Jour. Korean Inst. Mining Geol.
Vol. 22, No. 4, p. 381-393, 1989

Groundwater Flow Model of Igsan Area

Se Yeong Hamm* and Youn Ki Kim*

ATSTRACT : Hydrogeological modelling was performed to evaluate groundwater flow system in Igsan Area.
The study area extends over 790 km^2. The geology consists of Jurassic Daebo granite and gneissose granite and Precambrian metamorphic rocks. The capability of pumping yield is the highest in gneissose granite region among them due to comparatively thick weathered zone with thickness ranging from 10m to 25m.
The Colorado State University Finite Difference Model was used for the model simulation. The model was divided into 28 rows and 31 columns with variable grid spacing.
The model was calibrated under steady–state and unsteady–state conditions. In the steady–state simulation, the model results were compared with measured water table contours in September 1985 with determining hydraulic conductivities and net recharge rates during rainy season.
Unsteady state simulation was done to know the aquifer response due to groundwater abstraction. The non–steady state calibration was conducted to determine the distribution and magnitudes of specific yields and discharge/recharge rates during dry season as matching water level altitudes in May 1986.
The calibrated model was used to simulate water level vaiation caused by groundwater withdrawal and natural recharge from 1 October, 1985 until 30 September, 1995. The calibrated model can be used to groundwater development schemes on regional groundwater levels, but it cannot be used to simulate local groundwater level change at a specific site.

INTRODUCTION

Groundwater flow modelling is used to simulate an actual groundwater flow system. The groundwater flow can be modeled by physical(e. g., a sand tank), analog or mathematical methods. The mathematical model can be divided into analytical and numerical techniques. An analytical solution can only be adapted to simple aquifer problems, whereas the numerical approach can be widely applied to various complex aquifer problems. Two numerical methods are commonly used : the finite–difference and the finite element methods. Both methods need computing facilities. The finite difference approach has a simpler mathematical conception than the finite element method, and consequently is easier to program, however the finite element method is superior to the finite difference for simulating complex aquifer behaviours.

The partial differential equation of two–dimensional unconfined flow is

$$\frac{\partial}{\partial x}(K_{xx}h\frac{\partial h}{\partial x})+\frac{\partial}{\partial y}(K_{yy}\ h\frac{\partial h}{\partial y})=S\frac{\partial h}{\partial t}+q$$

where K_{xx}, K_{yy} : hydraulic conductivity in x– and y–direction, respectively (L/T)
S : storativity (dimensionless)
h : water head(L).
q : pumpage, infiltraticon, leakage from confining layer and evapotranspiration (L/T)
t : time(T).

The partial differential equation for two–dimensional confined flow is

$$\frac{\partial}{\partial x}(T_{xx}\frac{\partial h}{\partial x})+\frac{\partial}{\partial y}(T_{yy}\ \frac{\partial h}{\partial y})=S\frac{\partial h}{\partial t}+q$$

where T_{xx}, T_{yy} : transmissivity in x– and y–direction, respectively (L/T).

This study was done for simulating aquifer system in Igsan area(Fig. 1) as part of Groundwater Resources Survey of Korea sponsored by UNDTCD from 1984 until 1986. The Colorado State University Finite Difference Model (McWhorter and Sunada, 1977 ; Sunada, 1985, 1988)with slight modification was applied for the study to simulate steady–state and unsteady–state aquifer conditions. This model is two–dimensional and can be used for either confined or unconfined aquifers or both with leakage from the confining layer. The model is designed for the IBM–PC and IBM–PC compatibles.

HYDROGEOLOGY

The geology of the study area is mainly composed of Daebo granite, gneissose granite and

