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Application of Groundwater Vistas in Modelling Groundwater Flow in Keiyo Highlands

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Abstract

Mathematical models which are based on mathematical equations are normally used to describe groundwater flow in a given area. In Keiyo Highlands, 37% of residents use shallow water wells as the sole source of water. Some of these wells run dry during the dry season and therefore a study was conducted with the aim to construct a 2D groundwater of the unconfined aquifer in the area to understand the groundwater behaviour in the study area. Conceptual model showing the positions of rivers, wells and recharge was done. Other data included hydraulic conductivity, porosity for the numerical model which was developed using MODFLOW code. The Graphical User Interface used was Groundwater Vistas. A grid of 2160 cells of sizes 140m by 80m was constructed in Groundwater Vistas. The model was calibrated manually by trial and error method for a steady state conditions during the rainy season in the study area. The transient state showed how the depths of water reduced in the wells. The amount of recharge was 0.00045 m/day, porosity 50%. The hydraulic conductivity varies from 0.05 m/day to 0.09 m/day. The model predicted the heads well under both steady state and transient state conditions. The percent errors were 0.005% and -0.4% for both steady and transient state, respectively.

Key Words*:* Shallow Wells, Steady State, Transient and Prediction

Introduction

In Kenya, over 59%of the total surface water resources are found in the Lake Victoria basin (WRMA, 2009). One sub catchment that contributes water to this basin is Keiyo Highlands. In these highlands, 37% of residents use shallow water wells as the sole source of water (KNBS, 2010). Some of the wells run dry during the dry season which normally occurs between the months of January to March as the water in the wells feed the streams that drain into the Lake Victoria. This therefore, calls for estimation of quantity of groundwater resources for such an area for an effective integrated water resources management. There was need to construct a 2D groundwater of the unconfined aquifer in the area to understand the groundwater behavior in the study area.

Mathematical models which are based on mathematical equations are normally used to describe groundwater flow in a given area. These are empirical models which result from empirical equations that were developed after experiments in a specified area. Empirical models are usually site specific and are not of help in areas far from where they were developed. The other type of models are called deterministic models, which are the main models used in groundwater modelling (Winker, 2010).They provide a quantitative framework for understanding field information and for conceptualizing hydrological processes (Anderson, Woessner, & Hunt, 2015).

The mathematical model represents flow of a single phase fluid (water) at a constant density in a continuous porous medium under Darcy's law, which states that groundwater flows from high to low energy potentials. For 2D horizontal flow in an unconfined, heterogeneous, anisotropic aquifer, the differential equation is:

$$\frac{\partial}{\partial x}\left(K_x h \frac{\partial h}{\partial x}\right) + \frac{\partial}{\partial y}\left(K_y h \frac{\partial h}{\partial y}\right) = S_y \frac{\partial h}{\partial t} - R \ldots\ldots\ldots\ldots\ldots\ldots\ldots 1$$

Where K_x and K_y are hydraulic conductivities in the x and y directions.

S_y is the specific yield. R is recharge rate. Head (h) is equal to the elevation of the water table measured from the base of the aquifer (Anderson *et al*., 2015). The above equation can be solved analytically, but it will be time consuming and tedious owing to the irregular shapes of most aquifers. Therefore, a numerical solution with the help of a computer code was used.

The work of a groundwater model helps in interpreting and forecasting/ hindcasting(Anderson *et al*., 2015). Forecasting models are first tested by comparing model results to field measurements in a history matching exercise called model calibration. Forecasting helps us to understand the future behaviour on aquifer system while hindcasting helps to recreate the past conditions. Hindcasting applications are "uniquely challenging" (Clement, 2011), because it is not possible to collect additional observations to augment the existing historical data set, which is often meagre. It is worth to mention at this point that model validation/ verification is no longer important in groundwater modelling as per the recommendation of Anderson *et al*., (2015).

A groundwater model being a simplification of a real situation, is therefore, limited by the underlying approximations as well as uncertainty (Anderson, *et al*., 2015). Uncertainty arises because current and future hydrogeologic conditions represented in a model cannot be fully described or quantified as well as "unknown unknowns" (Hunt and Welter, 2010).

A deterministic groundwater model means a mathematical representation and is associated to input data for a specific problem. A code is a computer program that processes the input data for a specific model and solves the equations that describe the groundwater flow processes. A code is written in one or more computer languages. For example MODFLOW is written in Fortran (Anderson, *et al*., 2015). There are pre and post- processors for MODFLOW like Groundwater Vistas (GV), Processing MODFLOW for windows, visual MODFLOW, and Argus ONE.

