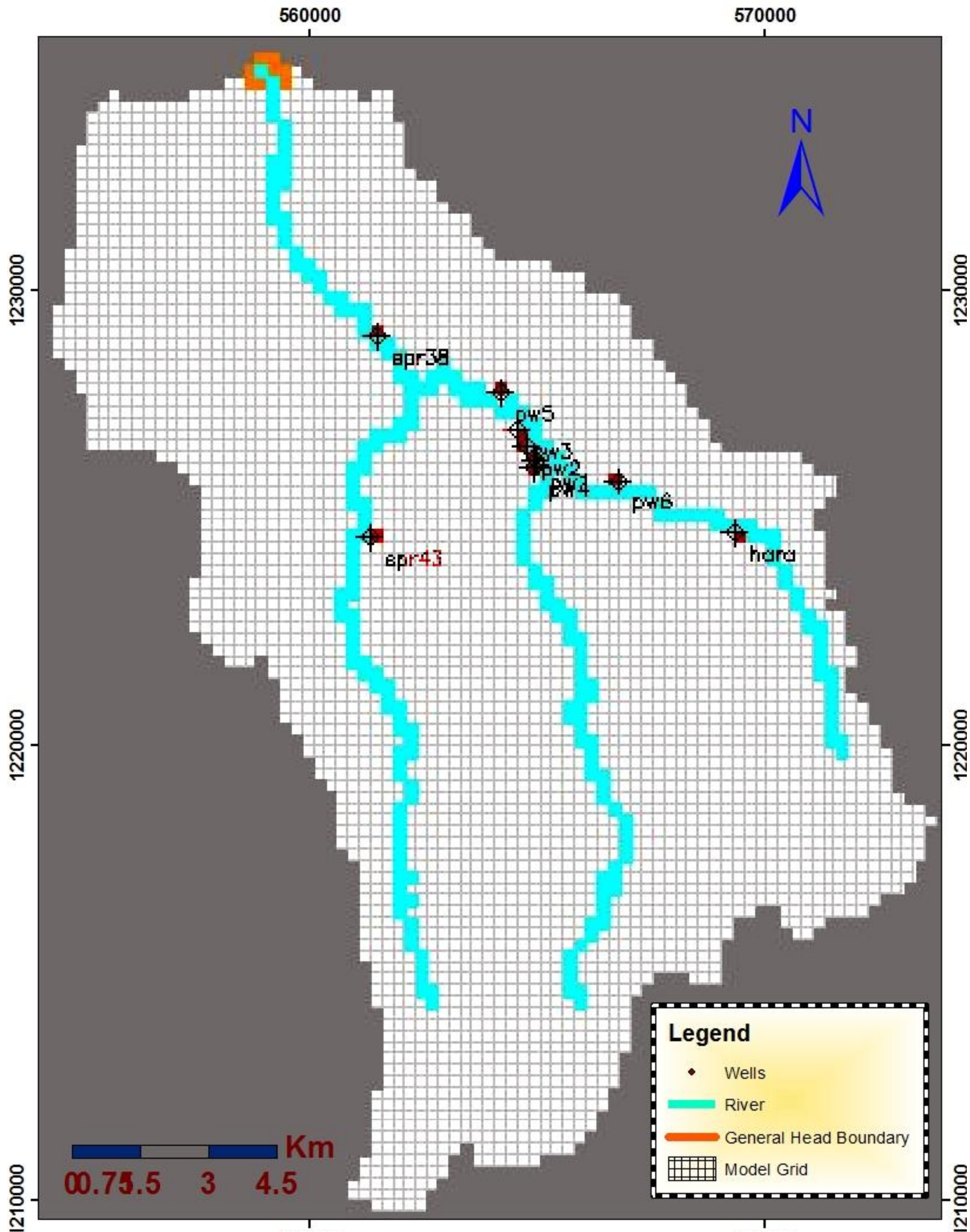

Figure 20:-Boundary condition of the study area

5.3.4 Model Stress and Fluxes

The fluxes into or out of the aquifer in the Gerado River Catchment as applied into the model and simulated by MODFLOW -2000/MODFLOW-2005(McDonald and Harbaugh, 1988) are summarized under the following sub sections. Various MODFLOW packages were used to simulate model stresses. The considered stresses include recharge to the aquifers, discharge to rivers, well and spring withdrawal and sub surface outflows.

Groundwater Recharge

As it was discussed under recharge conceptual modeling , the catchment was subdivided in to two zones of recharge (Figure 19).The recharge was simulated as specified flux by using the Recharge package of MODFLOW with the option recharge is applied to the top grid layer. The recharge value and zonation was modified during the calibration process.

