

Water resources assessment study in Kiambu District, Kenya

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ABSTRACT The hydrogeological features of the district consist of an almost closed basin with internal drainage, situated on the eastern rim of the main escarpment of the East African Rift Valley. Groundwater only occurs at depths between 150 and 200 m in fractured zones, weathered palaeo-land surfaces and thin alluvial deposits between volcanic layers. Various hydrogeological methods are applied to assess resources. Groundwater recharge was estimated from a surface water balance based on simplified assumptions for the actual evapotranspiration. A steady-state, numerical-difference flownet analysis was used to calculate the groundwater balance. Based on present and future water demand an integrated water-supply system using both surface water and groundwater, has been proposed. A forecast of its effects on the groundwater regime was made, using a numerical groundwater model.

Détermination des ressources en eaux dans la région de Kiambu, Kenya

RESUME L'étude analyse les caractéristiques hydrogéologiques de la région examinée. Cette région située au bord oriental de l'escarpement principal de la Vallée du Rift, est un bassin plus ou moins fermé à drainage interne. La nappe phréatique se trouve entre 150 et 200 m de profondeur dans des zones fracturées, d'anciennes surfaces d'altération et des dépôts alluvionnaires minces entre les couches volcaniques. Différentes méthodes hydrogéologiques ont été utilisées pour étudier la repartition des ressources en eau. L'alimentation de la nappe phréatique a été étudiée à partir d'un bilan d'eau de surface basé sur des hypothèses simplifiées en ce qui concerne l'évapotranspiration réelle. Une analyse du réseau d'écoulement en état d'équilibre par la méthode des différences numériques a été utilisée pour établir le bilan de l'eau souterraine. Un système d'approvisionnement en eau intégré a été proposé utilisant de l'eau de surface et de l'eau souterraine compte tenu des demandes en eau actuelles et futures. Une prévision de son influence sur le régime de l'eau souterraine a été faite à l'aide d'un modèle numérique.

No data were collected on the monthly variation of k_a . The values of k_a vary from 0.2, at the end of the dry season, to 1.0, during the rainy periods (Table 2). The actual evapotranspiration calculated this way amounts to 90% of the total annual precipitation, and consequently the annual groundwater recharge is in the order of 10% of the total rainfall. Applying this figure to the precipitation figures in the isohyetal map (Fig.2), the average annual recharge for the eastern part was estimated to be 100 mm and for the western part 60 mm.

TABLE 2 Actual evapotranspiration estimation

Month	P (mm)	E_{pan} (mm)	E_o (mm)	k_a	k_c	ET_a (mm)
January	59	193	154	0.5	0.69	53
February	40	202	162	0.2	0.6	19
March	72	201	161	0.75	0.85	103
April	242	140	112	1.0	0.93	104
May	173	103	82	1.0	1.11	91
June	43	90	72	1.0	0.83	60
July	25	83	66	0.75	0.69	34
August	24	96	77	0.75	0.6	35
September	28	138	110	0.5	0.85	47
October	52	176	141	0.75	0.93	98
November	123	139	111	1.0	1.11	123
December	76	161	129	0.75	0.83	80
Total	958	1726	1377			847

It should be realized, however, that all errors due to the various assumptions accumulate in the recharge figure. Therefore, the actual recharge may differ considerably from the calculated recharge. Nevertheless, the above mentioned recharge values were applied in studies using the groundwater model.

GEOLOGY

The area is part of the eastern border zone of the Rift Valley depression. It is infilled with volcanic and sedimentary deposits (Table 3), which rest directly on Pre-Cambrian basement rocks but are generally considered to be no older than Miocene (Fig.3).

Structurally the area belongs to the rift faulting zone. West of this fault system the topography descends gradually along a multitude of north-south oriented faults towards the main escarpment fault, which in turn drops towards the flow of the Rift Valley. There are three major faults and these divide the area into three morphological zones: the highland area, the grazing area and the western escarpment.

TABLE 3 Simplified lithostratigraphic table (after Gevaerts, 1964)

AGE	ROCK TYPE	LITHOLOGY	THICKNESS	HYDROGEOLOGY
QUATERNARY	RECENT	Alluvium	?	poorly draining
		Volcanic deposits	?	well draining
	PLEISTOCENE	Lacustrine deposits	?	no aquifers
		Rhyolites	?	no aquifers
		Limuru Trachytes	max. 200 m	impermeable,
TERTIARY	PLIOCENE	Kerichwa Valley Tuffs	min. 175 m	water-bearing in "palaeo-land surfaces" and
		Nairobi Trachytes	150 m	sandy lenses
			50 m	no aquifers
		Kandizi Phonolites	50 m	water-bearing in faults and
PRE-CAMBRIAN		Crystalline Basement	?	weathered zones

