

Article

# Use of $^{222}\text{Rn}$ and $\delta^{18}\text{O}$ - $\delta^2\text{H}$ Isotopes in Detecting the Origin of Water and in Quantifying Groundwater Inflow Rates in an Alarmingly Growing Lake, Ethiopia

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**Abstract:** Dual Radon ( $^{222}\text{Rn}$ ) and  $\delta^{18}\text{O}$ - $\delta^2\text{H}$  isotopes were utilized to (a) detect the origin of water, (b) pinpoint groundwater inflow zones and (c) determine rates of groundwater inflows in an expanding lake in central Ethiopia. The lake area expanded from 2 km<sup>2</sup> to 50 km<sup>2</sup> over the last 60 years, causing serious engineering and socio-economic challenge (inundation of urban utilities, irrigation farms, railways and roads; ecological changes in the lake; and threatening water salinization for water users downstream). Commensurate with the changes in volume, there was a change in salinity of the lake from a hypersaline (TDS 50 g/L) to a near freshwater (3 g/L) condition.  $^{222}\text{Rn}$  is powerful in pinpointing sites of groundwater inflows and determining groundwater inflow rates in lake systems with non-hydrologic steady-state conditions. The  $^{222}\text{Rn}$  method is complemented by the use of the stable isotopes of water ( $\delta^{18}\text{O}$ - $\delta^2\text{H}$  pair). The  $\delta^{18}\text{O}$ - $\delta^2\text{H}$  isotopes were used to discriminate the source of the water responsible for the expansion of the lake. The results show that the main source of water responsible for the expansion of the lake is the irrigation of excess water joining the lake through subsurface flow paths. The fast and voluminous flow is aided by a dense network of faults and by seismically induced modern ground-cracks that enhance the transmissivity of the aquifers to as high as 15,000 m<sup>2</sup>/day. The  $^{222}\text{Rn}$  mass balance shows the groundwater inflow rate is estimated at 4.6 m<sup>3</sup>/s. This is comparable with the 4.9 m<sup>3</sup>/s annual seepage loss from three large farms in the area. This work adds to the meager literature in the use of  $^{222}\text{Rn}$  in lake-groundwater interaction studies by demonstrating the capability of the method in addressing a practical engineering and socio-economic challenges.

**Keywords:**  $^{222}\text{Rn}$ ;  $\delta^{18}\text{O}$ - $\delta^2\text{H}$ ; lake-groundwater interaction; expanding lake; Ethiopia

## 1. Introduction

### 1.1. Background

Lake Beseka, a tectonic lake, located in central Ethiopia, is an exceptional lake where substantial increase in volume, area and depth, and decrease in salinity has been observed over the last 60 years [1,2]. The region is characterized by active tectonics related deformation [3,4], convergence of regional groundwater flow [5,6] and presence of intensive mechanized irrigation. Several hypotheses have been forwarded to explain the expansion of the lake including (a) seismically induced deformation of the basin which causes increased discharge of thermal springs feeding the lake [7], (b) change in regional

water balance leading to increased activity of groundwater discharge [6,8,9], (c) Irrigation return water from the adjacent farms joining the lake via sub surface paths [2,10] and (d) occasional accidental and unreported direct spill of excess irrigation water from farms through surface drains as a result of faulty drains arising from breakage and siltation related clogging [10].

Most of these hypotheses do not clearly identify the source of the water versus the mechanism that has led to increase in discharge of the water. For example, [7], using remote sensing method-based mapping of temperature of the lake, concluded the increase in discharge of thermal springs feeding the lake to be the principal cause of the swelling of the lake. However, the same authors indicated the source of the water that has led to the increment in thermal water discharge is unknown and is not related to the irrigation water.

The presence of competing possibilities makes it challenging to identify the direct and indirect causes of the lake level changes and to mitigate the impact of the growing lake [10]. The aims of this work are to investigate the source of water responsible for the lake level rise using  $\delta^{18}\text{O}$ - $\delta^2\text{H}$  and  $^{222}\text{Rn}$  tracers and to estimate groundwater inflow to the lake.

The  $\delta^{18}\text{O}$  and  $\delta^2\text{H}$  are proven tools in tracing movement of water around lakes [11–15].  $^{222}\text{Rn}$  isotope on the other hand is a proven tool in detecting specific site of groundwater discharge to open water bodies (lakes and rivers) and in quantifying the groundwater inflow to lakes without the need to reverting to detailed hydrological measurements [16,17] of groundwater flow into lakes.  $^{222}\text{Rn}$  is a radioactive daughter of radium-226, which in turn is a radionuclide in the U-Pb decay series.

### 1.2. Damages Caused by Lake Beseka Water Level Rise and Chemistry Changes

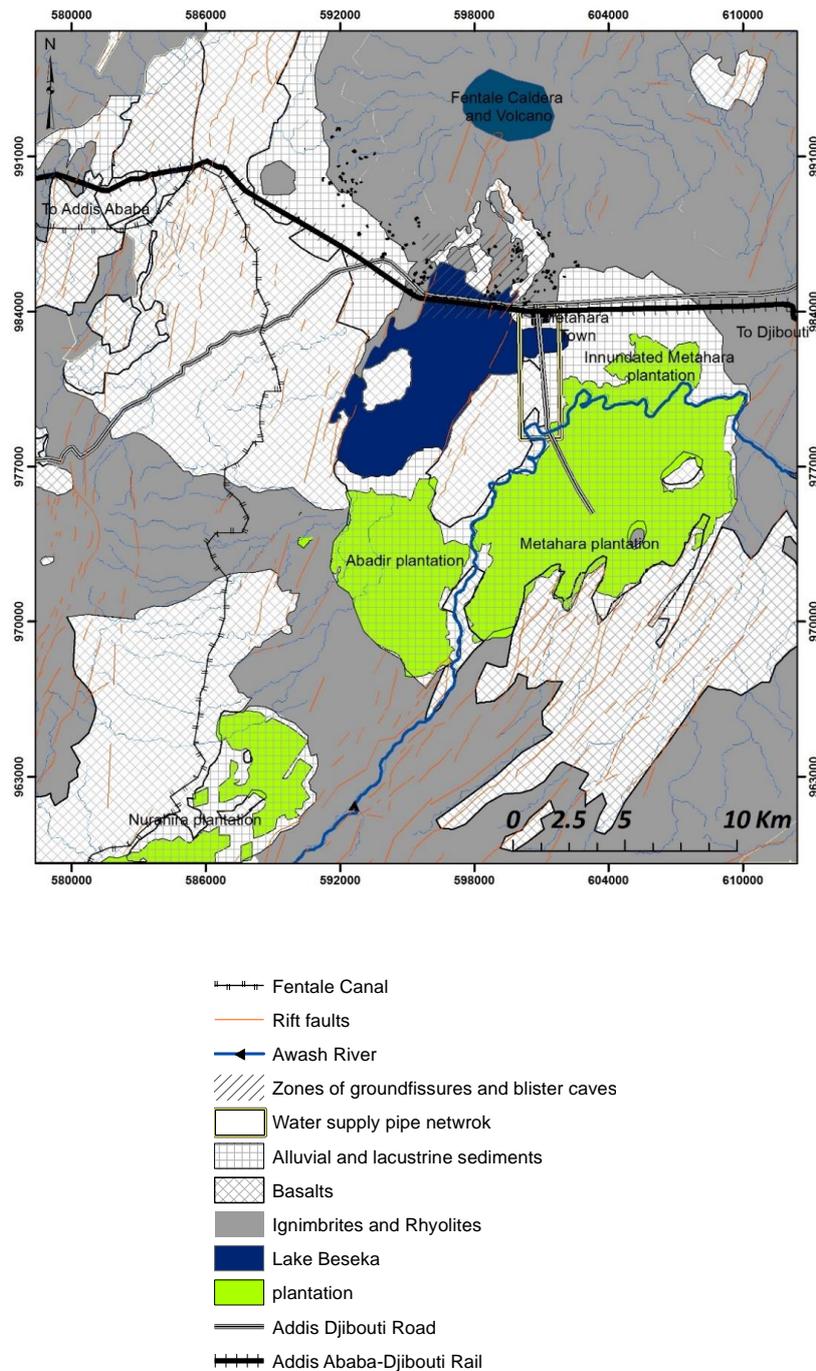
The economic damage caused by the expansion of Lake Beseka is enormous. The damage caused prior to 2008 is estimated in monetary value at 2.5 million USD [7]. Additional damage has been caused by the lake, which continued to expand since 2008 until it spilled into the nearby Awash River in 2010. The recent damages include (a) submergence under lake water of 8 km long drinking water distribution pipes and 5 km service pipes of the Metahara town water supply (20,000 people); (b) submergence underwater of urban utilities (clinics, schools and offices), residential areas and recreational facility; (c) submergence of 900 ha of sugarcane farm at Metahara plantation under saline water (Figure 1); (d) damage and submergence underwater of the Ethiopia-Djibouti railway and main express road connecting Addis Ababa to Djibouti port; (e) spillage of brackish water from the lake and in to an adjacent Awash River (Figures 2 and 3), degrading the quality of the river water for irrigation as well as drinking water use for millions of inhabitants downstream; (f) alteration/changes in ecology of the lake [1,18] and subsequent change in the faunal and floral composition of the lake; and (g) disruption of the movement path of wildlife in an adjacent national park.

