

Human impact assessment through a transient numerical modeling on the UNESCO World Heritage Site, Lake Naivasha, Kenya

Yohannes Yihdego^{1,3} · Gebrehiwet Reta^{2,3} · Robert Becht³

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Abstract Lake Naivasha was designated as a RAMSAR site in 1995. The lake environment is fragile and critically threatened by human-induced factors. The study presented a steady and transient numerical modeling. The long-term and system flux over time interaction between the lake and the surficial aquifer is represented in the Lake Package LAK3 with in the advanced 3-D simulation software (GMS). The model covers an area of 1817 km². Model calibration was constrained by the observed groundwater and lake levels using PEST. The effect of excessive abstraction was rigorously analyzed via scenario analysis. The simulation was evaluated “with abstraction” and “without abstraction” scenarios. The abstraction scenario was simulated using range of combination assuming that all the abstraction was from the lake or the groundwater and in the ratio of groundwater and lake water. The effect of the stress was evaluated based on the observed aquifer heads and lake stage at the end of the simulation time. The development of low groundwater-level anomalies in the well field is explained. The result indicates that the one of the well fields is not in direct hydraulic connection to the main recharging water body. Apparently, similar development of cone of depression was not generated in the other

two well fields, and this could have several reasons including due to the fact that these well fields are located relatively close by to the main recharging zones and concluded to have additional source of recharge, and this was supported by previous studies, whereby the isotopic composition of the boreholes has their source of recharge from precipitation and river and was also confirmed from the isotopic composition of unsaturated zone, which is a mixture of river and rain. The study reveals that seasonal variability of groundwater–surface water exchange fluxes and its spatially and temporally variable impact substantially on the water resource availability. Such analysis can be used as a basis to quantify the linkages between the surface water and groundwater regime and impacts in the basin. The model output is expected to serve as a basis via linking/coupling with others to incorporate the ecology and biodiversity of the lake to safeguard this high-value world heritage water feature.

Keywords Surface water–groundwater interaction · Water resources · Wetland · Abstraction · Irrigation · Impact assessment · RAMSAR

Introduction

Lake Naivasha is the only freshwater resources among many saline lakes in the Kenyan rift valley and was designated as a RAMSAR site in 1995 and has been considered as a highly significant national fresh water resource in Kenya (Fig. 1). Rapid industrial development and increase in agricultural production have led to freshwater shortages in many parts of the world, including Lake Naivasha and increased withdrawals from Lake Naivasha and the surrounding aquifers have the potential to affect water levels

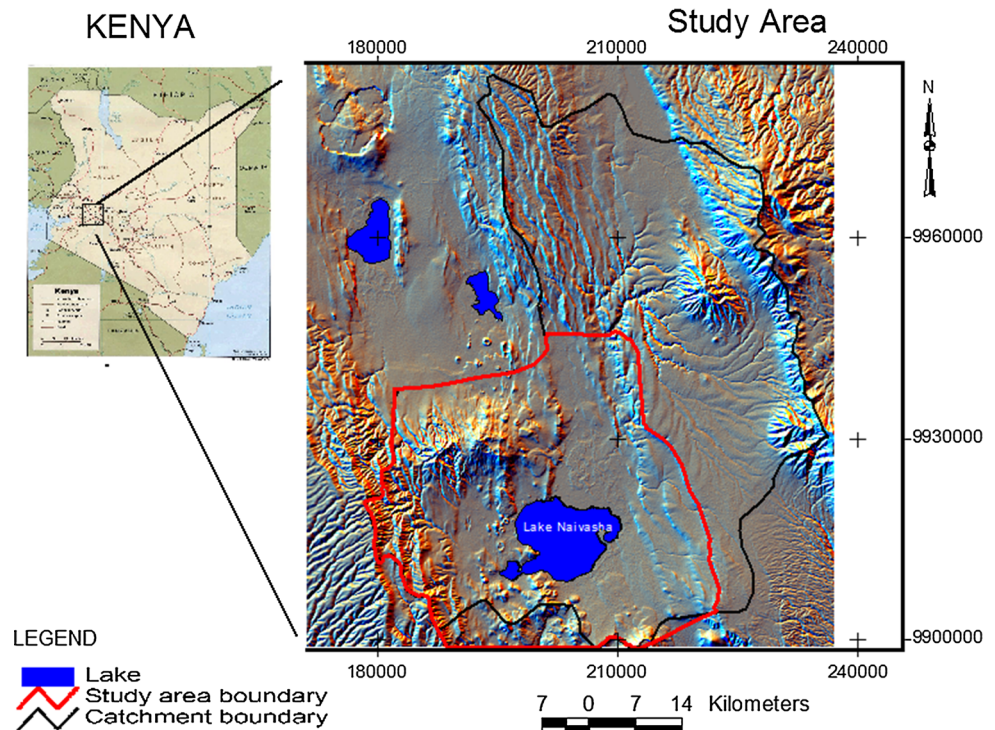
✉ Yohannes Yihdego
yohannesyihdego@gmail.com

¹ Snowy Mountains Engineering Corporation (SMEC),
Sydney, NSW 2060, Australia

² College of Hydraulic & Environmental Engineering, China
Three Gorges University, Yichang city 443002,
Hubei Province, China

³ Faculty of Geo-Information Science and Earth Observation,
University of Twente, P.O. Box 217, 7500 AE Enschede,
The Netherlands

Fig. 1 Location map of the study area (after Yihdego et al. 2016)



in these aquifers (Becht et al. 2005; Becht and Nyaoro 2006; Yihdego and Becht 2013). Greater emphasis is paid for optimal utilization of water resources due to increasing demand for agricultural, industrial and domestic consumptions (Becht and Harper 2002; Becht and Nyaoro 2006; Odongo et al. 2014, 2015; UNESCO 1999).

The study area is signaling water-level decline and water quality deterioration due to peak demand of lake and groundwater water usage. Sustainable development of water resources needs quantitative estimation of the available water resources (El-Kadi et al. 2014). Water balance studies have extensively been implemented to make quantitative estimates of water resources. Previous studies elsewhere (e.g., Alcolea et al. 2008; Ayenew and Becht 2008; Gupta et al. 2012; Simmons and Hunt 2012; Yustres et al. 2013) have addressed the extent to which model's parameters should be adjusted in order to allow it to replicate past system behavior as a precursor to its being used in management of future system behavior and how complex (or otherwise) needs to be when used in this capacity. In this context, a model's purpose is to predict the behavior of a system under a management regime. Selection of an appropriate level of model complexity is most difficult where predictions required of a model are only partially constrained by historical data (Doherty and Simmons 2013; Turner et al. 2015; Yihdego et al. 2015; Yihdego and Al-Weshah 2016a, b, c; Al-Weshah and Yihdego 2016; Yihdego and Drury 2016).

To understand the hydrogeological behavior of the rift lakes, it is essential to gain good conceptual view of the

geological and paleohydrologic processes (Hogeboom et al. 2015; Al-Weshah and Yihdego 2016; Turner et al. 2015; Yihdego and Webb 2012; Yihdego et al. 2014; Yihdego and Paffard 2016; Yihdego and Webb 2015, 2016). Lake Naivasha (Fig. 1) attracted researchers' interest in different aspects of the lake (e.g., Odongo et al. 2016; Becht and Harper 2002; van Oel et al. 2013; Yihdego and Becht 2013; Yihdego et al. 2016). Yihdego and Becht (2013) improved the understanding of the hydrogeology of the area by integrating the information on the groundwater flow, hydrochemistry and boundary through a groundwater model.

The aim of this paper is to explore the relation between lake water and nearby aquifers and assess the potential effect of increased abstraction rate in the well fields on Lake Naivasha water through three-dimensional transient numerical model. The data presented in this paper and interpretations have been mainly sourced from Yihdego (2005) and Reta (2011) thesis.

Site description

The study area is located 100 km northwest of Nairobi, Kenya, and covers an area of 3500 km². Lake Naivasha has a mean surface area of 145 km² at an average altitude of 1887.3 m amsl (Yihdego and Becht 2013) and has an average depth of 5 m. The major streams that drain the study area are the Malewa River and the Gilgil River.