GROUNDWATER CONDITIONS

Volcanic formations form nearly the entire area. These rocks do not themselves transmit water and may not even bear water: the weathered fault and fractured zones within them are the main water transmitting zones and weathered and sedimentary horizons provide the cross connections. Consequently, the water in the fault zones is in

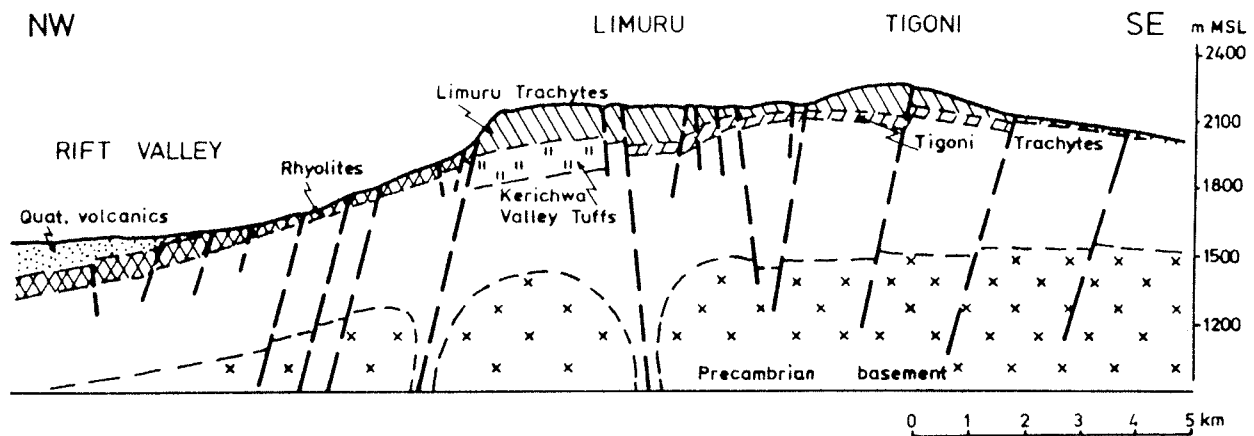


FIG.3 Geological cross section.

principle unconfined, whereas the water in the pervious horizontal layers is confined below impervious volcanic beds. The area can be divided into four groundwater zones (Fig.4):

- (a) the eastern boundary, characterized by the occurrence of many perennial springs feeding numerous small streams east toward the Athi plains;
- (b) the highlands, with high transmissivity values along the faults and groundwater levels at depths between 50 and 100 m;
- (c) the western grazing area, with groundwater levels at depths between 150 and 200 m and very low transmissivity values;
- (d) the Rift Valley floor with groundwater at a depth exceeding 320 m.

A piezometric map (Fig.5) was prepared using the data from 35 monitored boreholes. In areas where no boreholes are monitored, groundwater levels were measured when new boreholes were completed. It was assumed that all water-bearing strata are interconnected and that they have a common piezometric level. The piezometric map indicates flow eastwards, southwards and westwards. Because of the anisotropic transmissivity distribution (i.e. the values in the north-south direction are much higher than those for the east-west direction) the piezometric map does not give a picture of the groundwater flow distribution as it would were the aquifer isotropic.

The groundwater body is recharged by inflow of groundwater from the north, but the main source of recharge is percolation. As a consequence of the plentiful "long rains" (700 mm during March-May) of 1981, the groundwater levels rose in nearly all wells. In wells close to major faults, for example BH C4802 (Fig.6), the level rose less than 2 m. Thereafter, groundwater levels remained constant, showing little or no influence of the less abundant "long rains" of 1982 (500 mm). This behaviour indicates relatively large values for transmissivity and specific yield. In the central part of the area, most wells showed a rise in the groundwater of more than 25 m followed by a sharp decline in groundwater level until the onset of the "long rains" of 1982, for example BH C3642 (Fig.6). The transmissivity and specific yield in this part of the area are probably lower than those along major faults.

