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NUMERICAL GROUNDWATER FLOW MODELING OF THE CENTRAL MAIN ETHIOPIAN RIFT LAKES BASIN

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ABSTRACT: A three dimensional steady-state finite difference numerical groundwater flow model (MODFLOW) was used to study the groundwater-surface water interactions in the Central Main Ethiopian Rift lakes basin. Special emphasis is given to quantifying the groundwater fluxes and the subsurface hydraulic connection of the rift valley lakes. The result shows that despite the geographic proximity of the lakes the groundwater flux is highly variable, mainly owing to geologic factors. The net groundwater flux ranges from 7,011 m^3/day for Lake Abiyata to 651,022 m^3/day for Lake Shala. The groundwater flow converges from all sides of the basin to the center of the rift where the lakes are clustered, and ultimately to the terminal Lake Shala. Local, intermediate and regional flow systems were identified. The direction and magnitude of the faults have strong bearing on the occurrence and movement of the groundwater. The major flow systems and groundwater flux into the lakes are controlled strongly by rift faults.

Key words/phrases: Groundwater flow, groundwater modeling, hydrogeology, Rift lakes, water balance

INTRODUCTION

The studied area is a closed drainage basin bounded within the limits of 38°00'-39° 30' east longitude and 7°00'-8° 30' north latitude (Fig. 1). The total area of the basin is about 13,000 km^2, which includes part of the Ethiopian Rift system bordered by high altitude plateaux to the east and west. The rift floor is occupied by a series of large lakes fed by perennial rivers originating from the adjacent highlands.

The area is one of the most important sites in the rift from a water resources development point of view. The lakes and feeder rivers are being used for various purposes.

A few decades ago much of the area was covered by natural vegetation and the lakes were protected from large-scale human interferences (Street, 1979). However, the fast growing population has induced land degradation and increased abstraction of river and lake water. This trend is continuing. At present the lakes are being used for irrigation, fishing, recreation and the alkaline lake of Abiyata is the source of soda ash (Na_2CO_3). There are also future plans of expansion of irrigation and soda ash extraction.

Utilization of water in the Ethiopian Rift has proceeded without a basic understanding of the hydrogeologic system and the water balance of the lakes. This has become a critical problem in water resources management (Wenner, 1973; Makin *et al.*, 1976; HALCROW, 1989; Dereje Hailu *et al.*, 1996). Development plans were implemented with short-term interests, though it was understood that the recent increased abstraction of water from lakes and rivers caused changes in the chemistry of some of the lakes (Elizabeth Kebede *et al.*, 1994) and reduced their level significantly (HALCROW, 1989). Contrary to the general trend of lake level reduction in the East African Rift, some lakes remain stable (Tenalem Ayenew, 1998). This contrasting behavior requires assessment of the water balance, the relative importance of groundwater contribution and the hydraulic connection of the lakes.

Among the water balance components of lacustrine systems, groundwater is the most difficult to quantify. As a result, many hydrological studies of lake watershed systems have given little emphasis to groundwater (Crowe and Schwartz, 1981; Almenidinger, 1990; Crowe, 1990). When this study was started, little was known of the hydrogeological system and the role of groundwater in the water balance of the lakes. A basic question was whether a substantial part of the lake water depended on groundwater fluxes and, if so, from where and how much. This study has the objective of quantifying the groundwater fluxes of the major lakes, their subsurface hydraulic interconnections and the groundwater flow systems of the entire basin.

General overview of the basin

The basin can be divided into three physiographic zones: the rift floor, the rift escarpments and the highlands. The altitude ranges from around 1,600 meters above mean sea level (m.a.s.l) in the rift to over 3,000 m.a.s.l in most parts of the surface water divide. There are few volcanic summits having an altitude of more than 4,000 m.a.s.l. The lowest point is Lake Shala (1,550 m.a.s.l).

Four major lakes exist in the rift floor: Ziway, Abiyata, Langano and Shala. Table 1 summarizes the basic hydrologic data. Lakes Ziway and Abiyata are connected by the Bulbula river, and lakes Langano and Abiyata by Horakelo river. Shala has an independent catchment without surface water connections with the other lakes. There are also few small lakes in the rift and escarpment areas (between altitudes of 1,750 to 2,000 m.a.s.l) occupying volcano-tectonic depressions.

