



Numerical groundwater flow modelling under changing climate in Abaya–Chamo lakes basin, Rift Valley, Southern Ethiopia

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Abstract

Ongoing socio-economic developments and climatic change have been pressurizing the groundwater resource availability in the Abaya–Chamo lakes basin, Ethiopian Rift valley. The primary goals of the present study are: (1) to simulate the groundwater gradient and flow direction, (2) to calculate the groundwater balances and flux of the sub-major river basins under the water budget code of the MODFLOW, and (3) to predict the future groundwater levels of the lake's basin under a projected changing climate. The numerical groundwater flow of the Abaya–Chamo lakes basin aquifer system is simulated using the USGS three-dimensional finite-difference groundwater flow model MODFLOW-2005 with Block centered flow packages (BCF). The following datasets, such as aquifer properties, geology, recharge, discharge, topography, etc., were used to simulate the present model. The calibrated steady-state groundwater flow modeling simulation of the Abaya–Chamo lakes basin also confirmed the through-flow system in terms of groundwater gradient and flow direction, on which groundwater flow happens from the plateau toward the floor into the lakes from both directions with a high gradient exist in the escarpment. The present study provides a sound foundation for modern scientific direction in water resource evaluation by establishing integrated surface and groundwater models that change climatic conditions for sustainable water resources management.

Keywords Groundwater modelling · MODFLOW · Abaya–Chamo lakes basin · Rift valley · Ethiopia

Introduction

Worldwide, groundwater is an essential natural resource because it is well protected from surface contamination and is less affected by seasonal changes than surface water

(Woldeamlak et al. 2007; Mogheir and Ajjur 2013; Zam-mouri et al. 2007, 2014; Zektser and Everett 2006). The recharge process, which may be described as the rainfall portion that enters the aquifer and contributes to its replenishment, is the primary source of groundwater resources (Tilahun and Merkel 2009; Barron et al. 2012). The study of groundwater recharge is a critical step in assessing and managing the long-term usage of groundwater resources (Shaban et al. 2006; Dams et al. 2008; Chung et al. 2010; Chenini et al. 2009; Yeh et al. 2009; Yun et al. 2011; Wang et al. 2012; Scanlon et al. 2006; Tilahun and Merkel 2009; Rwanga 2013). It is impractical to evaluate groundwater recharge in situ over a broad region; it is typically calculated indirectly (Scanlon et al. 2006; Bakker et al. 2013). Due to its complexity, the actual groundwater system might be challenging to comprehend (Marnani et al. 2010; Moiwo et al. 2010).

The development of conceptual models is a valuable tool for simplifying the complicated underground water system. A reliable and accurate conceptual model is essential for a numerical groundwater flow model that decreases error

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(Chiang and Kinzelback 2001; Park et al. 2014; Mengistu et al. 2019). The numerical model was utilized to comprehend groundwater system change and flow direction. It was used to assess recharge, outflow, and aquifer storage. The model helps analyze the response of the groundwater system and future forecasting situations (Zhou and Li 2011). MODFLOW, a computer application, solves groundwater flow issues using numerical models (Wang et al. 2008; Tamal et al. 2014; Post et al. 2019). Many researchers have utilized MODFLOW, a groundwater flow simulation tool, to explore groundwater flow and behavior in various parts of the world (Idrissy and Smedt 2006; Khadri and Pande 2016; Bushira et al. 2017; Prasad and Rao 2018; Azeref and Bushira 2020; Gebere et al. 2020; Hassen et al. 2021; Jafarzadeh et al. 2021; Raazia and Dar 2021).

Groundwater is being extracted intensively in Ethiopia, including the present study area, without thoroughly examining the groundwater system (Azeref and Bushira 2020). Only limited hydrological/ hydrogeological studies were carried out in the present study area, the Abaya–Chamo lakes basin (Bekele 2001; Halcrow 2008; JICA 2012; Molla and Tegaye 2019; Molla et al. 2019). In the case of the Abaya–Chamo lakes basin, southern Ethiopia, the above studies assessed the quantity and identified water quality of surface and sub-surface compartments as a separate and independent system, sometimes at the river basin level; however, these studies provide a preliminary overview of the climate hydrology and hydrogeology of the region. None of these studies adequately address important groundwater integration and interaction with surface water and climate change issues as one unit basin system. With this background, the present study was conducted to simulate the groundwater gradient and flow direction, to calculate the groundwater balances and flux of the sub-major river basins under the water budget code of the MODFLOW, and to predict the future groundwater levels of the lake's basin under a projected changing climate.

Materials and methods

The present study has used a range of modeling techniques ranging from point-based, empirical, conceptual, soil water balance and lumped to more complex distributed models. The crucial parameters that affect the formation of water resources, such as time series meteorological and hydrological, hydrogeological data that cover all available measurement and periods, and geophysical spatial data, were collected, reviewed, and parameterized from different sources, including in the field. Models and Softwares used in the present study are the WetSpss models, MODFLOW-2005, and ArcGIS. MODFLOW is a computer program developed by the United States Geological Survey that numerically solves

the groundwater flow equation for porous media using the finite-difference method. Hydrogeologists commonly use the program to simulate groundwater movement and contaminant transfer. MODFLOW was designed to allow the user to pick and select which modules to use throughout a simulation. Each module examines a different facet of the hydrologic system (e.g., wells, recharge, and surface water bodies). The choice of analytical or numerical models is critical, and it is impacted by the complexity of the hydrogeological situation and the availability of field data. The current research developed a numerical model for groundwater flow solutions. In general, steady-state models operate quickly. They can be subjected more readily to computer-assisted history matching and uncertainty analysis.

Data used: long-term spatial hydro-meteorological data (available precipitation, temperature, potential evapotranspiration, wind), physical basin data (such as elevation, slope, land-use/land cover) from the Landsat-8 satellite image, and shuttle radar topographic mapper (SRTM) were downloaded using path/raw from the US Geological Survey website (<https://earthexplorer.usgs.gov/>). Hydrogeological data (includes groundwater depth, identification of aquifer systems, hydro-stratigraphic variations, hydrogeological properties, etc.). Well parameters such as specific storage, hydraulic conductivity, well yield, well lithological layer details, soil type, hydraulic heads, daily rainfall collected from National Meteorological Service (NMS), Ministry of Water, Irrigation, and Energy (MOWIE), and National Meteorological Agency (NMA), Hawassa, Ethiopia.

The study area

The study area, the Abaya–Chamo lakes basin, is located in the southern Main Ethiopian Rift valley, as given in Fig. 1. Geographically the region is bounded within 37° 15' 36" E to 38° 42' 58" E in the Eastern longitude and 5° 25' 5" N to 8° 07' 4" N in the northern latitude. The total basin area is 18905.5 km². The Great Rift Valley has dissected the country into western and eastern parts. The highlands have a height range of 2400–3700 m, with Mount Ras-Dashen reaching 4620 m above sea level. The Rift Valley has the lowest altitude of 120 m below sea level. The region's height ranges from 1107 to 3430 m above sea level. The basin is divided into three areas viz: the highland plateau (1833–3430 m), the transitional escarpment (1404–1833 m), and the rift floor (1107–1404 m). The drainage basin comprises two low-lying lakes: Abaya and Chamo, with 1109.9 km² and 316 km². This research focuses on the entire lake basin. Major rivers primarily end in Lake Abaya and overflow periodically to Lake Chamo via local grabens.