The transmissivity values calculated from 28 well pumping test records vary from 1 to $2000 \text{ m}^2 \text{ day}^{-1}$. High transmissivity values generally occur in boreholes drilled in major fractured zones. In

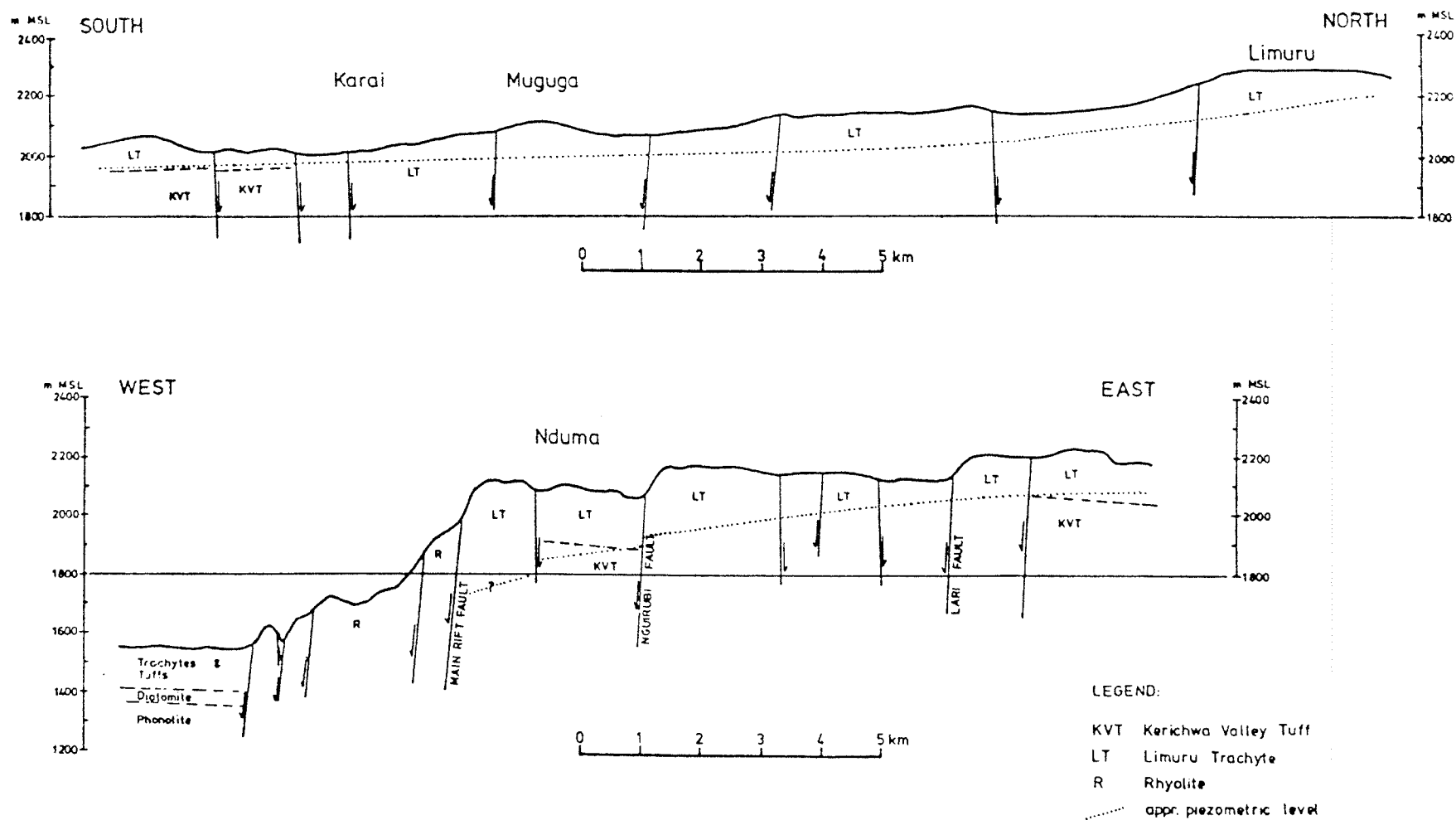


FIG.4 Schematic hydrogeological cross sections.