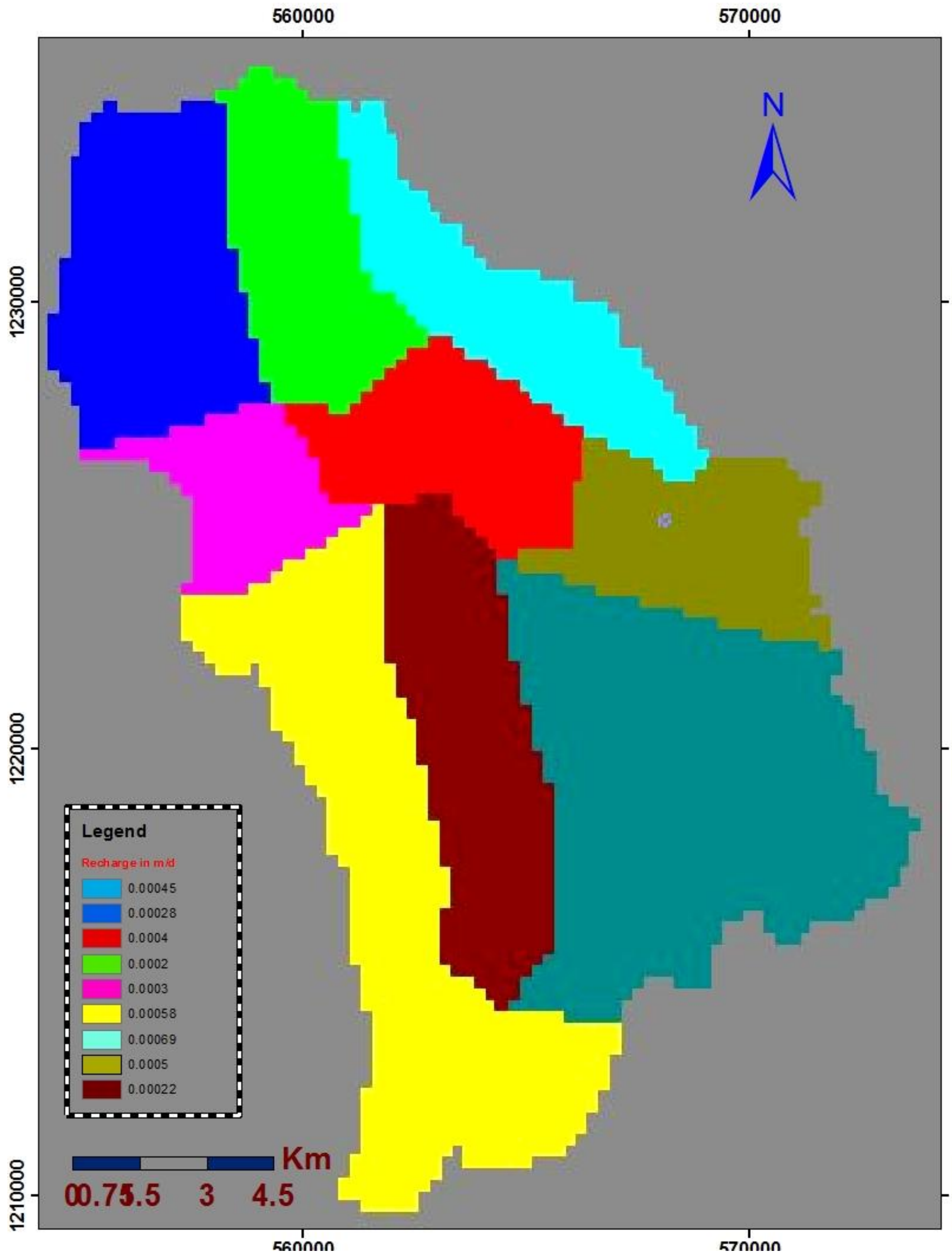

Figure 21 Simulation recharge zonation

Withdrawals

A pumping well is defined by using the Cell-by-Cell input methods. MODFLOW assumes that a well penetrates the full thickness of the cell. Data from 7 wells and 2 high yielding perennial springs (such as spring 38 and 43 supplying water to Kelina kabala and surrounding areas with a yield of 10 l/s were used to quantify daily and annual abstraction from the aquifer .Total daily abstraction based on 10 hours pumping from the aquifer of the Gerado River catchment is 7124.4m^3/d.

Base flow of Rivers

The main mechanism of groundwater discharge from the catchment is in the form of base flow along Gerado Perennial River and most perennial tributaries. Since, this river is not gauged and possesses both processes (gaining and losing) along its course within the catchment; a river package of MODFLOW was used to simulate the interaction between the river and the aquifer.

The river package is designed to simulate the effect of flow between rivers and aquifers based on the following relations:

$$QRiv=CRiv\ (HRiv-h),\ \text{when}\ h>RBot \qquad \text{Eq5.1}$$

$$QRiv=CRiv\ (HRiv-RBot),\ \text{when}\ h<RBot \qquad \text{Eq5.2}$$

Where QRiv: the rate of leakage through the river bed

HRiv: head in the river h: head in the aquifer

RBot: elevation of bottom of the river bed

CRiv: river bed hydraulic conductance

The River Package simulates the effect of loss from aquifer to rivers or from river to aquifers and uses stream bed conductance to calculate the flux.

Hydraulic conductance, head in the river and bottom elevations of river bed were approximated for river segments. Hydraulic conductance is incorporated into river package to account for the length (L) and width (W) of the river channel in a cell, the thickness of the river bed sediment (M) and the hydraulic conductivity of the river bed sediments (K). It can be expressed as:

$$C_{Riv}=KLW/M \qquad \text{Eq 5.3}$$

Streambed conductance terms were initially calculated for Gerado River using a streambed thickness and of 0.8 to 1.5 m, stream widths of 8m to 15m. The other tributary of Gerado river streambed thickness and stream widths are 0.2 to 0.6 m, and 5m respectively. The river bed sediment hydraulic conductivity was assumed to be 1 to 1/10 of the nearby formation hydraulic conductivity. The length of river segment was highly variable from cell to cell and was approximated for each cell. The hydraulic conductivity value was then adjusted during model calibration.

Subsurface Outflow

As it has been discussed under boundary condition, subsurface outflow occurs along the northern part of the catchment. This outflow was simulated using the General Head Boundary Package of MODFLOW.

A General-Head boundary simulates a source of water outside or inside the model area that either supplies water to, or receives water from, adjacent cells in proportion to the hydraulic-head differences between the source and model cell. The exchange rate of water between the model cell and the outside source or sink is given by the equation

$$Q = C\,(HB - h) \qquad \text{Eq5.4}$$

Where

- Q is the rate of flow into or out of the model cell [L^3/T],
- C is the conductance between the external source or sink and the model cell [L^2/T],
- HB is the head assigned to the external source or sink [L], and
- h is the hydraulic head within the model cell [L].

Values of C were initially calculated using the equation;

$$C = K \cdot L \qquad \text{Eq5.5}$$

Where

- K is the hydraulic conductivity between the model cell and the boundary head [L/T], and
- L is the length of the general-head boundary within a cell.

The initial values of C, ranging from 500–1,200 m^2/day, were distributed across the model cells adjacent to the general-head boundary. The boundary head in the adjacent catchment was assigned an initial head that was slightly less than the water table altitude of the model cells around the general-head boundary. The final values of C and head in the adjacent catchment, determined during model calibration are 800m^3/d and respectively. The total groundwater outflow across the boundary is 15377.1m^3/d (6302090 m^3/year).

5.4 Model Calibration

Model calibration is the process whereby model parameter structure and parameter values are adjusted and refined to provide the best match between measured and simulated values of hydraulic heads and flows. The model is calibrated by a trial-and-error process in which model parameters are adjusted within reasonable limits from one simulation to the next to achieve the best model fit. Model fit is commonly evaluated by visual comparison of simulated and measured heads and flows or by listing measured and simulated heads together with their differences and some type of average of the differences, which is then used to quantify the average error in the calibration. The objective of calibration is to minimize this average error which is called calibration criterion.

The manual inverse trial-and-error method provided satisfactory results for the steady state calibration and met the overall goals of the study. The Gerado river catchment regional steady-state groundwater model was calibrated to average conditions from 2002 to 2003. This period was chosen because all well drilled in this time. Water level measurements for 7 wells and 2 springs that were used for estimation of groundwater level of the aquifer were considered for calibration of the steady state model. These head observations were not evenly distributed throughout the model domain but were clustered in the central part of the study area that is on the alluvial deposit. The hydraulic heads were obtained from water level records measured during drilling. It should be noted that most of the measured head data and the uncertainty of the model are associated with errors due to the following reasons:

- All of the water level measurements were taken from the pumping wells, without monitoring well.
- The water level measurements are single time measurement (taken during drilling time).
- Measurement errors related to measuring device and operator/user.
- Errors due to averaging ground surface elevations from digital elevation models (DEM).

5.4.1 Evaluation of calibration

The results of the calibration should be evaluated both quantitatively and qualitatively (Anderson and Woessner, 1992). The mean of the observed and simulated heads differences was used to quantify the average error in the calibration process. The three ways of expressing the average difference between simulated heads (hs) and measured heads (hm) are the mean error (ME), the mean absolute error (MAE) and the root mean square error (RMS). The objective of the calibration is to minimize these error values.

❖ The mean error is the mean of the differences between measured heads and simulated heads.