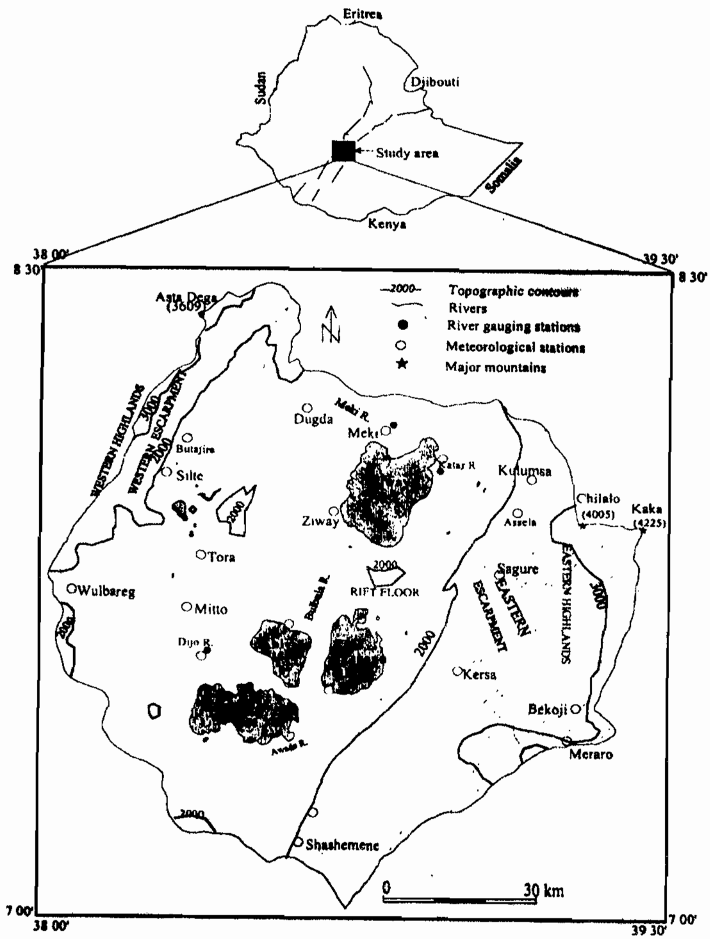


Fig. 1. Location map of the study area.

Table 1. Basic hydrological data of the lakes.

Lake	Altitude (m.a.s.l)	Surf. Area (km^2)	Catch.Area (km^2)	Max. Depth (m)	Mean Depth (m)	Volume ($10^6 m^3$)
Abiyata	1580	180	10740	14.2	7.6	954
Langano	1585	230	2000	47.9	17	3800
Shala	1550	370	2300	266	8.6	37000
Ziway	1636	440	7380	8.9	2.5	1466

Geologically all the lakes are situated on the tectonically and volcanically active segment of the rift floor (Fig. 2). The present day geological and geomorphological features are largely the results of Cenozoic volcano-tectonic processes (Di Paola, 1972). The Cenozoic geological history of Ethiopia is characterized by massive and voluminous mafic and silicic volcanism and domal uplift, Neogene rifting and volcanism, rift bound Plio-Pleistocene rifting and volcanism (Zanettin *et al.*, 1980). Although rift-related basins started to form during the late Oligocene to Early Miocene times, the Main Ethiopian Rift was fully developed by middle to late Miocene time (Giday Woldegebriel *et al.*, 1990). Faulting and consequent rift formation was accompanied by extensive basaltic and silicic volcanism restricted to separate centers aligned along the rift axis. Several shield volcanoes were developed on the plateaux (Di Paola, 1972; Giday Woldegebriel *et al.*, 1990). Various volcanic episodes formed thick volcanic rock sequences. The major rock types include basaltic, trachyitic and rhyolitic lava flows, ignimbrites, unwelded tuff and pumice fall deposits (Tesfaye Chernet, 1982). Lacustrine and volcani-clastic deposits are abundant in the rift.

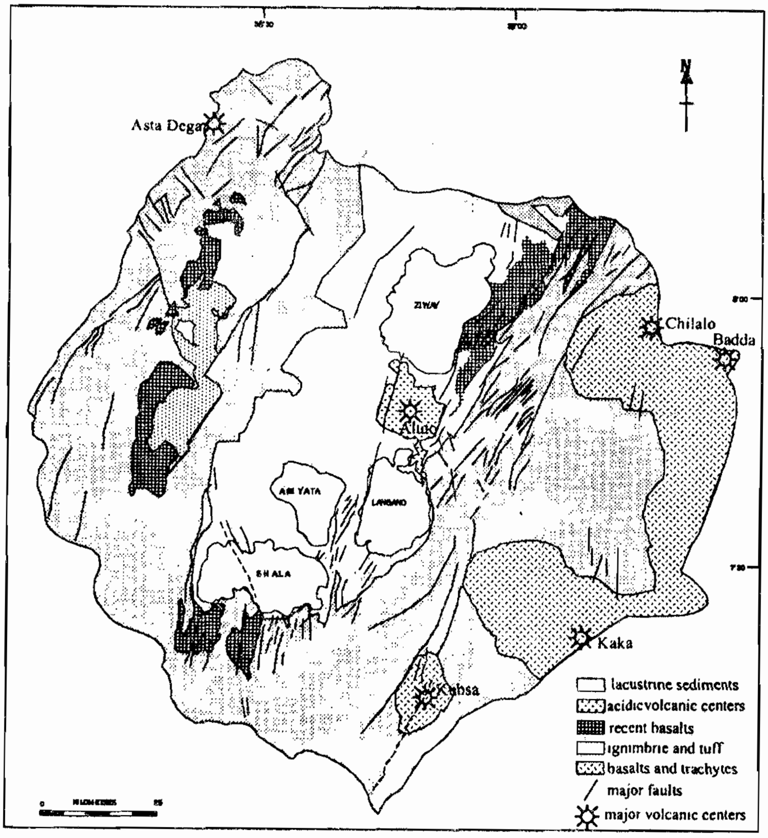


Fig. 2. Simplified geological map of the area (modified from Tenalem Ayenew, 1998).