*Korea Institute of Energy and Resources, Daejon 301–343, Korea

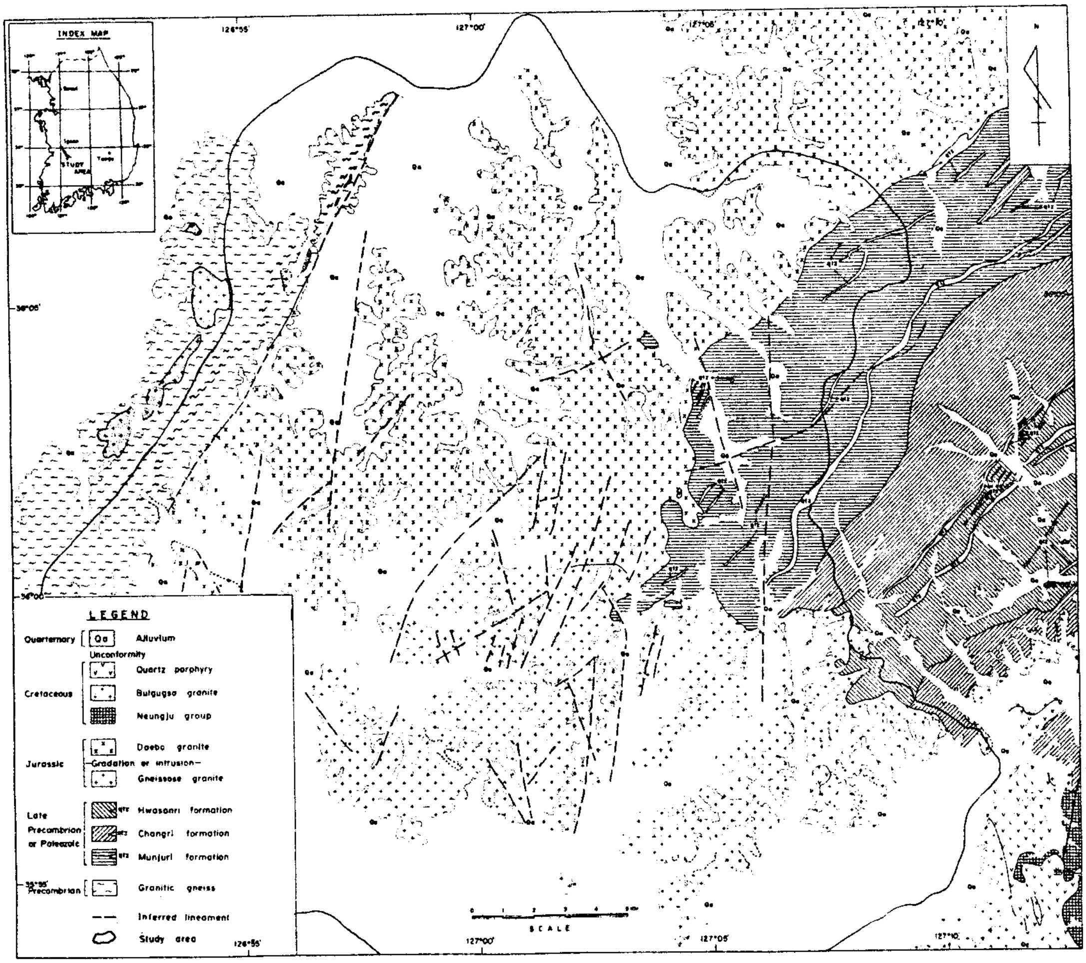

Fig. 1. Geological map of study area.

partly metamorphic rocks(Fig. 1).

Metamorphic rocks are oldest in this study area, and later intruded partly by granite porphyry. The potential of groundwater is very poor, possibly because the thickness of weathered zone is thin and the slope is steep, but a pumping well which is in contact with Daebo granite produces unusually high yield about 1,500 m^3/day.

Gneissose granite occupies the southern part of the study area. This rock is highly weathered near the surface and abraded down, thus shows flat topography. The thickness of the weathered zone ranges from 10m to 25m. The average pumping yield (690m^3/day) of the area in deep wells is higher due to weathering than other granite areas(223m^3/day) in Korea.

The Daebo granite outcrops in the northern part of the study area. The area of this rock shows low and flat topography but the thickness of the weathered zone is less than 5m. The average pumping yield (403m^3/day) is lower than in the area of gneissose granite.

STEADY STATE SIMULATION

The steady state simulation of the groundwater balance can be achieved by letting the storativity equal zero (Trescott et al., 1976 a) or by specifying a large time step with a given storage coefficient (Prickett and Lonnquist, 1971) which will result in no change of water level with time. In study area, a certain small value of storativity (10^{-12}) was used with reasonably large time steps.