### 1.3. Site Characterization

#### 1.3.1. Geology and Hydrogeology

The diversity of hypotheses made about the growth of Lake Beseka stems from the complexity of the geophysical and 'man-made' features of the region. Interplay of processes that can lead to change in hydrology and geometry of the lake coexists in the region. Firstly, the lake fills a tectonic depression in a seismically active region in East Africa. At around 1 mm/yr. extension, the fastest rift opening rate in the Ethiopian rift valley is centered on Lake Beseka area [4] leading to long-term and abrupt changes in lake geometry and geothermal activities. Reports show swarms of ground fissures (swarm of faults) have formed at different times over the last 100 years [3] following earthquake events. The recent swarms of fractures, six in number, are localized to the region north of the Lake over a 4 by 7 km belt [4]. The fractures generally appear as narrow cracks which may be several hundred meters long (Figure 1). Beyond this belt, they either disappear or grade into normal faults. Such cracks are believed to increase permeability structure of the rocks and thereby lead to increased discharge of springs feeding the lake [9].

Secondly, the area is underlain by volcanic rocks such as ignimbrites, volcanic ash, basalts and alluvio-lacustrine sediments. The basalts cover most of the western sector of the area. The basalts are highly vesicular with dense network of rift related fractures and joints leading to high transmissivity in the range of 7000 to 15,000 m<sup>2</sup>/day and hydraulic conductivity of 10 to 800 m/day [10]. The ignimbrites are densely dotted by blister caves, unique landforms that form when ashes fall on wet ground and subsequent vaporization of the moisture by heat inflates the overlying ash layer. This unique landform increases the permeability of the ignimbrites and ashes, which would otherwise have low hydraulic conductivity among the volcanic rocks in the region.



**Figure 1.** Characteristics of the study area: geology, inundated areas, extent of farms, irrigation canal location and infrastructure damaged by the lake expansion.

### 1.3.2. Hydrology, Chemistry and Irrigation Development

Lake Beseka was a small closed lake of 2.6 km<sup>2</sup> in the early 1960s before it expanded to its present size of 54 km<sup>2</sup> in area and (in 2014) 243,600,000 m<sup>3</sup> in volume. In the early 1960s, the lake was fed by saline springs discharging at the western lake shore [18,19]. These springs are now submerged underwater and discharge at more than 500 l/s. Commensurate with growth in size, electrical conductivity decreased tenfold from nearly 75,000 µScm<sup>-1</sup> in 1960s [19,20] to 3600 µScm<sup>-1</sup> in 2014. However, this decrease in EC was not regular. It markedly decreased in 1970s to 10,500 µScm<sup>-1</sup> over a period of 10 years and then decreased to around 6000 µScm<sup>-1</sup> over the next 40 years to 2009 (Belay, 2010). According to [8], the lake has an EC of 7100 µScm<sup>-1</sup> in 1998. Current EC values of the lake vary spatially—at 2600 µScm<sup>-1</sup>, it is the lowest in the South West bay area and at the 6400 µScm<sup>-1</sup> it is the highest in North West area. Detailed measurement by [7] shows that the temperature of the lake varies spatially. Maximum depth of the lake is recorded at 12 m while mean depth and mean volumes are 5 m and 243,600,000 m<sup>3</sup>, respectively.

Four large scale irrigation farms run since 1965, with a total irrigation area exceeding 20,000 ha. Net irrigation efficiency from the farms has been estimated at 40% [21] and return flow reaches 20% of the total intake. The detailed characteristics of the irrigation schemes are given in Table 1. Figure 1 shows the location of the farms. The Fentale irrigation scheme became operational in 2008 and the other three started in 1960s. The Fentale irrigation area is located in several small patches East of Fentale canal. Studies [10] MWIE (2014) shows that the main irrigation water loss takes place from the conveyance (primary) and distribution (secondary) canals and field level loss account for only 10% of total loss. Except the main Metahara farm, which is located at relatively lower ground (south of Awash River) and in a different catchment, three of the farms are located partially or completely within the surface and groundwater catchment of the lake (Figure 1). Compared to all the four irrigation sites, the lake is located at lowest elevation, forming a closed depression. A net total loss via seepage from the three schemes upstream of the lake is estimated at 4.89 m<sup>3</sup>/s [10]. In the year 2007, the conveyance loss from the Fentale main canal alone was estimated at 2.81 m<sup>3</sup>/s.

**Table 1.** Net loss m<sup>3</sup>/s of irrigation water to groundwater percolation over the period application [10].

Farm	Start Date	Farm Size (ha)	Uptake (m <sup>3</sup> /s)	Conveyance Loss (m <sup>3</sup> /s)	Distribution Loss (m <sup>3</sup> /s)	Field Level Water Application Loss (m <sup>3</sup> /s)	Net Seepage Loss (m <sup>3</sup> /s)
Main-Metahara-	1965	8000	6.02	-	-	0.60	1.41
Abadir-Metahara	1968	3500	3.51	-	-	0.35	1.06
Nura Hira	1970	3740	2.55	-	-	-	1.02
Fentale	2007	4520 *	4.61	1.2	0.3	1.3	2.81

The lake water balance and volume changes are accounted for by (a) a quasi-stable direct rainfall on the lake (0.6 m/yr.), (b) direct overland flow from wadi beds amounting to another 0.6 m/yr. [6], (c) an evaporation depth of 2 m/yr., (d) unknown amount of surface water outflow and (e) variable groundwater inflow and unknown groundwater seepage out of the lake.

## 2. Methodology: Theoretical Description

### 2.1. <sup>222</sup>Rn in Lake-Groundwater Interaction Studies

<sup>222</sup>Rn is a proven isotope in detecting surface water groundwater interaction and in quantifying groundwater flows to lakes and seas [16,17,22–24]. Qualitatively, <sup>222</sup>Rn can be used to pinpoint specific sites of groundwater discharge to lakes [16,22]. It is a good tracer because of its substantial quantity in groundwaters. Surface waters are devoid of <sup>222</sup>Rn. In surface water bodies, high <sup>222</sup>Rn concentrations are usually detected around the site of groundwater discharge allowing pinpointing groundwater advective inflow and seepage zones.