Geological and hydrogeological setting

Volcanic and sedimentary rocks dominate the study area shown with varying structural features. Depth of groundwater in the sediment aquifer ranges from 3 to 35 m. The sediment aquifer is predominantly unconfined and has relatively high specific yield, and the wells near the Lake Naivasha shore yield water from lacustrine-deposit aquifers and usually have higher specific discharge and transmissivities (Ojiambo 1992; Becht and Nyaoro 2006; Reta 2011). The structures of the rift valley are believed to play a key role on the groundwater flow systems of the area as shown with high hydraulic gradients across the rift escarpments (Fig. 1), indicating the effect of the major faults as zones of low permeability.

Methodology

The data analysis involved characterization of surface and subsurface parameter, including with a field work Geodetic-GPS survey program in order to accurately measure the height of the groundwater level and surface water levels. Time-series data including lake level, surface water inflow, evaporation and precipitation were analyzed on a monthly basis. Pump test data were analyzed for recently drilled boreholes. Recharge was estimated by relating monthly change in groundwater level and average recharge measured in the area. Water abstraction data mainly from the irrigated commercial farms were analyzed based on the irrigation area–depth relationship. A conceptual model was prepared and served in the three-dimensional finite-difference model setup. Impacts were assessed through scenario analysis.

Hydraulic properties

Sediment aquifers usually have relatively high permeability and are often unconfined with high specific yield. Transmissivity value estimate varies from 0.1 to 5000 m²/day. The corresponding hydraulic conductivity calculated from transmissivity values ranges from 14 to 750 m/day. The estimated hydraulic conductivities of the volcanic aquifer range from 0.1 to 1.1 m/day. Specific yield estimated 0.12–0.0001 for the unconfined/confined aquifer and 0.00146–0.00395 for the lake sediments.

Recharge and discharge condition

Direct recharge and indirect recharge from water bodies and lateral inflow from adjacent aquifers are the mechanisms of groundwater recharge. The area experiences an annual rainfall and evapotranspiration of 640–1700 mm,

respectively. Direct recharge is not a permanent process in the area since the potential evapotranspiration of the area was estimated to be higher than rainfall, except during rain seasons with high intensity.

Recharge estimates using direct recharge (Van Dam et al. 2008) and isotope mixing (Gat 2010) techniques are listed in Tables 1 and 2, whereby the techniques based on surface water and unsaturated zone data provide estimates of potential recharge, whereas those based on groundwater data generally provide estimates of actual/net recharge.

Estimated groundwater recharge from a basin is (Eq. 1):

$$R = A \times \sum_{i=1}^n \Delta hi \times Sy \tag{1}$$

where R = total groundwater recharge (m³/day), A = area (m²), Δhi = seasonal increase in groundwater level (m), Sy = specific yield of water yielding materials.

The study was taken as the steady-state average recharge for the basin. The seasonal increase in groundwater level is assumed to be directly proportional to the lake-level fluctuation and can be replaced by monthly lake-level increase. The specific yield (Eq. 2) was estimated from the relationship in Eq. 1.

$$Sy = \frac{avR}{\sum_{i=1}^n \Delta hi} \tag{2}$$

where Sy = “relating factor”, avR = average recharge in (m) of the area, Δhi = seasonal increase in lake level.

The transient recharge (Fig. 2) was calculated using the relation of Eqs. 1 and 2.

$$Ri = A \times \Delta hi \times Sy \tag{3}$$

where Sy = “relating factor” constant, Δhi = seasonal increase in lake level at time $t = i$ indicates the stress period $i = 1-942$, Ri = groundwater recharge at time $t = i$.

Assessment of the irrigation return flow was carried out based on the abstraction–return flow relationship (Eq. 4).

$$\text{Return flow} = C \times \text{abstraction} \tag{4}$$

where, C is the return flow coefficient ($0 < C < 1$).

The irrigation return flow depends on the soil characteristics, hydraulic properties of the aquifer, depth to water table, irrigation practice and type of crop. Based on the artificial recharge test, the contribution of irrigation return flow in the long-term water balance of the basin is assumed to be insignificant and will not be considered in this study.

Water abstraction in the study area occurs from irrigation and domestic wells. Abstraction for irrigation use is much larger than for domestic use. Water abstraction data mainly from the irrigated commercial farms were collected along with the irrigation techniques and the area of the

Table 1 Direct recharge estimated from SWAP model

Local name	Location		Recharge (mm/day)		Average recharge (mm/year)
	UTM_X	UTM_Y	Min	Max	
Kedong	209,691	9,908,544	0	7	43.75
Ndabibi	194,490	9,914,863	0	0.27	0.69
TPF	213,403	9,924,948	0	0.1	4.38
Marula	208,444	9,930,840	0	5.5	33.75

Table 2 Direct recharge estimates from 1-D mixing model

Type of recharge	Stable isotope	mm/day	mm/year	Average (mm/year)
Mixed recharge; lake: rain	18-Oxygen	1×10^{-5} – 7×10^{-5}	3.65–25.55	14.6
	Deuterium	3×10^{-5} – 7×10^{-5}	10.95–32.85	21.9

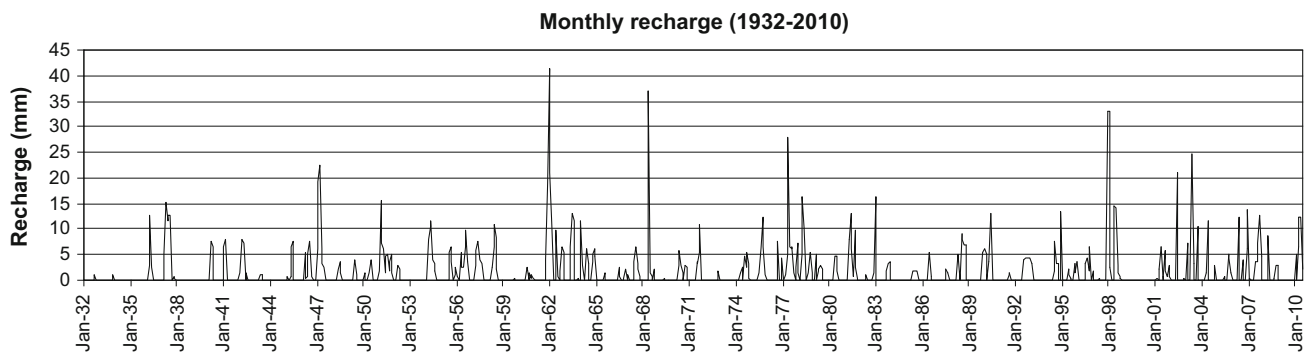
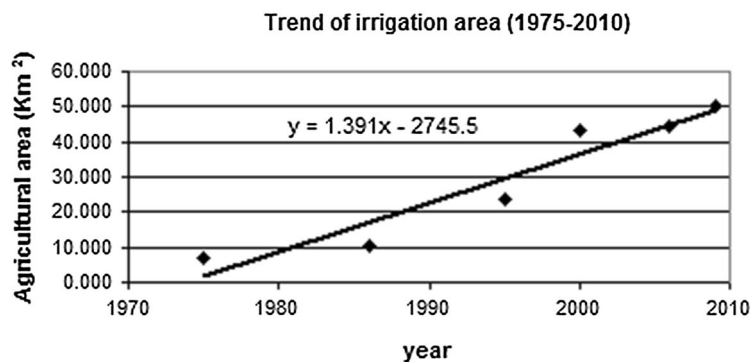


Fig. 2 Monthly recharge



Year	Agricultural area (km ²)
1975	7.147
1986	10.336
1995	23.383
2000	43.146
2006	44.672
2009	50.23

Fig. 3 Trend of irrigation areas (1975–2010)

farms (Fig. 3). Quantification of abstraction from different water bodies such as surface water and groundwater was made based on the abstraction record data sheet obtained from the Naivasha District officers (Fig. 4). Estimation of historical development of irrigation areas in the basin was made by processing time-series satellite images. Estimation of irrigation depth was estimated 1000 mm/year, based on previous estimations (Yihdego 2005).