The area modelled has an area of 790 Km^2 within which a netwook of 94 wells was used for water level monitoring(Fig. 2). Three types of wells are in use : galleries for domestic use, tube wells and open wells for irrigation purpose. Gal-

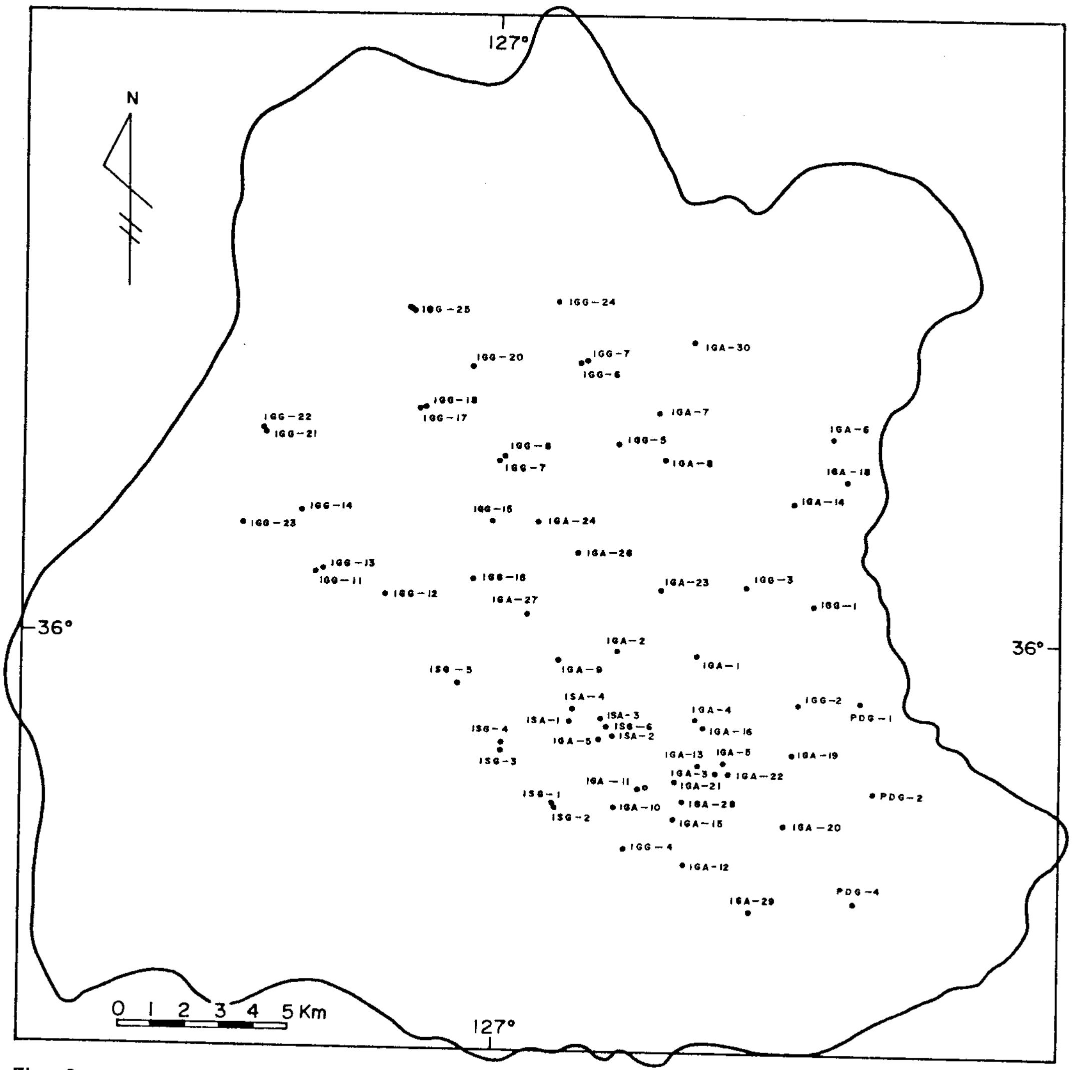

Fig. 2. Well inventory sites.

leries are largely distributed in the northern part. The open wells are drilled to several tens of meters by Agricultural Development Corporation(ADC) and are largely located in the southern part (Kim, 1987).

The model area is bounded by Kanggyoung stream in the north, Mankyong river in the south and highland along the east and the west, including Iri City. The model boundaries were slightly modified from earlier model domain (Lee, et al., 1987) in order to match the well defined natural hydrological boundaries.

Boundary Conditions

In general, boundary conditions are defined as constant head and constant flux boundaries. In the study area, the north and the south rivers are specified as constant head boundaries and the east and the west hills specified as impermeable boundaries. In addition, several internal boundries are also assigned to major drainage channels and watershed divides and three constant head nodes were given to major surface reservoirs. An impermeable area was given to meshes outside the model boundaries(Fig. 3).