Water budget calculations are standard features of most codes as it helps the modeller to assess the accuracy of the numerical solution. The water budget should show the total inflow being equal to the total outflow. The water budget should be less than 0.5% but an error as high as 1% may be acceptable (Anderson *et al*., 2015).

Groundwater Vistas is a unique groundwater model developed by Environmental Simulations Incorporation (ESI) for Microsoft windows. It is a pre and post processor for MODFLOW and other related models like MODPATH. It can draw both plan and section graphs of the aquifer being modelled. Student Version 6 was used in this study, which is a handy model when it comes to importing and exporting files from AutoCAD and Arcview. The other feature of GV is that it has an automatic calibration procedure which makes it easy to calibrate an aquifer once the target cells with known heads have been identified (Rumbaugh, & Rumbaugh, 2015).

To effectively use a given code, one has to construct a conceptual model which is site-specific. A conceptual model is a qualitative representation of a groundwater system that conforms to hydrogeological principles and is based on geological, geophysical, hydrological, hydrogeochemical and other relevant information like the amount of total dissolved solids (Anderson *et al*., 2015). The basic governing equation for groundwater flow assumes that the density of groundwater is constant and approximately equal to 1.0 gm/cm^3, which is a reasonable assumption for water with concentration of total dissolved solids (TDS) less than 10,000mg/l. Therefore, one has to get data on topography of the area, geology, hydraulic conductivities, precipitation, pumping and well hydrographs for one to construct a simple conceptual model (Alley, Reilly & Franke, 1999).

The main objective of this study was to examine the performance of a groundwater flow model system for Keiyo highlands, with a view of establishing the interaction between the groundwater in wells and streams in the Keiyo Highlands.

Methodology

Study Area

The elevation for the area varies from 2574 m above sea level at Kapkendabridge to 2770 in Nyaru trading centre. The soils in the study area are humic Nitisols overlaying acid igneous rocks (Kempen, 2007).

The study area is roughly rectangular and is bounded by two streams; Nyaru stream to the west and Kipsaina stream to the East. These streams converge at Kapkenda bridge. The water from the highlands contributes to the water that enters Lake Victoria which eventually becomes Nile River that drains into the Mediterranean sea. The plan and the sections for the study area are shown in Figure 1 to Figure 3.