In the study area, six major rivers flow into the two lakes. Billate (5659.5 km²), Gidabo (4199.1 km²), Gelana (3865.6 km²), Hamessa-Guracha (1006.7 km²), Kulfo-Gina

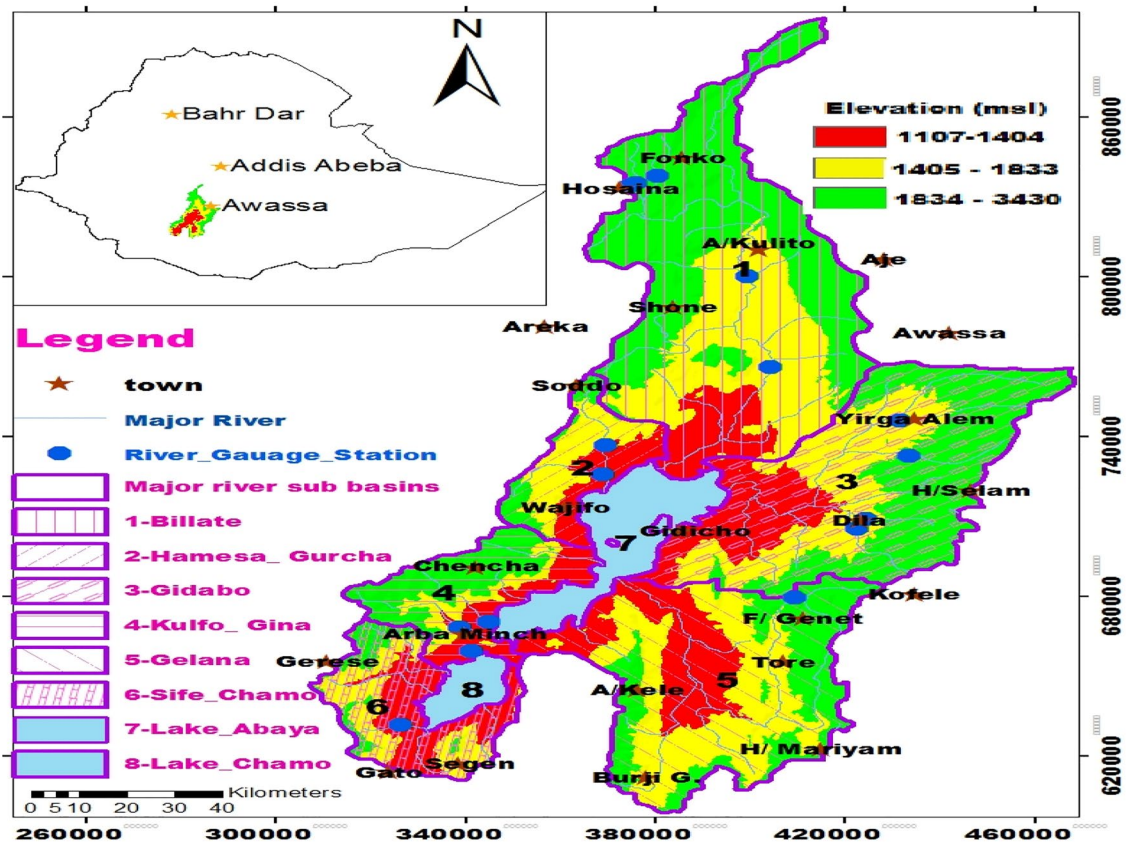


Fig. 1 Location, elevation, and major river basins of the study area

(1368.7 km²), and Sife-Chamo (1379.4 km²) are among the rivers with corresponding catchment areas (JICA 2012; Halcrow 2008). The study focuses on these major river sub-basins.

Hydrogeological conceptual model

A hydrogeological conceptual model is a graphical representation of a groundwater flow system, typically in a simplified hydrogeological diagram or cross-section. The first step in the groundwater modelling process is to create a database for modelling. The data is collected from various organizations, formatted, and input for a groundwater model. The data collected is processed in a GIS environment. The type of aquifer is also determined by the hydraulic properties of the water (Winkler et al. 2003). As a result, numerical groundwater flow modeling depends on a thorough understanding of the hydrostratigraphic system, conceptualization, and development. The occurrence of groundwater and flow consists of the relevant parts of the lakes basin such as highlands, escarpments, and rift floor, based on converging evidence from exploratory inventories of previous studies in the Abaya–Chamo lakes basin. This can be described by

dividing the lake into Eastern and Western compartments on the rift floor.

Conceptual hydrogeological model of the eastern compartment

The eastern highland ridge chain ranging from 2000 msl to over 3000 msl is associated with high rainfall distribution. The outcrop in the northeastern area is covered mainly with various volcanic rocks intercalated by volcano-clastic and sedimentary rocks. At the same time, the southeastern part consists of basement rocks. The recharge is only relatively good in areas where thick alluvial sediments cover the highland. Demis (2009), Halcrow (2008) described the groundwater flow system and direction in the escarpment area, where faults primarily control the flow. Groundwater contributions from springs in this Eastern region are relatively small in mountain areas, including hot springs at Yirgalem. Furthermore, eastern highland aquifers can also be recharged volcano-sedimentary rocks of the rift bottom partly through deep regional groundwater flow. Groundwater from the escarpment area recharges the aquifers on the rift floor in addition to the direct recharge from precipitation.

Conceptual hydrogeological model of the western compartment

The western highland is built by faulted basalts and ignimbrite intercalated by volcano-clastic and sedimentary rocks. Similar to the Eastern region, the outcrops form a gently undulating plain that receives adequate rainfall, resulting in the moderate runoff with good infiltration and the formation of moderate to highly productive fissured and mixed aquifers. Deeper fissured aquifers formed in underlying volcanic rocks are also recharged by aquifers outcropping in the area. In mountain areas, springs are typically small. Springs also at the foot of the escarpment represent deep local groundwater flow, including a group of Arba Minch-springs in Arba Minch. The existence of deep regional groundwater flow is also confirmed by a group of hot springs and hot water along the Bilate-river in the Abaya geothermal area.

Groundwater from the escarpment area recharges the volcanic and sedimentary aquifers of the rift floor in addition to the direct recharge by precipitation. The groundwater from the western escarpment and western part of the rift floor is drained by the Bilate-river. A simple conceptual hydrogeological model of the west and the eastern region is shown in Fig. 2. To summarize, the overall conceptual design principles of the Abaya–Chamo lakes basin can be clarified by the main recharge and discharge mechanisms. These are (a) direct recharge to outcropping aquifers, (b) vertical recharge from overlying aquifers into underlying aquifers, (c) horizontal recharge from neighboring aquifers and lakes (c) direct discharge by springs from outcropping

aquifers (cold and hot springs at the foot of the escarpment and the rift floor) (d) direct discharge to rivers and lakes (e) indirect discharge from one aquifer to another (vertical as well as horizontal). As a result, the groundwater flow is generally parallel to the surface water flow system initiated from the highlands through the escarpment to the rift floor. The relative elevations of static water levels between the individual sub-basin lakes govern groundwater flow direction on the rift floor.

Modeling approach

The USGS three-dimensional finite-difference groundwater flow model MODFLOW-2005 with Block centered flow packages (BCF) is used to simulate the numerical groundwater flow of the Abaya–Chamo lakes basin aquifer system. MODFLOW-2005, the most recent version of MODFLOW, was released in 2005 (Harbaugh 2005). MODFLOW-2005 versions that incorporate the use of parameters to define model input, the calculation of parameter sensitivities, and the modification of parameter values to match observed heads, flow, or advective transport using the observation, sensitivity, and parameter estimation processes described in detail by (McDonald and Harbaugh 2000; Harbaugh 2005).