Quantitative estimation of groundwater inflow to lakes is based the assumption that  $^{222}\text{Rn}$  sources balances for  $^{222}\text{Rn}$  sinks under steady state conditions [16,22]. Under steady state conditions, water and  $^{222}\text{Rn}$  balance equations for a well-mixed lake can, respectively, be written as:

$$G + S + PA_L = Q + EA_L \quad (1)$$

$$C_G G + C_S S + FA_b = QC_L + kAC_L + \lambda VC_L \quad (2)$$

Table 2 gives a detail description of what the individual symbols means in the equations. The first equation means that, under a hydrologic steady state, the sum of all inflowing waters (groundwater, surface water and precipitation on the lake) equals the outflowing components (aggregate surface and groundwater outflow and evaporation from the lake). Similarly, the  $^{222}\text{Rn}$  balance equation (Equation (2)) is obtained by balancing the total  $^{222}\text{Rn}$  entering and total  $^{222}\text{Rn}$  leaving the lake. The  $^{222}\text{Rn}$  balance in Equation (2) is a little bit complex. This is because there is additional source and sink of  $^{222}\text{Rn}$  other than the advective flux coming and leaving the lake along the water flow components in Equation (1). That is (a) additional  $^{222}\text{Rn}$  flux (F) joins the source terms from diffusion from lake bottom sediments covering an area ( $A_b$ ), (b) additional loss of  $^{222}\text{Rn}$  takes place from the lake,  $C_L$ , through radioactivity decay at rate of  $\lambda = 0.18 \text{ day}^{-1}$ , over the lake volume (V) and (c) additional sink of  $^{222}\text{Rn}$  takes place through exchange with overlaying atmosphere at velocity  $k = 0.16 \text{ m/day}$  from the surface area of the lake. Thus, Equation 2 balances all the diffusive and advective sources with combined advective, radioactive decay and gas exchange sinks of  $^{222}\text{Rn}$ .  $^{222}\text{Rn}$  coming from production within lake by decay of dissolved or suspended parent isotope ( $^{226}\text{Ra}$ ) is assumed to be negligible as measured and found so in many previous studies [17] because  $^{226}\text{Ra}$  bounds to sediments than to water.  $^{222}\text{Rn}$  concentration in atmosphere is negligible; thus, the  $^{222}\text{Rn}$  flux to the lake along with rainwater is not included in Equation (2).

Combining the two equations (Equations (1) and (2)), an equation for computing the groundwater inflow rate (G) without reverting to the need of data on aggregated ground and surface water outflow can be derived (Equation (3)). Details of the derivation of the equation and the explanations can be found in [16,21] and [22].

$$G = \frac{(S + PA_L + kA_L + \lambda V)C_L - FA_b - SC_S}{C_G - C_L} \quad (3)$$

**Table 2.** Description of terms used in Equations (1) and (2).

Variable	Unit	Explanation
G	m <sup>3</sup> /day	Groundwater inflow rate
S	m <sup>3</sup> /day	Surface water inflow from channelized and overland flows
P	m/day	Depth of precipitation on the lake
Q	m/day	Aggregate water loss to surface and groundwater outflows
A <sub>L</sub>	m <sup>2</sup>	Lake Area, m <sup>2</sup>
E	m/day	Depth of evaporation from lake surface
A <sub>b</sub>	m <sup>2</sup>	Area of Lake bottom surface contributing $^{222}\text{Rn}$ flux via diffusive pathways from sediments
V	m <sup>3</sup>	Lake volume
λ	day <sup>-1</sup>	$^{222}\text{Rn}$ decay constant

Table 2. Cont.

Variable	Unit	Explanation
k	mday <sup>-1</sup>	Gas exchange velocity (wind speed and temperature turbulent degassing rate)
F	Bq/m <sup>2</sup> /day	Diffusive <sup>222</sup> Rn flux from lake bottom sediments, F is computed as in [16,22,23] taking into consideration the lowest <sup>222</sup> Rn content in the lake corresponds to the sources from diffusive flux only $FA_b = kAC_L + \lambda VC_L$ where all the parameters except the diffusive flux F are known
C <sub>L</sub>	Bq/L	Mean <sup>222</sup> Rn activity in the lake
C <sub>G</sub>	Bq/L	Mean <sup>222</sup> Rn activity in inflowing groundwater obtained from measurement in a borehole
C <sub>S</sub>	Bq/L	Mean <sup>222</sup> Rn activity in inflowing runoff

The advantage of radon is that it reaches steady state condition over the matters of few days [16] because of its apparent short half-life (3.82 days) and low residence time. This justifies the assumption of radon steady-state condition [16,22,23] in Equation (2). Since nutrient loading and chemistry of groundwater depends on the depth of water circulation, identifying the source of the groundwater inflow to lakes may be essential. In this particular study separating the shallow groundwater from the deep groundwater inflows to the lake is essential in order to understand the source of the groundwater (deep circulating geothermally heated groundwaters versus shallow irrigation return water) responsible for swelling of the lake. To help separate the groundwater source to the lake, independent  $\delta^{18}\text{O}$ - $\delta^2\text{H}$  isotope tracers were used.

## 2.2. $\delta^{18}\text{O}$ - $\delta^2\text{H}$ in Lake-Groundwater Interaction Studies

The isotopes of water molecule ( $\delta^{18}\text{O}$ ,  $\delta^2\text{H}$ ) are proven tools in lake-groundwater interaction studies. These isotopes have been used in detecting direction of groundwater flow around lakes [13]. This is because pristine groundwaters unaffected by lake water mixing shows the isotope signal of local or regional groundwaters while when  $\delta^{18}\text{O}$ - $\delta^2\text{H}$  enriched lake water mixes with it, the groundwater isotopic signal changes towards positive value proportional to mixing ratio. This helps to identify groundwaters affected by lake water inflow. The  $\delta^{18}\text{O}$ - $\delta^2\text{H}$  isotopes have also been widely used in quantitative determination of evaporation rate and groundwater flow around lakes. Extensive literature exists on the use of  $\delta^{18}\text{O}$  and  $\delta^2\text{H}$  in quantitative determination of lake water balance. Because of the likely absence of isotopic steady state condition of Lake Beseka and lack of time series of  $\delta^{18}\text{O}$  and  $\delta^2\text{H}$  of the lake, the method has not been used in quantitative determination of the water balance of the lake. These isotopes are also used in determining the composition of the initial water that feeds the lakes. This indirectly helps in identifying the type of water that feeds the lakes. It has been proven in many studies that the  $\delta^{18}\text{O}$ ,  $\delta^2\text{H}$  composition corresponding to intersection point of the local evaporation line (LEL) and the local meteoric water line (LMWL) in a  $\delta^{18}\text{O}$ - $\delta^2\text{H}$  plot represents the average composition of the waters feeding lakes under consideration [11,14]. This allows the tracing back of the source of groundwater as shallow or deep, regional or local and geothermal or cold water if these sources are marked by different compositions.