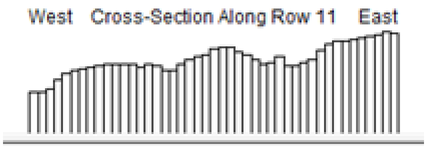

Figure 1. West to East Cross-Section

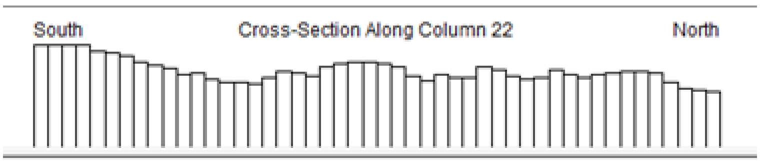

Figure 2. South-North Cross-Section of the Study Area

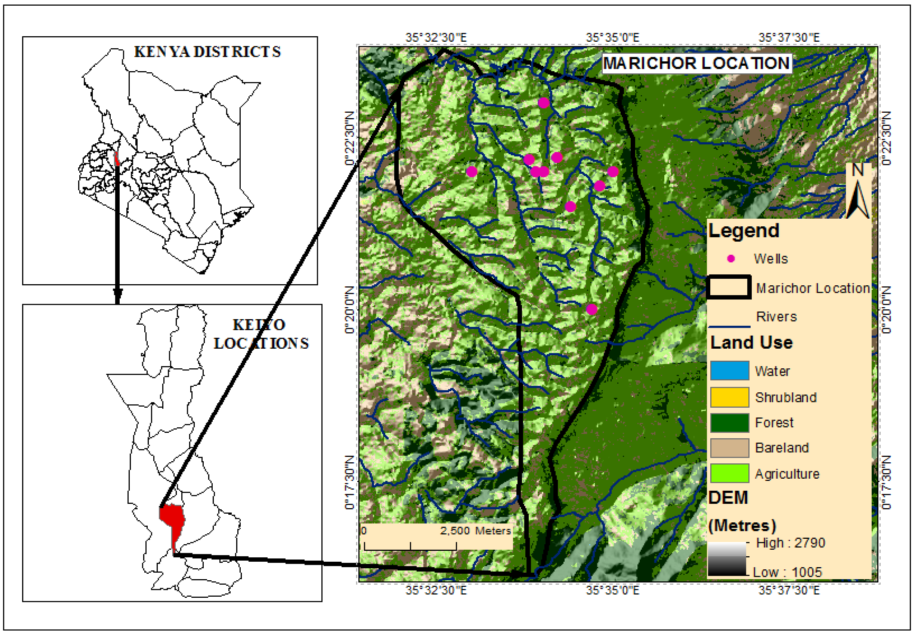

Figure 3. Plan Section of the Study Area

Model Requirements

Study Area

Google Earth Pro was used to capture the image of the study area so as to cover the areas of Nyaru, Chepkorio, Kapkenda and Chororget. The image was captured by using the snip tool to get a rectangular representation of the study area. The captured image was loaded onto the ArcGIS for digitization. In order to get the streams in the study area, a digital elevation model map was clipped from a DEM covering path/row (169/060) with a 30m resolution. The final map was prepared further by using editor tools and Grids in ArcGIS and later saved as a DXF file and exported to Groundwater Vistas.

Rainfall and Recharge

A long term annual rainfall for the period 1960-2014 for Kipkabus forest weather station was used to compute a rainfall event that has 90% exceedance. Since recharge from groundwater is normally from rainfall, the recharge was computed using the empirical equation suggested by Krishna Rao for areas with limited climatological data and receives an annual rainfall of between 600mm and 1000mm (Kumar, 2015).

The empirical equation used was

$$Recharge = 0.25(P - 400)$$

Where P is annual rainfall in millimeters.

Hydraulic Conductivity

Hydraulic conductivity (K) was determined using the Hvorslev slug-test method at time t_{37} when the depth rose to 37 percent of the initial change (Fetter, 2001). Since the ratio of depth of the saturated length (L) of the well to the radius (R) of the well was more than 8, the following equation was used as suggested by Hvorslev.

$$K = \frac{R^2 \ln\left(L/_R\right)}{2Lt_{37}}$$

Water was bailed out from three wells at different locations in the study area. The first well was located at Kipkwen area, the second one was located at Kapng'etik area and the third one at Chepkorio area.

Porosity

Porosity contributes to the water holding capacity. Porosity was determined by taking a sample of 100 cm^3. The samples were taken from a pit latrine that was been dug using sampling core cylinders. One sample was taken at 1m, the second at 2 m and the last sample was taken at 3 m depth from the surface. The sample was dried in an oven at 105°C until it reached a constant weight after 24 hours. The dried sample was then submerged in a 400 cm^3 of water and allowed to remain in a sealed chamber until it was saturated after 3 hours. The volume of the voids was equal to the original volume less the volume in the chamber after the saturated sample was removed.

Modelling

The model units were set as metres for length and days for time. The unconfined aquifer was modelled using MODFLOW. Groundwater vistas were used as the graphical user interface for MODFLOW. The top elevation was entered while the bottom elevation was taken to be 2555 metres above sea level. A grid of 2160 cells of sizes 140m by 80m where columns were placed as 80m and 140m for rows.

Model Calibration

Manual model calibration was done by trial and error method until a satisfactory match was observed between observed heads and simulated heads. The first calibration was done by changing the input recharge. After changing the recharge, the hydraulic conductivities were changed for different areas of the wells to ensure that the simulated heads matched observed heads. There were 55 head targets that were used to calibrate. The heads were obtained from the interview questionnaire.

As a measure of deviation between observed heads and simulated heads the Root Mean Square Error (RMS) was used. Residuals were defined as

$$Residual = Observed\ head - simulated\ head$$

$$RMS = \sqrt{\frac{1}{n}\sum_{1}^{n}(Residual)^2}$$

Other statistical measures that were used to the test goodness of calibration were residual mean, absolute residual mean, residual standard deviation, root mean square error and coefficient of determination, R^2.

$$R^2 = \frac{\left(\sum_{i=1}^{N}(S_i - \bar{S})(O_i - \bar{O})\right)^2}{\sum_{i=1}^{N}(S_i - \bar{S})^2 \sum_{i=1}^{N}(O_i - \bar{O})^2}$$

Where S_i is the simulated head, $\bar{S}$ is the mean of simulated heads, O_i is the observed head while $\bar{O}$ is the mean of observed heads.