The Eq. (1) used in the computer model to describe groundwater flow is:

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

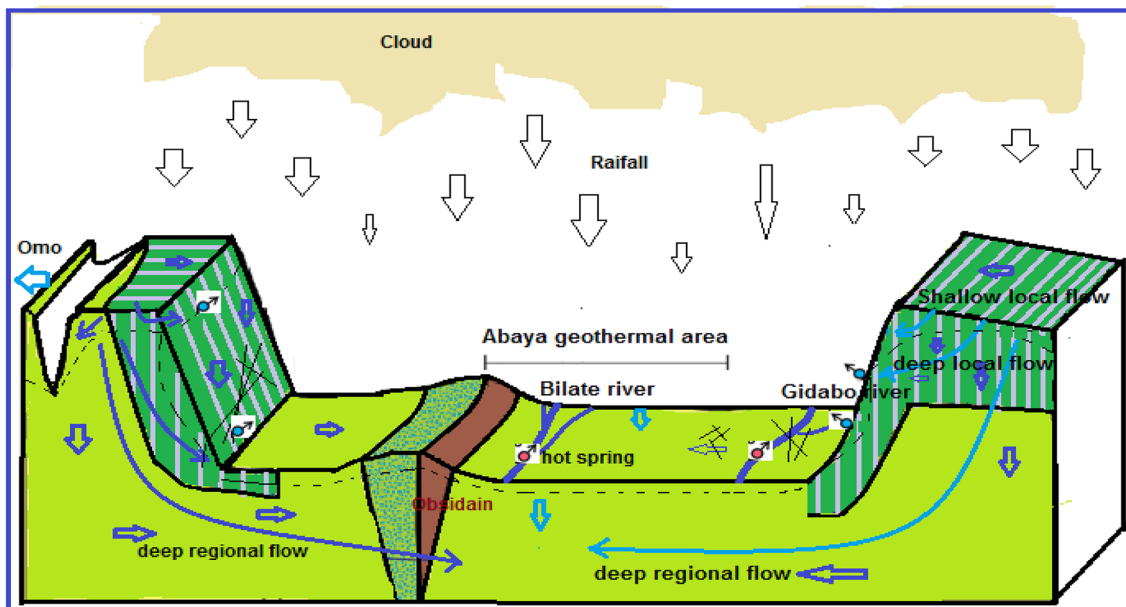


Fig. 2 The schematic diagram for the Western and eastern compartment

where: K_x , K_y , and K_z : are the values of hydraulic conductivity along x , y , and z coordinate axes, respectively and are assumed to be parallel to the major axes of hydraulic conductivity, in meters per day; h : is the potentiometric head, in meters; W : is a volumetric flux per unit volume and represents sources or sinks or both of water, such as well discharge, recharge and water removal from the aquifer by drains, per day; Ss : is the storage coefficient of the porous materials, per meter; t : is time, in days.

The Preconditioned Conjugate-Gradient with Improved Nonlinear Control (PCGN) package solves the flow Eq. (1) for heads in each cell as a method for solving matrix equations using an algebraic multi-grid solver in the finite-difference grid, via MODFLOW, which expresses the relationship between the head at a node (Harbaugh 2005). Determining boundary conditions for the aquifers, calculating recharge and discharge rates, and estimating aquifer attributes within the model all contribute to the definition of the flow system. To some extent, the model's accuracy is determined by this input data. The model was calibrated in the steady-state model with the water levels collected at different times during well-drilling, pumping test, and inventories. Hydraulic parameters were iteratively adjusted by trial and error until good matches to measured variables were achieved in the steady-state simulation with a constant input of spatial recharge estimated by the WetSpa model.

Model description

Grid design

The present model area covers the entire lakes basin on which five sub-major river basins were included, as shown in Fig. 3. The model uses a grid size of 500 m by 500 m and contains two layers, 343 columns, 615 rows, and 210,945 cells in each layer. The aquifer was discretized vertically into two layers (layer-1 and layer-2) (Fig. 4). Layer-1 corresponds to entire alluvium and lacustrine deposits, which range in thickness up to 45 m in the floor (near to lakes), and the thickness decrease in the escarpment. Using lithological well completion reports, a thickness of 302 m is determined for Layer-2, the volcanic and worn bedrock beneath the alluvial sediments of the area.

In the area, groundwater aquifers are multi-layered due to spatial geologic variability that is obvious from the pumping test and lithologic units and the presence of less permeable layers such as clay (JICA 2012). The lithological layer covering the upper aquifer is not homogenous, but most of the part of the aquifer is unconfined in the Abaya–Chamo lakes basin.

Using unconfined aquifer hydraulic parameters, Layer-1 represents the upper unconsolidated high-permeability aquifer thicknesses 20–45 m (Aquifer type-1). Layer-2 corresponds to the upper weathered, and fractured bedrock