Large-scale normal block faulting has disrupted the volcanic rocks. The rift valley is distinctly separated from adjacent plateaux by a series of step-faults usually trending parallel and sub-parallel to the rift axis. There are at least four sets of faults; NNE-SSW, N-S, NNW-SSE trending and arcuate faults associated with volcanic vent structures (Tenalem Ayenew, 1998). The first two are dominant and extend for long distances. Recent tectonic activity in the study area is restricted to the active Wonji Fault Belt, which is characterized by Quaternary volcanic centers and densely spaced normal faults. The extensive fault systems of the rift and the escarpment have increased significantly the permeability of the rock units. Fig. 3 shows the classified transmissivity map.

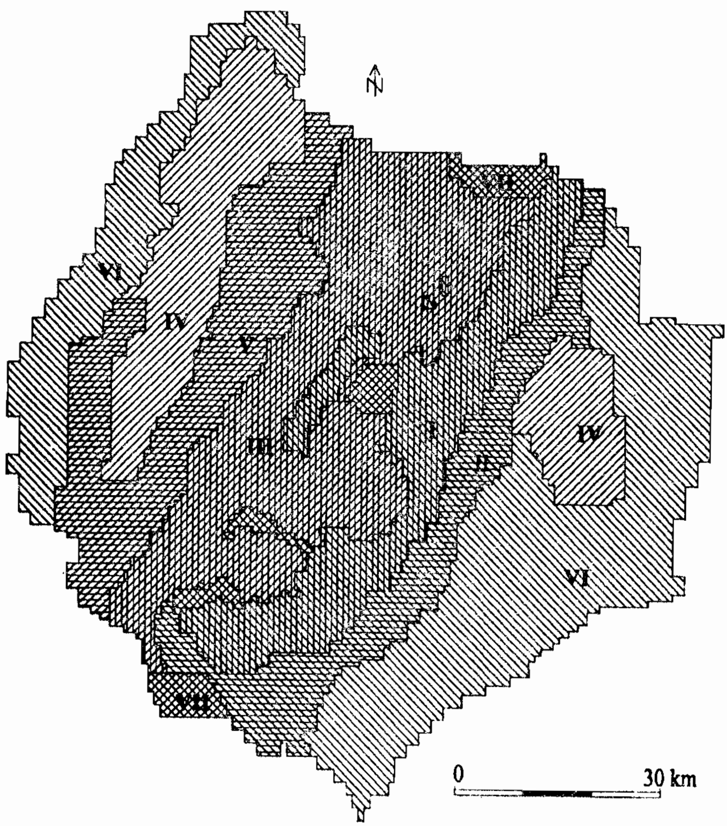


Fig. 3. Classified transmissivity map. The transmissivity range/calibrated values in m^2/day for the different zones are: I, 1700-2000/1820; II, 650-735/825; III, 300-400/320; IV, 240-280/264; V, 180-220/192; VI, 60-120/101; VII, 5-12/8.

METHODOLOGY

The entire basin has been modeled in two-dimensional aerial view using the standard United States Geological Survey groundwater flow model, MODFLOW, under steady-state conditions. This model is a modular three-dimensional finite difference groundwater flow code (MacDonald and Harbaugh, 1988), which simulates all aquifer types under both transient and steady-state conditions. The model has been widely applied in hydrgeological practices for simulating groundwater flow in various hydrogeological settings (Winter, 1978a, 1978b; Duan, 1996).

In addition to MODFLOW, two other models were used: the two-dimensional steady-state finite element groundwater flow model, Flonet (Guiguer and Molson, 1996) and the Stochastic Discrete Fracture model, SDF (Rouleau, 1988). The former has been used to understand the mechanism of flow at depth along selected cross-sections and the latter to study the role of faults in the movement and occurrence of groundwater. Cross-sectional and discrete fracture simulations were made first at catchment level and justified before assembling the basin-wide conceptual model for two-dimensional aerial simulation. It should be noted that these models were used only to obtain ancillary information for developing the conceptual model for MODFLOW simulation.

Modeling groundwater flow through discrete fracture networks is difficult and data intensive (Snow, 1969; Sharp and Maini, 1972). For describing groundwater flow in fractured media, porous media models or a continuum approach have been used widely by increasing the hydraulic conductivity or transmissivity values of cells where fracture flow occurs. One such model is MODFLOW, which uses the porous media approximation for fractured media. However, possible errors introduced by using the porous media approximation in fractured media is either rarely addressed or glossed over in many groundwater modeling practices (Bradbury and Muldoon, 1993). The SDF model provides valuable information on how groundwater flow behaves in discrete fractures and helps in estimating the equivalent hydraulic conductivity or transmissivity for porous media models.