The absolute residual mean was calculated after making all residuals positive and is thus an average error in the model.

Modelling Steps

The running of the model was done in two steps. First, it was done by specifying the pumping rates of different wells and recharge. The initial calculated recharge was increased slowly until the simulated heads matched the observed heads in 55 selected wells in the study area. Initially the levels of wells in these wells were set as targets and the program run. The simulated heads and observed heads were compared in the calibration process, and proper calibration was achieved when the heads lined along the 45° line.

The second step was the transient state, in which six wells within the boundaries of the rivers were observed for a period of 120 days. The observed drawndowns in these wells were compared with the modelled drawndowns.

The wells were changed to transient by unchecking the steady stead button and clicking transient states. For transient state, start and end of stress period is given as 1, unlike 0 for steady state.

Initial heads from the steady state model was used janDraft.hds. During this period recharge was taken as zero and evapotranspiration was taken as -0.0073 m/day because evapotranspiration removes water from the aquifer termed as sinks in groundwater modelling.

Results

Recharge from Rainfall

The recharge for the Keiyo highlands was taken as 135.9 mm/year based on annual rainfall of 943.6 mm (Table 1). This recharge value is 14.4% of the annual rainfall which lies between 5% and 30% for Kenya (WRMA, 2009). The recharge that was used as an input in the model was 3.72 x 10^{-4} metres per day based on the rainfall observed in Kipkabus area.

Table 1. Rainfall in Kipkabus near the Study Area from 1960 to 2014

	Year	Amount (mm)		Year	Amount (mm)		Year	Amount (mm)
1	1960	1071	19	1978	1596.7	37	1997	1774.3
2	1961	1185.4	20	1980	1153.5	38	1998	1582.8
3	1962	1051.8	21	1981	1442.7	39	1999	1555
4	1963	1426.7	22	1982	1925.1	40	2000	1228.4
5	1964	1319.8	23	1983	1349.3	41	2001	1503.2
6	1965	956.3	24	1984	619.2	42	2002	1124.4
7	1966	1376.4	25	1985	1270.1	43	2003	974.9
8	1967	1521.4	26	1986	976.6	44	2004	1178.2
9	1968	1586.9	27	1987	1118.6	45	2005	885.5
10	1969	1125	28	1988	1377	46	2006	1815.3
11	1970	1747.2	29	1989	1461.5	47	2007	1659.5
12	1971	1293.3	30	1990	966.7	48	2008	1513.5
13	1972	1228.6	31	1991	1595.2	49	2009	723.8
14	1973	986.2	32	1992	1700.4	50	2010	1284.8
15	1974	837.8	33	1993	1418.8	51	2011	1371.8
16	1975	1355.4	34	1994	1551.6	52	2012	1262.5
17	1976		35	1995	1191	53	2013	1435.7
18	1977	1615.3	36	1996	1090.2	54	2014	1019.4
							Mean	1309.089
							Standard error	39.54

Hydraulic Conductivity Determination

For Kipkwen area, the experiment was done on 27th December, 2015 in Well number 8. The initial depth to water table was measured before and after bailing out 500 litres of water and thereafter depths were measured every hour during the day, at 9.30 pm at night and the last reading was done at 6.30 am the following day i.e. 21 hours of observing the rise of water table. The depth of the well was 38 feet or 11.4 metres and has a radius of 0.52 m. The saturated length L was found to be 11.4 – 3.95 = 7.45 m.

Table 2. Determination of Hydraulic Conductivity for Kipkwen Well

Elapsed time (Hours)	**Depth to water table for surface(inches)**	**Change in water level (inches) h**	**h/h_o**
Static level	158 (or 3.95m)		
0	182	24 (h_o)	1.00
1	181	23	0.96
2	179	21	0.88
3	176	18	0.75
4	175	17	0.71
5	175.5	17.5	0.73
6	175.5	17.5	0.73
7	171	13	0.54
8	170	12	0.50
9	170	12	0.50
12	168	10	0.42
21	161	3	0.13

The time t_{37} was determined by plotting h/h_o on the vertical axis on logarithmic scale and elapsed time on the horizontal axis. The regression equation was found to be h/h_o = 0.92-0.04t with a coefficient of determination R^2 of 0.93.