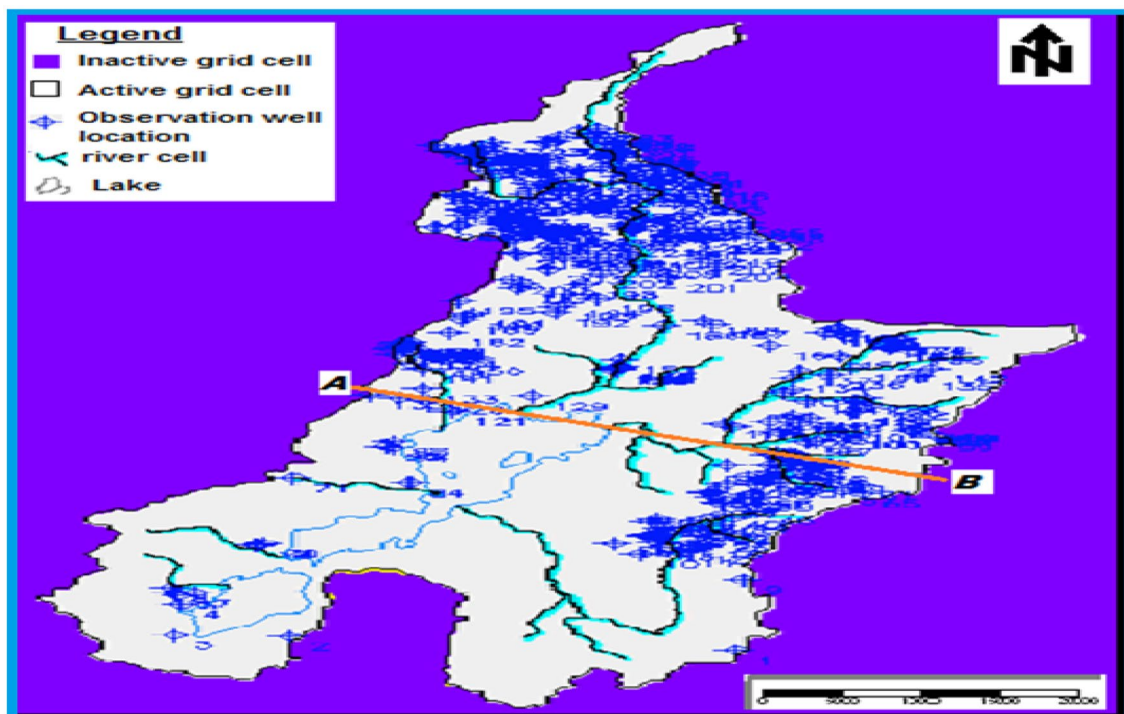


Fig. 3 Model area, model grid, and observation well in 2D Plan

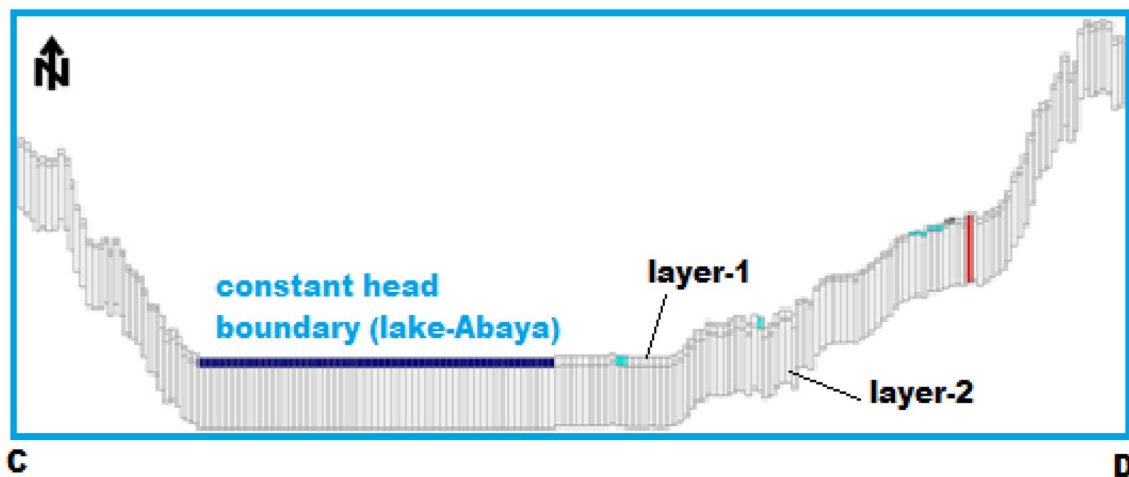


Fig. 4 Cross-sectional view of the model area

underlay the alluvial aquifer. This layer was assumed to be confined aquifer hydraulic properties (Aquifer type-0). In the studied area, the two layers of the aquifer system have varying thicknesses due to lithological type and permeability variances. As a result of the SRTM (DEM) 30 m resolution of the land surface and available lithologic wells, the estimated formation thickness was used to establish the height of each cell in the model. Ground surface elevation above sea level can be determined by looking at the top altitude of layer-1. In the study region, more than 240 drilling lithology log samples were used to calculate the bottom altitude of this layer, which corresponds to the bottom of alluvial material. The lower altitude of the Layer-1 bottom is also taken as the top elevation of the Layer-2. The bottom of layer-2 is taken from the average volcanic and weathered bedrock thickness estimated from the depth of lithological well log available in the study (Fig. 4).

Boundary conditions

Boundary conditions represent quantitative specification of dependable variables such as head and related flux variables at the boundaries of the model domain (Anderson and Woessner 1992). In common sense, model boundaries are supposed to be the actual hydrologic boundaries of the problem domain. In principle, generally, boundary conditions are categorized as (a) a specified-flux boundary or no flow boundary, (b) a specified-head boundary, and (c) a head-dependent flux boundary, for which the boundary flux is the product of a specified factor and the difference between the simulated head at the boundary and specified head of an external source/sink.

Both lakes (Abaya and Chamo) were considered as specified constant head boundary cells and limited to the first model layer for simplification since the depth of the

lakes does not exceed the model thickness. The water levels for constant head boundary cells to lakes were their mean water level elevation, about 1176 m and 1109 for Abaya and Chamo Lake, respectively. Whereas, the major rivers were simulated by river package where the water heads in the river-cells were assumed nearly more remarkable by 0.5–2 m toward downstream from the average elevation in each cell.

Volcanic mountains bound the lateral basin boundaries of the model area are assigned to be no-flow boundaries at these surface water divides. So that at this condition, there is no groundwater flux entering into the modeled area was assumed from the ridges of surrounding basins of Abaya–Chamo lakes basin. Topographically, the lower area, particularly in the south part, studies the area along which surface water and groundwater outflow are also considered as the head-dependent flux boundary. At these localities in Fig. 5 (yellow shades), the model is simulated with the General-Head-Boundary (GHB) module for both model layers.

The model top boundary or layer-1 or the upper boundary was considered a free surface boundary, including head-dependent flux and specified-flux at boundary cells. Here, spatially distributed groundwater recharge estimated by the WetSpa model was given as the specified-flux boundary, and the head-dependent boundary represents springs and groundwater seeps from central river beds. Recharge was specified and simulated with the recharge (RCH) module. The bottom boundary of the model is a defined no-flow boundary. This no-flow boundary is where the aquifer comes into contact with massive bedrock.

Model input parameters

Aquifer properties and stresses are essential in the simulation, which varies horizontally and vertically. Thus, these properties are subjected to a range of values since the basin is

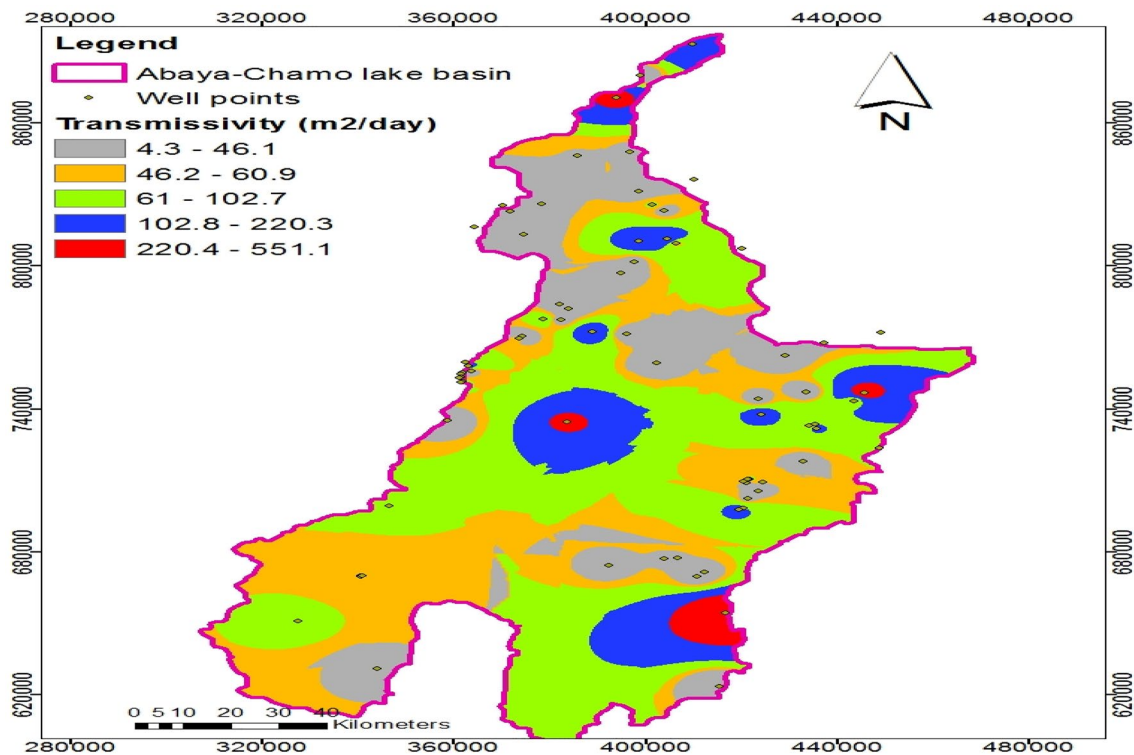


Fig. 5 Regionalized transmissivity map from available pumping test data

characterized by high topographic, geologic, and hydrogeological variation as a complex rift margin zone. Therefore, the properties cannot be precisely parameterized in a numerical flow model. The aquifer properties and fluxes specified for each active cell in the model estimate the average conditions in the area represented by the cell. The initial aquifer properties were calculated from the pumping test analysis and previous hydrologic and hydrogeological studies conducted in the area. Indeed, the aquifer properties were modified in the possible range during groundwater flow calibration.