## 3. Data requirements, Sampling and Analysis

### 3.1. <sup>222</sup>Rn and $\delta^{18}\text{O}$ - $\delta^2\text{H}$ in Waters

<sup>222</sup>Rn survey has been conducted using portable DURRIGE-RAD7 electronic <sup>222</sup>Rn counter fitted with RAD aqua module. A total of 16 <sup>222</sup>Rn inventory points were taken. The measurement sites within the lake were chosen, taking key questions into consideration, including: (a) what is the role

of the seismically induced ground fissures and blister caves in the northern sector of the lake in channeling groundwater to the lake, (b) what is the role of the irrigation return water from irrigation farms located in southern, western and south western sector of the lake and (c) is there upwelling of groundwater inflow throughout the bottom of the lake. Considering these, most of the measurements were conducted in all areas except in the south eastern and south-central part of the lake. Groundwater  $^{222}\text{Rn}$  content was measured in one borehole and one spring. The measurements were conducted from 21 March, 2014 to 30 March, 2014. The counting is based on the principle of liquid–gas–membrane extraction [25] extraction module, which consists of hollow vinyl fibers, allows  $^{222}\text{Rn}$  stripping from the water of interest into a connected closed air-loop.  $^{222}\text{Rn}$  counting was conducted over a period of 4 to 5 h depending on the stabilization of the reading. Subsequent to the process water temperature was measured using HoBo temperature sensor. The temperature record along with the  $^{222}\text{Rn}$  record is entered into CAPTURE software provided by the supplier of the DURRIGE equipment to obtain the final  $^{222}\text{Rn}$  result accounted for water temperature variation. The overall standard deviation varies between 15 and 25%, higher for low  $^{222}\text{Rn}$  contents. The first read is discarded and the average of the last four reading was taken as the mean  $^{222}\text{Rn}$  composition of the specific water point. Accompanying  $^{222}\text{Rn}$  measurement survey of electrical conductivity was also conducted. The measured data and the measurement sites are shown in Table 3 and Figure 2. The assumption is that the measured  $^{222}\text{Rn}$  content in the well represents the mean composition of  $^{222}\text{Rn}$  in inflowing groundwater. This assumption is not strictly valid in cases where diffusive flux dominates the groundwater inflow to lakes [22]. In the study, because of the fact that discreet fracture flow is the main mode of groundwater inflow to the lake, the assumption that  $^{222}\text{Rn}$  in groundwater well equals the  $^{222}\text{Rn}$  concentration in the inflowing water gives sense. In order to show the uncertainties associated with this, a sensitivity run was conducted. The mean of the measured lake  $^{222}\text{Rn}$  content was used as input into the Equation (3). Since all surface water inflows are seasonal and occur from dry wadi beds with no contact with groundwater the  $^{222}\text{Rn}$  concentration in surface runoff to the lake is considered negligible. Table 4 enumerates the list of variables, parameters and associated values utilized in this work and in  $^{222}\text{Rn}$  mass balance calculations.

**Table 3.**  $^{222}\text{Rn}$  data from Lake Beseka and groundwater around the lake (standard deviation of individual measurement varies between 15 and 25%). UTM coordinates are for 37N region.

Code	UTM, East	UTM, North	Mean Bq/m <sup>3</sup>	Highest Bq/m <sup>3</sup>	Lowest Bq/m <sup>3</sup>
LBRn-12	593,967	982,945	241	336	97.7
BARn-1	596,605	973,681	3910	6290	1600
LBRn-19	594,903	981,620	54	86.3	18.9
LBRn-OL	600,000	982,426	31	44.7	12.6
LBRn-CO	594,663	975,450	294	549	124
LBRn-11	592,610	981,505	798	1200	138
LBRn-14'	596,650	983,509	35	38.4	31.5
LBRn-13	594,748	984,042	453	641	104
LBRn-21	599,118	983,737	18	29.6	9.44
LBRn-3	594,651	975,441	191	231	158
LBRn-9	591,564	979,251	2060	2670	864
LBRn-20	597,419	982,370	41	48	10
LBRn-5	591,653	976,332	2340	3620	684
LBRn-17	598,483	980,546	34	51.1	9.44
LBRn-14	596,809	983,113	27	38.9	18.9
LBRn-15	599,385	982,689	22	34	12

**Table 4.** Description of terms used in  $^{222}\text{Rn}$  mass balance equation.

Variable	Unit	Value and Source
S	$\text{m}^3/\text{day}$	81,500
P	$\text{m}/\text{day}$	0.00151, meteorology station
E	$\text{m}/\text{day}$	0.00548, meteorology station
Q	$\text{m}/\text{day}$	Aggregate water loss to surface and groundwater outflows (not required in Equation (3))
$A_L$	$\text{m}^2$	49,000,000 (excluding the area of the island)
$A_b$	$\text{m}^2$	49,000,000
V	$\text{m}^2$	243,000,000 (estimated from bathymetric survey [10])
$\lambda$	$\text{day}^{-1}$	$^{222}\text{Rn}$ decay constant
k	$\text{mday}^{-1}$	0.16 [16]
F	$\text{Bq}/\text{m}^2/\text{day}$	5.78
$C_L$	$\text{Bq}/\text{L}$	0.4 (mean of $^{222}\text{Rn}$ in central and northern part of the lake)
$C_G$	$\text{Bq}/\text{L}$	3.9 (measured in a borehole in Abadir farm)
$C_S$	$\text{Bq}/\text{L}$	0

Samples for  $\delta^{18}\text{O}$  and  $\delta^2\text{H}$  were taken intermittently between 1997 and 2014 from the lake, the Awash River and from adjacent groundwaters. The samples were acquired under the IAEA-TC projects entitled ETH8003, ETH8005, ETH8007 ([www.iaea.org](http://www.iaea.org)). During this time, a total of 60 samples were collected and measured for  $\delta^{18}\text{O}$ - $\delta^2\text{H}$  content in the laboratory of Krakow and the IAEA Isotope Hydrology Laboratory Vienna Austria (ETH8/003, ETH8005/ ETH80060 and in the Laboratory of Nuclear physics University of Krakow, Poland (ETH8/007). All the  $\delta^{18}\text{O}$ - $\delta^2\text{H}$  results are reported in per-mil (‰) notation calibrated against V-SMOW. The groundwater samples were collected from two distinct zones. The first zone represents groundwater in the catchment of the Lake Beseka area and particularly from piezometers and water supply wells located inside the irrigation schemes as wells as in upstream and downstream zone of the anticipated groundwater flow direction to and from the lake. The second group of groundwater samples represent deep groundwaters (>100 meters) collected from drilled deep wells along an east-west transect stretching from lake Koka to Lake Beseka (Figure 2).

### 3.2. $^{222}\text{Rn}$ in Diffusive Flux

Diffusive  $^{222}\text{Rn}$  flux (F) in Equation (3) is  $^{222}\text{Rn}$  joining the lake from the lake bed sediments. This is the function of the bulk density of the sediments,  $^{222}\text{Rn}$  diffusion coefficient in sediments and  $^{222}\text{Rn}$  production rates from  $^{226}\text{Ra}$  in the sediments. Ideally the diffusive  $^{222}\text{Rn}$  flux is determined from consideration of bulk density of sediments, bulk density of the grains, the  $^{222}\text{Rn}$  diffusion coefficient and the  $^{222}\text{Rn}$  production rate from  $^{226}\text{Ra}$  in water using diffusion equation [16] or using empirical relations [26] (Burnett et al., 2003). In cases where groundwater advection is significant, which is true for the lake under investigation, inputs via diffusion can usually be ignored [17,23]. The near absence of lake bed sediments [7] underneath the lake and the rocky bottom of Lake Beseka could justify this. An alternative approach is to solve for F in Equation (2) by assuming the lowest  $^{222}\text{Rn}$  content in the lake corresponds to region of only diffusive  $^{222}\text{Rn}$  flux [16,22]. This means Equation (2) reduces to  $FA_b = kAC_L + \lambda VC_L$  for that part of the lake. Plugging the lowest lake  $^{222}\text{Rn}$  concentration in the equation, the diffusive flux (F), can be solved for. In this specific work we followed this approach to compute F and to later show the sensitivity of the estimated groundwater inflow to the variation in the F.

### 3.3. Lake Geometry (Bathymetry, Mean Depth, Area)

Bathymetric survey of the lake has been conducted in a number of occasions [7,10]. The geometric data (mean depth, lake volume, lake bottom area and lake surface area) required in the  $^{222}\text{Rn}$  balance equation is retrieved from [10]. It is assumed that the surface of the lake bottom sediments over which diffusive flux of  $^{222}\text{Rn}$  is taking place is equal to the lake surface area. The size of the island in the

middle of the lake is accounted for in the lake area computation. Because the lake is spilling water since 2010 through an artificial canal in the eastern sector of the lake, the area of the lake is pretty much stable since that time.

