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APPLICATION OF ISOTOPE AND WATER BALANCE APPROACHES FOR THE STUDY OF THE HYDROGEOLOGICAL REGIME OF THE BISHOFTU CRATER LAKES, ETHIOPIA

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ABSTRACT: The major Ethiopian rift lakes have been studied for various purposes in the last few decades. However, crater lakes situated elsewhere in the country are some of the poorly understood hydrologic systems. This study focuses on quantifying the groundwater fluxes and assessment of the hydrogeological regime of the Bishoftu Crater Lakes using conventional water balance and stable environmental isotope (^{2}H and ^{18}O) techniques. The convenient geological setting of these lakes has provided a suitable base for determining groundwater exchange rates using stable isotopes coupled with hydrologic balance. The result indicates that these lakes have highly variable groundwater fluxes. Evaporation and groundwater fluxes are the major components of the hydrologic balance. Unlike the other lakes, lake Hora appears to have extremely low groundwater outflow. The lakes are highly evaporated with respect to the present day precipitation. Other groundwater and surface water sources are relatively depleted in the heavy isotopes. The isotope and hydrochemical signature indicates that the circulation of groundwater and recharge is fast and the regional groundwater flow direction is from north to south.

Key words/phrases: **Crater lakes, Ethiopian rift, groundwater, isotopes, water balance**

INTRODUCTION

Water balance study is an integral part of water resources assessment of lacustrine systems and river basins. In most cases such studies consider surface water hydrology only. This is due to the fact that quantification of groundwater fluxes and understanding the mechanism of subsurface hydrodynamics is difficult. As a result many hydrological studies of lake watershed systems have given little emphasis to groundwater (Crowe and Schwartz, 1981; Crowe, 1990; Almenidinger, 1990). In some cases groundwater flux is obtained as a residual of the other water balance components or quantified as a net flux (Winter, 1978; Freeze and Cherry, 1979; Tenalem Ayenew, 1998).

With hydrochemical (Winter, 1978; Seifu Kebede, 1999) and environmental isotope (Fontes *et al.*, 1979; Karbbenhoft, 1994; Darling, 1996; Tenalem

Ayenew, 1998; Seifu Kebede, 1999) techniques, coupled with classical hydrogeological investigation approaches, groundwater flow components can be constrained and groundwater-surface water interactions can be studied in detail.

In the faulted volcanic terrain, where the chain of Bishoftu Crater Lakes (BCL) are situated, quantifying groundwater fluxes to and from the lakes by conventional methods and unraveling subsurface hydraulic connections is very difficult. This is due to the heterogeneity of the surrounding aquifer systems, the complexity of rift tectonic structures and the lack of relevant hydrogeological data. When this study was started, little was known of the hydrogeological regime and the role of groundwater in the water balance of the lakes. A basic question was whether a substantial part of the lake water depends on groundwater fluxes and, if so, from where and how much?

In this study the net groundwater flux of the lakes is determined based on conventional hydrological balance using limited hydrometeorological data. The net groundwater flux is separated into its outflow and inflow components by combining the water balance and isotopic balance equations. The isotope data is also used to assess the interaction among the crater lakes and their hydraulic relation with the surrounding groundwater system.

Study area

The BCL are located some 45 km southeast of Addis Ababa within the Main Ethiopian Rift (Fig. 1). The altitude ranges from 1800 to 2000 meters above sea level. Five permanent natural lakes occupy a chain of craters: Hora, Babogaya, Bishoftu, Kilole and Arenguade (Table 1). There are other artificial lakes, swamps and ponds: Lake Kuriftu, an originally dry crater depression filled by diverting the tributary (Balbala river) of the perennial Mojo river; Lake Cheleleka is an extensive shallow pond that has remained a permanent body of water since the early 1970's; Lake Balbala, an artificial reservoir constructed on one of the tributaries of Mojo river in the early 1980's. In estimating the hydrologic balance, only the first three permanent lakes have been considered, since they have convenient hydrogeological setting for the study and the availability of relevant hydrometeorological data.

Table 1. Basic hydrological data of the lakes (modified from Prosser *et al.*, 1968).