In the present work, the SDF model has been applied only to a small highly fractured area (Fig. 7) to find the equivalent aquifer hydraulic parameters for the porous media model. The SDF model simulation involved a detailed study of the fracture parameters (orientation, density, spacing, aperture, roughness, etc.). This model was used to study the flow behavior in the discrete fractures and to obtain information on the fractured area where pumping test data on aquifer parameters is not available.

A Geographic Information System (GIS) software known as the Integrated Land and Water Information System (ILWIS) was used to develop the Digital Elevation Model (DTM) of the basin and to prepare the maps for the various model input variables by overlay operations. The DTM served as a very important tool for model input of the top and bottom of the aquifer.

All boundary conditions were determined, before model simulations, from existing hydrometerological data and extensive field hydrogeological survey, which took more than three years. Extensive field measurements on river discharge were made and river morphology was studied to assess the aquifer-river relations and to estimate the variables for the river package (RIV) of MODFLOW (Equations 2-4). Wellhead measurements were made using a sensitive deep meter in few open wells and used for model calibration. Aquifer thickness has been estimated from well lithologic logs and startigraphic survey.

All the collected relevant data were assembled in a systematic hydro-meteorological and hydrogeological database in the form of attribute tables and maps. Conceptualizing the flow system and boundary conditions, the steady-state groundwater flow of the entire basin has been simulated based on the following differential equation developed for unconfined aquifers for two-dimensional aerial view:

$$\frac{\partial}{\partial x}(T_x \frac{\partial h}{\partial x}) + (\frac{\partial}{\partial y} T_y \frac{\partial h}{\partial y}) \pm R \qquad (1)$$

where h is hydraulic head; T_X and T_y are components of transmissivity in the x and y direction respectively, and R is sink/source term.

The exchange between aquifers and rivers has been handled using the river package. The river package is designed to simulate the effect of flow between rivers and aquifers based on the following three equations:

$$Q_{riv} = C_{riv}(H_{riv} - h), \quad \text{when } h > R_{bot} \qquad (2)$$

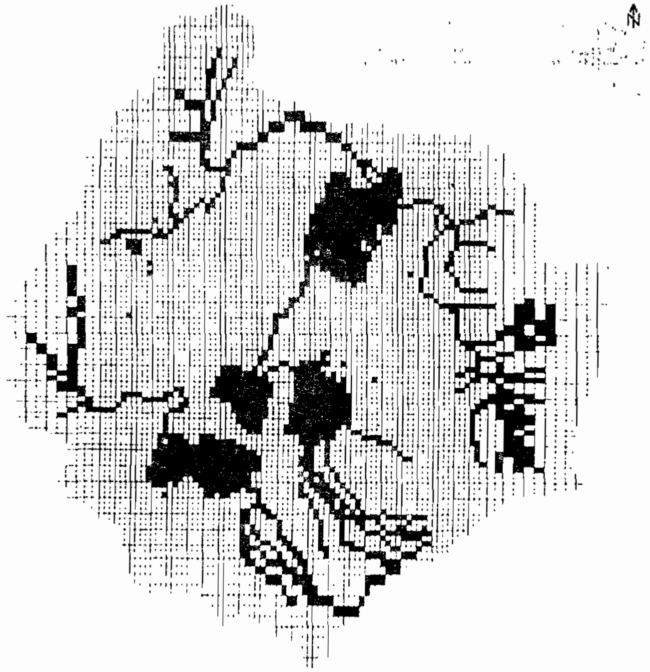
$$Q_{riv} = C_{riv}(H_{riv} - R_{bot}), \quad \text{when } h \pm R_{bot} \qquad (3)$$

$$C_{riv} = \frac{KLW}{M} \qquad (4)$$

where Q_{riv} is the rate of leakage through the river bed, H_{riv} and h are head in the river and in the aquifer respectively, R_{bot} is the elevation of the river bed bottom, C_{riv} is river bed hydraulic conductance, K is hydraulic conductivity of the river bed, L is the length of the river reach contained in the cell under consideration, W and M are average width of the river and thickness of the river bed in the same cell.

Boundary conditions and model input parameters

The model domain is bounded, on almost all sides, by volcanic hills and mountains. Model simulations under different scenarios and isotopic studies have revealed that limited groundwater enters into the study area from Lake Awassa basin adjacent to the south (Darling *et al.*, 1996; Tenalem Ayenew, 1998). Compared with the size of the basin and the fluxes of the lakes this inflow is very small (12,345 m^3/day). The inflow has been accounted in the model by a specified flow boundary. Substantial outflow to the Awash basin adjacent to the north is not evident from a number of cross-sectional Flonet simulations. Except the small inflow from the south, the basin can be considered as a closed system. Therefore, all the other outer model boundaries are considered as a no flow boundary (Fig. 4). The lakes are treated as constant head boundary taking the average elevation of the lakes (Table 1).