$ME=1/n\ \Sigma\ (h_m-h_s)$ Eq5.6

Where h_m is measured head

h_s is calculated head and

n is number of head measurements.

❖ The mean absolute error is the mean of the absolute values of the differences between measured heads and simulated heads.

$MAE=1/n\ \Sigma|\ (h_m-h_s)|$ Eq5.7

❖ The root mean square error is the square root of the averages of the squared differences between measured heads and simulated heads.

$RMS=\{1/n\ \Sigma\ (h_m-h_s)^2\}0.5$ Eq5.8

The above error measures can only be used to evaluate the average error in the calibrated model. The RMSE is usually thought to be the best measure of error if errors are normally distributed. The maximum acceptable value of the calibration criterion depends on the magnitude of the change in heads over the problem domain (Anderson and Woessner, 1992).

Model generated scatter diagram showing the calibrated fit between the observed and simulated heads is shown in Figure 22. The scatter plots are visually examined whether points in a plot show deviation from the straight line in a random distribution or have systematic deviation, where systematic deviation of the plots can indicate systematic error in adjusting the parameter values. The scatter plot shows a correlation coefficient of 0.88 between measured heads and simulated heads is also one of good indicator of calibration quality (Figure .22).

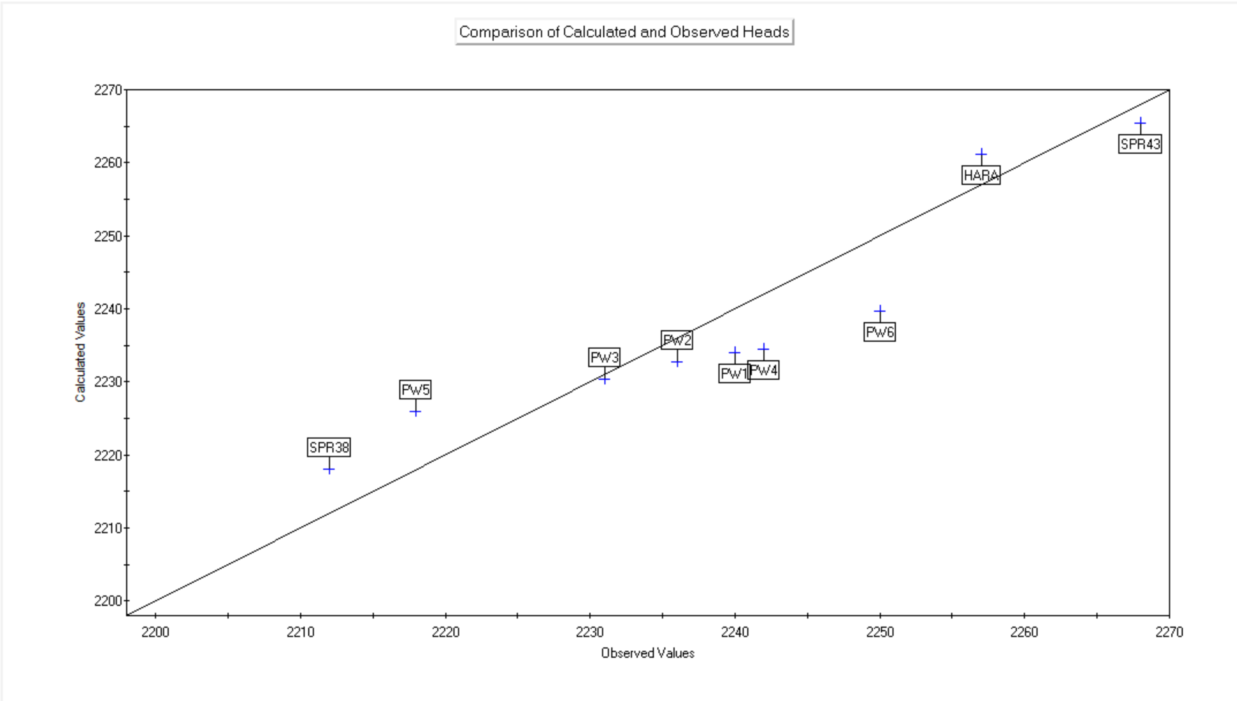

Figure 22:- The scatter diagram showing the comparison of measured and simulated heads

The differences in simulated and observed heads are shown in Table 15

Table 15:-Lumped sum calibration errors

BH ID	Simulated value (hs),	Measured value (hm)	(hm-hs)	∣ hm-hs ∣	$(\text{hm-hs})^2$
PW1	2233.41	2240.00	6.59	6.59	43.43
PW2	2231.98	2236.00	4.02	4.02	16.18
PW3	2229.66	2231.00	1.34	1.34	1.80
PW4	2233.76	2242.00	8.24	8.24	67.90
PW5	2224.87	2218.00	-6.86	6.86	47.13
HARA	2260.66	2257.00	-3.66	3.66	13.36
PW6	2239.64	2250.00	10.36	10.36	107.39
SPR38	2216.99	2212.00	-4.99	4.99	24.94
SPR43	2269.54	2268.00	-1.54	1.54	2.36
			ME=1.5	MAE=5.29	RMS=6

Qualitatively, calibration results are evaluated using pattern matching of measured and simulated head which gives very good match. The contour map of the hydraulic heads was assumed to represent the top of the groundwater surface in an aquifer. It helps to observe the groundwater flow direction. The groundwater flow directions more or less agree with the conceptualized flow direction.

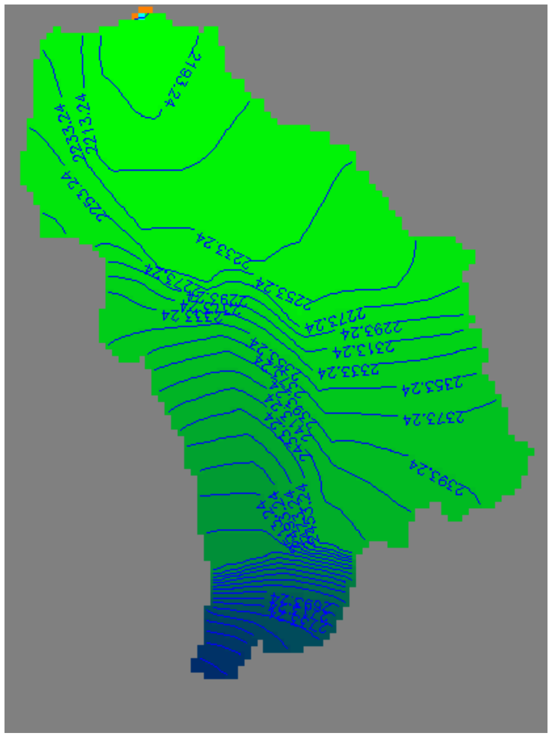

Figure 23:- Model simulated head patterns

5.5 Water budget of the model domain

In direct way of checking, the amount of residual error in the solution is to compare the total simulated inflows and outflows as computed by the water budget of the entire domain. The water budget modeling is one of the methods to quantify the amount of water that flows through the aquifer system. The basic equation for a water budget under steady state condition is that sum of inputs to the aquifer equals sum of outputs from the aquifer.

$$\sum \text{Inputs} = \sum \text{Outputs} \quad \text{Eq5.9}$$

In the model area, the inflow term includes the recharge and river leakage whereas, the outflow term includes wells and spring withdrawal, head dependent boundary and river leakage under natural condition. The water budget is calculated by water budget tool in MODFLOW. The model result shows both inflow and outflow are in balance which is consistent with the steady-state modeling theory.