Lake	Altitude (m)	Lake Area (km^2)	Max. Depth (m)	Mean Depth (m)	Volume ($10^6 m^3$)	Catchment Area (km^2)
Kilole	2000	0.771	6.4	2.6	2	3.103
Arenguade	1900	0.541	32	18.5	10	0.808
Babogaya	1870	0.579	65	38	22	0.293
Hora	1850	1.029	38	17.5	18	0.628
Bishoftu	1870	0.929	87	55	52	0.386

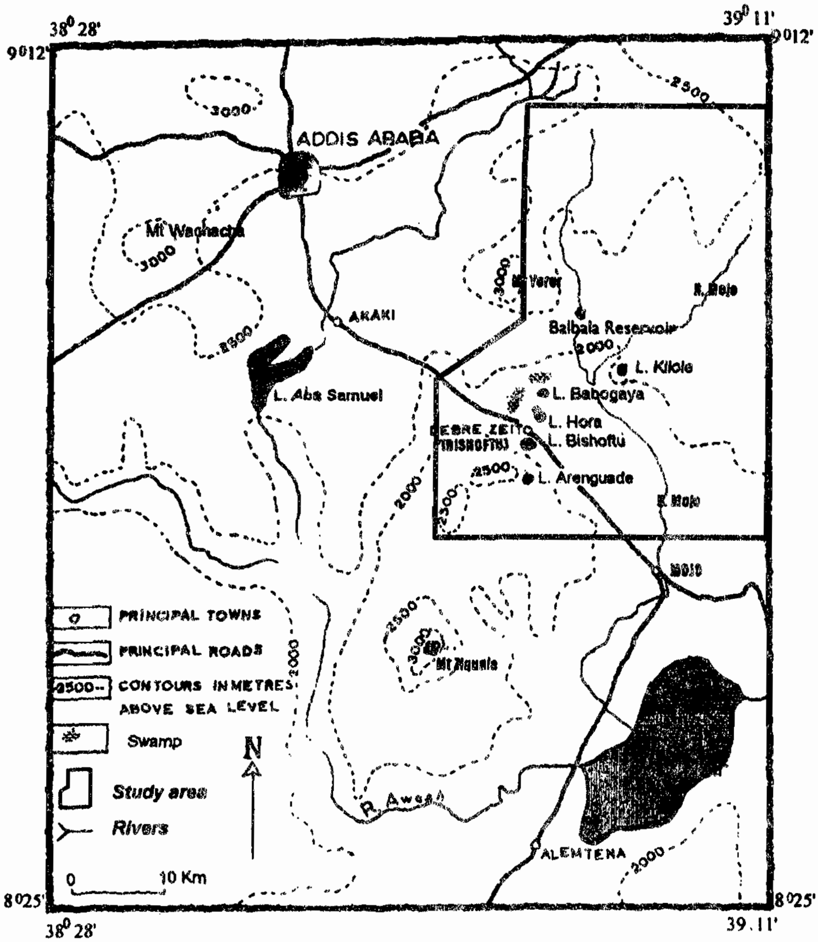

Fig. 1. Location map of the study area showing drainage and relief.

All of the crater lakes are maars, *i.e.*, volcanic collapse structures above zones of fractured rock that extend to an igneous dike at depth (Lorenz, 1986). The explosion craters are roughly circular in shape, and their sizes range from 0.5 to 1 km^2. The bathymetry of the lakes has been studied by Prosser *et al.* (1968). Their depths vary from 6.4m (Kilole) to 87m (Bishoftu). Three of the lakes (Babogaya, Bishoftu and Hora) are closed to any channeled surface water inlets or outlets (Fig. 3). Compared with many small terminal lakes in the Ethiopian rift, such as Abiyata, the range of lake level fluctuation in these crater lakes is lower. This hydrological behavior has created convenient conditions for the assumptions considered in isotopic balance calculations.

The surrounding region is covered by old basaltic flows (7.5–4.5 Ma) and acidic volcanics (4 Ma) (Fig. 2). Maars, cinder cones, and lava flows represent more recent (10 Ka) basaltic activity (Gasparon *et al.*, 1993). The lakes lie on a NNE-SSW trending fault-belt, which is parallel to the axis of the Main Ethiopian Rift. The old basalt, which forms the base of the volcanic units is the main aquifer of the area. The transmissivity of this rock unit varies over a wide range, from 389 m^2/day to 21,600 m^2/day (Aynalem Ali, 1999). The overlying young basic pyroclastic rocks interbedded with minor acidic products make up the largest part of the study area and have a transmissivity value ranging from 1,100 m^2/day to 18,000 m^2/day. The abundant scoria cones and the acid volcanic domes and pyroclastics constitute the major recharge areas. The static groundwater level in most of the area is well above the level of the lakes.