#### 3.4. Timing of $^{222}\text{Rn}$ and $\delta^{18}\text{O}$ - $\delta^2\text{H}$ Data Collection

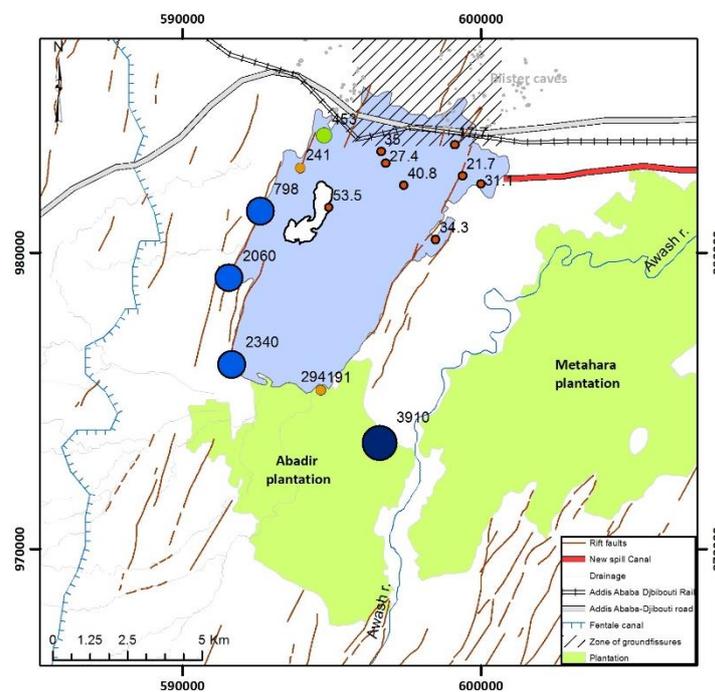
The primary approach of this work is complementing results obtained from  $^{222}\text{Rn}$  with the results obtained from  $\delta^{18}\text{O}$ - $\delta^2\text{H}$  regarding the magnitude and the sources of water entering the lake. However, it should be noted that neither the location nor the timing of the samples for each measurement are the same.  $^{222}\text{Rn}$  data has been collected over a period of 9 days in March 2014, while the stable isotope data has been collected over a period of 17 years (1997–2014). Since the two isotope sets were used for different but complementary purposes, the difference in timing of the data collection should not be a problem. The  $\delta^{18}\text{O}$ - $\delta^2\text{H}$  dataset were used to detect the composition of the initial water that feeds the lake. This is based on the theory that the intersection between the Local Evaporation Line (LEL) and the local meteoric water line (LMWL) represents the composition of the initial water that feeds open water bodies. The slope of the local evaporation line depends on the humidity and the isotopic composition of the ambient vapor. These two parameters, humidity and the  $\delta^{18}\text{O}$ - $\delta^2\text{H}$  composition of the ambient air, can reasonably be assumed have remained unchanged since 1997. This means the difference in time of the data collection will not erase the complementarity of the two tracers.

## 4. Results and Discussion

### 4.1. Causes of Lake Level Rise: $^{222}\text{Rn}$ Evidence

The first management question about the lake is the reason behind the alarming expansion of the lake. Since there is no change in the trends in rainfall and therefore runoff in the catchment, the suspect is thus an increase in groundwater discharge. We followed the approaches proposed elsewhere in relating the  $^{222}\text{Rn}$  pattern with pinpointing groundwater discharge zones. Pattern in the  $^{222}\text{Rn}$  has been used for example to interpret the role of the lake bed geology in governing  $^{222}\text{Rn}$  pattern [22] or in localizing sites of groundwater discharge [16,24]. A closer inspection of the  $^{222}\text{Rn}$  pattern presented in Figure 2 shows anomalous  $^{222}\text{Rn}$  content along the shoreline in the South (S), South West (SW) and Western (W) part of the lake. The shoreline in the north and the east as well as the central part of the lake shows no such anomaly. The  $^{222}\text{Rn}$  pattern indicates groundwater inflow is restricted to the SW, S and W part of the lake. The low  $^{222}\text{Rn}$  content in the northern portion of the lake precludes presence of substantial groundwater discharge emerging from the volcano in the North. An increase in discharge of groundwater coming from the northern sector of the lake catchment was one of the primary suspects [10]. This is because, over the last six decades, a number of new seismically induced ground-fissure swarms have occurred. This process was suspected to enhanced the hydraulic conductivity of the aquifers (Figures 1 and 3) and channeling the groundwater stored in the abundant buried and exposed blister caves in that region. The very low  $^{222}\text{Rn}$  concentration in the northern part of the lake means the ground fissures and the caves are not contributing significant volume of groundwater to the lake in that sector. The lack of high  $^{222}\text{Rn}$  content in the center of the lake also indicates the absence of substantial advective groundwater inflow from the lake bed precluding the fact that the entire groundwater regime in surrounding the lake has been affected. The pattern in  $^{222}\text{Rn}$  variation mirrors the pattern observed in surface temperature variation as reported by [7]. High lake surface temperature recorded by the authors corresponds to the sites where we recorded high  $^{222}\text{Rn}$  content in the lake water. The authors used the surface temperature anomaly of the lake water to pinpoint sites of groundwater inflow, by assuming that the high temperature areas are where anomalous groundwater inflow is happening. There is a very good match between temperature anomaly data and the  $^{222}\text{Rn}$  values at the three sites with the highest  $^{222}\text{Rn}$  content (2340 Bq/m<sup>3</sup>, 2060 Bq/m<sup>3</sup>, 798 Bq/m<sup>3</sup>). Typical values of surface water affected by groundwater would range in  $^{222}\text{Rn}$  content in the order of tens to few hundreds of Bq/m<sup>3</sup>. As can be observed, the  $^{222}\text{Rn}$  content of

the lake near the island a km from the shore shows a value of 53 Bq/m<sup>3</sup> confirming the rapid loss of the <sup>222</sup>Rn in the lake water over short distance from the shore. The anomalously high <sup>222</sup>Rn and temperature values at the three sites correspond to sites of the submerged springs. Both <sup>222</sup>Rn and temperature methods thus confirm increase in discharge of these springs are the primary cause of the lake volume changes. A visual observation of the groundwater inflow sites in the SW shore of the lake during the sampling shows spring waters forcefully pushing the overlying lake water column over an area of nearly 500 m<sup>2</sup> [7] proposed the increase in discharge of these springs is caused by increase in geothermal activity triggered by the recent tectonic activities. [7] precludes the role of the irrigation return water from the main possible causes on the assumption that irrigation return water should have lowered the lake temperature in that zone. This is because irrigation uses cold river water from nearby Awash River. The temperature-based evidence did not confirm the actual source of the water itself. The <sup>222</sup>Rn method is insufficient to confirm sources of groundwater. [16] for example proposed use of other tracers if such information is sought in <sup>222</sup>Rn based lake groundwater interaction studies.