Monthly averaged potential evaporation on the floor of the basin exceeds rainfall by a factor of 2–8 for every month except April when the potential evaporation still exceeds rainfall save for the wettest years (Yihdego 2005). The natural vegetation surrounding the lake is mainly papyrus swamp vegetation. Natural vegetation outside of the lake surroundings is shrub, cactus, savannah and acacia. Acacia trees can be attributed to the shallower groundwater

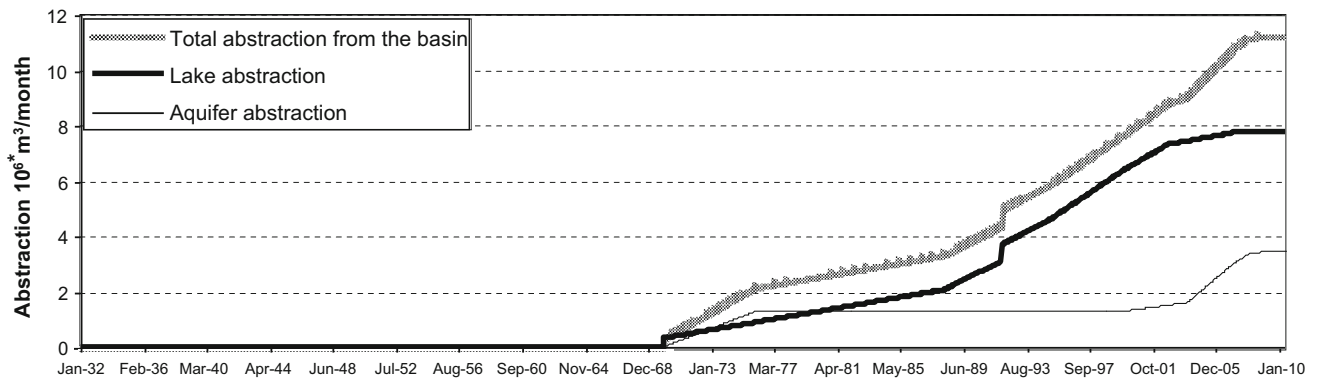


Fig. 4 Analysis result of abstraction from different water bodies

table. The depth of root zone in the study area is assumed to be 3–12 m.

Model development

The existing well database was updated and reorganized. Available data were screened and preprocessed, field survey points mapped out, mapping units delineated and appropriate field materials and tools identified. The general topographical and hydrologic background was described from available records, maps, aerial photographs and satellite imagery. The Landsat images band composite 543 was processed using ILWIS software. The study area boundary includes the simulation of subsurface hydrogeological boundaries and influence of major structures in addition to the catchment boundary.

A Geodetic-GPS survey program was made during the field work to accurately measure the groundwater level and surface level at each well, river stage and lake level. Hydrogeological observations were taken with the help of ASTER satellite images, geological maps and cross sections and previous studies of the area.

The water level of Lake Naivasha has been monitored since 1908. Lake-level series was recorded from 1932 to 2010. In modeling of lake Naivasha–aquifer interaction, a key step forward was the incorporation of the lake bottom bathymetry into the digital surface model in order to the map the aquifer out crops on the bottom of lakes and to simulate the aquifer–lake interaction in detail (Yihdego and Becht 2013).

Model setup

The conceptual model approach (Yihdego 2005; Yihdego and Becht 2013), which is the most efficient approach for building realistic, complex models, was used for the numerical model simulation. The lake–groundwater

interaction was three-dimensionally modeled using GMS-MODFLOW 2000 (Harbaugh et al. 2000). GMS incorporates the lake package LAK3, which was utilized to estimate the water budget of the lake Naivasha.

The modeled domain covers an area of 1817 km² with two non-conceding aquifers. Layer one is unconfined with a thickness that varies from few meters to 100 m. Layer two is semi-confined aquifer with thickness varies from 30 to 220 m. The first layer contains the lake. For simplification purpose, the two layers are simulated as a horizontal continuous layer with an average thickness 60 m for the first layer and 100 m for the second layer. In order to reflect the non-conceding nature of the layers, a low hydraulic conductivity was assigned to the boundaries of the first layer. A contrast of two orders of magnitude in hydraulic conductivity between the aquifer and a non-permeable unit at the boundaries results in ignoring the horizontal flow in the layer. To simulate the semi-confined nature of the second aquifer, a quasi-three-dimensional modeling approach was assumed (Yihdego and Becht 2013). The model grid contains 104 rows, 120 columns and two layers. The horizontal spacing is uniformly equal to 500 m. A total number of 24,960 cells were designed as 14,562 active cells and 10,398 in active cells.

Boundary conditions

The western part of the watershed boundary that peaks at the Mau scarp was taken to be a no-flow boundary. The northwestern boundary spanned by the Eburru hills beneath which are acknowledged to be outflow to the Elmenteita lake basin (Yihdego and Becht 2013). In order to show the conceptual implication of the outflow through this boundary, the boundary condition is located behind the Eburru hills. This hydraulic boundary was simulated by General Head boundary with head elevation specified at 1830 m. The eastern boundary is specified as no-flow boundary, because the fault is considered to impede most of the

inflow from the Kinangop Plateau. Most of the flux is considered to take place in the deeper horizons. Minimal in flow through this area has been considered negligible. The southern part of the model boundary was specified as General Head boundaries ranging from 1800 to 1850 m, being flow through the Olkaria and Longonot volcanic complexes has been considered to be the conduit for most of the lake outflow from the basin. The bottom of the system has been considered to be composed of undifferentiated volcanic materials that have a very low.

The lake surface at the center of the domain is a time-variant boundary whose boundary has been defined within the lake package. The maximum depth of the lake is 30 m, located at the Crescent land, with an average depth of 6 m. The two main rivers, Malewa and Gilgil, are time-variant boundaries whose boundaries were defined within the river package.

Representation of the lake

The current coverage of the lake is approximately 110–120 km². However, the initial historical lake surface area was about 181 km² (from historical data at the year 1932). The lake surface area was set to the initial coverage as surveyed in 1957 by the Ministry of water Works (Kenya). The initial lake bathymetry was represented by the lake package triangular networks (Yihdego and Becht 2013). The lake was assumed to have an initial stage of 1891 (stage at the initial condition of the year 1932). The lake represented with a minimum elevation of 1874 m (at the crescent lake) and maximum of 1896 m (maximum stage that can be achieved during the simulation).

MODFLOW lake package (LAK3)

The lake–aquifer interaction was simulated in both transient- and steady-state flow conditions through lake package. The interaction between the lake and the surficial aquifer was represented by updating at the end of each time step a water budget for the lake that is independent of the groundwater budget represented by the solution for heads in the aquifer including the computation of lake volume and stage. The lake stage was crucial in making the estimates of groundwater seepage to and from the lake.

Steady state

Two types of observation data were used in the calibration process: water table elevations from observation wells and observed lake level in the year 1932. The initial conditions have been considered to be the hydrologic stresses (lake levels, river flows, evaporation and precipitation) at the 1932 period. The initial hydrologic stresses (lake levels,

river flows, evaporation and precipitation) at the 1932 period were used for the steady-state calibration because levels within this duration correspond to the natural stresses that were acting in the system then and lack of enough data to adequately describe the piezometric surface at the start of the simulation period.

Model parameterization

Most of the aquifer parameters collected from pump test result that were basically derived from the geologic map and pump test result of the area were in transmissivity values; however, the model requires hydraulic conductivity values. These values were used to map the initial estimates for the zones. In the estimate for recharge, an initial value of 20 mm/year equivalent to about 4% of average precipitation was assumed for the lake sediments. The western scarp area of Mau stretching north to the Eburru area was considered to have a higher recharge. Previous work estimates recharge at 50 mm/year equivalent to 7% of the average precipitation. It is assumed that the Pleistocene pyroclastic that flank escarpments appears relatively absorptive and transmits infiltrating precipitation and runoff to the underlying fracture and fissure systems of less absorptive and permeable rocks.

Parameter zonation was made for recharge and hydraulic conductivity (layer 1: Fig. 5), based on the 3-D conceptual model developed by Yihdego (2005) and Yihdego and Becht (2013). The vertical hydraulic conductivity of the formations is assumed 10% of the horizontal hydraulic conductivity.