The CSUFDM uses three boundary codes to denote any boundary conditions :

$10{,}000 \leq H(I, J) < 20{,}000$–impermeable boundary
$20{,}000 \leq H(I, J) < 30{,}000$–underflow boundary
$30{,}000 \leq H(I, J) < 40{,}000$–constant head boundary

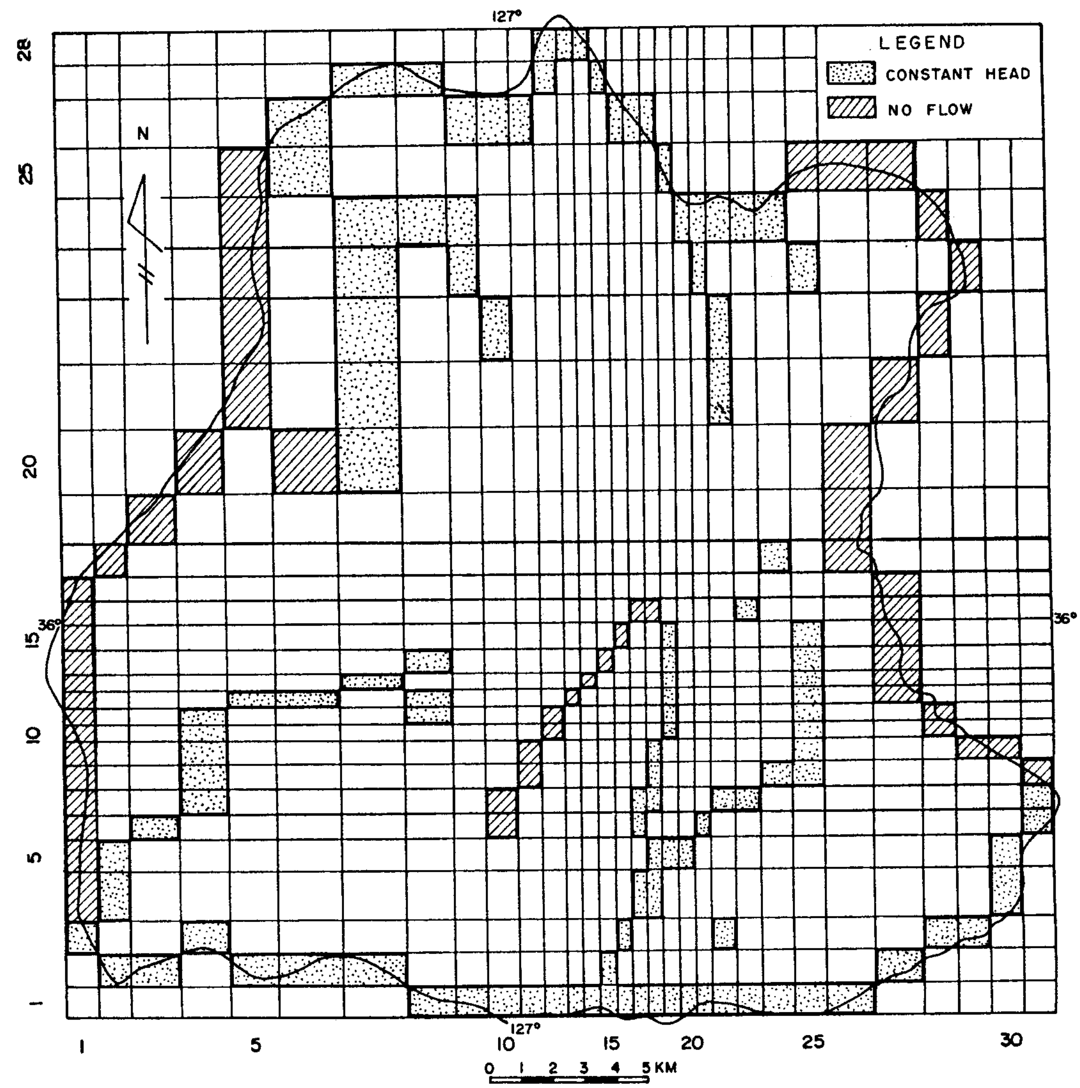

Fig. 3. Model grid network and boundary condition.