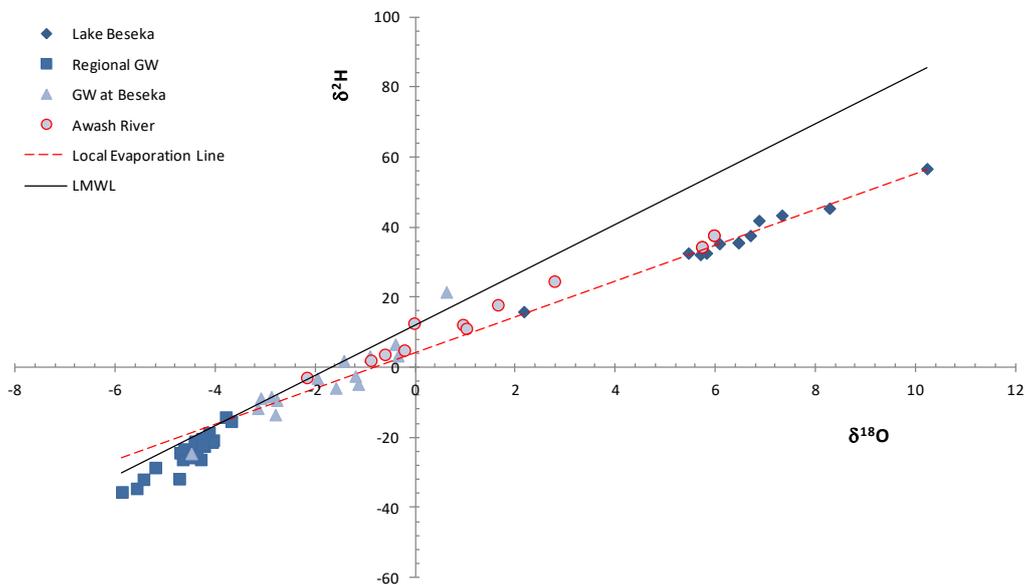
**Figure 2.** <sup>222</sup>Rn composition map of the lake and groundwater, the size of the circles are proportional to the <sup>222</sup>Rn content, the numbers marked near the circles indicate the actual <sup>222</sup>Rn content of the site.

#### 4.2. Cause of Lake Level Rise: $\delta^{18}\text{O}$ - $\delta^2\text{H}$ Evidence

The stable isotope of water ( $\delta^{18}\text{O}$ - $\delta^2\text{H}$ ) can provide complementary evidence to the <sup>222</sup>Rn and provide evidence on the actual sources of groundwater that is feeding the lake. It is long established that isotopes of water are good tracers of origin of groundwater in lakes. Early on, [11] suggested that the  $\delta^{18}\text{O}$  and  $\delta^2\text{H}$  composition corresponding to the intersection point between the Local Evaporation Line (line that best fits the  $\delta^{18}\text{O}$  versus  $\delta^2\text{H}$  plot of evaporated water bodies) and the Local Meteoric Water Line represents the net composition of initial water that feeds the evaporating water body. The same approach has been used to trace the  $\delta^{18}\text{O}$ - $\delta^2\text{H}$  composition of the groundwaters feeding the various Boreal lakes in Canada [14], lakes in Ethiopia [15] and lakes elsewhere [12]. In most studies, the net  $\delta^{18}\text{O}$ - $\delta^2\text{H}$  composition of the waters feeding the lakes is sought for water balance study purposes. This is because the  $\delta^{18}\text{O}$ - $\delta^2\text{H}$  composition of inflowing water is an input parameter in the lake isotope balance models [11,12,27,28]. However, in the current study, the information is sought to first identify the  $\delta^{18}\text{O}$ - $\delta^2\text{H}$  composition of the inflowing water and then seek further what kind of groundwater this



regional deep groundwaters originating from the plateau and they plot along the Local Evaporation Line and on enriched end of the LMWL.



**Figure 4.**  $\delta^{18}\text{O}$ - $\delta^2\text{H}$  plot of waters from Lake Beseka Surrounding and regional deep groundwaters.

The intersection point between the LEL and the LMWL shows an enriched  $\delta^{18}\text{O}$ - $\delta^2\text{H}$  value compared to the deep regional groundwaters, testifying the deep regional groundwater is not contributing significant water to the lake. The LEL points to or originates from water with  $\delta^{18}\text{O}$  and  $\delta^2\text{H}$  composition similar to local groundwaters within the Lake Beseka surroundings (including groundwaters from Metahara state farm) testifying the origin of the Lake Beseka is from shallow groundwaters linked to the irrigation farms. A closer look at the comparison between the composition of shallow groundwaters around Lake Beseka area and the Awash River waters shows slight compositional difference. The shallow groundwater has slightly depleted  $\delta^{18}\text{O}$ - $\delta^2\text{H}$  composition compared to the Awash River water testifying the shallow water is derived from some slight admixture (around 10%) of regional groundwaters and more than 90% irrigation water drawn from the Awash River. This means the major source of the water that is responsible for Lake Beseka expansion has its composition similar to the shallow groundwaters in the irrigation farms. Deep geothermal and deep groundwaters hypothetically cannot produce the composition of the Lake Beseka water. This precludes the fact that the spring discharge increase hypothesized by previous authors [2,7] has been as the result of the upwelling of the deep geothermal groundwater. We hypothesize the increase in the spring discharge is caused by irrigation excess water that enters the groundwater and returns to surface after heated by the local thermal activity.

One of the other hypotheses about the cause of the lake level rise was the disturbance in the regional groundwater regime (water level) following the construction of the Koka dam (Figure 3) and thereby increase in groundwater discharge to Lake Beseka [8]. In a farm further upstream (the Wonji farm) (Figure 4) groundwaters shows enriched  $\delta^{18}\text{O}$ - $\delta^2\text{H}$  content similar to the Awash River. At around +4 ‰ in  $\delta^{18}\text{O}$ , lake Koka is enriched too [32]. Inspection of Figure 3 shows almost all deep well waters between Lake Koka and Lake Beseka show little isotopic enrichment with little sign of admixture from the Koka reservoir. The effect of Lake Koka on adjacent groundwaters is limited to few hundred meters within the outflow zone [32]. Thus, these observations preclude the role of the Koka dam construction in affecting the water volume in Lake Beseka.

Groundwater outflow is already detected downstream of Lake Beseka from the spatial plot of  $\delta^{18}\text{O}$  in groundwaters. The waters downstream (Figure 4) show enrichment in  $\delta^{18}\text{O}$  as a result of mixing of the regional groundwater with groundwater leakage from the lake.

#### 4.3. Quantifying Rate of Groundwater Inflow- $^{222}\text{Rn}$ Balance Method

Using the steady state solution for the  $^{222}\text{Rn}$  flux (Equation (3)) and considering the input parameters in Table 2, the groundwater inflow to the lake has been estimated. The value is presented in Table 5.

Regardless of the inherent uncertainties in some of the input parameters and the various assumptions made, the estimated groundwater inflow rate ( $4.6 \text{ m}^3/\text{s}$ ) is comparable with the net irrigation water loss to seepage  $4.89 \text{ m}^3/\text{s}$  from the three irrigation schemes upstream of the lake (Table 1). Compared with previous estimates [6,7], the  $^{222}\text{Rn}$  based method gives higher groundwater inflow rate. This probably relate to the fact that the year 2014 represent a different hydrologic regime with the introduction of the Fentale canal and the associated irrigation schemes and the resulting substantial leakage loss from the Fentale canal. In the year 2010 for instance water loss to subsurface leakage has been estimated at  $2.85 \text{ m}^3/\text{s}$  from the Fentale irrigation scheme [10].

Another independent check on the validity of the result is the observation made on the outflow from the lake. The outflow from the lake during the week of sampling was estimated at around  $2.5 \text{ m}^3/\text{s}$  and it was comparable to the difference between the estimated groundwater inflow rate and the evaporation rate from the lake. A more complete independent check of the inflow rate however can be obtained from a simple water balance computation. If one assumes the water balance of the lake during the hydrologic condition of March 2014 represent the steady state hydrologic condition; Equation (1) can be solved for total surface and groundwater outflow rate (Q) plugging in the groundwater inflow rate (G) obtained from the  $^{222}\text{Rn}$  balance method. This resulted in a total outflow rate of  $2.2 \text{ m}^3/\text{s}$ . This is comparable with the surface water outflow via the artificial canal connected to the lake. The challenge with this assertion is the fact that some of the outflow could go through subsurface path as also observed from  $\delta^{18}\text{O}$  plot (Figure 5). The presence of groundwater outflow from the lake can be confirmed from a closer look of the  $\delta^{18}\text{O}$  pattern of groundwater around lake Beskea (Figure 5). The enrichment in the groundwaters east of the lake (some of the wells are now buried) is indicative of groundwater outflow in eastern direction of the lake. The isotope pattern mirrors the general west to east groundwater flow pattern described in [10] from piezometric evidence. Given the low permeability of the ignimbrite and ash beds under the very thin lacustrine sediments east of Lake Beseka zone, the groundwater outflow rate can be considered small.