, constant head cells; , river node cells;
, inactive cells (out of the boundary); , active cells.

Fig. 4. Model grid and boundary conditions.

Based on the amount and distribution of rainfall and soil water balance approach the net recharge has been estimated and categorized into three zones: the rift valley (0.00001 mm/day), the rift escarpment (0.0001 mm/day) and the highlands (0.0003 mm/day). Since the groundwater level is not very close to the surface, evaporation from the groundwater system is assumed to be negligible. The abstraction of water from wells is insignificant. There are less than 20 productive boreholes used for limited local water supply purposes. The withdrawal from these boreholes and shallow dug wells does not affect the groundwater flow regime in a significant manner.

The average hydraulic conductivity ranges between 0.09 m/day in the homogeneous less fractured highland volcanic terrain and around 25 m/day in the highly fractured rift floor volcanics and permeable lacustrine deposits. Hydraulic conductivities have been averaged out for the different layers based on lithologic log information obtained from borehole drilling reports and field hydrogeological observations. The maximum aquifer thickness ranges from around 150 m on the highlands to 250 m in the rift floor. Local exceptions exist in the highly faulted zones and recent volcanic centers.

The steady-state model was calibrated using limited well-head observations (Fig. 5). The calibration parameters (aquifer thickness and hydraulic conductivity or transmissivity) were varied within a reasonable range based on these well-head observations. After many trial simulations reasonable fit has been achieved between the observed and simulated heads. The difference between the observed and calculated heads for the limited observations ranges between -1.2 and 3.4 meters.

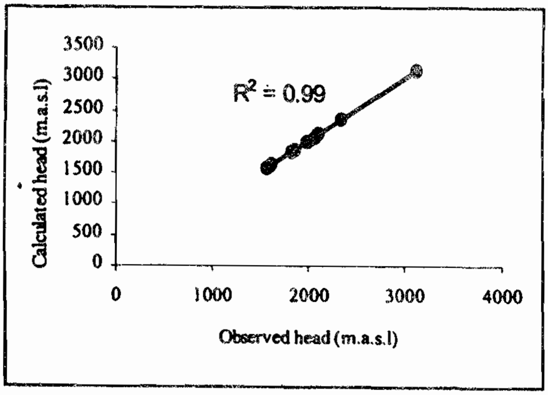


Fig. 5. Relation between observed and model calculated heads at the selected points for calibration.

A sensitivity analysis has been performed in order to establish the effect of uncertainty on the calibrated model. The purpose of sensitivity analysis is to quantify the uncertainty in the calibrated model, which may potentially arise from variations in the estimates of aquifer parameters, stresses and boundary conditions. Sensitivity analysis is typically performed by changing one parameter value at a time and assess the impact of these changes on the model output. The transmissivity and recharge have been used as the main calibration parameters to note the changes on the fluxes of the lakes and the head distribution.

Model discretization

In a finite difference numerical model, the continuous problem domain is replaced by a discretized domain consisting of an array of nodes and associated finite difference blocks or cells. The nodal grid forms the framework of the numerical model. Irregular model grid has been chosen (Fig. 4). The length and width of the model domain are equal (165,000 meter). A cell length of 2500, 1875 and 1250 meters were chosen for the different areas accounting a total number of 104 columns and 99 rows. Since more emphasis has been given to the rift lakes, the grid size is relatively smaller in the rift so as to get more precise values of model estimated groundwater fluxes and to study the hydraulic interconnection of the lakes and their relation with geologic structures.

The aquifer can be broadly divided into two layers: an upper more permeable layer, characterized by deeply weathered and fractured rocks overlain by predominantly permeable soils, alluvial, colluvial and lacustrine sediments, and a lower less permeable layer consisting of massive volcanic rock sequence. The upper layer is not continuous in some places, such as those covered by recent lava flows not affected by extensive weathering and fracturing and in volcanic cones and hilly areas. Due to high degree of fracturing and the absence of conceivable extended confining layer, good hydraulic connection is assumed to exist between the two layers. Therefore, at basin-wide scale the aquifer has been treated as an unconfined system. The two layers have been treated as a single unit by averaging the hydraulic conductivity values with depth. However, in profile modeling using Flonet, the aquifer has been discretized as a two-layer system. The bottom of the aquifer is treated as impermeable in both cases.

In a two-dimensional areal model, anisotropy in transmissivity is represented by the difference between transmissivity in the x and y (T_x and T_y) directions. Input to the model may consist of the two transmissivity arrays or the T_x array and an anisotropy factor (the ratio of T_y and T_x) to compute the T_y array from the T_x array. Based on previous hydrogeological investigations (Tesfaye Chernet, 1982), experiences of modeling in fractured terrain (Bradbury and Muldoon, 1993) and field hydrogeological observations, an average anisotropic factor of 0.5 was used.