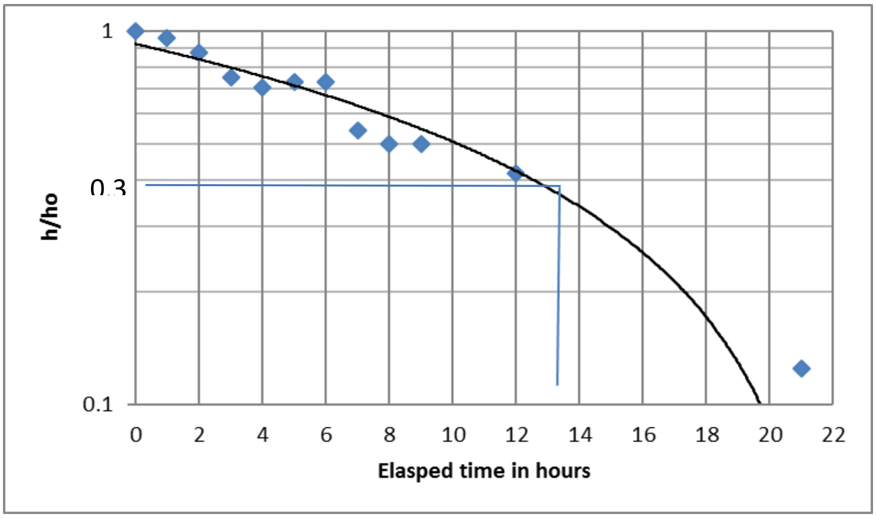

Figure 4. Determination of t_{37} for Kipkwen

From the equation, t_{37} was found to be 13.26 hours or 0.55 days (Figure 4). A value of the hydraulic conductivity of;

$$K = \frac{0.52^2 \ln\left(7.45/_{0.52}\right)}{2 \times 7.45 \times 0.55} = 0.09\ m/day$$

For the well located in Kapng'etik area, the experiment was done on 31st December, 2015 in well number 3. The initial depth to water table was measured before and after bailing out 750 litres of water and thereafter depths were measured every hour from 7.00am and the last reading was done at 4 pm the same day meaning 10 hours of observing the rise of water table. The depth of the well was 70 feet or 21 metres and had a radius of 0.40 m. The saturated length of water in the well L was found to be 21 – 5.4 = 15.6 m (Table 3).

Table 3. Determination of Hydraulic Conductivity for Kapng'etik Well

Elapsed time (Hours)	**Depth to water table for surface(inches)**	**Change in water level (inches)**	**h/h_o**
Static level	216 (or 5.4 m)	(h)	
0	276	60 (h_o)	1.00
1	279	63	1.05
2	280	64	1.07
3	284	68	1.13
4	264	48	0.80
5	260	44	0.73
6	258	42	0.70
7	240	24	0.40
8	246	30	0.50
9	232	16	0.27

The time t_{37} was determined by plotting h/h_o on the vertical axis on logarithmic scale and elapsed time on the horizontal axis. The regression equation was found to be $h/h_o = 1.18-0.09t$ with a coefficient of determination R^2 of 0.85.

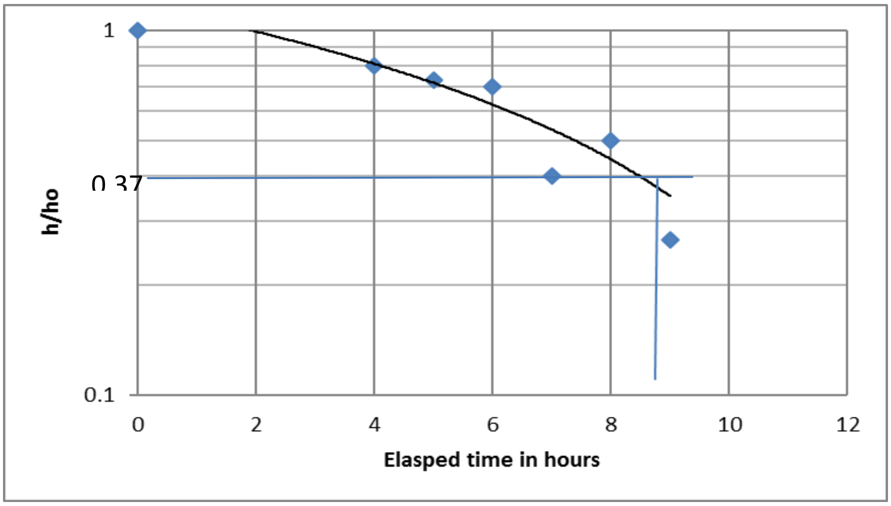

Figure 5. Determination of t_{37} for Kapng'etik

From the equation, t_{37} was found to be 8.8 hours or 0.37 days (Figure 5). A value of the hydraulic conductivity of;

$$K = \frac{0.4^2 \ln\left(15.6/_{0.4}\right)}{2 \times 15.6 \times 0.40} = 0.05\ m/day$$