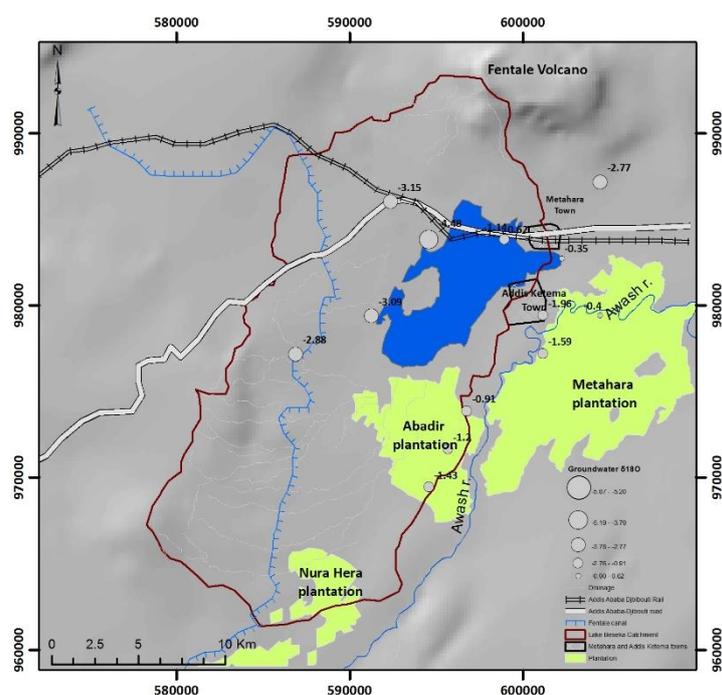


Figure 5. Spatial variation in  $\delta^{18}\text{O}$  of groundwaters around Lake Beseka.

**Table 5.** Annual groundwater inflow rate in  $10^6 \text{ m}^3$  to the lake, estimates from previous studies are included.

Current Study	Belay, 2010 [6]	Goerner et al., 2009 [7]	MWIE, 2014 [10]	HALCROW, 1989 [33]	Tesemma, 1998 [8]
170	33	17	103 *	1.5	37

\* The value is for base year 2010.

#### 4.4. Sensitivity Analysis

The inherent uncertainty in the model input parameters necessitates sensitivity analysis run so as to evaluate the validity of the results. Figure 6a–f shows sensitivity of the model output (groundwater inflow rate) to the various input parameters. The sensitivity was run on selected input parameters. The selection is based on the level of confidence on the value of the input parameters, i.e., parameters on which we made assumptions were chosen for sensitivity run and parameters for which exact measurement exist were not chosen (e.g., lake geometry parameter, area and volume were not chosen and the measured evaporation rate, precipitation rate). In the sensitivity test a parameter value was forced to vary by certain amount keeping all other values as in the model input value except for the one under consideration. The chosen variables were (a) the  $^{222}\text{Rn}$  concentration of the lake ( $C_L$ ), (b) the diffusive  $^{222}\text{Rn}$  flux ( $F$ ), (c) the gas exchange velocity ( $k$ ), (d) the concentration of the groundwater end member ( $C_G$ ) and (e) the surface water inflow rate which was estimated from runoff coefficient consideration ( $S$ ) and its  $^{222}\text{Rn}$  concentration ( $C_S$ )

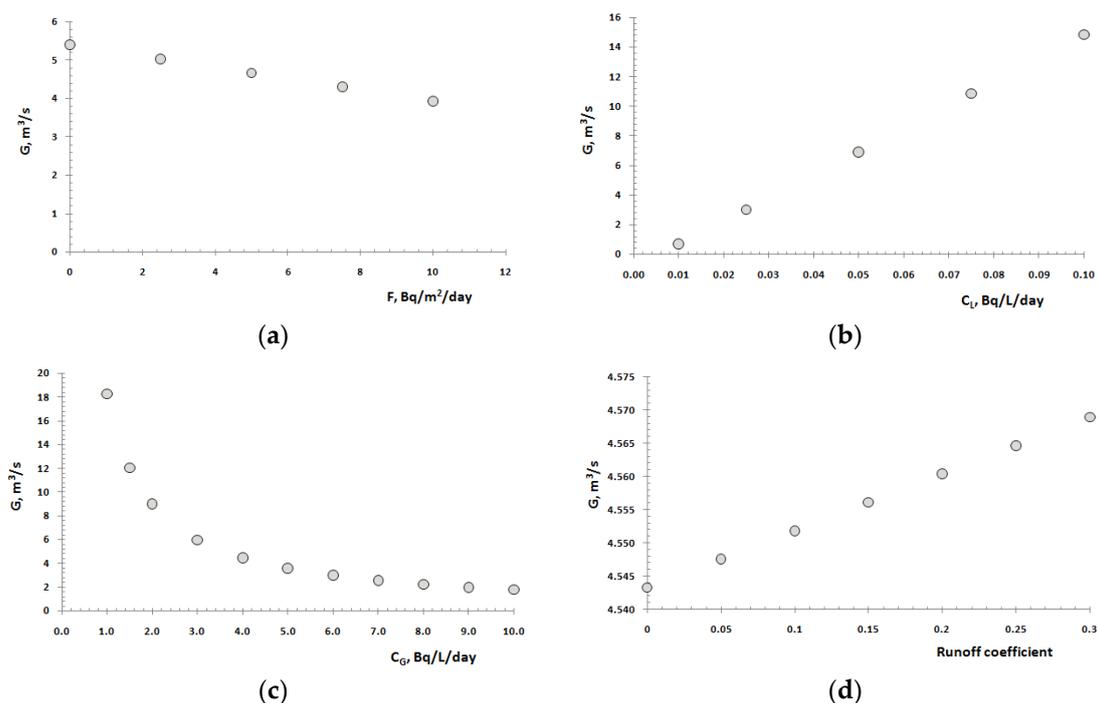
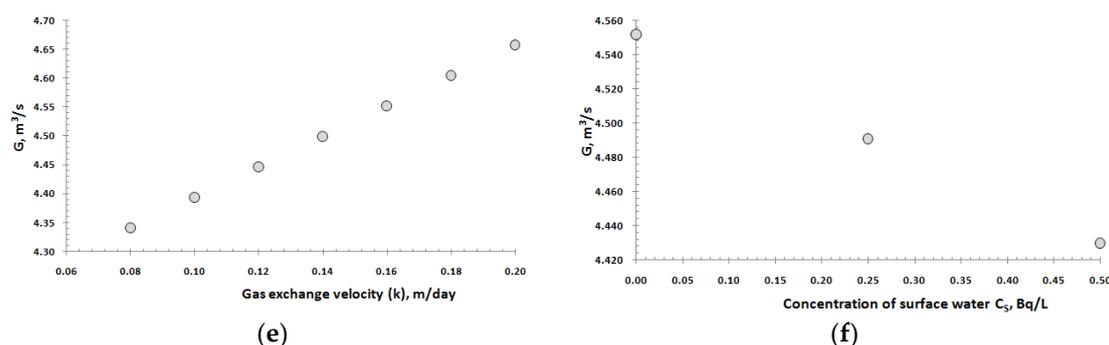


Figure 6. Cont.



**Figure 6.** Sensitivity run, ran individually by fixing other parameters as in Table 2. (a) sensitivity to diffusive radon flux, (b) sensitivity to radon concentration in Lake, (c) sensitivity to radon concentration in groundwater, (d) sensitivity to runoff coefficient, (e) sensitivity to gas exchange velocity, (f) sensitivity to radon concentration of surface waters.