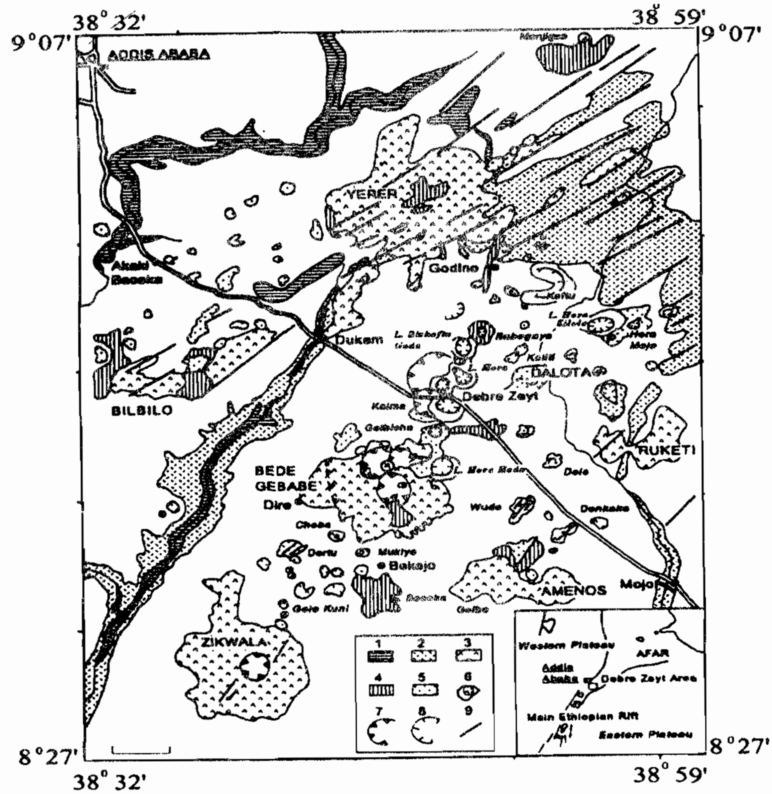

Fig. 2. Schematic geological map of Debrezeit area and surrounding regions. 1, Addis Ababa basalts; 2, Ignimbrites of the Upper Miocene-Pliocene Nazret Group; 3, Trachytes and rhyolites of central volcanoes; 4, Younger basaltic lavas; 5, Younger basaltic cinder cones; 6, Younger hydromagmatic pyroclastic deposits(tuff cones); 7, Crater rim; 8, Maar rim; 9, Fault. (After Gasparon *et al.*, 1993.)

The climate in the study area is semi-arid (Daniel Gamechu, 1977). The rainfall has large seasonal and inter-annual variability. The average annual rainfall is around 830 mm. According to Nicholson (1996) the East African rainfall derivation, variation and seasonality is very complex. However, one generalization that can be made for the studied region is that there are two rainfall periods: The 'Little Rains' which come from the southeast from the Indian Ocean (March to May) and the 'Big Rains' from the southwest coming from the Atlantic Ocean (Mid June to Mid September). The dry season lasts between October and February (Table 2). Temperature shows large diurnal but small seasonal changes with an annual average of 19° C. The mean annual humidity is 0.6 and varies from 0.53 to 0.70. Analysis of open water evaporation has been made, based on Penman (1948) method, energy balance and mass transfer approaches. Based on these three approaches, the long-term (30 years) average open-water evaporation was estimated at 1710 mm (Seifu Kebede, 1999).

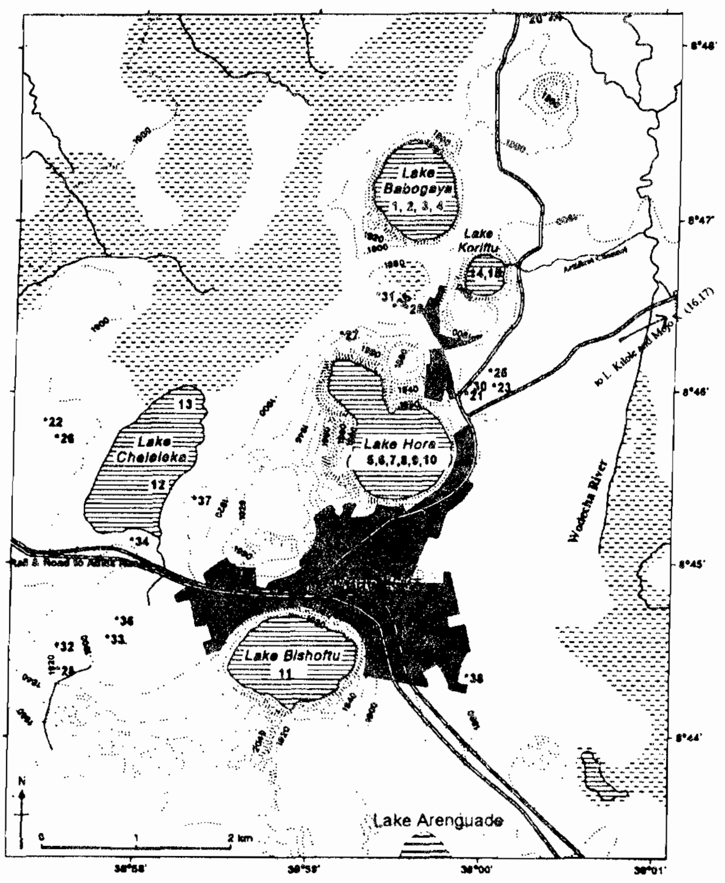

Fig. 3. Physiographic map and water sampling points with sampling points indicated.