Table 16:-Water budget of the entire model domain in $Mm^{3}y^{-1}$

FLOW TERM	IN	OUT	IN-OUT
Wells	0	2.6	-2.6
Recharge	39.3	0	39.3
River leakage	11	42.1	-31.1
Head dependent boundary	0	5.6	5.6
Total	50.3	50.3	0

Discrepancy [%] 0.00

Table 17:-Water budget of the entire model domain in mmy^{-1}

FLOW TERM	IN	OUT	IN-OUT
Wells	0	10.4	-10.4
Recharge	157.2	0	157.2
River leakage	44	168.4	-124.4
Head dependent boundary	0	22.4	-22.4
Total	201.2	201.2	0

Discrepancy [%] 0.00

The model simulated water budget result shows that the main groundwater outflow components are discharge to the river (Table 17). The model calibrated total recharge rate of the catchment is 157.2mmy^{-1}, whereas the recharge estimated using water balance method was 148.95mmy^{-1}. Despite of the limited data used to estimate the recharge using the water balance method it agrees with the recharge estimated by the model calibration.

5.6 Model Sensitivity Analysis

Sensitivity analysis is the measure of uncertainty in the calibrated model caused by uncertainty in aquifer parameters and boundary conditions. Sensitivity analysis was performed by systematically changing the calibrated values of conditions (Anderson and Woessner, 1992). The main objective of a sensitivity analysis is to understand the influence of various model input parameters and hydrological stresses on the aquifer system and to identify the most sensible parameter(s), which will need a special attention in future studies. By running the calibrated model for the respective changed values of the input parameter and comparing the result with the calibrated head, the parameter(s) sensitive to the model was established. The parameter values were varied within a reasonable range. Thus, it is important step in modeling studies.

The response of the calibrated model to changes in model parameters of hydraulic conductivity and recharge was examined. During model run, when the effect of one parameter was being tested, the other parameters were kept unchanged from the calibrated value. The amount of changes in heads from the calibrated solution was used as a measure of the sensitivity of the model to that particular parameter.

The calibration values of recharge and hydraulic conductivity were varied by 10%, 20%, 30%, 40% and 50% increase and decrease at different times to test the sensitivity of the model to the parameter. A total of twenty model runs have been made by changing the hydraulic conductivity and recharge by the specified percent and the root mean squared head changes from the calibrated value are shown in table18. As observed from the values, the model was most sensitive to both decrease in the recharge and hydraulic conductivity values, mainly beyond 50 percent and less sensitive to both recharge and hydraulic conductivity increasing of the value.

Table 18:-Results of sensitivity analysis Test on water level

Change in parameters (%)	RMS Values for the Selected Parameters	
	Recharge	Hydraulic conductivity
-100	23.17	31.56
-75	15.61	15.87
-50	10.43	10.36
-25	7.43	8.02
0	0	0
25	6.21	7.04
50	7.84	8.57
75	9.51	9.62
100	10.36	10.83

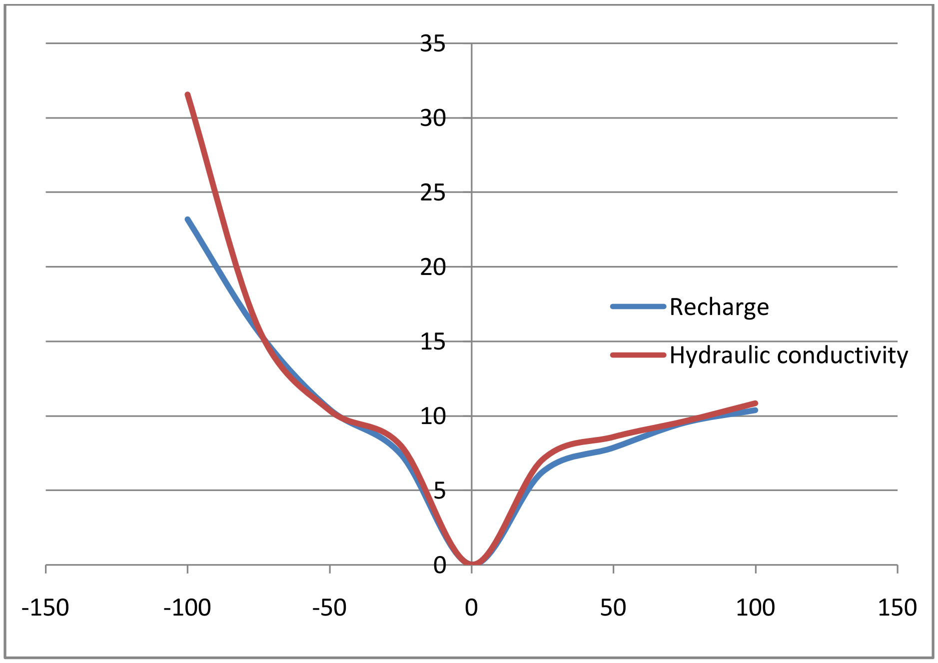

Figure 24:-Plot of the Results of Sensitivity Analysis Test on Heads

5.7 Scenario Analysis

The calibration of groundwater flow model can be used to simulate the potential effect of alternative water management plan on hydraulic head and groundwater movement in the study area. It can also be used as a tool to evaluate and compare the responses of an aquifer system to potential future stresses. One of the aims of this working was intended to test the responses of the hydrologic system to different scenarios. So, alternative scenarios were developed to test the response of the hydraulic system to changes in water uses or hydrologic stresses under steady state condition. System response was evaluated by using fluxes and heads of the calibrated model as a baseline and compared with resulting changes in base flow, changes in water table elevation and change in groundwater outflow in the new scenario simulation.

Change in water levels and fluxes caused by increased groundwater withdrawal in the whole catchment were simulated using the model it affect caused by decreased recharge due to less than normal precipitation that may result from weather modifications was also simulated.

It should be noted that the results of the scenarios depend on future land use, population growth, weather conditions, hydrologic stresses etc, and may not be used as a predictive tool to generate absolute amounts in the future, but used primarily to test the response of the system. In this study, the model was used to simulate the aquifer system response to increased withdrawals and decreased recharge, for both scenarios, all model parameters were unchanged from those specified in the steady state simulation.

The first scenario, which is increment of pumpage by 50%, was selected to represent possible future changes in water use in the catchment, or to investigate the effects of water-management practices that could mitigate potential adverse effects of increased water withdrawals. The heads calculated for this scenario shows a maximum decline of the water level by 0.866 m and a minimum of 0.131m, where high yielding wells get the bigger decline of water level. The increase in pumpage decreases the groundwater outflow through river leakage and general-head boundary.