Initial and prescribed hydraulic heads

MODFLOW requires initial hydraulic heads to solve the numerical solution for the grid cell. In principle, the initial hydraulic heads at constant head cells are used as specified head values and remain consistent throughout the flow simulation. Since the simulation is a steady-state flow, the initial heads are used only to solve the iterative equation using block-centered finite-difference. Thus, initial heads at the constant head cells were assigned to the actual values while all other

values were set at an arbitrary level within the thickness of the layer. Therefore, the initial hydraulic head of a constant head cell was prescribed in between the top and bottom layer elevation for both assigned unconfined and confined aquifer layer types. The elevation difference below 15 m was assumed as the initial and prescribed hydraulic heads for corresponding layers. This helps fill the incapability of MODFLOW in converting a dry fixed-head cell.

Horizontal/vertical hydraulic conductivity and transmissivity

In the development of groundwater flow simulation, the specification of horizontal and vertical hydraulic conductivity values of the aquifer system is essential because groundwater flow within the flow system across the model layers was theoretically assumed to be flat. This means combining this aquifer property with a hydraulic head gradient determines the groundwater flow direction. Materials with a higher hydraulic conductivity are coarser and more porous. Aquifer system hydraulic conductivity on the horizontal

plane was distinguished from earlier investigations in the study area. When evaluating litho-hydro-stratigraphic distribution in the model domain, the information gathered from the most readily available transmissivity data are shown in Fig. 5. Finally, the conductivity was grouped into different zones, and each of them was assigned to the active cells of model layers.

The complication of hydraulic conductivity, such as aquifer types and saturated thickness, increases with transmissibility in different locations for the same aquifer. In addition to the specific data in the study area, previous hydrogeological studies (Halcrow 2008; JICA 2012; Kefale and Jiri 2013) summarized transmissivity values for the major aquifers as converging information based on all available wells in the main Ethiopian rift valley, which includes the study area (Table 1).

Based on the literature evidence, the vertical hydraulic conductivity to horizontal hydraulic conductivity ratio is between 0.01 and 0.1 (Anderson and Woessner 1992). Therefore, the initial vertical hydraulic conductivity of the aquifer units for the study area was estimated at 10% of the horizontal hydraulic conductivity for calibration purposes. These values are assigned to the active cells of the model layers.

Model stresses

Potentiometric heads distribution in the groundwater flow system responds to the external flux stressed on the aquifer system. The significant water fluxes can be related to the amount and distribution of recharge to the study area's groundwater system from the infiltration of precipitation and sub-surface inflow to the model area at the head-dependent flux boundaries and river beds. At the same time, discharge from the system is through pumping well, groundwater outflow at the river beds, springs, and subsurface outflow from the modelled area at the general head boundary.

Result and discussion

Groundwater recharge

Groundwater recharge to the ground system results from the infiltration of precipitation left from actual evapotranspiration, interception, and surface runoff (Abu-Saleem et al. 2010; Al Kuisi and El-Naqa 2013). The spatial distribution of recharge rate in the study area was determined by spatially distributed surface water balance model (WetSpss) as given in Fig. 6 by considering different basin features. The calibrated surface model output estimated about 113.7 mm/yr as an average groundwater recharge for the study area. This spatially distributed recharge was then applied to the top active cell of the model as a spatially varying specified flux to the uppermost active layer.

Groundwater withdrawals

The steady-state simulation of the Abaya–Chamo lakes basin considers outflow flux from the groundwater system. These outflows include abstraction from pumping wells of the aquifer system for water supply and irrigation in the area, spring discharges from the system. The available measured flow was collected, which shows the general spatial distribution and relative amount of all outflows in the basin. The well-package represented this water pumping from groundwater system through wells in the processing ModFlow. The river leakage systems were also simulated using a river package to flow into/out the groundwater system. The general head boundary condition stimulated the lateral subsurface inflow/outflow into/out from the model area southeast of Lake Chamo. Furthermore, losses from the groundwater system through groundwater evaporation were also assumed around the northeast of Lake Abaya in the rift floor. The evaporation module represented this parameter in ModFlow as the outflow from the aquifer system.

Table 1 Range of aquifer transmissivity values of some major aquifer types

Aquifer types	Transmissibility (m ² /day)		
	Minimum	Maximum	Average
Fine sand (unclassified fluvial deposits)	90.2	388.0	239.1
Lacustrine sediment	10.0	2080.0	1045.0
Rhyolite	2.5	376.5	189.5
Volcanic (trachyte, tuff, volcanic sediments)	1980.0	3801.0	3801.0
Volcanic & sedimentary rocks	2.0	3801.0	1901.5
Scoria	158.7	158.7	158.7
Basalt	64.0	79.0	71.5
Welded tuff	12.5	914.0	463.3

Fig. 6 Total annual groundwater recharge map (mm)