The sensitivity run shows the model output is less sensitive to the diffusive  $^{222}\text{Rn}$  flux ( $F$ ), the gas exchange velocity ( $k$ ) and the runoff from the catchment ( $S$ ) and its  $^{222}\text{Rn}$  composition. The model is highly sensitive to the  $^{222}\text{Rn}$  concentration of the lake water ( $C_L$ ) and that of the groundwater inflow ( $C_G$ ). The fact that groundwater inflow to the lake comes through discrete fractures increase our confidence in the assumption that concentration of the inflowing water is similar to the concentration of the wells water, some better constraint on the composition of the lake water particularly (a) depth profiling of the  $^{222}\text{Rn}$  content and (b) area and or volume integrated averaging of the  $^{222}\text{Rn}$  content of the lake as well as detailed and dense surveying of the  $^{222}\text{Rn}$  content would improve the confidence on the model output.

#### 4.5. Mitigation and Water Resources Implication

Socio-economic and engineering problem caused by a simple hydrologic change creates a condition that is extremely complex to mitigate. This is because the mitigation measures that can be taken will have socio economic and environmental tradeoffs by themselves. Quantifying and modeling the unintended consequences of mitigation measures is not the objective of the current work. However, there are three choices to be taken to mitigate the problem associated with the lake level rise. The first is to improve irrigation efficiency to minimize the groundwater inflow through seepage loss from the farms, the second is to regulate lake outflow and the third is to intentionally divert more river water into the lake to dilute it and use it as a fresh water reservoir. Each of these measures has its own advantages and disadvantages as discussed below.

*Improving irrigation efficiency:* Improving irrigation efficiency in the farms would reduce the seepage loss from the farms to the groundwater. This will ultimately reduce the groundwater inflow to the lake. However, any reduction of groundwater inflow to the lake would result in a decrease in lake outflow and thereby an increase in salinity of the lake. Thus, the saline lake will remain a permanent threat. Given that it starts to flow out again because of other hydrological reasons, such as exceptional heavy rains or other geological reasons, the lake can cause salinization of the Awash River system downstream of the lake. Improving irrigation efficiency will have also its cost implication to the state farms.

*Regulating release of lake outflow:* As of the year 2010, the lake has been spilling to the Awash River. Since the Awash River has sufficient discharge and dilute chemistry during wet season, it can buffer the effect of the saline water outflow from the lake to the level that can be used for drinking and irrigation water use. However, during the dry season when the discharge of the river is low, the buffering capacity of the Awash River is low. As already observed, unregulated spilling of the lake into the River during the dry season causes high level of salinity and high level of other undesirable elements (such as fluoride) in the river. A regulated release of the lake water to Awash River, taking

into consideration the discharge and the chemistry of the Awash River, can be used as an immediate measure to mitigate the impact of the Lake on dry season river salinity.

*Intentionally diverting the Awash River into the Lake:* Disregarding the ecological changes that would come along with additional dilution of the lake water, diverting the dilute Awash River into the lake can cause additional reduction of salinity of the lake to the level it can be used as a fresh water reservoir. This requires detailed accounting of chemical and hydrological mass balance and engineering considerations. Given that the dilution of the lake Water is successfully conducted, the lake water can be used as a fresh water reservoir of substantial volume.

## 5. Conclusions

The few cases of  $^{222}\text{Rn}$  use in lake studies focus on methodological development as well as on quantifying groundwater inflow to lakes. This study shows a case where the combined  $^{222}\text{Rn}$  and  $\delta^{18}\text{O}$ - $\delta^2\text{H}$  study could yield an important information in a practical water management question. The current work shows how the approach could be successfully used in addressing practical field problems regardless of the uncertainties associated with estimating  $^{222}\text{Rn}$  balance model input parameters. The deficiencies encountered in using  $^{222}\text{Rn}$  in quantifying groundwater inflow to the lake have been filled by using the combined  $\delta^{18}\text{O}$ - $\delta^2\text{H}$  tracers. The capability of  $^{222}\text{Rn}$  in (a) detecting groundwater inflow zones and (b) quantifying groundwater inflow means that, if utilized along the  $\delta^{18}\text{O}$ - $\delta^2\text{H}$  lake balance studies, the uncertainties associated with individual approaches could be substantially reduced. In this particular work, we demonstrated the combined use of  $^{222}\text{Rn}$  and  $\delta^{18}\text{O}$ - $\delta^2\text{H}$  tracer for detecting source of groundwater to the lake as well as quantifying its flux. The deficiency in  $\delta^{18}\text{O}$ - $\delta^2\text{H}$  data and the non-steady state hydrology of the lake did not allow quantitative estimation of the groundwater inflow from  $\delta^{18}\text{O}$ - $\delta^2\text{H}$  application. Application of  $\delta^{18}\text{O}$ - $\delta^2\text{H}$  for estimating water flux around the lake could be subject for future research.

The dual  $^{222}\text{Rn}$  and  $\delta^{18}\text{O}$ - $\delta^2\text{H}$  isotope systematic effectively demonstrate the origin of water and causes of growth of Lake Beseka. It is concluded that the cause of the lake swelling is related to irrigation return water joining the lake via subsurface flow paths. The regular horizontal  $^{222}\text{Rn}$  pattern (decreasing from west to east and SW to NE) is an indication of a simple geologic setting where by discharge of groundwater to lake is taking place in restricted spots (in SW, S and W) ruling out the importance of lake deformation and the role of the ground fissures located in the northern sector of the lake. The ground fissures in the S, SW and W plays a role in enhancing the permeability of the rocks.

Although  $^{222}\text{Rn}$  is a proven tool in detecting sites of groundwater discharge into surface waters and lakes the type of that groundwater (shallow or deep) could not be identified from use of  $^{222}\text{Rn}$  alone [16]. The insufficiency of the  $^{222}\text{Rn}$  method in deciphering the source of the groundwater (deep or shallow) has been overcome in this study by using the  $\delta^{18}\text{O}$ - $\delta^2\text{H}$  tracer in pinpointing the type of groundwater (shallow or deep) feeding the lake. The two types of evidence show that the water that led to the changes in hydrology of the lake is the increase in discharge of shallow groundwater (though springs), which in turn is related to irrigation return water from the expanding irrigation fields. The dense fractures leading to high permeability and porosity of the basaltic aquifers underlying the western sector of the catchment has aided rapid transfer of the infiltrating water to reach the lake. The other option of running unsteady state  $\delta^{18}\text{O}$ - $\delta^2\text{H}$  balance model was not achieved because of paucity time series data over the lake's historical change in volume and isotopic composition.

The conclusions made about the origin of water do not contradict the conclusion made by [7] in that the increase in discharge of springs in the western and south western sector of the lake is responsible to increase the lake water gain. However, the current work pinpoints the origin of the water to be from irrigation return water entering the geothermal reservoir and returning to the lake as the main source of water. While the mechanism was clearly supported with evidence, [7] left an open question about where the actual water that leads to the increase in the discharge of the spring comes from. The isotopic evidence in this work reveals the source of the actual water is irrigation return water following the subsurface flow path. Because of the shallow geothermal gradient related

to the Fentale volcano [4], the shallow groundwaters can easily be heated, thereby imparting high temperature to the seepage waters leading to the temperature profile observed by [7].

The water balance of the lake computed using  $^{222}\text{Rn}$  can represent the annual average values. This is the water balance of the lake that is dominated by groundwater flow, and the seasonality in rainfall and surface flows is small. However, improvement in the computed groundwater flux could be made if monthly time series of  $^{222}\text{Rn}$  has been obtained from the lake and the groundwaters.

A more recent view of irrigation efficiency suggests the irrigation loss to seepage should not be counted as a loss as far as that water returns to the river system and service downstream users [34]. In the case of Lake Beseka, the irrigation loss has led to environmental, engineering and socio-economic challenges, necessitating a more unified definition of irrigation efficiency accounting in not only water quantity but also quality.

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