Lastly, for the well number 72, located in Chepkorio area, the experiment was done on 1st January, 2016. The initial depth to water table was measured before and after bailing out 500 litres of water, and thereafter depths were measured every half an hour from 7.00 am and the last reading was done at 12.30 pm the same day meaning 5.5 hours of observing the rise of water table. The depth of the well was 65 feet or 19.5 metres and has a radius of 0.43 m. At the time before the experiment studied, the saturated length of water in the well L was found to be 19.5 – 6.425 = 13.075 m (Table 4).

Table 4. Determination of Hydraulic Conductivity for Chepkorio Well

Elapsed time (Hours)	**Depth to water table for surface(inches)**	**Change in water level (inches)**	**h/h_o**
Static level	257.0 (or 6.425 m)	(h)	
0.0	292.0	35.0(h_o)	1.00
0.5	283.0	26.0	0.74
1.0	272.5	15.5	0.44
1.5	266.5	9.5	0.27
2.0	264.0	7.0	0.20
2.5	262.5	5.5	0.16
3.0	261.5	4.5	0.13
3.5	261.0	4.0	0.11
4.0	260.5	3.5	0.10
4.5	260.5	3.5	0.10
5.0	260.0	3.0	0.09
5.5	260.0	3.0	0.09

The time t_{37} was determined by plotting h/h_o on the vertical axis on logarithmic scale and elapsed time on the horizontal axis. The regression equation was found to be $h/h_o = 0.66-0.14t$ with a coefficient of determination R^2 of 0.70.

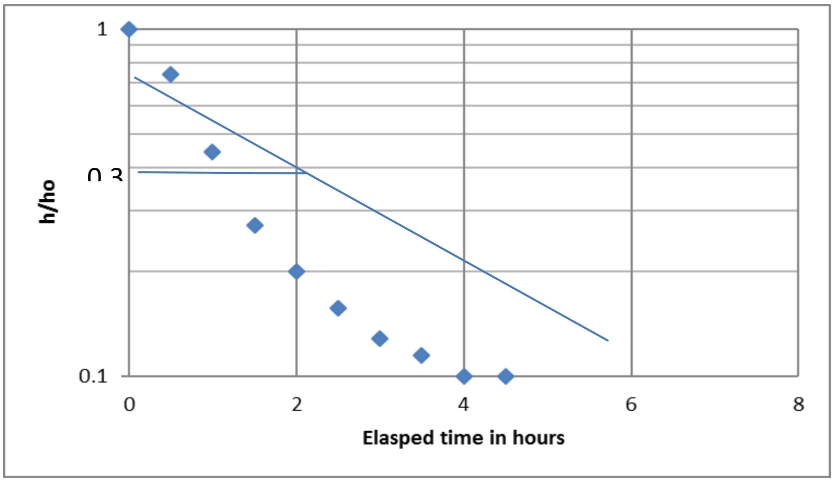

Figure 6. Determination of t_{37} for Chepkorio

From the equation, t_{37} was found to be 2.13 hours or 0.09 days (Figure 6). A value of the hydraulic conductivity of;

$$K = \frac{0.43^2 \ln\left(13.075/_{0.43}\right)}{2 \times 13.075 \times 0.43} = 0.056\ m/day$$

Therefore, the hydraulic conductivity for the study area ranged from 0.05 m/day to 0.09 m/day. The different values of hydraulic conductivity can be attributed to different topographies in the location of different wells. The obtained value is close to 1.06 m/day observed in sandy loam soils of Tandojam in Pakistan (Qureshi, Sarki, Mirjat, Mahessar, & Kori., 2014)

Porosity

The calculation of porosity was done in a tabular form as shown in Table 5.