The second scenario simulates a case of decreased recharge to aquifers by 25 % that may results from lower than normal precipitation. It is real that changes in climate conditions from time to time are affecting precipitation amount in the county adversely and reducing recharge to groundwater, as the main source of recharge is precipitation. The heads calculated for this scenario shows a maximum decline of the water level by 10.623m and a minimum of 0.583m.

Table 19 shows the resulting water balance from the two different scenarios with the steady state model calculated water balance. Values of the difference between the simulated water level in the scenarios and those specified in the steady state simulation are shown in Table20.

Table 19:-Water balance of the two scenarios and the steady state model in m3/day

	IN			OUT			IN-OUT		
	Calculated	Increased pumpage	Decreased recharge	Calculated	Increased pumpage	Decreased recharge	Calculate	Increased pumpage	Decreased recharge
Wells	0.0	0.0	0.0	7124.4	10686.6	7124.4	-7124.4	-10686.6	-7124.4
Recharge	107906.3	107906.3	80929.7	0.0	0.0	0.0	107906.3	107906.3	80929.7
River leakage	30138.8	30212.9	31034.0	115423.0	112067.8	90271.7	-85284.2	-81854.9	-59237.7
Head dependant bounds	0.0	0.0	0.0	15377.1	15364.2	14569.7	15377.1	-15364.2	-14569.7
SUM	138045.1	138119.1	111963.7	137924.5	138118.6	111965.9	30874.8	0.6	-2.2
Discrepancy [%]	0.00E+00	0.00E+00	0.00E+00						

Table 20:-Water level difference between simulated and those resulting from scenarios

			Deferent between calibrated and scenario water levels	
Well ID	Observed Value	Calibrated Value	Pumpage Increase by 50 %	Recharge Decrease by 25%
pw1	2240	2233.326	0.526	0.81
pw2	2236	2231.861	0.508	0.905
pw3	2231	2229.526	0.348	0.817
pw4	2242	2233.688	0.525	0.817
pw5	2198	2224.657	0.204	0.583
pw6	2270	2239.574	0.137	0.781
Harrow	2257	2260.627	0.486	3.099
spring43	2268	2269.112	0.866	10.623
spri38	2212	2215.909	0.131	1.01

5.7 Model Limitations

As a model is a device that represents an approximation of a field situation, it is true that there are a number of limitations associated with it. Numerical groundwater flow models are only approximations of complex natural systems and have uncertainty. Therefore, it is essential that for any groundwater model to be interpreted and used properly these limitations be understood. In the numerical groundwater flow simulation of Gerado River Catchment some of the associated limitations are:

- ❖ Simulation of the groundwater system was based on various assumptions regarding the real natural system being modeled. In this study, some of these assumptions were that the system was represented as single layer, the aquifer is unconfined and simulation was made assuming that the system is under steady state condition, which can never be known in the absence of long term water level data.
- ❖ The whole study area was discretized in to a number of cells of equal size (250m by 250m) and input into the numerical model for simulation. The level of discretization used was too coarse to incorporate the effects at local scale, like the effects of structures. In addition, the grid size used was not compatible with well diameters or river channel

widths that are represented to have homogeneous properties in a cell. Their exact locations were approximated by the centers of the cells in which they occur. Hydrologic parameters and aquifer unit geometry in portions of the model area are not well known at this scale. For instance, aquifer thickness and hydraulic conductivity can change at intervals smaller than the current model resolution.

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

Because of this the model results should not be interpreted as a perfect simulation, rather as a system response within fairly realistic model input parameters. The model may not be readily used for the detail groundwater management purposes because of the stated limitations. The results should be interpreted and applied considering all the limitations and drawbacks associated to the input parameters.

CAPTER SIX

6 Conclusions and Recommendations

6.1 Conclusions

A one dimensional groundwater flow model under steady state condition was constructed as a tool to understand the aquifer system and to analyze the response of the system to future change in stresses. In doing so, conceptual model was developed based on the geology and hydrogeology of the area. The conceptual model was used as input into the numerical model to examine system response. From this study the following conclusions have been made:

- The model was calibrated using heads measured in 7 wells and 2 perennial springs. Calibration results are evaluated using statistical lumped sum description of the average differences of residuals between simulated and measured heads (mean error, mean absolute error, root mean square error & R2) were calculated for the residuals and found to be 1.5m,5.29m,6m and 0.88m respectively.
- The average recharge obtained by the model is about 157.2mm/year which closer to the manually calculated value of 148.95mm/year. The total groundwater in Gerado catchment is 39.3 Mm^3/year. This volume of water compared with the total abstraction in the basin (about 2.6 Mm ^{3}/year) is good storage and can sustain well abstraction at the existing rate or more than the existing in the future.

- Model simulated heads most sensitive to both decrease in the recharge and hydraulic conductivity values, mainly beyond 50 % and less sensitive to both recharge and hydraulic conductivity increasing of the value.

- The model was used to simulate the response of the aquifer to different scenarios, which includes increased withdrawals and decreased recharges. The effects of these scenarios were evaluated with respect to changes on groundwater heads, Base flow and sub-surface outflows compared to the steady state simulated values. The effects of increased in groundwater withdrawal rate by 50% in withdrawal rate cause maximum decline of the water level by 0.866 m and a minimum of 0.131m, and decreasing groundwater discharge through general head boundary and Base flow to river. The second scenario is effect of decrease in groundwater recharge by 50 % due to less than normal precipitation cause maximum decline of the water level by 10.623m and a minimum of 0.583m , decrease of groundwater discharge through the general head boundary & Base flow.

- The overall accuracy of the results of these simulations depends on future land use/ cover, hydrologic stress conditions, and change in climate. In addition, such scenario results will be applied for practical purposes if and only if the assumptions on which the simulations were based are valid, therefore the results should not be interpreted as perfect predictions, rather as system response projections. Moreover, the results should be interpreted and applied by considering all the limitations and drawbacks associated with the numerical model and knowledge gap of the modeler.

6.2 Recommendations

Based on the overall numerical simulation process of the Gerado River Catchment and the results of the model, the following recommendations were forwarded:

- Sufficient groundwater level monitoring wells should be placed in the whole catchment in order to understand the general fluctuations in ground water levels. This helps to carry out transient ground water flow modeling, so that system response can be predicted with greater confidence.
- The thickness of the alluvial aquifer and volcanic aquifer was inferred from the existing borehole data except at some part of volcanic aquifer and the alluvial aquifer was assumed to be hydraulically connected with the underlying fractured volcanic aquifer. This assumption needs further field verification through drilling.
- In order to determine rational water budget of the basin that can be used for independent calibration of the model simulation result, the necessary data gaps over the entire basin have to be filled. Accordingly, the following hydro-meteorological stations have to established within the basin;

- River discharge measuring stations have to install at les five places: where the Gerado rivers, the three tributary of Gerado river (Yito, Kelina and Negeweli) that drain the highland join the valley floor and at the outlet of the catchment.
- Depending on the criteria set for the establishment of gauging stations that can be required for a given catchment at least two stations have to be installed for the northwest direction and inside the catchment.