Steady-state calibration

The steady-state calibration was accomplished by minimizing the difference between model-predicted steady-state aquifer water levels and measured groundwater level and lake stage during the time of simulation. PEST (Doherty 2004; Doherty et al. 2010) was used to calibrate the model. The model parameters of aquifer hydraulic conductivity and riverbed conductance and lake leakage were estimated during steady-state calibration. The steady-state calibration was accomplished using 41 observations and three adjustable parameters: hydraulic parameters layer 1 and layer 2, recharge layer 1, river bed conductance and lake leakage.

Steady-state simulation results indicate a wide range in horizontal hydraulic conductivities from 0.1 to 75 m/day for the lacustrine sediments and 0.001–0.1 m/day for the volcanic aquifer. The steady-state model was calibrated to the initial states of the lake (lake stage = 1891 m); this means that the initial aquifer water level around the lake had to be higher in order to have the same head as the lake

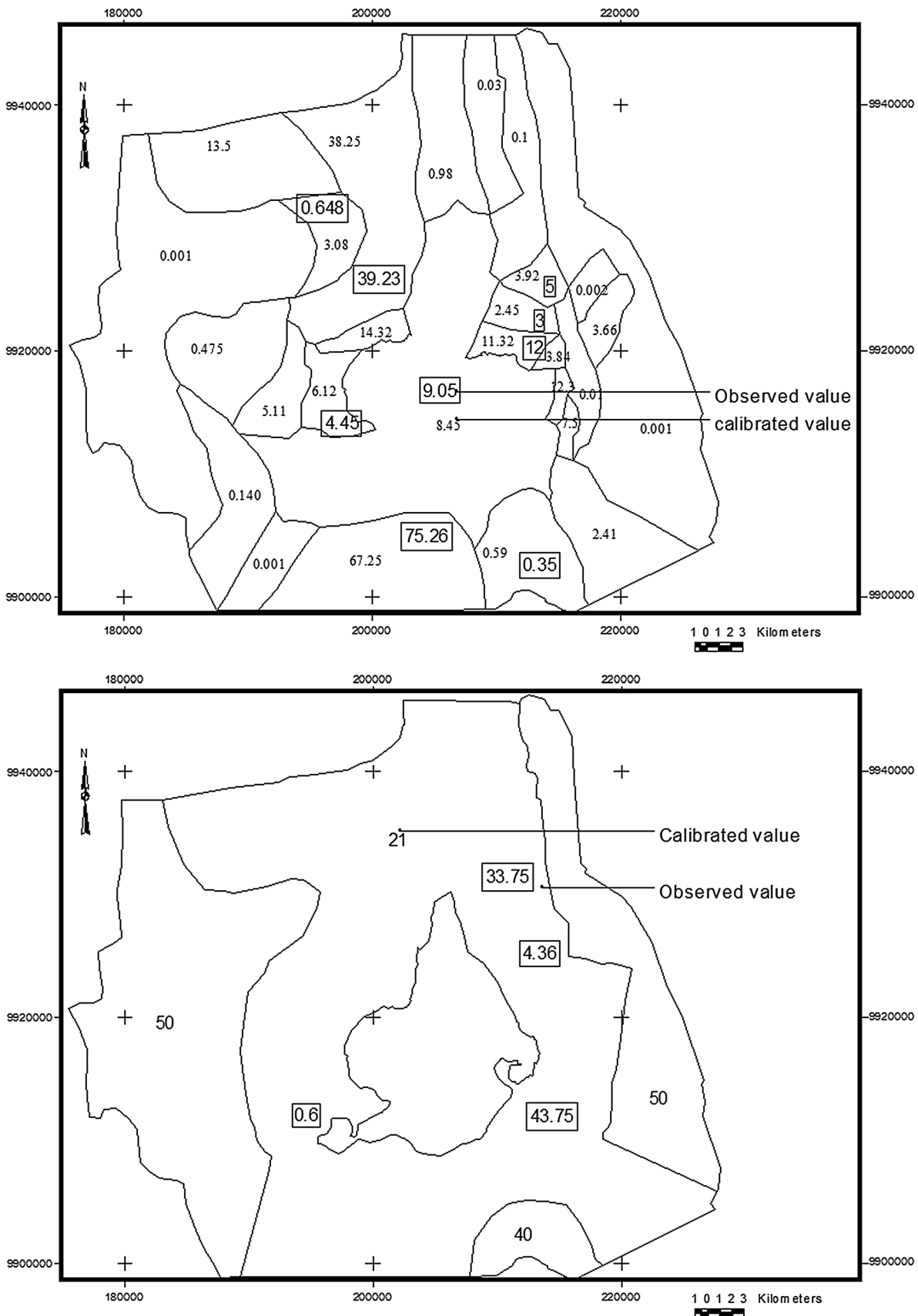


Fig. 5 Parameter zones for hydraulic conductivity and recharge (layer 1)

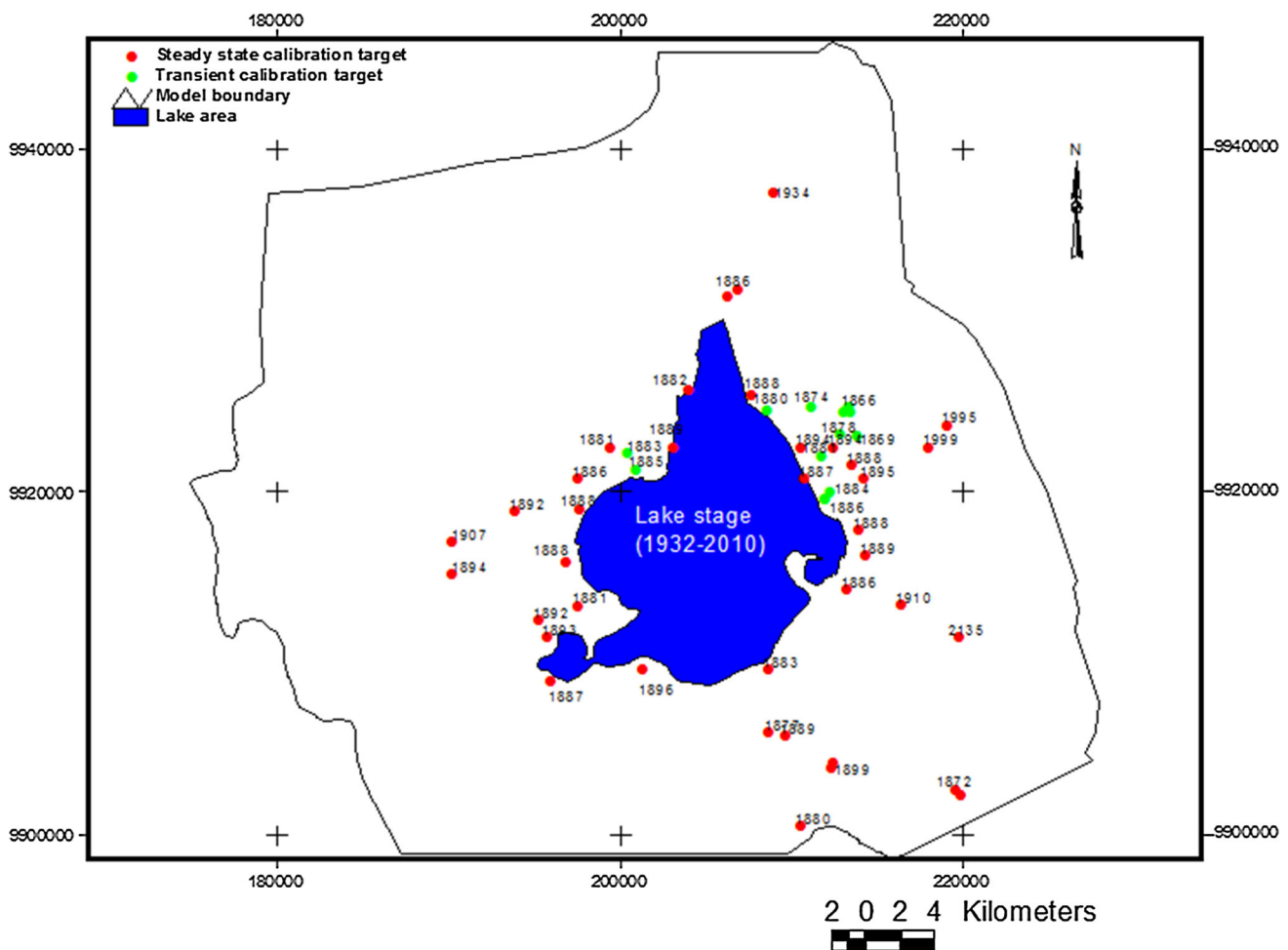


Fig. 6 Location map of calibration wells

itself. But later during the transient simulation, when the lake stage starts to decrease, the hydraulic head in the surrounding aquifer was not immediately respond to the stress on the lake. In other words, change in the lake flux does not bring the some effect on the lake shore aquifers. This implies that the preliminary steady-state hydraulic conductivity at the immediate vicinity of the lake had to be scaled up in order to allow lower hydraulic head equivalent to the lake stage during the transient simulation.