Table 2. Basic hydrometeorological data of the area (modified from Seifu Kebede, 1999).

Parameters	Jan.	Feb.	Mar.	Apr.	May.	Jun.	Jul.	Aug.	Sep.	Oct.	Nov.	Dec.	Annual
Mean monthly precipitation (mm)	9.0	26.0	41.3	57.5	51.0	87.7	211.5	225.8	103.7	16.8	4.9	2.9	838
Mean monthly PET (mm)	131	150	173	164	185	141	118	110	113	145	130	126	1687
Mean Monthly temperature (° C)	18	19	21	21	21	20	19	19	19	18	17	17	19
Relative humidity in %	53	54	54	58	51	64	74	77	74	57	52	52	60

METHODOLOGY AND THEORITICAL CONSIDERATIONS

Hydrometeorological data (average 30 years recording) have been collected to estimate the surface water balance components (Seifu Kebede, 1999). To support the conventional water balance approach, stable environmental isotopes of ^{2}H and ^{18}O have been used to determine the groundwater inflow and outflow components of the lake on the basis of isotopic balance.

Water samples were collected carefully and kept in a polyethylene sample bottles and completely filled and tightened with double-sealed caps. Samples from wells were collected using a Klyen Downhole Sampler. Water samples from the lakes were collected using a water sampling apparatus designed to collect samples at different depths. Water samples for isotope analysis were collected from January 1998 to January 1999 in tight polyethylene bottles, additionally sealed with PVC tape. Temperature, pH and electrical conductivity of water samples were measured *in situ*.

The samples were stored in a refrigerator in the Geochemistry Laboratory of the Department of Geology and Geophysics prior to analysis at National Environmental Research Council Geosciences Laboratory, England. The water samples were treated with zinc turnings at 500° C to generate hydrogen for the D/H analysis, and equilibrated with CO_2 for $^{18}O/^{16}O$ analysis. Mass spectrometric determinations were performed on a VG SIRA+ISOPREP 18 in conjunction with laboratory standards calibrated against VPDB and VSMOW-SLAP scales. The results are reported in the usual δ notation in per mil (‰) versus these standards; reproducibility is better than 0.02 and 2‰ for $\delta^{18}O$ and δD, respectively.

The water balance and isotopic balance equations have been combined as follows to constrain the contribution of groundwater to the hydrologic budget of the lakes.

The general water budget equation of a closed lake can be given by:

$$\Delta V/\Delta t = P + R + Gi - E - Go \quad (1)$$

where, P, R, E, Gi and Go are precipitation, runoff, evaporation, groundwater inflow and groundwater outflow, respectively. V is the volume of water in the lake and t is the time.

The expression for the isotope balance of a closed lake (Karbbenhoft *et al.*, 1994) is:

$$\Delta(\delta_L V)/\Delta t = \delta_p P + \delta_p kP + \delta_{gi} Gi - \delta_{go} Go - \delta_E E \quad (2)$$

where each term in Equation (1) is multiplied by its respective isotopic composition expressed in delta notation in units of per mil relative to the Standard Mean Oceanic Water (SMOW). The kP term in the equation is the total rainfall that is converted to runoff, k is runoff coefficient. The runoff coefficient has been estimated based on field slope and hydrogeological observations of the craters containing the lakes and other similar studies made on runoff processes in the rift valley (HALCROW, 1989; Tenalem Ayenew, 1998).

Under isotopic steady-state conditions and a well mixed lake ($\delta_{go}=\delta_L$), Equation (2) can be simplified as:

$$\Delta(\delta_L V)/\Delta t = \delta_p P + \delta_p kP + \delta_{gi} Gi - \delta_L Go - \delta_E E. \quad (3)$$

Combining Equations (1) and Equation (3) gives:

$$Gi = [P(1+k)(\delta_L - \delta p) + E(\delta_E - \delta_L)] / [\delta_{gi} - \delta_L]. \quad (4)$$

Combining water budget and isotope budget equations, it is possible to arrive at equation (4). The advantage of equation (4) is that it is viable to quantify the groundwater inflow independent of the groundwater outflow unlike when the water budget equation is used alone. In using Equation (4), however, difficulties arise mainly because, δ_E (isotopic composition of evaporating water) is difficult to measure. The δ_E term can however be obtained from the following relation formulated by Craig and Gordon (1965):

$$\delta_E = [\alpha^* \delta_L - h\delta_A - \varepsilon] / [1 - h + 10^{-3} \Delta\varepsilon] \qquad \text{with } \alpha^* = 1/\alpha \quad (5)$$