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APPENDICES

Annex 1. Meteorology data

Annex 1.1 Tita monthly total rain fall

Year	jan	Feb	mar	apr	may	jun	jul	Aug	sep	oct	nov	dec
2002	59.7	0	51.6	173.6	46.3	5.9	198.9	367.6	107.4	24.2	0	26.5
2003	23.7	37	48	68.5	2.4	37.6	280.2	410.6	105.9	2.1	5.1	41.7
2004	8.3	6.2	69	122.6	20.7	59.1	232.1	217.9	99.8	100.6	46	5.5
2005	75.2	27.5	162.9	74.3	99	68.1	341.8	294.2	0	49.1	29.2	0
2006	8.5	26.1	133.5	84.3	69.2	54.3	346.8	301.6	143.05	48.6	1.6	51.9
2007	46.4	97.8	67.8	121.2	67.9	41	348.1	206.4	79.83	23.1	13.1	0
2008	11.1	0.8	0	16.4	56.3	18.5	308.1	217	120.4	54.3	46.6	0
2009	26.4	72	31.4	71	21.8	22.1	311.7	265.2	53.7	47.1	13.6	70.2
2010	0	62.2	131.1	182.1	129.4	14.5	380.5	509.2	51.2	26.5	8.6	3.9
2011	14.6	2.1	83.2	64.4	175.4	13.3	273.9	331.7	47	11.6	23.6	0
2012	0	0	55.5	126.6	46.7	67.8	337.3	305.3	64.7	0.4	4.1	2.5
2013	19.4	0.6	55.4	64.4	66.6	14.4	330.4	283.5	78.2	120.3	6.5	0

Annex 1.2 Kutaber monthly total rain fall

year	Jan	Feb	mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nev	Dec
1992	30	7.6	88.7	79.4	36.8	1.7	164.9	343.2	167.1	44	31.3	6.9
1993	14.5	10	77.5	85.9	60.7	2.3	245.3	195.6	214.4	62.6	0	16.5
1994	0	0	21	73.4	103.2	22.3	579.3	347	177.5	0	73.1	22.2
1995	0	69.3	21.2	121.3	85.5	38.1	292	360.6	105.2	31.9	0	14.4
1996	34.5	0	63.3	0	200.6	68.9	319.8	456.8	45.2	0	23.5	2.3
1997	7.8	0	44.7	24.4	29.5	160	343.5	334.1	83.2	100.5	89.8	0
1998	28.7	13.7	29.9	32.4	17.5	5.4	584.8	383.9	141.1	35.2	0	0
1999	19.9	0	3.7	39.1	19.9	41	484.1	397	182.1	167.5	0	1.8
2000	0	0	5.8	51.8	50.2	34.7	431.4	560.9	184.5	109.3	18.2	11.5
2001	0	0	112.1	5.2	87	40.6	457.3	389.9	147.5	9.5	0	7.8
2002	23.3	17.1	58.7	76.1	24.7	43.4	361.5	405.1	151.2	10.4	0	13.1
2003	14.6	34.9	66.3	46.3	17.1	43.9	244.5	354	99.1	3	22.5	6.5
2004	2.5	0	28.4	70.8	5.4	62.1	252.2	259.8	88	67.5	36.5	0
2005	7.5	4.2	98.2	54.1	103.1	47.9	319.4	341	82	23.6	43.4	6.9
2006	3.6	0	148.7	39.9	53	24.2	329.9	330.4	118.3	32.5	0	0
2007	16.6	34.3	43	98.9	8.5	45.2	341.8	87.8	6	0	0	0
2008	0	0	0	27.6	45.4	39.7	258.1	320.7	193.6	10.9	4.4	0
2009	7.3	10.5	35	22.5	0	66	377	284.7	23.4	62.8	0	0

Annex 1.3 Guguftu monthly total rain fall

year	jan	Feb	mar	apr	may	jun	jul	Aug	sep	oct	nov	dec
2002	70.5	20.6	90.1	50.5	25.4	38.5	364.6	371.2	116.8	21.5	0	36.3
2003	16.2	46.1	37.5	76.3	28.5	85.4	368.3	426.7	175.3	2.5	1.5	33.1
2004	39.5	16.5	114.7	95.5	2.4	133	412.4	348.9	141.9	49	29.8	5.1
2005	66	4.9	68	98.7	163.6	74.1	402.4	463.2	76.9	10.9	7.9	0
2006	16.5	9.5	102.1	36.7	28.3	25	439.9	493.7	82	46.7	0	2
2007	15.3	45.5	40	38.9	45.7	93.8	338.8	244.5	1.5	8.5	25.4	0
2008	1.9	22	0	52.2	84.3	64.8	345.1	301.5	81.4	32.6	45	0
2009	13.2	10.5	42.3	1.2	4.5	69	412.13	235.4	67.6	59.5	2.5	94
2010	7.1	72.5	82.3	125.2	119.7	22.2	541.5	603	51.6	29.9	15.8	19.4
2011	24.3	6.2	88.8	88.6	172.7	31.2	297.4	352.8	97.8	31.8	27.7	0
2012	0	0	47.2	203.7	81.6	139.4	531.2	353.5	42.2	0	1.5	1
2013	4.2	0	112.5	26.4	30.5	28	686.4	497.8	191	117.4	8.2	0

Annex 1.4 Dessie monthly total rain fall

year	jan	feb	mar	apr	may	jun	jul	Aug	sep	oct	nov	dec
2002	34.3	0	0	0	21.8	0	0	372.4	0	20.8	0	42.4
2003	42.8	40.8	42.9	183.2	9.3	49.5	179.8	353	190.2	2.3	10	77.8
2004	0	14.9	11.5	70.3	5.1	46.3	326.2	264.9	92.6	94.1	50.9	7.6
2005	16	12.9	124.3	116.5	117.3	46.2	355.1	408	79.4	36.5	11.8	0
2006	2.3	3.5	117.1	85	102.2	30.6	532.3	411.1	222.7	65.8	0	40
2007	48	118.5	67.2	64.9	64	34.3	356.4	196.1	87.7	56.8	12.8	0
2008	9.5	0.6	0	22.9	54.2	33.8	252.8	198.8	136.1	51	50.8	41
2009	24	16.8	17.9	88.2	9.1	37.4	353.7	340.4	85.8	67.1	24.4	0
2010	0	62	92.7	130.8	126.8	36.7	372.6	465.4	43.3	46.8	7.8	12.3
2011	13.8	1	74	79.8	146.1	15.5	284.6	295.7	36.3	49	39.3	0
2012	0	0	27.4	112.8	42.4	84.5	302.1	272.9	51.1	0	0	18.3
2013	30.6	3.3	31.4	57	29.8	11	327.5	310.4	92.5	135.1	14.9	0