In Prickett and Lonnquist's model, zero transmissivity is assigned to a barrier boundary and very large storage coefficient, e. g. 10^{21} is set for a recharge boundary.

Grid System

The block–centered grid system was discretized into 868 lattices(=28 rows × 31columns) with variable grid spacing(Fig. 3). Finer grids are made in areas having sharp variation in heads and along the external boundaries to incorporate their irregular pattern. It is felt that finer grids were necessary for constant head boundaries rather than for noflow boundaries. Grid spacings were employed to satisfy $\Delta X_i/\Delta X_{i-1} \leq 1.5$ or $\Delta X_{i-1}/\Delta X_i \leq 1.5$(Trescott et al., 1976), where ΔX_i and ΔX_{i-1} are the grid spacings of the (i)th and (i−1)th nodes, respectively.

Initial Head Levels

Initial head values are based on the water level contour map of September 1985(Fig. 4). The contour map incompletely covers the study area due to absence of the observation wells in the southwestern portion and the north–western portion. Consequently guessed water levels were used for those areas.

Bedrock Elevation

Bedrock elevations and water levels are required to estimate the saturated thickness of the unconfined aquifer. Since lithological logging data given by the Agricultural Development Corporation(ADC, 1981) gave higher bedrock elevations than the water table altitudes in many locations, the following assumptions were made : thicknesses of 10 to 15 meters at high land of the central eastern part and 40 meters in the lower eastern part were given, for the remainder, 20m to 30m were assigned(Fig. 5).

Ground Surface Elevation

In an unconfined aquifer, the ground surface should be higher than or equal to water table and higher than bedrock elevation. Ground surface elevations are taken from the topographic map.

Discharge from the Aquifer

In the study area, groundwater is abstracted throughout the year for domestic use and only from June to August for agricultural purpose. Due to insufficient field data abstraction for domestic and agricultural purposes has been considered as distributed discharge.

Qdis=Total domestic use/Total area for domestic use
=9,689.25m^3/day/404,675,000m^2
=0.0000239 m/day
=0.024 mm/day
where Total domestic use
=Total population of Igsan Gun Area × Water consumption per man
=77,514 × 125 1/day
=9,689.25 m^3/day
(National Institute of Environment, 1983, Basic study of Main Rivers in the Country)

Groundwater is discharged for irrigation during June, July and August. Table 1 shows computed groundwater withdrawal from the irrigation wells during June, July and August. Total withdrawal from each well in paddy season was calculated as Q=Specific capacity× Drawdown × Pumping days in irrigation season
where drawdown and pumping days in irrigation season were uniformly guessed 10m and 30days, respectively.

Recharge into the Aquifer

Recharge takes place mainly from rainfall and some portion of recharge is derived from return flow from irrigation during June, July and August. Since most of the recharge comes from precipitation, the relationship between rainfall and water level rise in the aquifer can be estimated. Then, the net recharge can be computed as follows :

$Qinf=(\Delta H \times S)/\Delta t$

where ΔH is water level rise in the aquifer during the rainy season (Δt) and S is specific yield or storage coefficient. If we get rainfall–water level fluctuation data for a certain period, we can establish a rainfall–recharge relationship.

Tables 2 and 3 show the rise in water level in irrigation wells and galleries from May to September. From the tables, it is seen that some wells do not indicate a rise or fall and that irrigation wells commonly show higher rise than the galleries. The average rise in both types of the wells is about 2m. Using this value and a specific yield of 0.01, the net recharge was calculated as

$Qinf=(2m \times 0.01)/120$ days$=0.167$ mm/day.

The net recharge is the sum of precipitation recharge and irrigation return flow less irrigation and domestic pumpage.

Hydraulic Conductivity Distribution

Pumping test was conducted at the test site IGA–11 with observing drawdowns at wells IGK–1, 2 and 3 on 29 April, 1986(Fig. 2). Boulton's method, Jacob's method, recovery method and Theis curve fitting analysis were used to determine hydraulic conductivities by analyzing the pumping test. These data were also analysed by a radial model (Rushton and Redshow, 1979). The results are summarised in Table 4.

Based on the pumping test analysis, the hydraulic conductivity was calculated as follows :
Hydraulic conductivity=Transmissivity/(Top elevation of aquifer -Bottom elevation of aquifer)
1.2m/day=57.7 m^2/day/(50m–1.48m)

This value would be reasonable as considering the permeability value ranging 1 to 10m/day for fractured igneous and metamorphic rocks even though massive igneous and metamorphic rocks have very low values between 10^{-4} and 10^{-5}m/day (Todd, 1980). Morris and Johnson's table(1967) gives 1.4m/day of hydraulic conductivity for weathered granite. This initial value still will be changed during the steady state runs.