RESULTS AND DISCUSSION

Table 2 summarizes the groundwater fluxes of the four major lakes averaged on daily basis. Lake Shala gets the largest net groundwater flux and Abiyata the lowest. Over a wide range of likely model simulation scenarios with variable aquifer thickness and hydraulic conductivity or transmissivity values, amounts of the flow to Shala and Langano remain very high while Lake Ziway gets moderate inflow. The largest amounts of the flow come through the Meki and Katar river deltas.

Many fault-controlled springs are aligned along NNW-SSE direction east of Ziway, indicating the importance of these major rift fault systems in controlling the movement and occurrence of groundwater. Groundwater migrates to the south to lakes Abiyata and Langano (Fig. 7) following the N-S and NE-SW trending faults. Ziway and Abiyata lakes are hydraulically interconnected. The presence of groundwater barrier west of the Langano fault limits the hydraulic connection between Langano and Abiyata (Fig. 2). Shala is separated from Abiyata by impermeable rock units forming the caldera rim. This has minimized the outflow from Abiyata to Shala (3,350 m^3/day). Lake Shala remains terminal from both surface water and groundwater perspective.

Table 2. Model estimated groundwater flux of the lakes (m^3/day).

Lake	Inflow	Outflow	Net flux
Abiyata	73464	3350	70114
Langano	370891	51853	319038
Shala	651022	none	651022
Ziway	220788	40047	180741

The total inflow from the adjacent Awassa basin to the study area has been estimated as 12,345 m^3/day. Most of this flow enters Lake Shala. Compared with the size of the model domain and other fluxes to and from the lakes, this value is quite small. Apart from this flow, all evidence indicates that inflow into the basin in other areas or substantial outflow from the basin is not evident.

This steady-state groundwater balance of the whole basin shows that the rivers and lakes receive large quantities of groundwater from aquifers while there is limited flow from lakes to aquifers. Subsurface outflow from the rivers and the lakes are 16.5% and 7.8% of the total groundwater inflow, respectively.

The model-estimated head distribution reasonably agrees with the field-measured heads from limited wells. Figure 6 shows the model-simulated groundwater contour map, which is almost similar to the regional groundwater contour map reconstructed from limited well-head observations. Water table elevation lies slightly below the land surface altitude, varying within 20 to 30 meters in large parts of the model domain. The depth to the regional water table approaches land surface in the rift floor around the major lakes. It decreases sharply on the rift shoulders and close to lake shores and along major river valleys. It tends to be deeper beneath the volcanic mountains on the highlands and other positive topographic features. Dominant groundwater flow occurs in the upper few tens of meters in most parts of the basin.

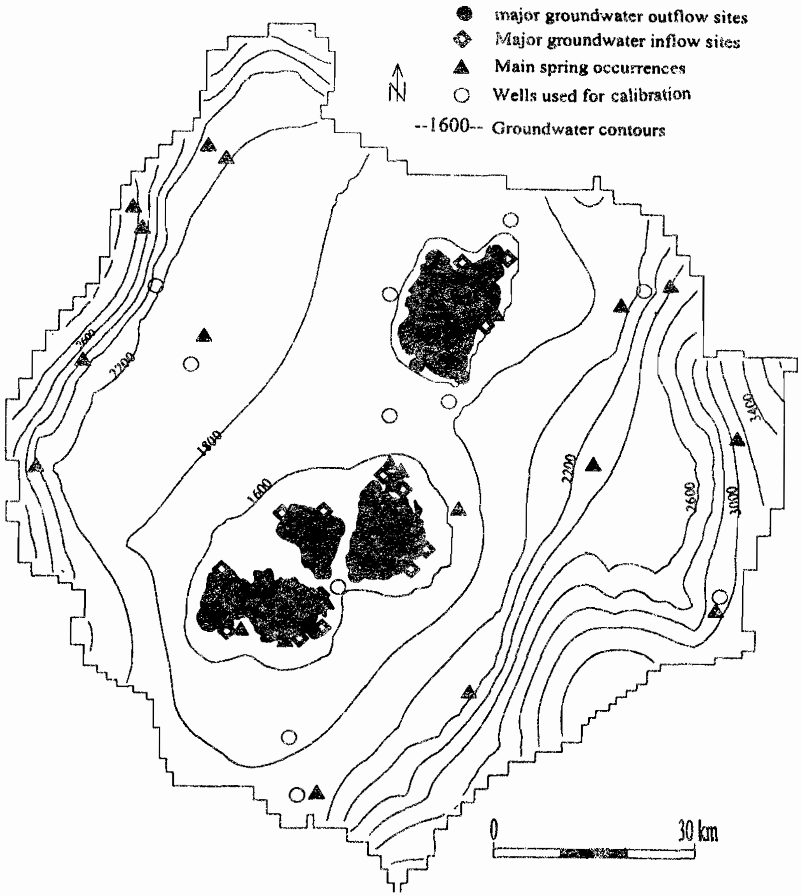


Fig. 6. Model estimated groundwater contour map.