where α is the equilibrium isotopic fractionation factor which is equal to 1.0098 for $\delta^{18}O$ at 25° C, h is the relative humidity normalized to water surface temperature, δ_A is the isotopic composition of local atmospheric water vapor, ε is the total fractionation factor, and $\Delta\varepsilon$ is kinetic fractionation factor. The total fractionation factor is the sum of the equilibrium fractionation factor (ε^* which is 1000 times α^*) and kinetic fractionation factors and is given by:

$$\varepsilon = \varepsilon^* + \Delta\varepsilon \tag{6}$$

$$\varepsilon = 1000\,(1-\alpha^*) + \Delta\varepsilon \tag{7}$$

$$\Delta\varepsilon = K\,(1-h) \tag{8}$$

The K term has been empirically determined to be 14.3 ‰ for ^{18}O and 12.5 ‰ for ^{2}H (Gonfiantini, 1986).

Equation (5) introduces a new unknown namely the atmospheric composition of local vapor (δ_A). It is the lack of measurement of δ_A and good accuracy in determining ε and h that poses problems in isotope balance methods. The method is suitable either where measurements of δ_A and h are available, or where extrapolation can be made from a lake whose isotopic balance has been carefully determined from an index lake which can be terminal from both groundwater and surface water points of view. The latter approach has been used in various studies (Dincer, 1968; Gat and Levy, 1978; IAEA, 1981).

The BCL provide a unique opportunity to solve the isotope balance without directly measuring all of the ambient environmental parameters of δ_A, h and ε. Based on hydrological and chemical mass balance approach, Hora has been selected as an index terminal lake due to the negligible groundwater outflow (Seifu Kebede *et al.*, 2001).

The isotopic composition of a terminal lake in a balanced state of total inflow and outflow will approach a steady-state value (IAEA, 1981), so that,

$$\delta_E = \delta_I \tag{9}$$

where δ_I is the weighted isotopic composition of total inflow. The information obtained from the intersection of local evaporation line with the Local Meteoric Water Line (Fig. 4) gives a δ_I value of -2.8 ‰ for ^{18}O.

For Lake Hora, as it is assumed to be a terminal lake, the isotope balance approach can be solved by directly substituting $G_o = 0$ in Equation (1) and calculating the unknown G_i. On annual basis, the change in volume of the lake is assumed to have no influence on the water balance and isotopic balance calculations.

The total mechanistic separation factor ε can be determined from Equation (7), *i.e.*, from equilibrium separation factor (ε^*) and diffusion controlled fractionation factor ($\Delta\varepsilon$). The average relative humidity record from the meteorological station close to the lake is 0.6. The kinetic temperature, which is important to calculate the open water evaporation, is extrapolated from the Thematic Mapper band six satellite image analysis (Tenalem Ayenew, 1998) from the nearby Ziway-Shala lakes basin.

Using Lake Hora as an index and substituting ($\delta_E = \delta_I = -2.8$ ‰) into Equation (5), the unknown δ_A has been estimated to be -12‰. The isotopic composition of atmospheric vapor determined from the empirical relation (-12‰) is close to that of the Afar region determined by Gonfiantini *et al.* (1973) which is -15‰ and that of lake Bosumtwi (Ghana in West Africa) which is -9‰ (Turner *et al.*, 1996).

These two environmental parameters (ε and δ_A) have been used to calculate the limiting isotopic composition δ^* for the Bishoftu area. This limiting isotopic composition is the isotopic enrichment value which residual waters of a shrinking water body approach (IAEA, 1981), and is given by:

$$\delta^* = \delta_A + \varepsilon / h. \tag{10}$$

The limiting isotopic composition for the region is calculated to be 14.2 ‰. This value is slightly higher than the maximum oxygen isotopic composition reported for the lakes of the Ethiopian rift, which is 10.7 ‰ (Craig *et al.*, 1977).

A more useful and simplistic isotopic balance equation for non-terminal lakes (Babogaya and Bishoftu) can be derived from the above equation and be rewritten as in equation (11). This expression was originally developed by Gat and Levy (1978) in the isotope budget studies of sabkhas in Sinai Desert.

$$\delta_{Lss} = \frac{\left[\left(\frac{\varepsilon}{h} + \delta_A\right) + \frac{I}{E}\left(\frac{1-h}{h}\right)\delta_I\right]}{\left[1 + \frac{I}{E}\left(\frac{1-h}{h}\right)\right]} \tag{11}$$

where, δ_{Lss} is the steady-state isotopic composition of a lake with inflow and outflow and I is the total inflow ($Gi + P + R$).

Equation (10) and (11) can be rewritten in a simpler form as:

$$\frac{I}{E} = \frac{h}{1-h}\left(\frac{\delta^{*} - \delta_{Lss}}{\delta_{Lss} - \delta_{I}}\right) \tag{12}$$

The only unknown in equation (12) is the total water inflow (I) and implicitly the groundwater inflow as P and R can be determined from hydrometeorological data.