Transient model

The initial condition for the transient simulation was the steady-state head solution generated by the calibrated steady state. Use of model-generated head values ensures that the initial head data and the model hydrologic inputs and parameters are consistent.

The model was designed with a total of 942 stress periods to match the data available from 1932 to 2010 spanning over 79 years that included monthly lake levels, stream flow, and evaporation and precipitation

data over the whole duration. The transient recharge was calculated based on the previous study (Yihdego 2005) to match the existing data in transient simulation. Transient water abstraction was estimated based on the irrigation area–depth relationship. A total of 14 water-level data and the lake stage were assumed to constraint the model solution. The well location is shown in Fig. 6.

Transient calibration

The steady-state model parameters, boundary conditions, aquifer hydraulic conductivities, riverbed conductance and lake leakance were kept constant during the transient-state calibration. The steady groundwater outflow and inflow which are indication of the groundwater–lake interaction were calculated as a function of the head difference between the lake and the aquifer. The storage parameters, specific storage and specific yields were adjusted by try and error in each calibration attempt. The transient model required a warm-up period of about five stress periods

because observations during the initial stress period (1932) are partly dependent on events that occurred years prior to 1932. By using the ending steady-state heads as the transient starting heads, the transient state was run as a steady state for five stress periods.

The transient simulation required initial estimates for the storage coefficients of the aquifer. In the modeling conceptualization above, the two layers were simulated as convertible layer, implying that the layer can be either confined or unconfined depending on the elevation of the computed water table. A calibrated values of specific yield (0.01–0.1) for the lacustrine sediments and (0.01–0.001) for the volcanic aquifer were used for the model simulation. The corresponding specific storage values were in the range of 0.001–0.0001 adjusted during calibration. The storage coefficient values in this study were generally lumped into two main zones to the lake sediments/alluvial materials around the vicinity of the lake and the reworked volcanic materials.

Results and discussion

Steady-state model result

The result of the calibration was evaluated both qualitatively and quantitatively. Model statistics for the steady-state calibration indicated an overall $R^2 = 0.985$ between measured and modeled aquifer water levels. The standard error for the aquifer water level estimates was 0.738 m, indicating that about 95 percent of the modeled aquifer water levels are within about -0.2 and 2.6 m of observed values. The R^2 value was computed as 0.98 using 42 observation wells, with the mean error (1.24 m), mean absolute error (4.74 m) and the root-mean-square error (5.73 m^2). The scatter plot was examined and the plot showed deviations from the straight line are merely a random distribution (i.e. without any sign of systematic deviation). The result is in agreement with Yihdego and Becht (2013), using “high hydraulic conductivity” method for the lake–aquifer interaction.

Steady-state model contour map of simulated heads

The contour map of simulated aquifer heads is shown in Fig. 7. Both layers have a similar spatial piezometric configuration. The piezometric heads around the vicinity of the lake mapped exactly as the observed values. Away from the lake horizons, the aquifer hydraulic properties have been affected by different factors like heterogeneity and lithostratigraphy effects. The head-dependent boundary specified at the north, south and southeastern was calibrated as 1830, 1800 and 1850 m, respectively.

Transient-state model result

The transient model simulation results were evaluated by observing the measured and calculated aquifer and lake level (Figs. 8, 9). Summary of the transient calibration is given in Table 3.

Sensitivity analysis

Groundwater recharge and hydraulic conductivity and the storage parameter were each varied separately. In this procedure, the calibrated hydraulic conductivity and recharge values and aquifer specific yield were increased and decreased by a magnitude equivalent to 20, 40 and 60% of the calibrated values. The analysis result is presented in Fig. 10. The steady-state model is sensitive to both recharge and hydraulic conductivity. The model is highly sensitive to increasing and decreasing of recharge. The model also shows strong sensitivity to a decreasing than increasing in hydraulic conductivity. On the other hand, the transient model shows equal sensitivity to increasing and decreasing of the specific yield but with a slow response.

For the some reason, hydraulic conductivities along major outflow directions were also increased by 15–50% of the steady-state calibration values. The adjustment does not bring a significant effect on the expected model solutions. Considering the sensitivity hydraulic conductivity in the model (Fig. 10), the result indicates that the model is less sensitivity to an increasing in hydraulic conductivity. This could provide insight for future data collection and model improvements (Hogeboom et al. 2015).

Simulation of lake–aquifer abstraction from the basin

The amount of abstraction was initially estimated by considering the relation between abstraction and area of irrigation lands. Area of irrigation was calculated by calculating time-series Landsat images, and depth of irrigation was estimated using different statistical analysis. Simulation of abstraction in this study aimed at estimation of combined abstraction (abstraction from the aquifer and withdrawal from the lake). The simulation was evaluated “with abstraction” and “without abstraction” scenarios. The effect of the stress was evaluated based on the observed aquifer heads and lake stage at the end of the simulation time.

Scenario one: without abstraction

Agricultural activities in the basin before 1980 were negligible. Owing to this reason, all water loses from the lake during this period save for loses due to evapotranspiration and groundwater outflow. This scenario assumes that the