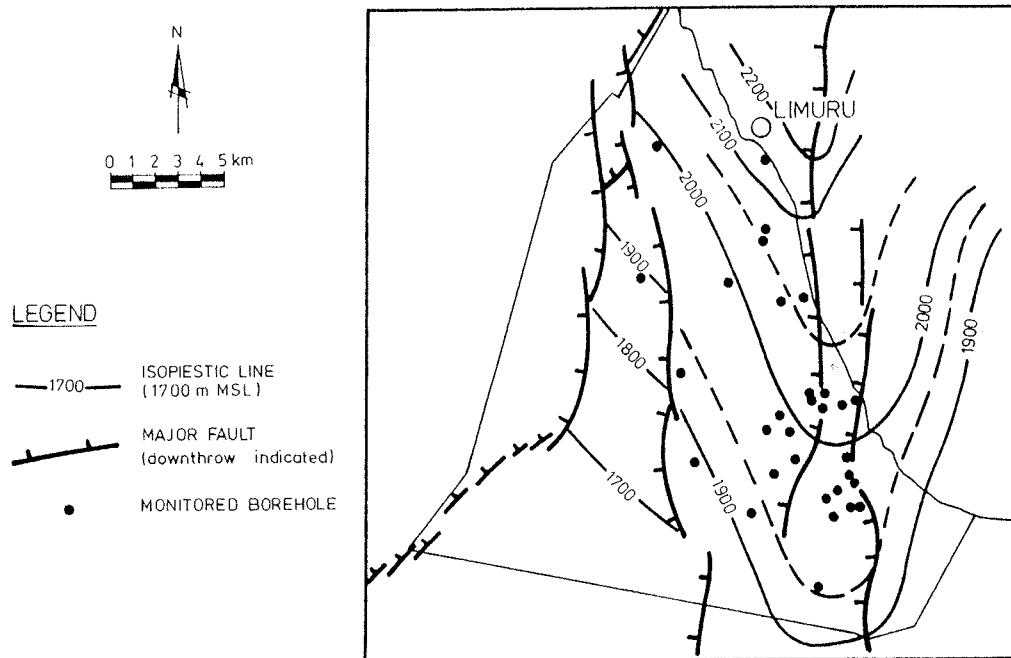


FIG.5 Groundwater zones and piezometric map.

less fractured zones the transmissivity values are less than $25 \text{ m}^2 \text{ day}^{-1}$. The transmissivity only has a local value because of the anisotropic and heterogeneous character of the water-bearing zones.