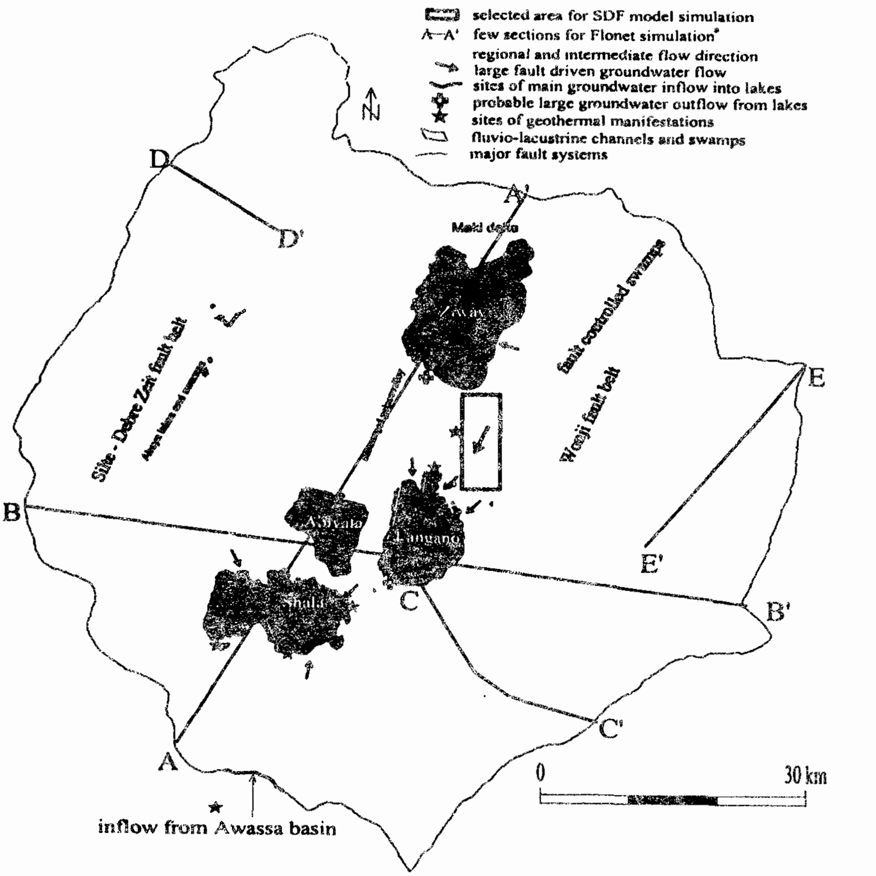


Fig. 7. Schematic map showing groundwater flow directions, geological structures and cross-section directions.

Groundwater flow converges from all sides towards the center of the basin and ultimately to the lowest elevation in Lake Shala. The largest contribution to groundwater flow in the rift comes from the eastern highlands. In the eastern half of the rift floor, much of the groundwater is directed towards lakes Langano and Shala following the N-S and NNE-SSW trending faults (Fig. 8). The SDF model clearly demonstrated the important role of these faults in diverting large quantities of groundwater to the south to Lake Langano, ultimately flowing to the lowest elevation in Lake Shala. The dominant groundwater flow line in the model follows the major fault systems oriented parallel and sub parallel to the axis of the rift. Flow towards lakes is dominantly through these discrete fractures; diffused flow has limited importance.

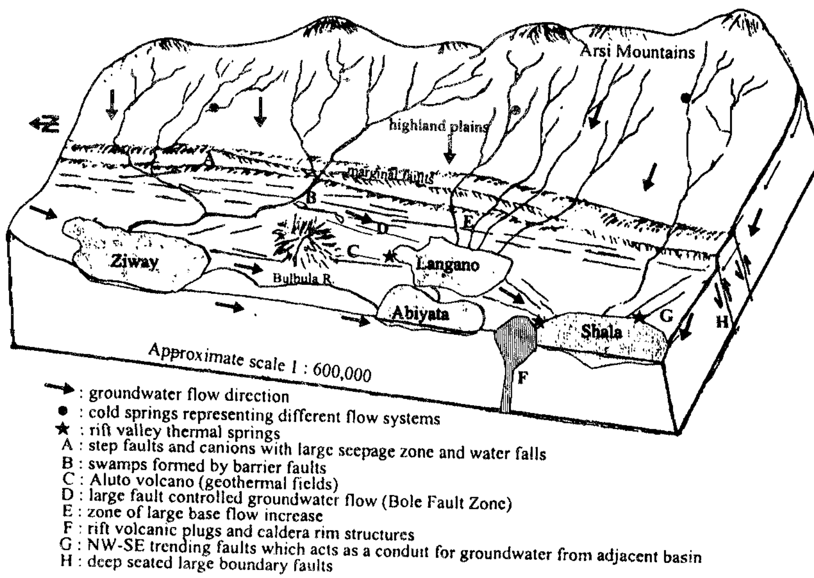


Fig. 8. Schematic view of groundwater flow system in the eastern half of the basin.