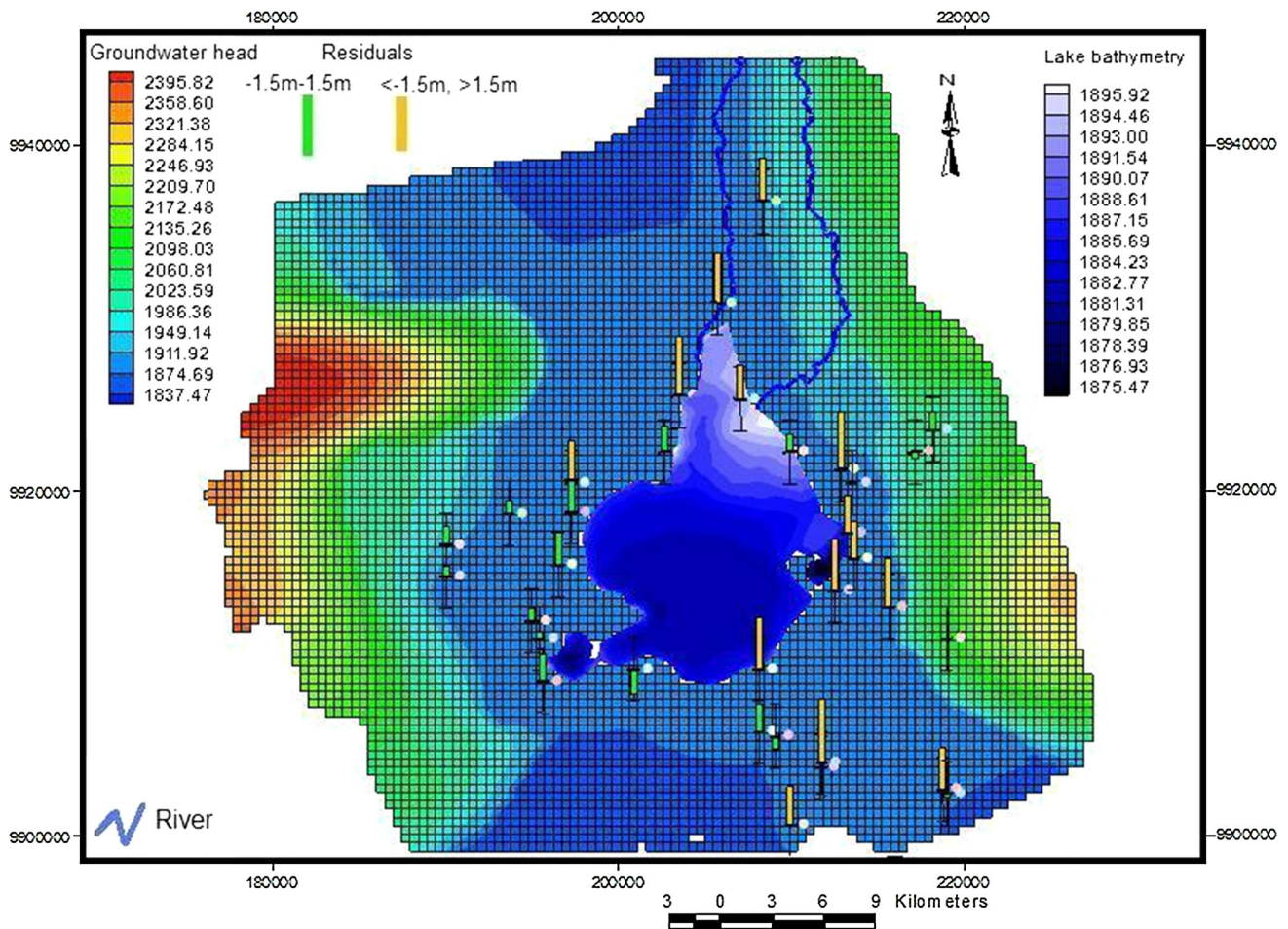


Fig. 7 Contour map of simulated aquifer heads

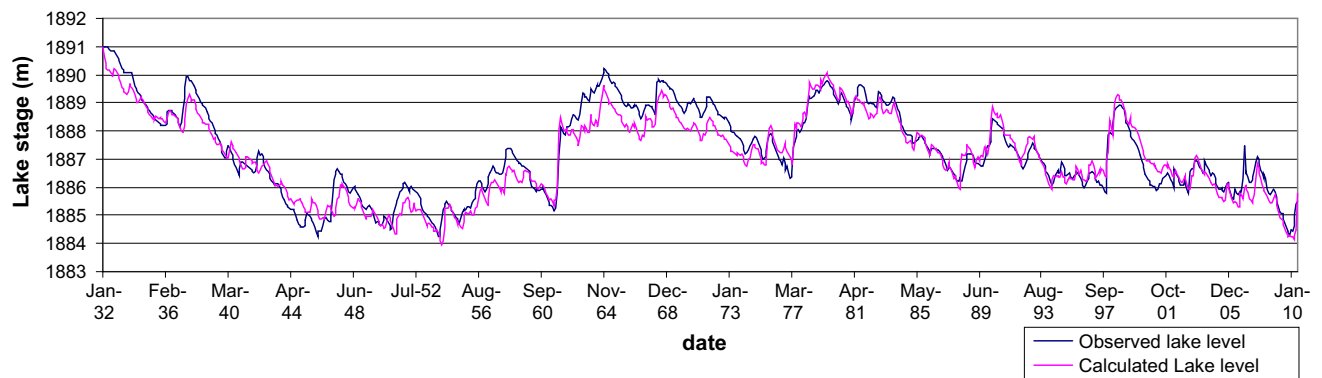


Fig. 8 Time-series plot of observed and calculated lake levels (1932–2010)

there was no abstraction at all before or after 1980. The model was made to run from 1932 to 2010, and the result was evaluated.

The lake-stage graph result shows a divergence around the 1980 (Fig. 12). The effect is also apparent in the aquifer water levels where the calculated heads are

higher (Fig. 11). The difference between observed and calculated aquifer and lake level after 1980 (Fig. 12) was taken as important indication of the effect of abstractions from the basin for agricultural, industrial and domestic purposes. Figure 13 shows the simulated contour heads spatially.



Fig. 9 Observed and calculated aquifer head (observation, June 2010)

Table 3 Summary of transient calibration error

Evaluation criteria	Lake level	Aquifer heads
Mean error	-0.161	-0.932
Mean absolute error	0.387	1.54
Root-mean-square error	0.478	2.449
R^2	0.905	0.732

Scenario two: with abstraction

The abstraction scenario was simulated using range of combination, assuming that all the abstraction was from the lake or the groundwater and in the ratio of 30% (groundwater) and 70% (lake water).

Lake abstraction from the lake was implemented by applying the abstraction stress directly on the lake. Well abstraction in the area is everywhere. However, not all the single wells have measurable stress on the aquifer of the area. The main well field is found on the northeastern part of the lake. The site is highly affected by groundwater abstraction since the history of agriculture activities in the basin.

In this scenario, the total stress of abstraction was applied to the aquifer. Implementation of the scenarios is made on seven wells. The wells are indicated on the ground by the main pivot wells found in three well fields as shown in Fig. 15, with a depth up to 120 m. Abstraction stress applied to all wells was the same, the total abstraction divided by the seven wells.

Well field A includes Three Point, Manera and Delamer sites, and the well field is the most exploited site in the area, represented by four wells. Well field B includes all wells around Marula sites, and the well field is represented by two wells. Well field C includes all wells very near to the lake and represented by one well (assuming that there could be an option to pump water directly from the lake).

Response of the lake stage to the different abstraction schema is presented in Fig. 14. The deviation of simulated from observed water levels in the past three decades is distinct and indicates the magnitude of industrial abstraction. The deviation between the different curves, (as a result of the different abstraction stress in the area), indicates the degree of interaction between the surface water bodies and the groundwater.

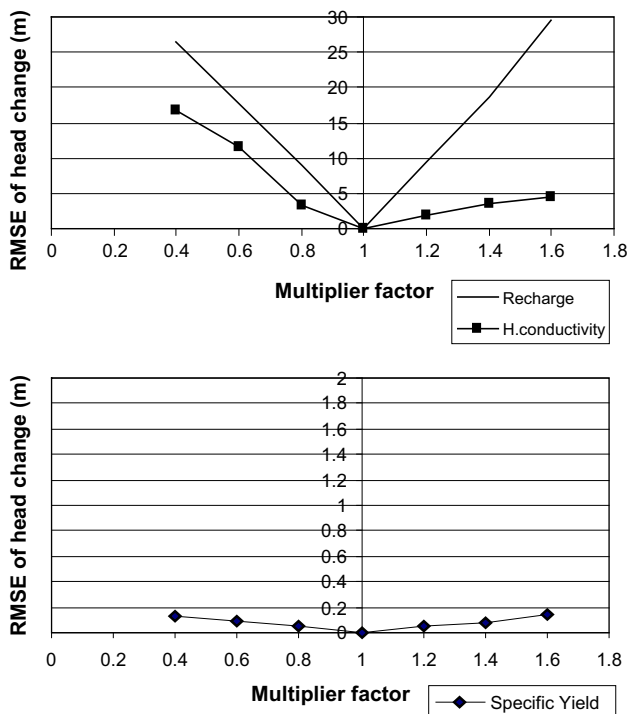


Fig. 10 Sensitivity analysis

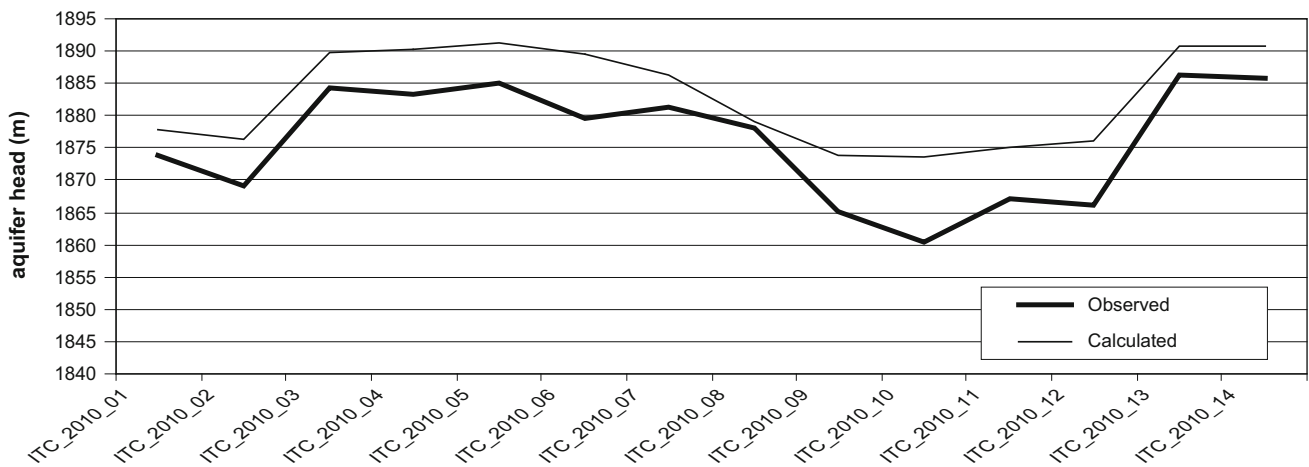


Fig. 11 Observed and calculated aquifer head (abstraction not included)