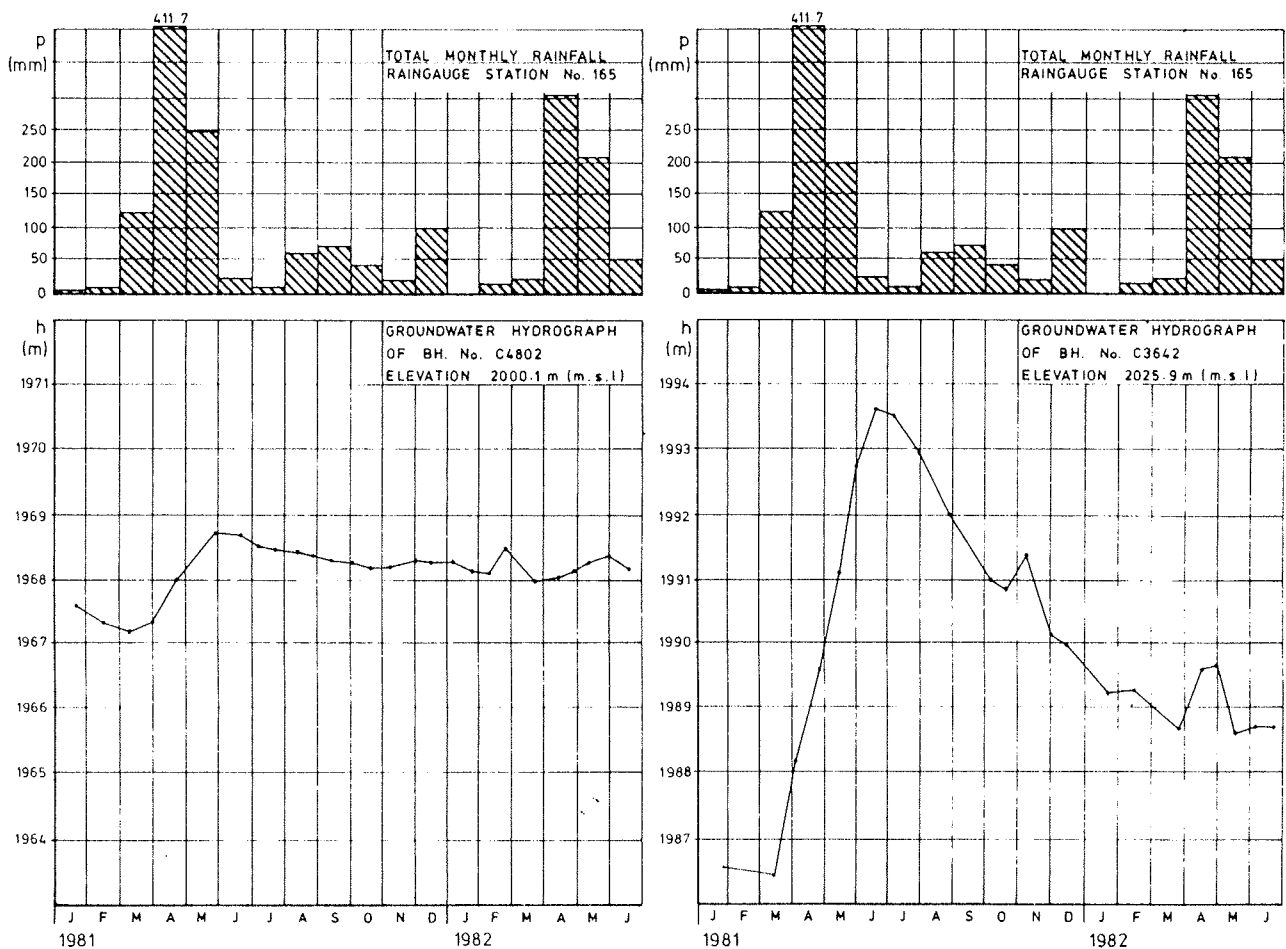


FIG.6 Two examples of groundwater level hydrographs.

To calculate regional flow, T values much smaller than the maximum values have to be adopted. When the transmissivity (T) and the specific gravity (Q/s) were correlated, the result was the relation $T = 0.87 Q/s$ with a correlation coefficient $r = 0.97$.

DRILLING

Five reconnaissance boreholes have been drilled to collect additional hydrogeological information for the project and to obtain proper geological logs. The borehole sites were selected on the basis of the existing geological map and aerial photo-interpretation. Because of the complex geological conditions and the groundwater occurring in relative thin aquifers at great depth, resistivity sounding proved to be unsuitable to detect these aquifers. However, resistivity profiling has been applied successfully to locate faults and fracture zones more accurately.

At completion of the drilling programme, one borehole was reported dry, two had very little water and two were considered suitable for exploitation. Because no appropriate equipment was available at that time, none of the boreholes have been properly well tested. By bailer test it was checked whether these boreholes could be emptied, which was not the case. Just recently one borehole was successfully test pumped; a pumping rate of $4.5 \text{ m}^3 \text{h}^{-1}$ resulted in a drawdown of 6 m after 24 h.

The relevant borehole data are summarized in Table 4.

TABLE 4 Borehole data

Bore-hole no.	Total depth (m)	Water struck level (m below surface)	Water rest level (m below surface)	Estimated safe yield ($\text{m}^3 \text{h}^{-1}$)
C-4886	212	156	134	10.0
4893	296	?	240	<2.0
4909	300	270	213	10.0 (tested)
4971	320	dry	dry	-
4974	320	?	219	<2.0
Total drilled depth = 1448 m				

The drilling costs amounted to approximately 1000 Kenya Shillings per metre. Depending on the depth and the other local conditions, the total well cost for a 200-m-deep borehole and an average pumping rate of $10 \text{ m}^3 \text{h}^{-1}$ will vary between 0.75 and 1.25 million Kenya Shillings.

Although in the western part of the area groundwater could be found at great depth, groundwater development was recommended in the

eastern part where the water table is between 40 and 60 m below the surface.

GROUNDWATER MODEL STUDIES

Steady-state numerical-difference method

To obtain a tentative overview of the groundwater flow pattern and to establish a groundwater balance, the steady-state, numerical-difference method was applied. Although this method is not suitable for studying different groundwater development strategies and their influences on the water table, its advantage is that it is easy to apply and does not require a computer. The calculated groundwater flows are the result of the piezometric levels, the transmissivities and recharge values and therefore this method is an excellent means of verifying the relationship between the various parameters.

A grid consisting of nodes in a square of 2 x 2 km was used (Fig.7). Using the piezometric map (Fig.5), each node was assigned a groundwater level. Then the water level difference Δh between each pair of nodes was calculated. The area between each pair of nodes was assigned a transmissivity value based on the available data on transmissivity and specific capacity. The amount of recharge - 10% of the annual rainfall - for each node was established using the isohyet map for 1981. Groundwater abstraction values for each nodal area were obtained from field data.

Figure 8 shows a schematic flow diagram. A node i is surrounded by four other nodes (nodes 1, 2, 3, 4). The subsurface flow q between node i and each of its surrounding nodes is given by Darcy's