Table 5. Calculation of Porosity at Different Depths for a Site in the Study Area

Depth	**Volume of sample cm^3**	**Mass of dry sample g**	**Bulk density g/cm^3**	**Initial Volume of water cm^3**	**Volume of water after removing sample**	**Volume of void cm^3**	**Porosity**
1	100	123	1.23	400	350	55	0.55
2	100	115	1.15	400	350	50	0.50
3	100	116	1.16	400	350	50	0.50

The bulk density ranged from 1.15 to 1.23 g/cm^3. The porosity was between 50% and 55%. The porosity at 1 m depth was close to 57% observed in soils in Kabete Campus of University of Nairobi (Karuku *et al.*, ,., 2012). These values are close to weathered rocks of plutonic origin have porosities of between 30 and 60%; and those of clay the porosities of between 33 to 60% (Fetter, 2001). The water holding capacity of any soil is due to the porosity

Steady State with Pumping and Recharge

The estimated recharge value of 0.00037 m/day when first used in the model did not produce the desired results that matched the observed heads in the wet season and therefore the recharge was adjusted to 0.00045 m/day. The contours of the heads are shown in Figure 7. The wells are the boxes in red while the blue and continuous lines indicate the water levels. Where it is light yellow, it indicates higher grounds while the deep blue areas at the north eastern corner of the figure indicate low areas.

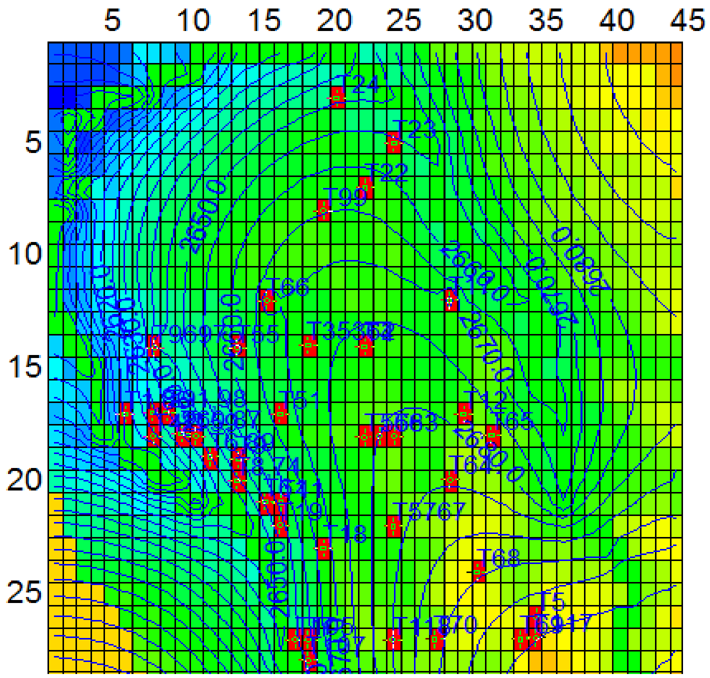

Figure 7. Head Distribution in the Study Area

Model performance was assessed by using calibration statistics shown in Table 6.

Table 6. Calibration Statistics

Statistic	**Value**
Range of observations	121.1 m
Residual mean	0.05 m
Absolute residual mean	6.44 m
Residual standard deviation	8.78 m
Minimum residual	-28.5 m
Maximum residual	9.41 m
RMS error	8.78 m
Coefficient of determination R^2	0.89

From Table 6, root mean square error (RMS) was 8.78 m and the absolute residual mean was 6.44 m. The residual standard deviation was 8.78 m, residual mean was 0.05 m and the coefficient of determination R^2 was 0.89. The residual mean used both negative and positive values and should be close to zero if the positive and negatives balance one another.

The residual standard deviation divided by the range of observed heads was found to be 7.25 %, and the absolute residual 5.32 %. The two values were less than 10 percent, and therefore, the model predicted the heads correctly. The residual mean divided by the range of observed heads was found to be 0.04% which was less than 5%.

Mass Balance for the Entire System

Water entering the system from the rivers is 1,363.4 m^3/day. Water entering the system from recharge is 10,835.4 m^3/day. Water leaving the system from wells and rivers are 17.4 and 12181.9 m^3/day respectively meaning that much of the recharge flows to the rivers. Total recharge can be calculated analytically, which is approximately equal to $0.0045 \times (45 \times 80) \times (48 \times 140) = 10{,}886.4\ m/day$. The total inflows and outflows are all equal to, 12199.83 m^3/day.

Figure 8 shows that most of the observations were on or near the 45° line. The scatter plots showed a good fit between observed and simulated heads.