Annex 1.5 Kombolcha monthly total rain fall

year	Jan	Feb	mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nev	Dec
2002	18.1	5.8	58.9	82.1	20.7	10.6	278.5	254.8	86.8	7.1	0	62.2
2003	43.7	26	75	80.9	5.8	48.1	214.9	269.3	149.3	0.2	12.9	60.2
2004	12.6	11.7	39.7	142.3	14.8	28	220.3	239.4	82.5	66.8	40.1	5.5
2005	19	4.5	76.1	67	86.4	20.9	356.9	321.6	40.8	25.7	6.7	0
2006	8.8	0	119.1	87.1	0	14.7	365.8	277.2	144.6	62	1.2	6.4
2007	30.5	34	43	115	14.9	36.1	320.7	175.5	92.3	21.6	6.6	0
2008	25.5	0	0	18.1	44.7	29.4	286.9	208.2	68.7	47.1	75.6	0
2009	20.6	11.4	17	13.4	6.4	34.6	370.1	320.3	58.4	63.2	11.5	32.4
2010	0	31.6	70.7	102.7	107.6	15.9	343.9	524.8	18.6	11.7	15.2	11.6
2011	15	0	50.4	25.5	130.9	20.4	278	338.6	90.5	16.6	41.4	0
2012	0.3	0	109	177.1	34	50.1	284	261.6	52.9	0.5	0	0.2
2013	36.8	0.3	54.2	55.2	40	1.8	336.3	284.9	67.2	132.7	0.6	0

Annex 1.6 Guguftu tmax

YEAR	jan	feb	mar	apr	may	jun	jul	aug	Sep	oct	nov	dec
2002	14.95	15.99	15.42	15.74	16.25	15.08	14.21	13.45	13.93	13.80	14.22	14.48
2003	15.15	15.73	15.98	15.62	15.55	16.15	13.89	13.37	14.27	13.61	13.71	14.49
2004	15.26	14.69	15.07	14.87	15.66	15.39	13.78	13.44	14.65	13.97	13.97	14.39
2005	14.62	15.64	15.27	15.52	14.81	15.75	14.12	13.45	14.10	13.94	14.21	15.27
2006	15.61	15.08	15.21	15.16	15.46	15.63	13.75	14.16	13.62	13.83	13.84	14.84
2007	14.96	14.94	14.99	14.93	15.61	15.16	15.30	14.92	15.06	16.22	16.98	16.75
2008	15.66	16.87	16.25	16.67	16.38	16.44	15.23	15.98	15.28	15.30	15.36	15.36
2009	15.66	15.39	15.58	15.55	15.23	15.52	15.24	15.22	14.86	14.87	15.11	14.50
2010	15.16	15.43	15.07	15.37	15.42	15.78	15.99	15.94	15.86	15.46	15.33	15.81
2011	16.07	15.79	15.36	15.99	15.36	15.78	16.35	16.25	15.79	15.95	15.57	15.58
2012	15.97	15.67	15.64	15.61	15.46	16.11	14.78	15.34	15.80	15.28	15.80	16.18
2013	15.80	15.75	16.35	15.79	16.42	15.83	14.49	15.03	15.09	14.89	15.04	15.50

Annex 1.7 Guguft tmin

YEAR	jan	feb	mar	apr	may	jun	jul	aug	Sep	oct	nov	Dec
2002	3.3	2.4	2.7	4.6	4.1	3.8	4.3	4.2	4.9	3.2	1.4	3.3
2003	2.8	3.4	4.5	4.6	4.8	5.0	4.3	5.8	4.7	2.4	1.8	0.8
2004	3.3	2.7	4.1	5.3	4.6	4.0	3.8	4.1	4.4	3.0	2.5	2.6
2005	2.7	2.7	4.1	5.6	5.4	5.3	4.6	3.5	4.8	3.4	1.2	-0.8
2006	1.3	2.7	4.9	5.1	5.4	5.3	4.3	4.3	3.8	2.9	2.9	4.4
2007	5.0	4.7	5.0	5.2	5.0	4.7	5.4	5.3	5.0	3.7	3.6	3.5
2008	3.5	4.1	4.7	4.8	4.6	4.8	4.8	4.7	4.9	3.9	3.2	4.7
2009	4.7	4.7	4.8	4.7	4.8	4.9	4.6	4.4	4.5	3.9	2.3	3.7
2010	3.1	4.5	4.5	5.3	6.1	5.3	5.1	5.1	5.3	4.7	0.8	2.7
2011	3.8	3.1	4.7	5.5	5.5	5.2	4.9	4.6	4.5	4.5	5.6	5.0
2012	5.5	2.7	3.8	5.2	5.7	4.9	5.4	5.8	4.9	2.2	3.2	4.3
2013	4.5	3.6	5.5	5.4	5.9	5.4	4.3	4.7	4.7	3.6	2.8	0.7

Annex 1.8 Tita tmix

year	jan	feb	mar	apr	may	jun	jul	aug	sep	oct	nov	dec
2002	18.1	20.9	21.6	22.6	24.4	25.4	24.6	22.2	21.2	21.4	22.3	19.8
2003	20.5	22.7	22.3	23.0	24.1	25.0	22.7	21.8	21.3	21.1	20.6	18.9
2004	20.6	20.4	21.6	20.9	24.6	24.8	23.2	22.3	21.8	20.3	20.7	18.9
2005	19.1	22.4	22.4	22.7	22.3	25.0	22.7	22.4	20.0	21.6	20.8	19.5
2006	20.8	21.4	21.1	20.9	23.4	25.2	22.7	22.2	22.1	21.4	20.3	19.2
2007	17.9	20.1	22.3	21.7	24.3	25.0	21.8	22.3	22.5	21.2	20.4	19.9
2008	20.5	20.9	23.6	23.0	24.3	25.8	24.1	22.4	22.2	21.1	20.1	19.6
2009	19.4	21.3	22.8	22.9	24.7	26.5	23.1	29.3	23.0	21.5	22.1	19.2
2010	19.5	20.4	20.3	22.3	23.8	26.0	23.3	22.2	22.2	21.3	20.3	21.1
2011	19.8	22.2	21.1	24.3	23.2	25.3	24.2	21.9	22.6	21.0	19.7	19.6
2012	20.6	22.4	22.8	21.5	23.9	25.9	23.4	22.6	22.8	22.4	21.7	20.8
2013	21.1	22.1	23.1	23.9	24.7	25.6	23.0	21.3	22.8	28.6	21.1	19.8
2014	20.1	20.6	22.5	22.6	22.9	26.0	25.7	23.4	22.9	21.6	20.9	19.5
2015	19.9	22.8	23.1	24.1	24.1	25.2	26.2	23.8	30.6	23.9	22.2	20.5