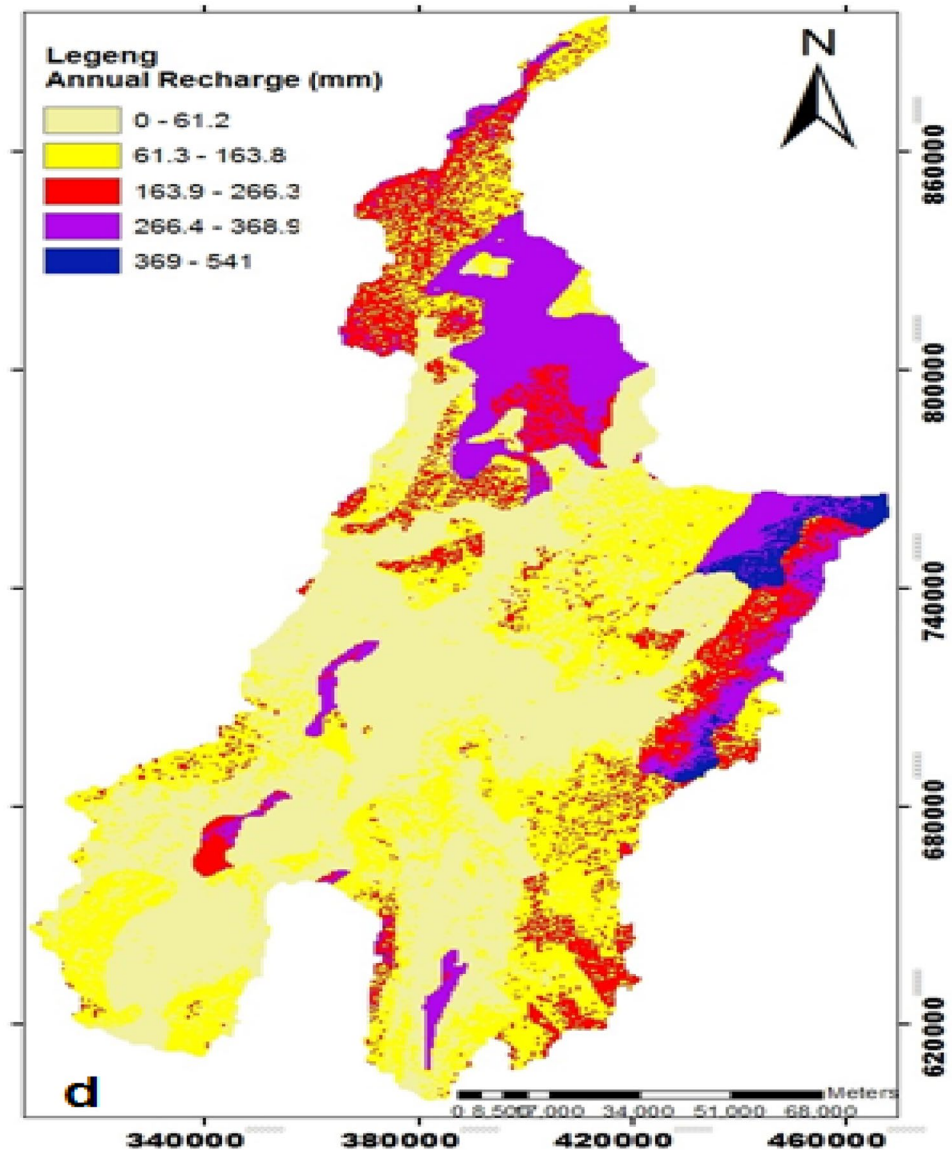


Table 2 Simulation result of long-term groundwater balance of the study area (m³/day)

Water budget of the whole model domain			
Flow term	In	Out	In–out
Storage	0.00E+00	0.00E+00	0.00E+00
Constant head	1.97E+04	4.99E+06	– 4.97E+06
Wells	0.00E+00	8.21E+05	– 8.21E+05
Drains	0.00E+00	0.00E+00	0.00E+00
Recharge	5.80E+06	0.00E+00	5.80E+06
Et	0.00E+00	7.34E-01	– 7.34E-01
River leakage	1.05E-00	6.63E+01	– 6.53E+01
Head dep bounds	5.20E+02	5.49E+03	– 4.67E+03
Sum	5.82E+06	5.82E+06	– 6.90E+01
Discrepancy [%]	0.00		

Model calibration

The model simulation includes the entry of organized input data and interpretation of the model results. Calibration of the model is achieved when simulated results are compared with the observation in an acceptable range with a minimum error. The error level is permitted when the input parameter values are adjusted within reasonable ranges through an interactive and iterative method. Indeed, trial and error calibration was applied (Anderson and Woessner 1992). Input parameters are manually tweaked so that the model simulates the observed heads within the error criterion's range. The calibration was conducted by keeping out recharge constant with varying aquifer hydraulic conductivity, well discharge, and riverbed material conductance values as given in Table 2. The hydraulic conductivity zones were

systematically extrapolated following converging information and hydrogeological map to achieve the best fit results.

Steady-state model simulation

The steady-state of the groundwater flow model provides the quantified groundwater balance, fluxes, and flow direction of the system and the water flux between surface and groundwater. The balance between the inflow and outflow can be used to express the steady-state flow conditions with null aquifer storage (Eq. 1 the term: $S_s \frac{\partial h}{\partial t} = 0$). The groundwater flow modeling of the Abaya–Chamo lakes basin using steady-state simulation was based on the water level measurement of 376 observation wells. Qualitative and quantitative evaluations of the calibration data were also conducted (Anderson and Woessner 1992). It was determined by comparing the observed and simulated heads' average differences. A closer simulated water level result against observed water level should be a quantitative statistical error description criterion. Qualitatively, the simulated head distribution result was supposed to be comparable to the conceptual model's groundwater gradient and flow direction. The water table is known to follow the topography with gentler

fluctuation generally. The simulated steady-state calibration results in 2D/3D-diagram as given in Fig. 7) illustrate that the highest groundwater level will be found along the basin boundary (chain of mountain ridges), flows toward low altitude into the lakes from all directions.

As predicted, the contours and gradients of the potentiometric surfaces were simulated. As discussed in the previous chapters, hydraulic head from the observation wells, groundwater flow generated from the plateau toward the floor into the lakes from both directions with a high gradient in the escarpment. Similarly, the simulated potentiometric surface in the two and three dimensions given in Fig. 8 confirms the flow gradient and direction. In other words, the simulated steady-state potentiometric contour generally appeared similar to the measured average potentiometric surface. The quantitative comparison was made between the simulated and measured hydraulic heads using statistical methods to test the calibration match. Measured hydraulic heads and simulated hydraulic heads are compared to evaluate their overall goodness-of-fit (Mean Error ME and Mean Absolute Error MAE) and the same was calculated by using the following Eqs. 2 and 3.