RESULTS AND DISCUSSION

The isotope analysis results of rainfall, lakes, swamps, rivers and groundwater samples are given in Table 3. Water sampling points are shown on Fig.3. Fig. 4 shows $\delta^{18}O$ Vs $\delta^{2}H$ plots of these water bodies with respect to the Local Meteoric Water Line (LMWL). Table 4 shows the water balance components of the lakes obtained from the combined hydrological and isotope balances. The result of the hydrologic balance, the isotopic composition of each of the water types and their importance in indicating groundwater-surface water interactions is discussed.

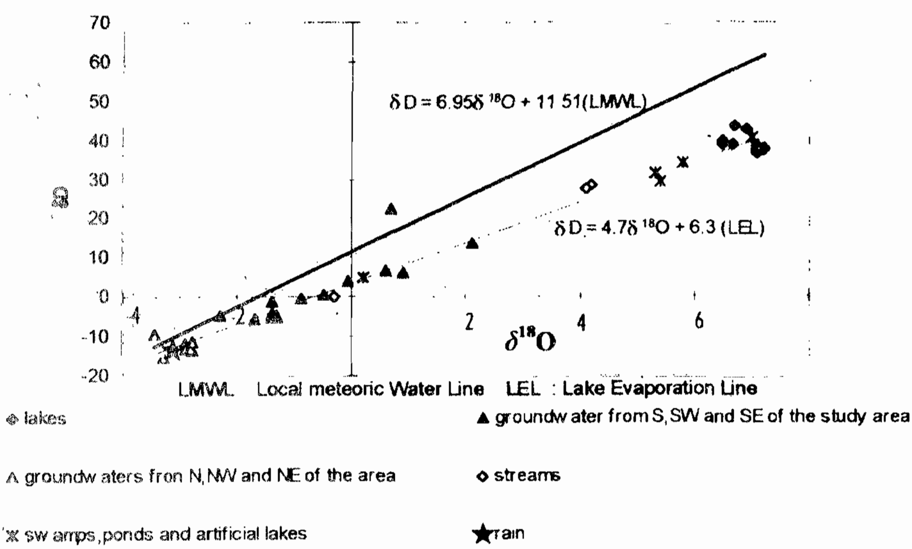

Fig. 4. $\delta\,^{18}O$ versus $\delta\,^{2}H$ plots.

Table 3. Isotope analysis data (EC electrical conductivity; blank where data not available).

No	Sample no.	Sampling Date	Sampling site	EC in μScm^{-1}	$\delta^{18}O$	δD
			Lake Waters			
1	BAB98-1	Jan-98	Babogaya, margin	816	6.5	38.9
2	BAB98-2	Jan-98	Babogaya, 60m depth	-	6.7	42.5
3	BAB98-4A	Jan-98	Babogaya, surface, center	848	6.7	38.9
4	BAB98-3	Jan-98	Babogaya,10m depth	-	6.5	39.6
5	HBM98-1	Jan-98	Hora, marginal (weeds)	2200	7.2	37.6
6	HBM98-2		Hora, marginal open water	2290	7.1	37.0
7	HBM98-3	Jan-98	Hora, lake bottom (30m)	2290	7.2	37.6
8	HBM98-4	Jan-98	Hora, 20m depth	2270	7.1	38.4
9	HBM98-5	Jan-98	Hora,10m depth	2250	7.2	37.4
10	HBM98-6	Jan-98	Hora, surface lake center	2180	7.1	37.8
11	B98-1	Jan-98	Bishoftu, margin	1570	6.9	42.9
			Swamps, ponds and artificial lakes			
12	CH98-1	Jan-98	Cheleleka, south shore	404	7.0	40.6
13	CH98-2	Jan-98	Cheleleka, east shore	293	5.8	33.4
14	KU98-1	Jan-98	Kuriftu, margin	206	5.4	29.5
15	KU98-2	Jan-98	Kuriftu, center	-	5.3	31.4
16	K98-1	Jan-98	Kilole, margin	1560	0.2	4.9
			Rivers			
17	MOJO98-1	Jan-98	Mojo river	418	-0.3	0.1
18	WR98-2	Jan-98	Wedecha river	205	4.2	28.4
19	WT98-1	Jan-98	Wedecha tributary	171	4.1	27.5
			Groundwaters from NW, NE and N of the area			
20	DZ98-3	Jan-98	Shimbra meda well field	518	-2.8	-13.5
21	DZ98-2	Jan-98	Management institute well	604	-2.9	-11.6
22	66	Jan-99	Dragados DZ well		-3.3	-15
23	63	Jan-99	Management institute well		-2.9	-13.1
24	DZ98-1	Jan-98	Shimbrameda well/tap	534	-2.9	-13.4
25	60	Jan-99	Agri-Res center well 2		-2.8	-11.5
26	57	Jan-99	Elfora		-3.4	-9.7
			Groundwaters from S, SW and SE of the area			
27	58	Jan-99	Hora Tannery well		-1.7	-5.6
28	69	Jan-99	Almaz Doro well		-1.4	-5.2
29	70	Jan-99	Slaughter House well		-1.4	-4.3
30	62	Jan-99	Agri. Res. Inst. well 1		-1.4	-1.1
31	61	Jan-99	N. Franco well		-0.9	-0.2
32	67	Jan-99	B. Nile Plastic Factory well		-0.5	0.5
33	65	Jan-99	Defense Eng. well 1		0.9	6
34	59	Jan-99	Ada Pasta Factory well		0.6	6.5
35	68	Jan-99	Hora Agro Industry well		2.1	13.8
36	64	Jan-99	Defense Eng. well 2		0.7	22.2
37	CHW98-2	Jan-98	Royal Hotel hand dug well	500	-1.3	-5.0
38	73	Jan-99	Ato Gizaw hand dug well		-2.3	-4.9
			Rainfall			
40	Rain	Jan-99	Close to Lake Hora		-3.0	-13.1