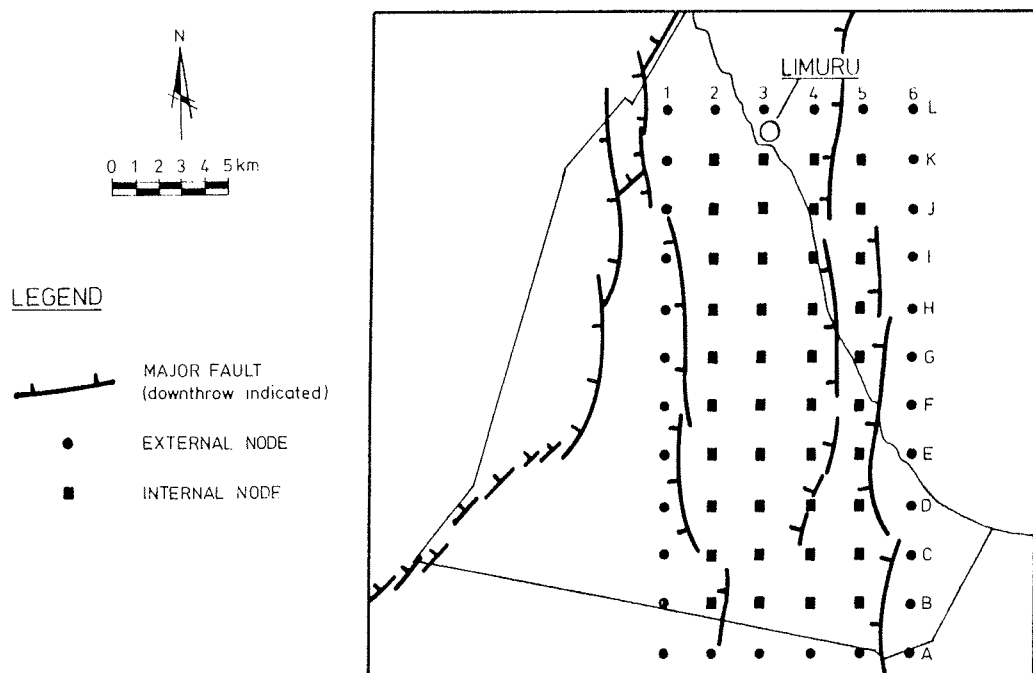


FIG.7 Nodal network steady-state, numerical-difference method.

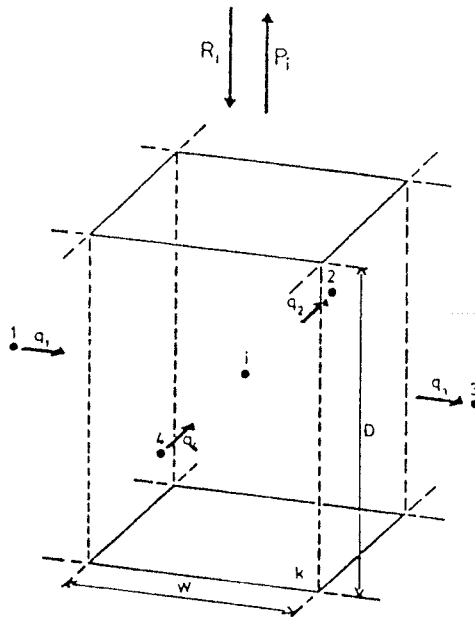


FIG.8 Schematic flow steady-state, numerical-difference method.

law:

$$q = K \cdot (\Delta D/L) \cdot WD \quad (2)$$

where K is hydraulic conductivity (m year^{-1}); D is thickness of the aquifer (m); L is distance between two nodes (m); W is width of nodal square (m); and h is $h_{1,2,3,4} - h_i$ (m). As the width of the nodal squares equals the distance between two adjacent nodes (i.e. $W = L$) it follows that:

$$q = KD \cdot \Delta h \quad (3)$$

As steady-state conditions were assumed, i.e. there is no change in storage, from the law of continuity it follows that for each node the total inflow equals the total outflow. For node i in the example of Fig.8, this gives:

$$q_1 - q_2 - q_3 + q_4 + R_i - P_i = 0 \quad (4)$$

where R_i is the vertical recharge over nodal area i ($\text{m}^3 \text{year}^{-1}$); and P_i is the vertical discharge over nodal area i ($\text{m}^3 \text{year}^{-1}$). This calculation was made for each node point.

The condition that the sum of the calculated flow components is exactly zero, is seldom met. As long as the difference between total outflow and inflow is less than 10%, no further adjustments are made. If the difference is larger, the assumed transmissivity values have to be adjusted, in order to reduce the difference. Note that a change, for example, in the transmissivity value T , has a repercussion on the inflow to node i and on the outflow from node 1. The resulting water balance (Table 5) showed only little difference between total inflow, including vertical recharge, and total outflow, including the vertical discharge (pumping).