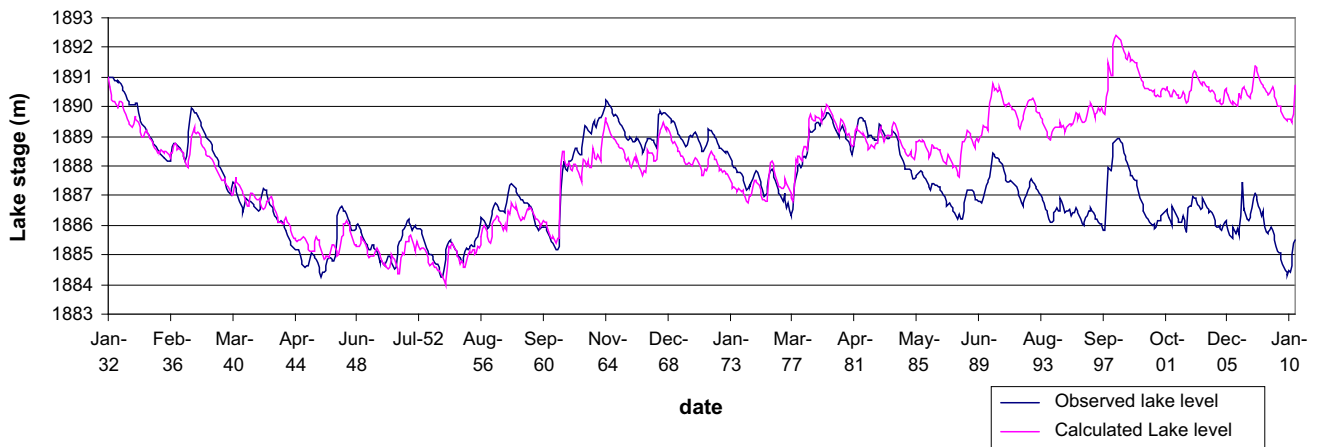


Fig. 12 Time-series plot of observed and calculated lake level (abstraction not included)

Abstraction estimation was needed to optimize in order to match the observed values. Optimization was made by adjusting the initial abstraction input values until the simulated aquifer and lake stage match the observed values. The result is in agreement with the estimates calculated from abstraction analysis in this study (Yihdego et al. 2016). The little increase in the current estimate supports the current condition that the groundwater level at the well field and the lake level are currently in the state of decreasing while abstraction is continuously increasing.

The final simulation contour map of aquifer heads and lake stage is shown in Fig. 15. The long-term groundwater-level fluctuation of the area was between minimum of 1800.2 m and maximum of 2337.8 m. The long-term lake-stage fluctuation was between minimum 1885.8 m and a maximum 1891 m. The final lake stage is calibrated as 1885.8 m, and the result was very close to the observed value, which was 1885.45 m. Low water level at the

northern, southern and southeastern limit shows the level specified at the boundary condition (Fig. 15). Interestingly, the development of low groundwater-level anomalies in the well field (Fig. 15) is evident in comparison with the no-abstraction scenario (Fig. 13), while no enhanced abstraction effect was observed in the other two well fields (Fig. 15). Also, the cone of depression in this well field remains confined to that specific location. Dry cells were also observed only in this particular well field (Fig. 15), indicating that the applied abstraction stress is the maximum stress that the well field could carry. The result indicates that the well field is not in direct hydraulic connection to the main recharging water body, which is predominantly the lake. Apparently, similar development of cone of depression was not generated in the other two well fields most likely due to the fact that these well fields are located relatively close by to the main recharging zones. The well field in the high abstraction area is suspected to have additional source of recharge, namely from rivers, and

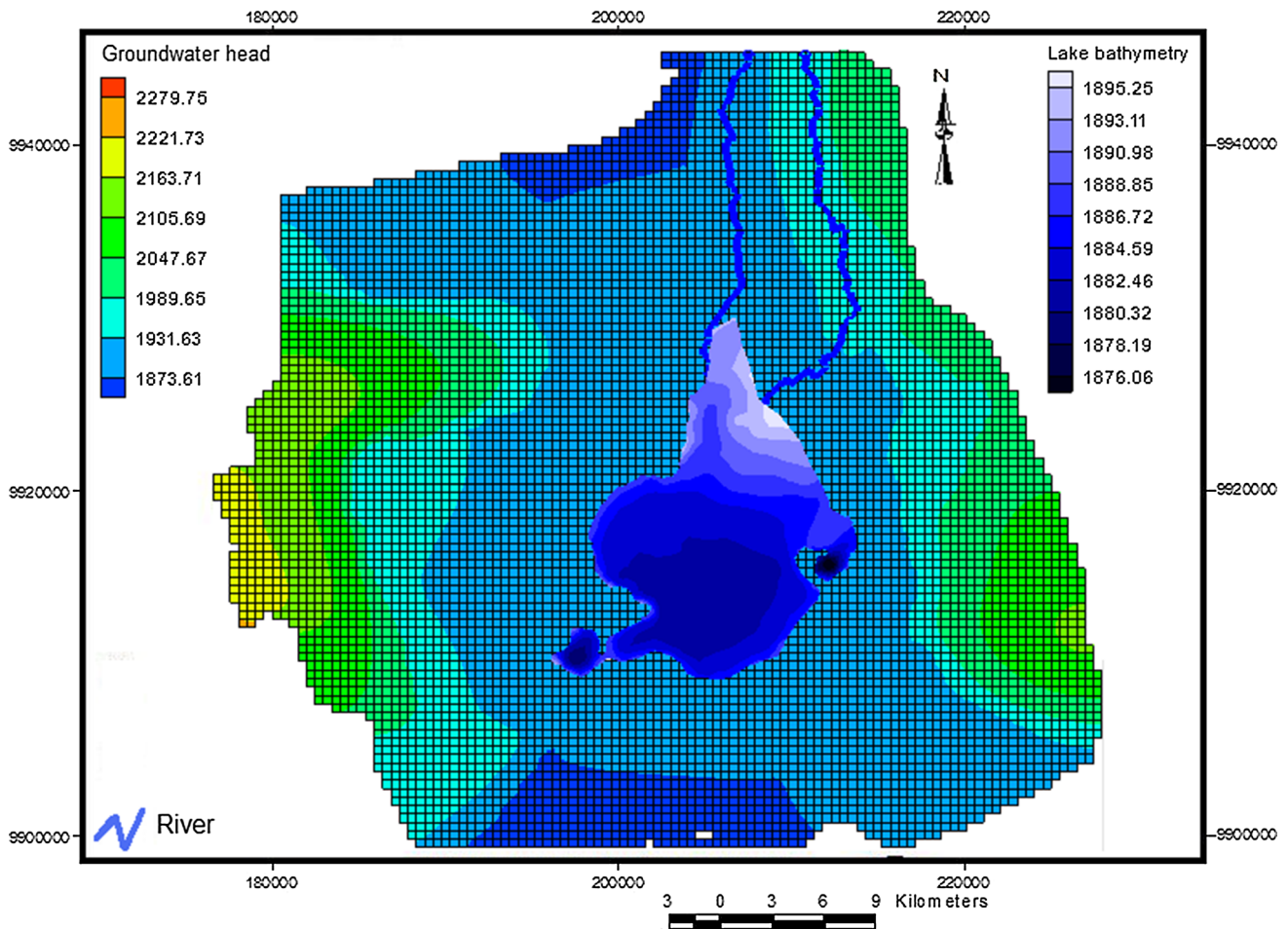


Fig. 13 Groundwater contour map of simulated heads (abstraction not included)