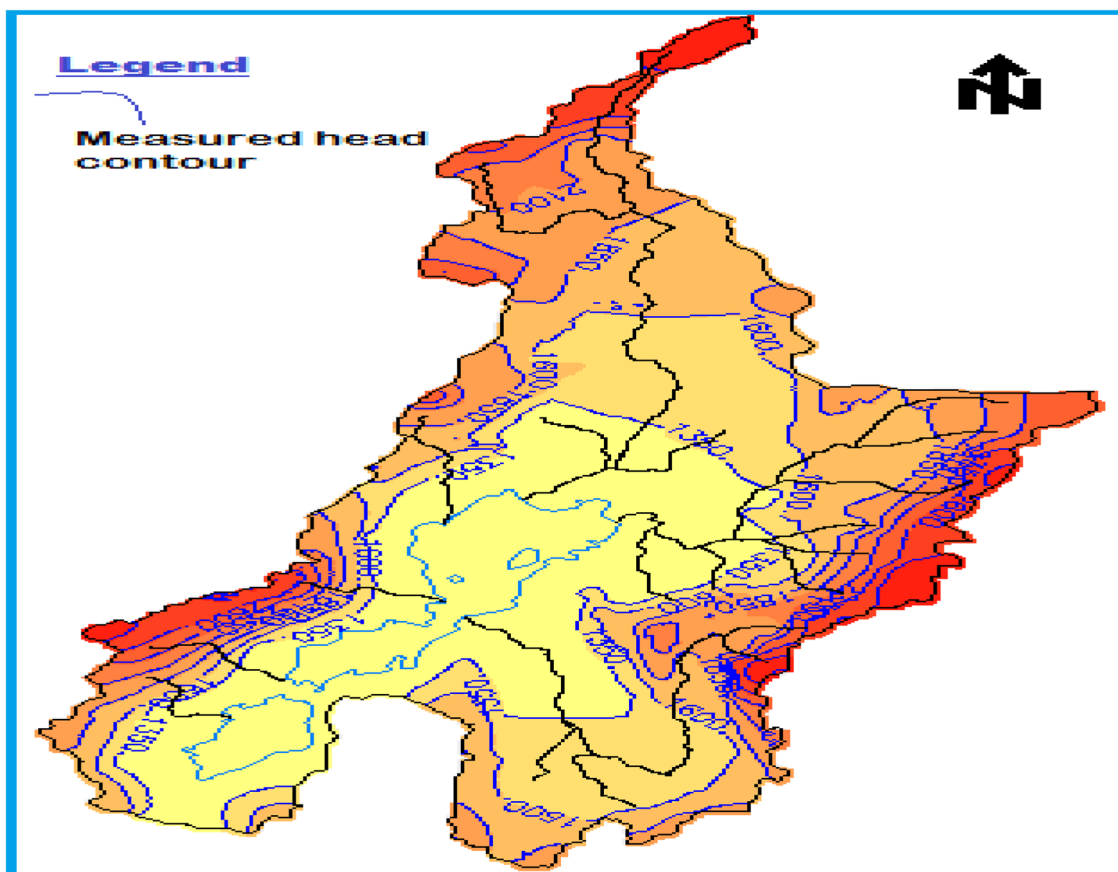


Fig. 7 Two-dimensional map showing the simulated hydraulic head contour

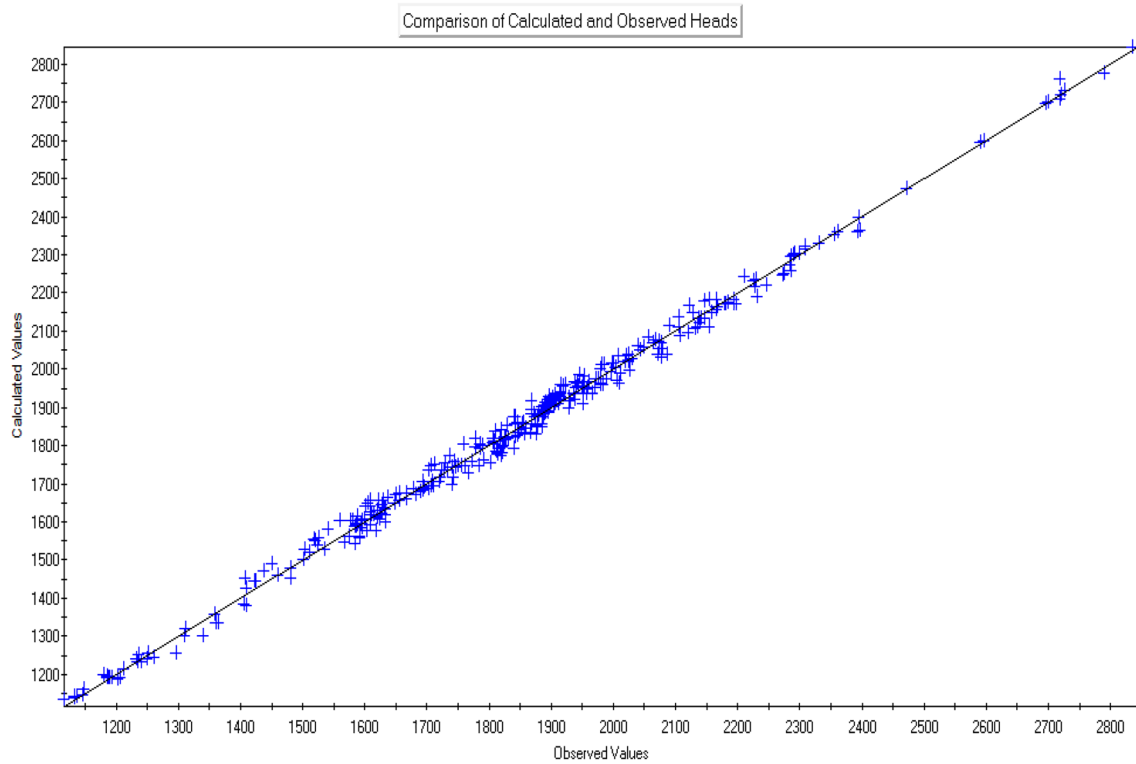


Fig. 8 Observed and calculated head comparison graph

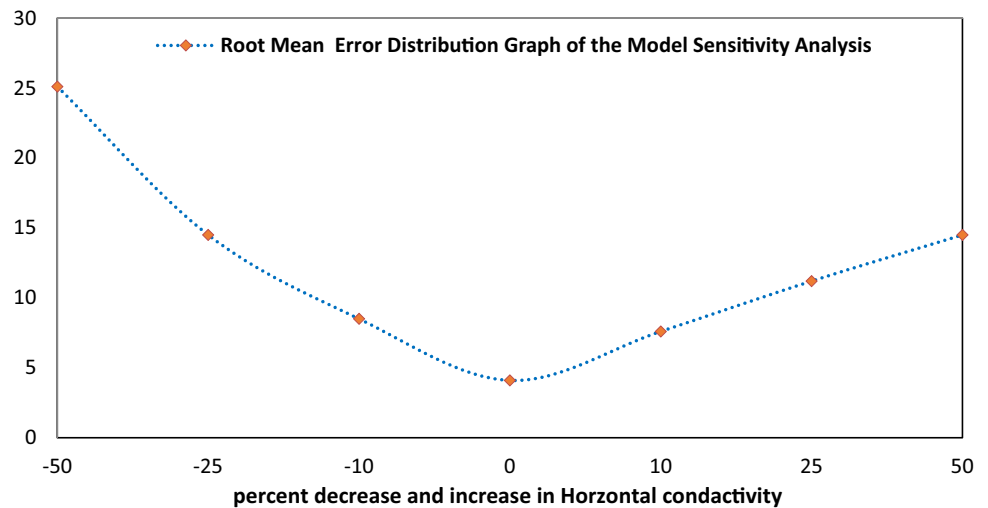
$$ME = \frac{1}{n} \sum_{i=0}^n (h_o - h_s)_i \tag{2}$$

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

$$MAE = \frac{1}{n} \sum_{i=0}^n |(h_o - h_s)_i| \tag{3}$$

where: n = the number of observation points, h_o = the observed heads, h_s = the simulated heads (Anderson and Woessner 1992). After steady-state calibration for residual heads between simulated and observed static water values, the summary of result statistics was calculated for 376 well-observations. The computed mean error (ME) for all water level measurements in both aquifers is around -0.8 m. This implies that the observed and estimated water levels have negative skewness. The root means average error for all

Fig. 9 Root mean error distribution graph of the model sensitivity analysis



wells is 4.1 m. This satisfies 90% of all simulated heads matches with observed hydraulic heads within ± 15 m. The mean error (ME) is the mean difference between measured and simulated heads.

Model sensitivity of hydraulic conductivity

The primary goal of sensitivity analysis is to characterize the range of errors in the calibrated model generated by unknown aquifer parameters (Anderson and Woessner 1992). The present study determined the model's sensitivity to hydraulic conductivity in steady-state by systematically increasing and decreasing the model parameter values by 10, 25, and 50%. Figure 9 shows the sensitivity of horizontal conductivity against the hydraulic head, which can explain the effects of the adjustments on the conductivity to much up simulated to observed water level distribution in terms of root mean error. The result reveals, the model was sensitive to the horizontal hydraulic conductivity. For example, a 25% increase or decrease in horizontal hydraulic conductivity in the aquifer system increase or decrease the simulated hydraulic head by 11–14.5%, respectively. Here the sensitivity of the model is amplified for the percent decrease in horizontal hydraulic conductivity compared to the percent increase. The hydraulic head behaves relatively more unstable in the highland than in the rift floor at a low altitude of the study area.