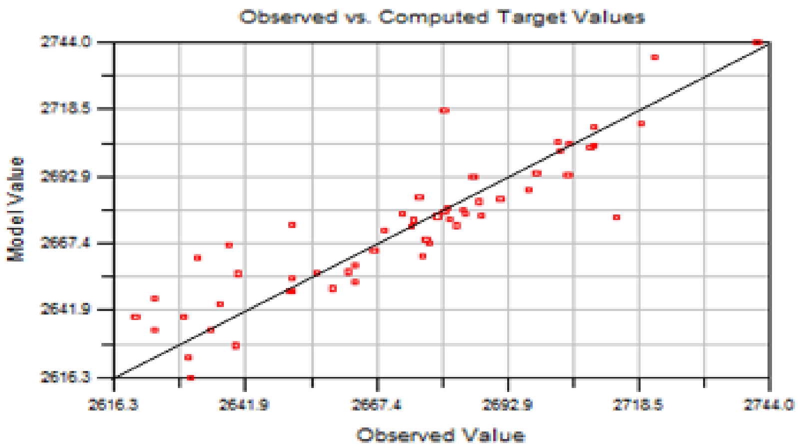

Figure 8. Model Values versus Observed Values

In general, 69% of the residuals were positive meaning that the observed heads were higher than the simulated meaning that the model underestimated the heads for 69% of the wells (Figure 9). The most negative residual was -28.5 m. This well was located near the escarpment and it is thought that the well was getting its water from a perched water table. The discrepancy can also be associated with the digital elevation model which was less accurate because of the resolution of 30m by 30m. The wells in red have negative residuals meaning that the model overestimated the observed heads. The wells in blue have positive residuals meaning that the model underestimated the observed heads.

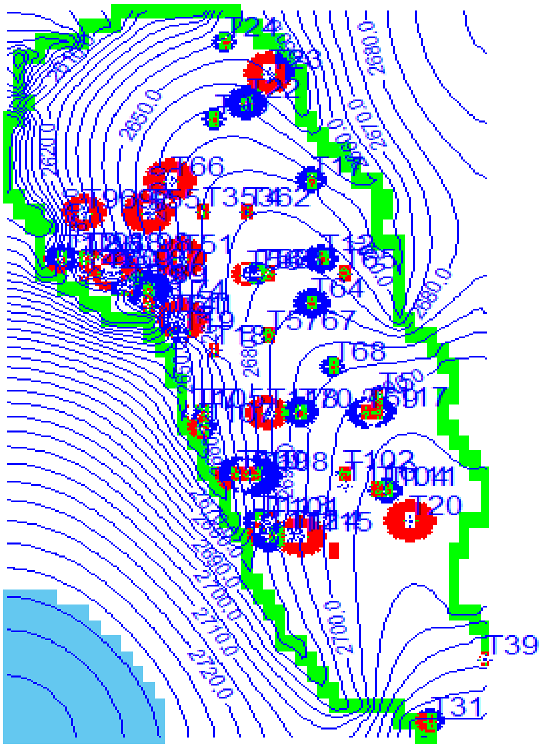

Figure 9. Plots of Wells that were Overestimated or Underestimated by the Model

Table 7 gives the mass balance for recharge, river and wells.

Table 7. Mass Balance Model Summary for Steady State

	Inflow m^3/day	**Outflow m^3/day**
Wells		17.4
Recharge	10836.4	
River	1247.0	12065.4
Total	12083.4	12082.8
Error 0.005		

Transient State

The model predicted the drawdowns for 4 out of six wells well. The discrepancy between the observed and the simulated drawdowns was attributed to the heterogenity of the aquifer and the possibility of a perched water table in the two wells. The mass balance for the entire model is shown in table below.

Table 8. Mass Balance Model Summary for Transient State

	Inflow m^3/day	**Outflow m^3/day**
Storage	170,012.07	
Wells		19.95
Evapotranspiration		161,746.3
River	1608.6	10551.8
Total	171620.6	172318.1
Error -0.4%		

Table 8 showed a percent error less than 0.5%. It showed that the aquifer has enough storage of water when compared to the amount of water withdrawn during the period. So there is no danger of groundwater depletion in the area due to over-abstraction which has been a problem in the past decade (Wada, Van Beek, Van Kempen, Reckman, Vasak, & Bierkens., 2010).Table 8 showed that the water which was gained from the river by the aquifer upstream is lost back to the river downstream which agrees with Bencala, Gooseff, & Kimball., (2011).

Conclusions

The Groundwater Vistas model predicted well the heads in different wells in the study area under steady state conditions. The model also worked for transient state since the percent errors were within the limits. The area has enough storage and therefore this water can be used without danger of mining of water. There is, however, need to collect more hydrogeological data in the area to help in improving the model results.

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