The large hydraulic gradient and the presence of numerous step-faults have resulted in the formation of different discharge areas at different elevations, mainly along the rift escarpment and at the feet of highland volcanic summits. The large elevation difference between the rift and the highlands favor the formation of local, intermediate and regional flow systems. Different springs and seepage zones represent these flow systems.

The cross-sectional model (Flonet) simulation revealed that the groundwater flow system can be schematized with depth into three zones: a shallow zone of active and fast flow, a medium zone of relatively slower flow and a deep flow zone with long residence times. The shallow systems are confined to the upper permeable soil, sediment and weathered rock zone (usually less than 50 meters). This zone is considered to be the phreatic near surface aquifer with high permeability and storage capacity. Immediately below this zone, there exists an intermediate fractured rock zone (50 to 60 m thick). Groundwater in the intermediate zone is the source of consistent base flow to rivers, large springs and lakes. The lower zone is a massive volcanic sequence, locally fractured by deep-reaching faults. On the highlands flow in this zone is very low due to the rarity of large faults. Furthermore, moderate flow exists within the rift faults, associated locally with geothermal systems. Figure 9 shows a simple schematization of the flow in cross-sectional view, simulated by Flonet, along the rift floor.

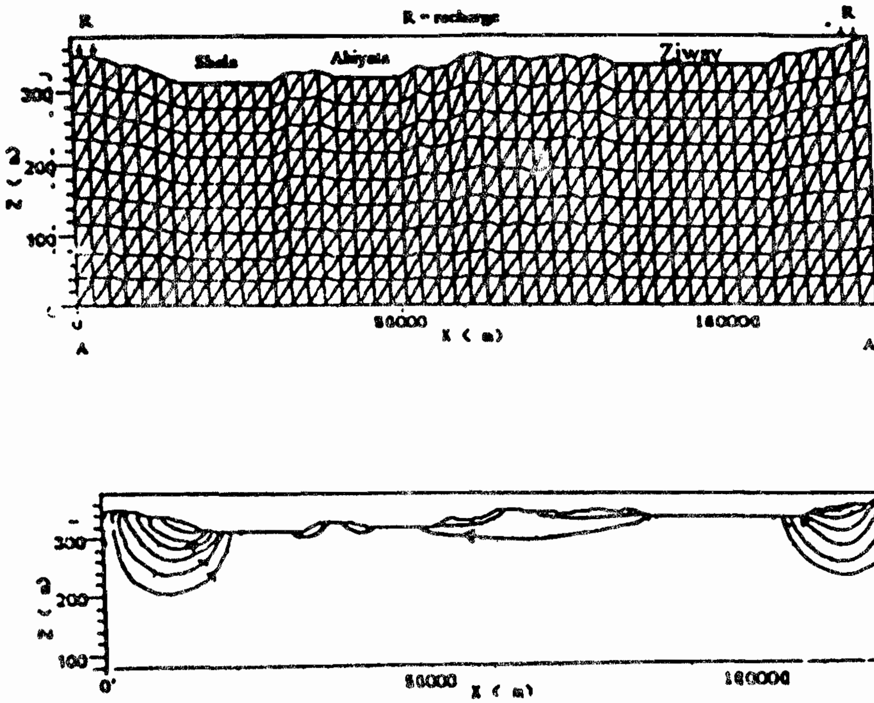


Fig. 9. Flonet simulated flow lines and finite element grid along section A-A′.

The sensitivity analysis revealed that the head remains more stable in the rift floor than on the highlands when the transmissivity and the recharge change. Those lakes with a high groundwater flux (Shala and Langano) are more sensitive than lakes characterized by lower groundwater flux (Abiyata and Langano). For example, a 10 % increase in transmissivity will result in a change of flux of 10.5 % for Lake Shala but only 4.9% for lake Abiyata.

CONCLUSIONS

The present work has shown that groundwater plays a very important role in the water balance of the lakes. In all the lakes the total inflow is greater than the outflow, which ultimately leaves mainly as river discharge. The net annual steady-state groundwater flux to lakes Abiyata, Ziway, Langano and Shala is 26, 66, 117 and 238×10^6 m^3 respectively. This result is in good agreement with the estimates made independently using conventional water balance approach.

Despite their proximity there is no significant subsurface hydraulic connection between Langano and Abiyata and between Abiyata and Shala. Shala and Langano are groundwater-controlled lakes.

There is no strong evidence that supports substantial groundwater outflow from the basin; the inflow is also low. Groundwater flow in the rift is primarily controlled by geologic structures, either in a direct way by flow in the tensional faults or through fluvial and lacustrine deposits whose occurrence is influenced by tectonism. The major groundwater conduits are faults parallel and sub-parallel to the rift axis, *i.e.*, NNE-SSW trending.

Recharge and groundwater flow in the weathered upper zone on the highlands are the driving forces for much of the hydrology of the basin. As revealed by cross-sectional simulations, the flow is dominantly shallow and operates in the upper weathered and fractured zone. Deep upwelling groundwater flow systems are limited.

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