Isotopes and water balance

Evaporation and groundwater play very important roles in the hydrologic cycle of all the lakes (Table 4). The maximum groundwater inflow goes to Hora, located at relatively lower topographic position. It is believed that Babogaya supplies the major flux to the other lakes situated at lower elevation, following the regional hydraulic gradient. Lake Bishoftu has an intermediate value between the terminal Lake Hora and Lake Babogaya, which has groundwater outflow.

Table 4. Estimated water balance components of the lakes. (Expressed in m^3/yr and percentage of each fraction of the water balance components) where Gi is groundwater inflow to the lake, Go is groundwater outflow form the lake, P is annual volume of precipitation of the lake, E is evaporation flux from the lake and R is runoff flux to the lake.

Lake	P (m^3)	Gi (m^3)	R (m^3)	E (m^3)	Go (m^3)
Hora	854070	552355	353165	1759590	0
	49%	31%	20%	100%	0%
Babogaya	480570	587514	121595	990090	199590
	40%	50%	10%	83%	17%
Bishoftu	771070	804586	160190	1588590	147256
	44%	46%	9%	92%	8%

Table 5. Annual water budget of Bishoftu Crater Lakes from chloride mass balance approaches (from Seifu Kebede *et al.*, 2001).

Lake	P (m^3)	Gi (m^3)	R (m^3)	E (m^3)	Go (m^3)
Hora	854070	703882	260413	1759590	58775
	47%	39%	14%	97%	3%
Babogaya	480570	710678	121388	990090	322545
	37%	54%	9%	75%	25%
Bishoftu	771070	749859	160813	1588590	93151
	46%	45%	10%	93%	6%

The accuracy of the water budget estimate for Babogaya and Bishoftu depends on the validity of the assumption that Hora is terminal. One way to check this is to compare the result with estimates from another independent method. The result of this study is in good agreement with an independent estimate made using chloride mass balance approach as shown in Table 5 (Seifu Kebede *et al.*, 2001). The chloride mass balance shows that the groundwater outflow from Lake Hora is negligible, only 3%.

The slight difference in the results between the two approaches may be attributed to the degree of precision in calculating δ_E using the terminal lake approach and the validity of considering Lake Hora as a strictly terminal lake. The close similarity between the chloride mass balance and isotope budget calculations therefore shows that the BCL provide a means of calculating an isotope budget without depending on direct measurements of environmental parameters (h,δ_A, and ε). More precise results can be obtained when these environmental parameters are measured, particularly the relative humidity and isotopic composition of ambient vapor, are measured.

Isotopic variability and groundwater – surface water interactions

Rainfall: The isotopic composition of precipitation over Ethiopia and East Africa has been summarized by Rozanski *et al.* (1996) based on the Addis Ababa station data established in 1961 by the International Atomic Energy Agency (IAEA). The long-term weighted mean value for $\delta^{18}O$ is -1.3 ‰ and of ^2H is 1.8‰ for Addis Ababa station (2360m above sea level). The long-term record of Addis Ababa gives a LMWL defined by $\delta^2H = (6.95 \pm 0.22)\ \delta^{18}O + 11.51 \pm 0.58$. As the studied area, is close to Addis Ababa, this relation is used as a LMWL. A single rainfall sample collected from the area has isotopic composition of -3.13‰ for $\delta^{18}O$ and -13‰ for ^2H, and plots close to the LMWL.