TABLE 5 Groundwater balances

Recharge/discharge component ($\times 10^6 \text{m}^3$)	I	II	III
<i>INFLOW</i>			
Subsurface inflow from the north	5.6	4.1	4.4
Vertical recharge (i.e. percolation)	15.9	13.8	13.8
Total inflow	21.5	17.9	18.2
<i>OUTFLOW</i>			
Subsurface outflow eastward	4.5	4.7	3.9
Spring outflow eastward	1.7	1.7	1.7
Subsurface outflow westward	6.2	4.2	4.1
Subsurface outflow southward	7.4	5.9	5.8
Vertical discharge (i.e. pumping)	1.9	1.5	2.9
Total outflow	21.7	17.9	18.4

I : Groundwater balance from steady-state, numerical-difference method.

II : Ditto, from standard groundwater model.

III : Groundwater balance with increased production of groundwater as proposed (WRAP, 1983).

NUMERICAL GROUNDWATER MODEL

To calculate the effects on the groundwater regime of additional groundwater abstraction, a numerical groundwater model was applied. For this purpose the standard groundwater model of Boonstra & de Ridder (1981) was selected. This model is of the "integrated finite difference" type, which has various advantages that make it suitable for use in irregularly shaped areas with anisotropic hydrogeological conditions and limited data.

The model is composed of an irregularly spaced network of grid points (nodes) around which Thiessen polygons are constructed, i.e. nodal areas (Fig.9). By carefully selecting the grid points, the polygon sides will follow the borders of the area to be studied. Grid points can also be selected to correspond with borehole locations, so that during the calibration process calculated groundwater heads can be compared directly with the field measurements.

In each polygon, both Darcy's law and the equation of conservation of matter are valid. Combination of the two results in:

$$\frac{\partial}{\partial x} \left(KD \frac{\partial h}{\partial x} \right) + \frac{\partial}{\partial y} \left(KD \frac{\partial h}{\partial y} \right) = - N \quad (5)$$

where $K(x,y)$ is hydraulic conductivity for horizontal flow (m day^{-1}); $D(x,y,t)$ is saturated thickness at time t (m); $h(x,y,t)$ is hydraulic head at time t (m); and $N(w,y,t)$ is source or sink term at time t .

The left-hand side of equation (5) represents the horizontal flow, the right-hand side the vertical flow. Vertical flow (N) consists of different flow components, depending on the type of

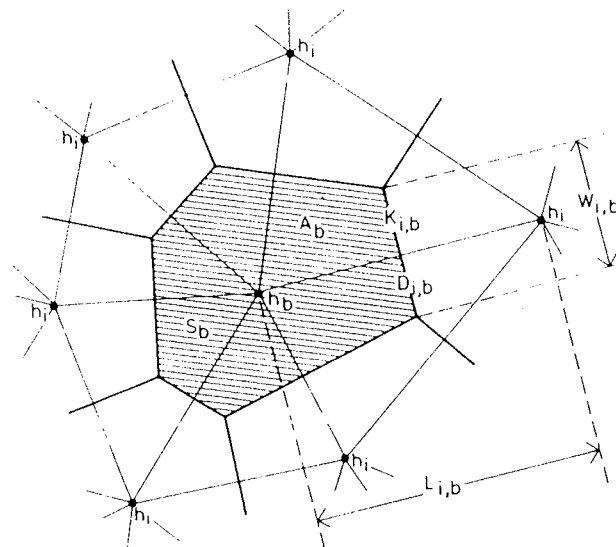


FIG.9 Nodal geometry.

aquifer. For practical reasons, i.e. lack of data concerning the hydraulic head in the overlying layers, the aquifers are assumed to be unconfined. For unconfined aquifers N is given by equation (6):

$$N = R - P - \mu \frac{\partial h}{\partial t} \quad (6)$$

where $R(x,y,t)$ is net rate of recharge (mm year^{-1}); $P(w,y,t)$ is net rate of abstraction (mm year^{-1}); $\mu(w,y,t)$ is specific yield (dimensionless); and t is time.

For an arbitrary node b of the nodal network (Fig.10) the groundwater flow can be calculated by combining equations (5) and (6), and discretizing the time. The result is:

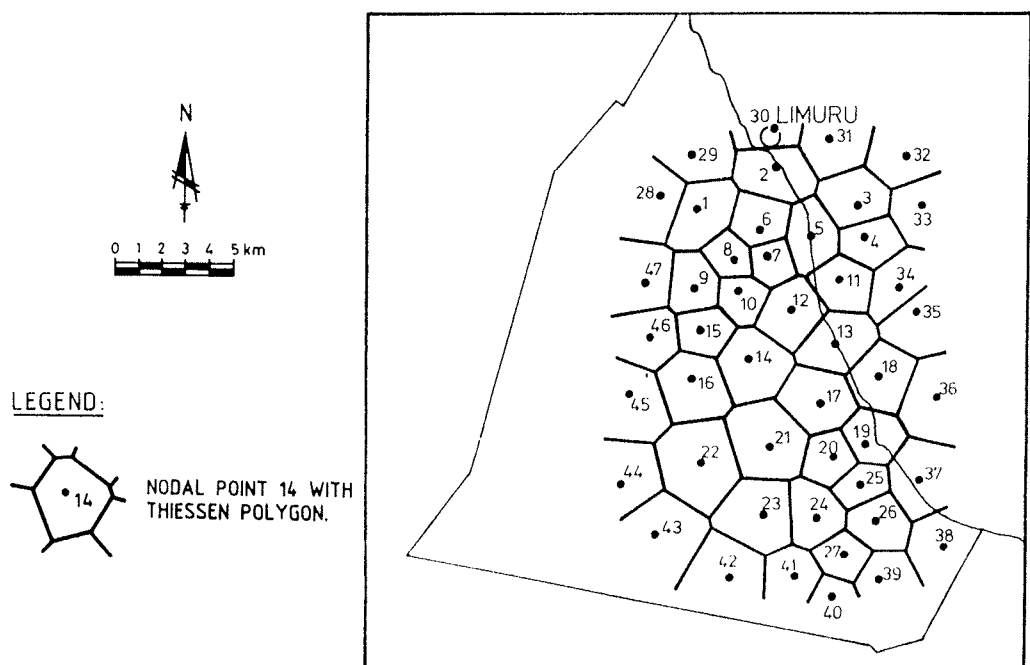


FIG.10 Nodal network: standard groundwater model.

$$\sum_i^n [h_i(t + 1) - h_b(t + 1)] \frac{w_{i,b} K_{i,b} D_{i,b}}{L_{i,b}} =$$

$$A_b P_b(t + 1) + A_b R(t + 1) + \frac{A_b \mu}{\Delta t} [h_b(t + 1) - h_b(t)] \quad (7)$$

where $t + 1 = t + \Delta t$; $w_{i,b}$ is length of polygon side between nodes i and b (m); $L_{i,b}$ is distance between nodes i and b (m); and A_b is area associated with node b (m^2).

The hydrogeological parameters of each node were assigned a value that was supposed to be representative for the polygon under consideration. Because the faulting is predominantly north-south the transmissivity for north to south is much greater than from east to west. In this way the anisotropy of the area was taken into account.

The model was calibrated by adjusting the transmissivity in each node in such a way that the calculated head corresponded as well as possible with the actually measured head. The differences between measured and estimated head varied between +7 and -8 m, which, given the large differences in elevation and the steep hydraulic gradient, is an acceptable result.

The resulting water balance (II in Table 5) showed a remarkable agreement with the water balance from the steady-state, numerical-difference method (I in Table 5). The differences between I and II mainly result from the fact that the areas covered by the models do not coincide.

After the calibration of the model it was used to calculate the response of the groundwater table to the proposed increased abstraction of groundwater. It was calculated that the groundwater level would fall regionally by less than 10 m, although locally the level could fall more. The groundwater balance with the proposed doubling of groundwater production is presented (III in Table 5).

CONCLUSIONS AND DISCUSSION

Water balance studies are essential when assessing available water resources, both in terms of surface water and of groundwater. The method applied to calculate the actual evapotranspiration from pan-evaporation and general crop data gave satisfactory results. However, because little is known about the amount of water consumed by natural vegetation, this method is not yet suitable for use in arid and semiarid regions, where there is more natural vegetation than crops. Further applied research on the evapotranspiration of natural vegetation could be very beneficial for water balance studies in such areas. The steady-state, numerical-difference method proved to be a useful method for a rapid first approximation of the order of magnitude of the groundwater flow, the water balance and the relationship between recharge, transmissivity and piezometric levels. The results from a standard groundwater model indicate that the additional groundwater production proposed (WRAP, 1983) will not cause unacceptable lowering of groundwater levels. Furthermore it confirms the results from the steady-state, numerical-difference method. The great advantage of this model is the flexibility of the node network and the applicability to areas

with anisotropic hydrogeological conditions. The density of the nodal spacing can easily be adapted to the shape and size of the area and the borehole distribution. The transmissivity values applied in both models have no physical meaning in the sense of Darcy's law. However, the applied transmissivity values are useful for regional comparison.

ACKNOWLEDGEMENTS The authors wish to thank Mr D.M.Kirori, Director Water Resources Department, Ministry of Water Development, for permission to publish the results of this study. They acknowledge the contribution made by the International Ground Water Modelling Centre, Delft, The Netherlands, especially by Mr W.I.M.Elderhorst, who assisted with the calibration of the model.

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