Annex 1.9 Tita min T

year	jan	feb	mar	apr	may	jun	jul	aug	sep	oct	nov	dec
2002	9.7	11.3	10.9	11.8	11.3	12.0	11.9	11.4	11.0	8.9	6.0	8.4
2003	7.7	8.8	9.9	9.5	10.0	10.8	11.6	11.8	10.8	6.3	5.5	5.5
2004	9.5	7.7	9.2	10.9	10.0	11.6	11.6	11.5	10.3	6.6	6.1	6.8
2005	7.6	6.4	8.3	10.1	15.0	11.1	11.8	11.7	10.0	7.0	5.8	3.6
2006	6.8	9.1	10.2	10.7	10.5	11.6	12.3	12.1	10.4	8.5	7.6	8.7
2007	9.0	9.6	9.2	10.7	11.1	12.8	11.9	11.7	10.7	7.2	6.1	4.2
2008	6.7	6.5	7.3	9.7	12.1	12.5	12.4	11.7	10.9	8.4	6.4	5.8
2009	7.5	8.2	9.3	10.1	10.2	12.1	12.1	12.1	10.3	7.3	5.6	8.9
2010	7.1	10.2	9.5	11.6	11.8	11.7	12.1	11.8	11.1	8.2	6.5	6.0
2011	7.6	6.4	8.4	10.2	11.2	11.8	12.2	11.8	13.4	7.1	8.1	5.5
2012	6.0	4.9	7.7	10.6	10.6	11.6	11.8	11.3	10.4	6.0	6.4	5.6
2013	6.5	6.6	10.1	11.0	10.6	11.9	11.8	11.1	9.6	7.8	6.3	4.4
2014	6.6	8.9	9.5	9.8	10.3	10.7	11.1	10.6	10.1	8.2	6.8	4.6
2015	5.3	6.3	7.8	8.3	11.8	12.2	12.3	12.3	11.4	8.5	7.6	8.9

Annex 1.10 Dessie T max

Year	Jan	Feb	mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nev	Dec
2008	21.1	22.0	24.9	24.1	25.2	26.6	24.8	23.4	22.7	22.8	20.3	21.0
2009	20.9	22.4	24.2	23.9	25.5	26.9	24.3	23.7	24.7	24.2	23.4	23.0
2010	22.1	22.9	22.3	23.4	24.9	27.0	25.7	26.3	25.5	24.0	22.9	21.5
2011	21.0	23.6	22.4	24.8	25.0	26.5	24.9	23.5	24.1	23.3	22.0	22.0
2012	22.9	23.1	21.3	23.4	24.8	26.6	24.9	24.0	23.9	24.2	23.9	23.4
2013	23.1	23.9	24.9	25.6	25.7	26.7	24.8	23.1	24.4	23.6	24.2	22.1
2014	22.5	22.7	24.4	24.8	24.2	27.3	27.5	26.5	23.0	24.8	24.1	22.4
2015	20.7	23.6	24.2	24.4	24.6	25.8	26.5	25.0	23.7	23.4	22.3	20.6

Annex 1.11 Dessie T min

Year	Jan	Feb	mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nev	Dec
2008	5.6	4.6	6.3	8.1	10.0	9.9	11.5	11.0	10.1	7.4	7.5	4.5
2009	6.0	7.6	8.7	9.1	9.2	11.0	11.5	11.2	9.5	8.2	4.2	5.0
2010	7.3	9.4	11.4	10.7	9.7	10.5	10.6	10.6	9.1	6.2	5.0	5.5
2011	6.7	5.2	6.4	8.1	9.5	9.7	10.3	10.0	8.4	6.0	6.6	3.2
2012	3.6	3.2	8.4	9.0	8.3	8.6	9.5	9.3	8.4	4.5	4.2	2.9
2013	4.1	4.4	8.2	10.9	11.2	12.7	12.4	10.5	9.8	8.2	6.8	5.4
2014	6.4	7.2	8.6	10.2	10.0	10.4	11.3	10.9	9.5	7.5	6.8	5.5
2015	5.6	6.5	8.0	8.3	10.7	11.0	11.0	11.4	10.6	8.2	8.3	8.9

Annex 1.12 Kombolcha T max

2002	24.1	27.1	27.5	27.7	30.2	30.7	28.1	26.4	26.3	26.1	25.8	24.5
2003	26.1	25.5	27.2	26.5	30.1	29.9	28.7	27.5	26.4	25.2	25.5	24.9
2004	25.0	28.1	28.1	28.1	27.8	30.7	27.8	27.3	26.9	26.0	25.8	25.3
2005	25.6	27.2	27.1	26.3	29.0	31.4	28.0	27.1	26.4	26.7	25.4	24.5
2006	23.6	26.3	28.2	27.6	30.4	30.4	27.1	27.3	27.3	26.2	25.3	25.5
2007	26.3	26.5	29.3	29.0	30.2	31.1	29.2	27.3	27.2	26.3	24.8	25.1
2008	25.2	27.2	28.6	29.3	30.4	32.1	28.2	27.6	28.0	26.2	26.7	25.0
2009	25.6	26.6	26.6	28.5	29.2	31.5	28.6	27.1	27.0	27.0	25.9	24.3
2010	24.8	27.1	26.4	29.5	28.4	30.1	28.6	26.7	27.2	26.3	24.7	23.2
2011	26.4	27.6	28.3	26.9	28.6	30.2	28.3	27.5	27.6	26.7	26.9	26.1
2012	26.3	27.5	28.5	29.2	29.7	30.8	28.2	26.4	27.4	26.1	25.8	24.8
2013	25.6	26.0	27.7	28.7	28.3	30.9	29.6	27.1	26.7	25.7	25.8	24.8
2014	25.6	26.0	27.7	28.7	28.3	30.9	29.6	27.1	26.7	25.7	25.8	24.8
2015	25.1	28.1	28.6	29.6	29.6	30.6	31.4	28.5	27.6	27.6	26.1	24.9

Annex 1.13 Kombolcha T min

2002	10.7	11.9	13.3	14.1	14.2	15.0	15.0	15.4	14.6	9.4	8.5	8.1
2003	12.1	10.5	11.2	14.4	13.0	13.5	14.8	14.6	17.9	9.5	8.6	10.3
2004	13.6	10.0	13.3	13.8	15.1	15.1	15.2	15.2	14.4	9.7	8.1	5.7
2005	9.6	12.5	12.8	13.9	13.9	14.8	15.3	15.2	14.0	11.8	10.1	12.1
2006	11.3	12.5	12.0	14.0	14.5	16.5	15.4	15.1	14.4	9.7	8.6	6.5
2007	9.2	9.0	8.9	12.9	15.4	15.7	15.8	15.1	14.3	11.3	9.4	8.0
2008	10.1	11.1	12.3	13.5	14.0	16.2	15.4	15.4	14.0	11.5	8.7	12.2
2009	10.4	13.6	13.2	15.1	15.5	16.0	15.7	15.1	14.5	11.6	9.3	9.5
2010	11.0	9.8	11.7	13.8	15.1	15.4	15.7	15.3	13.7	10.9	11.9	8.4
2011	8.8	7.6	10.5	13.8	13.9	15.2	15.4	15.2	14.5	9.8	9.6	9.4
2012	10.1	10.0	14.2	15.3	15.3	16.8	16.0	15.4	14.1	12.0	10.7	7.5
2013	9.9	13.3	13.6	14.5	14.8	15.2	15.3	14.9	14.0	11.5	10.6	7.9
2014	9.9	13.3	13.6	14.5	14.8	15.2	15.3	14.9	14.0	11.5	10.6	7.9
2015	8.8	9.5	11.7	12.2	15.1	15.4	15.9	15.4	14.8	11.8	11.3	12.3