Lakes and Other Evaporated Water bodies: The isotopic compositions of the three natural lakes show strong positive value. The enrichment of the lakes in the heavy isotopes indicates that they are in a state of evaporation with respect to the present day precipitation. The average $\delta^{18}O$ composition of Hora, Babogaya, and Bishoftu are 7.2‰, 6.5‰ and 6.9‰, respectively. Hora is the most enriched, as it is the case for terminal lakes, with enrichment of 0.7‰ and 0.3‰ as compared to Babogaya and Bishoftu respectively. In terms of ionic concentration, the lakes have higher values than the other water types. The electrical conductivity (EC) ranges from 816 µS/cm for Babogaya to 2290 µS/cm for Hora. As in the case of isotopic composition, the ionic concentration is by far lower than the main alkaline lakes of the Ethiopian Rift.

The isotopic variability of the lakes with depth is also very low. This suggests that the lakes are well mixed. Relatively, the variation with depth for Babogaya is slightly higher than Hora. This is perhaps due to the slightly higher depth that may lead to little isotopic stratification.

Generally the artificial lakes, swamps and rivers are isotopically more enriched ($\delta^{18}O$ is 0.2-7.0‰ and δ^2H is 4.9-40.6‰) than the groundwaters but less enriched than the three major natural lakes. The influence of evaporation on Cheleleka, being a very shallow lake, is clearly noticeable. The artificial lakes Kilole and Kuriftu are depleted due to flushing by the river waters. Samples from the perennial Mojo and Wedecha rivers and the tributary stream show isotopic values more enriched than the groundwater and rainfall. This is most likely due to *en route* evaporative enrichment as these tributaries are held in a reservoir (Balabala) upstream of the sampling point.

The isotopic composition of the evaporated water bodies (natural lakes, artificial lakes, swamps, and river waters) plot on or close to an evaporation line defined by $\delta^2H = 4.7\delta^{18}O + 6.3$. This equation resembles the East African lakes evaporation line (Craig *et al.*, 1977). The lakes within the Ziway-Shala basin show an evaporation line defined by $\delta^2H = 5.5\delta^{18}O + 8.5$ (Tenalem Ayenew, 1998). The LMWL intersects this evaporation line at a point of -2.8‰ for $\delta^{18}O$ and -13‰ for δ^2H. The intersection point of LMWL and evaporation line at a given site has been used to estimate the average isotopic composition of inflow waters to lakes in various isotope budget studies of lakes (Dincer, 1968).

Groundwaters: The $\delta^{18}O$ isotopic composition of groundwater in the study area shows variability ranging from -3 to 4 ‰. Samples from the north, northwest and northeast of the lakes (up gradient) plot close to the LMWL, indicating limited secondary enrichment. Therefore, the ground water is most likely recharged rapidly through fractures, faults, joints and scoria cones with little or no evaporation. However, the slight variation in isotopic composition among groundwater samples could be attributed to differences in source and temporal variations of recharge.

The isotopic compositions of wells to the south, southeast and southwest of the study area (down gradient) are slightly enriched and plot on a groundwater-lake mixing line. The isotopic enrichment in these areas may indicate mixing with evaporated waters (lakes and swamps). Mixing in this locality may be facilitated by fractures and faults that underlie the maars. A similar observation of mixing of groundwater in Bishoftu area with lake water is reported by Darling (1996).

Based on their topographic position and distribution of faults, lakes Bishoftu and Babogaya are believed to contribute the largest percentage of evaporated-type water to the groundwater system. However, when the overall isotopic composition of the subsurface water is considered, the influence of the lakes in the groundwater system is quite low. The groundwater isotopic composition is more of non-evaporated meteoric water than the concentrated water of the lakes. The EC value of these wells (518–599 μS/cm) reveal also rapidly circulating meteoric waters with low rock-water interactions. Hence, the residence time of the groundwater of the region is low. This is in good agreement with the study made around Akaki close to the study area (Aynalem Ali, 1999).

From the spatial isotopic variation, it can be deduced that groundwater flows generally from north to south in the study area. The chemistry and isotopic composition also reveals that most groundwater converges to the terminal Lake Hora.

CONCLUSIONS

The varying hydrology and particular hydrological setting of the BCL provides a good opportunity for stable isotope and hydrologic balance studies without reverting to the measurements of environmental parameters which are lacking in many isotope budget studies. The result of water budget calculations shows that the lakes have variable hydrology and that groundwater plays a significant role in the hydrology of the BCLs. This strengthens previous ideas on the importance of groundwater in the hydrology of the BCLs (Prosser *et al.*, 1968; Wood *et al.*, 1976; Wood and Talling, 1988; Zinabu Gebremariam, 1994). The varying amount of groundwater outflow might be responsible for the difference in the salinity of the BCLs which have fairly similar catchment geology. The isotopically more depleted groundwater with low ionic concentration indicates fast recharge process in the highly fractured volcanic area and the groundwater systems have low residence time.

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