Groundwater balance model

Water balance is a valuable assessment tool since it calculates the proportional significance of each component to the overall budget. Similarly, as mentioned in the preceding section on surface water balance, the water balance of the subsurface is based on the conservation of mass principle for defined limits in space and time. It indicates the rate of change in the amount of water stored in a system, such as a lake basin, and is balanced by the rate at which water flows into and out of the system. It may be expressed in the following Eq. 4:

$$\text{Inflow} - \text{Outflow} = \pm \Delta S \quad (4)$$

where ΔS is the change in water storage, when the change in storage is positive, the water content increases and, when negative (i.e., there is depletion instead of storage), it decreases, under steady-state simulation, the overall water budget is balanced (inflow minus outflow) within a percent discrepancy of 0.00. Table 2 summarizes the comparison of the steady-state water budget for the numerical simulation and estimate for the conceptual model.

Table 3 Water budget of Abaya Lake

Flow term	In	Out	In–out
Storage	0.00E+00	0.00E+00	0.00E+00
Constant head	1.97E+04	4.42E+06	– 4.40E+06
Horiz. exchange	1.15E+06	1.17E+04	1.14E+06
Exchange (lower)	3.27E+06	9.04E+03	3.26E+06
Recharge	3.77E+02	0.00E+00	3.77E+02
Sum of the layer	4.44E+06	4.44E+06	1.00E+00
Discrepancy [%]	0		

Table 4 Water budget of Chamo Lake

Flow term	In	Out	In–out
Storage	0.00E+00	0.00E+00	0.00E+00
Constant head	0.00E+00	5.71E+05	– 5.71E+05
Horiz. exchange	8.60E+04	4.75E+01	8.59E+04
Exchange (lower)	4.85E+05	0.00E+00	4.85E+05
Recharge	8.71E+01	0.00E+00	8.71E+01
Sum of the layer	5.71E+05	5.71E+05	0.00E+00
Discrepancy [%]	0		

Surface and groundwater interaction

About 66.3 m³/day (0.02 MCM/year) groundwater leakage is estimated by the model as a groundwater contribution to river flow from the flow system are shown in Table 2. While only 1.05 m³/day as a river leakage into the groundwater system. This number indicates that about 0.02 million cubic meters (MM³/year) net water flux is out from the groundwater through river beds. About 1814.1.MCM net outflow is simulated annually at the study area's constant head boundary (both lakes). However, the inflow rate into the groundwater system is lower than only 7.2MCM/yr.

Water balance of Lake-Abaya and Lake-Chamo

The Abaya and Chamo Lakes inflow/outflow is the principal component that plays a significant role both as part of groundwater and also as a discharging media to the surrounding flow system. The water balances of the lakes were treated separately to get insight into the interaction of the lakes and adjacent surface flow features in terms of flux contribution, as shown in detail in Tables 3 and 4.

The water balance of the lakes and the entire model domain has been simulated under steady-state conditions using spatially distributed recharge (by WetSpas surface water balance model) in conjunction with a numerical groundwater flow model for estimating groundwater fluxes. Tables 3 and 4 show the fluxes of the lakes used to represent groundwater balances of the lakes. The result indicates

Table 5 Presents the groundwater flux among major-river basins and lakes in the study area

Flow matrix	Lake-Abaya	Lake-Chamo	Billate	Hamessa-Guracha and Kulfo-Gina	Gidabo	Galena	Sife-Chamo
Lake_Abaya	0.00	0.00	2.49E+06	4.46E+05	1.44E+06	9.48E+05	0.00
Lake-Chamo	0.00	0.00	0.00	1.87E+05	0.00	1.13E+05	2.96E+05
Billate	1.06E+05	0.00	0.00	5.07E+05	1.00E+06	0.00	0.00
Hamessa-Guracha and Kulfo-Gina	1.44E+05	6314	5.37E+05	0.00	0.00	5.41E+04	2.80E+04
Gidabo	2.95E+05	0.00	6.53E+05	0.00	0.00	2.63E+05	0.00
Galena	3.81E+05	0.00	0.00	4.58E+04	5.38E+04	0.00	0.00
Sife-Chamo	0.00	1.77E+04	0.00	2.07E+04	0.00	3.41E+04	0.00

lake-Abaya receives the most significant groundwater influx at the constant head (about 1613.3 MCM/year.) than lake-Chamo (208.4 MCM/year). Groundwater flux estimation from the model demonstrates that groundwater controls the inflow of lakes.

Groundwater flux's within the model domain

The entire water balance of the Abaya–Chamo lakes basin in the model domain was divided into five subregions based on surface water divide in terms of major river basins, as shown in Table 5. Such major river basins are Billate (1), Hamessa–Guracha and Kulfo-Gina (2), Gidabo (3), Galena (4), and Sife-Chamo (5) river basin are assigned as region 1, 2, 3, 4, and 5, respectively. For the total water influx to the lake-Abaya, groundwater of Galena (3.81E+05) and Gidabo (2.95E+05) sub-basins contribute much greater than Kulfo-Gina and Hamessa-Guracha; while Bilate contributes the least (1.06E+05). Whereas, Chamo-Sife sub-basin contributes the largest in Lake Chamo compared to Kulfo-Gina groundwater flow shown in Table 5.

Groundwater outflow

The groundwater outflow from the study area domain is simulated using a general head boundary to the south and east of Lake-Chamo. A discharge is expected when the static water level increases. The model result in Table 2 shows that a 0.19 MCM water influx is added annually into the model area from south of the basin boundary. And about 2 MCM out water flux is also taken away from the system annually. Thus here is about – 1.7 MCM/year. A net water outflow is estimated as the outflow from the study area at the south boundary to Konso and Segen basin.

Conclusion

Fragmented scientific studies and poor resource management practice in the existing knowledge and traditional way of future water resource estimation need modern scientific prospects and approaches to support the overwhelming rapid growth of population and satisfy increasing water demand. A comprehensive and integrated evaluation of expected changes is fundamental for a whole range of scales of water resource management and planning under projected global changes. This research is believed to provide a sound foundation for modern scientific direction in water resource evaluation as a decision support tool by establishing an integrated surface and groundwater model under changing the climate for sustainable water resources management and is only viable by considering the complete water cycle in the lake basin. Thus, the projected changes were evaluated using the latest CMIP5 GCM model for future periods (the 2020s, 2050s, and 2080s) under rcp4.5 and rcp8.5. These changes were then used to force independently modelled spatially distributed WetSpss model and ModFlow under the steady-state condition to represent the surface and sub-surface hydrologic system. A parameter integration of surface and groundwater model has an essential advantage in the inclusion of various cases in an integrated manner since the lakes basin is characterized as a complex rift margin zone due to its diverse variation in topography hydro-climate feature, and geology and associated geologic structure. Being the first integration work (CMIP5-WetSpss-MODFLOW) for the lake basin, many challenges and constraints were encountered, mainly in the groundwater flow modelling of the whole area as one unified system.

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Declarations

Conflict of interest The authors declare that they have no competing interests.

Ethical approval No formal approval is required for this study since this article does not contain any studies with human participants or animals performed by any of the authors.

Consent to participate Not applicable.

Consent for publication Not applicable.

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