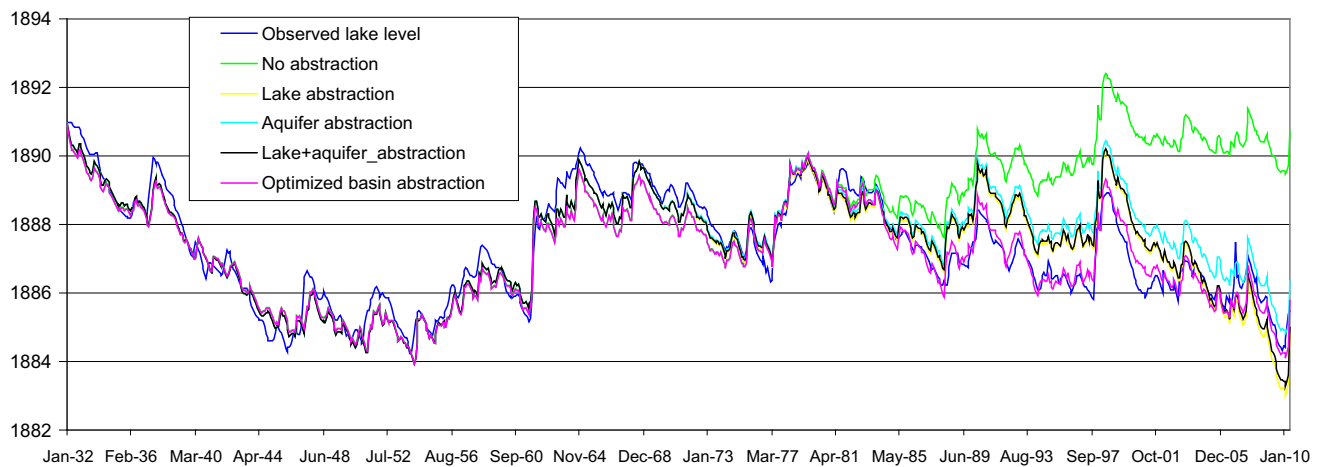


Fig. 14 Response of the lake stage to the different abstraction schemes

this was supported by previous studies, whereby the isotopic composition of Marula (well field: B) signifies that these boreholes have their source of recharge from precipitation and river Malewa and was also confirmed from

the isotopic composition of unsaturated zone of Marula ($d18O = -2.80\text{‰}$, $d2H = -23.6\text{‰}$), which is a mixture of river Malewa and rain (Ojiambo et al. 2001; Yihdego and Becht 2013).

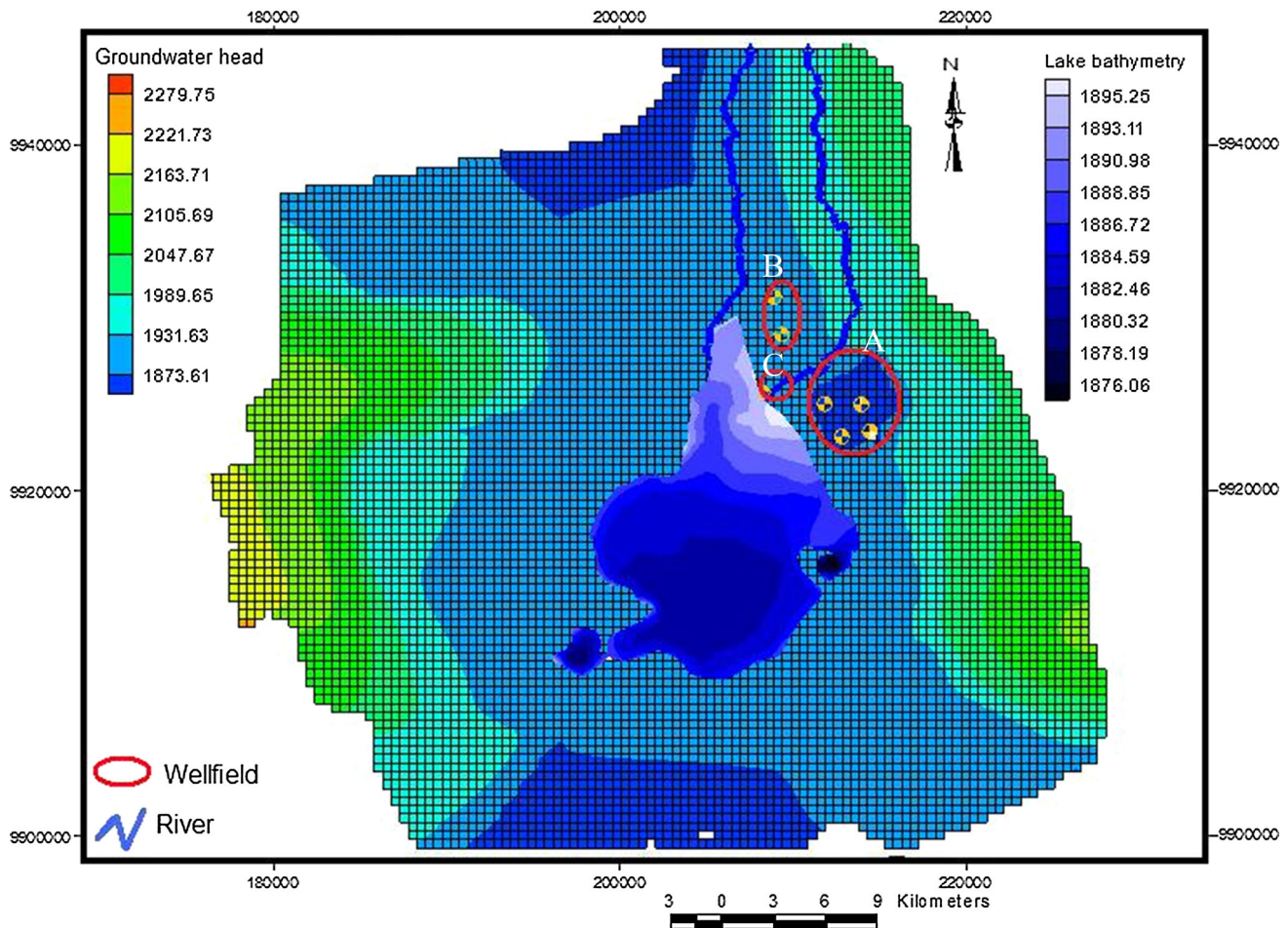


Fig. 15 Groundwater contour map of simulated heads (2010; abstraction included)

Conclusion

The current study reveals that seasonal variability of groundwater–surface water exchange fluxes and its spatially and temporally variable impact substantially on the water resource availability. Current research findings show that the lake cannot sustain further development activities on the scale seen over the last decade. Such analysis can be used as a basis to quantify the linkages between the surface water and groundwater regime and impacts in the basin. Model sensitivity analysis shows that the model is highly sensitive to increasing and decreasing of recharge. The model is also highly sensitive to decreasing in hydraulic conductivity but less sensitive to increasing in hydraulic conductivity. On the other hand, the transient model shows increasing sensitivity to increasing and decreasing of the storativity values but with a slow response.

Lake Naivasha is freshwater among the east Africa rift valley lakes, which is currently as the center of industrial development. The lake environment is fragile but dynamic

and supports tourism and geothermal power generation from deep-rooted stream jets among other economic activities. Lake Naivasha's biodiversity is critically threatened by human-induced factors, including habitat destruction, pollution (from pesticides, herbicides and fertilizers), sewage effluent, livestock feeding lots, acaricide and water abstraction. This high population has encroached wetlands and converted them into agricultural lands, residential areas and tourist hotels. The model is best used for prediction of impacts to groundwater abstraction at a regional scale due to the uprising groundwater lake water usage in the basin. The stake holders and the regulatory organization are among the sectors to get benefit from this model output, for sustainable utilization of the water resource in the lake basin. Basin-wise water resource management strategy can be designed by integrating the regional groundwater model with other water evaluating soft wares to incorporate the ecology and biodiversity of Lake Naivasha designated as a RAMSAR site (high-value world heritage).

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