Boundary Head Levels

To detrmine the external and internal fixed–head boundary levels, the surface elevations close to the river and approximate extension of water table contours to the boundaries were used.

Steady State Model Calibration

Stready state simulation is a calibration procedure to match computed heads with the field values as changing hydraulic conductivity, recharge rate and even boundary conditions. Initial data with uniform hydraulic conductivity

Table 1. Computation of ground water withdrawal from irrigation wells.

Well No.	Grid location	Specific capacity m^2/d	Draw-down in m	Pumping day in paddy season	Total amount of withdrawal in paddy season in m^3	Average withdrawal during the simulation period in m^3/d	Remarks
IGA 1	14, 21	21.36	10	30	6408	53.4	
2	15, 16	10.17	10	30	3051	25.43	
3	8, 21	30.03	10	30	9009	75.08	
4	11, 20	10.00	10	30	3000	25.0	
5	9, 22	28.72	10	30	8616	71.8	
6	20, 26	20.47	10	30	6141	51.18	
7	21, 18	18.70	10	30	5610	46.75	
8	20, 19	13.67	10	30	4101	34.18	
9	14, 21	10.17	10	30	3051	25.43	
10	7, 15	29.42	10	30	8826	73.55	
11	8, 17	29.26	10	30	8778	73.15	
12	5, 20	10.47	10	30	3141	26.18	
13	9, 21	7.19	10	30	2157	17.98	
14	19, 25	6.77	10	30	2031	16.93	
15	7, 19	13.08	10	30	3924	32.70	
16	11, 21	18.08	10	30	5424	45.2	
17	22, 20	16.60	10	30	4980	41.5	
18	20, 26	7.44	10	30	2232	18.6	
19	9, 24	4.94	10	30	1482	12.35	
20	6, 24	9.81	10	30	2943	24.53	
21	2, 19	20.10	10	30	6030	50.25	
22	8, 22	10.23	10	30	3069	25.58	
23	17, 18	26.89	10	30	8067	67.23	
24	19, 11	8.4	10	30	2520	21.0	
25	20, 24	30	10	30	9000	75.0	sp. cap.assumed
26	18, 13	7.0	10	30	2100	17.5	sp. cap.assumed
27	16, 11	8.0	10	30	2400	20.0	sp. cap.assumed
28	7, 20	18.70	10	30	5610	46.75	sp. cap.assumed
29	4, 23	7.44	10	30	2232	18.6	sp. cap.assumed
30	22, 20	5.8	10	30	1740	14.5	
ISA 1	11, 30	7.57	10	30	2271	18.93	
2	10, 15	8.14	10	30	2442	20.35	
3	11, 15	7.5	10	30	2250	18.75	sp. cap.assumed
4	12, 13	7.50	10	30	2250	18.75	sp. cap.assumed
Total		489.62			146,886		

and uniform recharge rate were inputted for the model.

The hydraulic conductivity was adjusted according to the topography and the distribution of lineaments. Low hydraulic conductivity (K) value 0.5m/day was given to higher region which has thin weathered zone with fresh weathering grade, while high K values between 3.1 and 3.5m/day to major lineament zones which may be important groundwater passages in the bedrock(Fig. 6).

Higher recharge rate 0.1mm/day was imposed to the eastern high land and north-western boundary portion and 0.05mm/day was given to the other part(Fig. 7).

Seventeen trial-error runs achieved the steady state simulation(Fig. 8) as comparatively well matching with groundwater contours of September 1985.

Sensitivity analysis evaluates the effect of change of input data, i. e. storativity, hydraulic conductivity, natural recharge and so on, on simulated water levels and reveals that the water table is more sensitive to changes in one kind of input data rather than another (Boonstra and de Ridder, 1981). In this study, precise

Table 2. Rising in water level in irrigation wells from May to September

Well No.	Water level rising(m)	Well No.	Water level rising(m)
IGA- 1	3.13.	IGA- 17	3.15
2	0.23	18	Overflow
3	4.61	19	2.12
4	-	20	0.89
5	3.91	21	-
6	-	22	-
7	0.41	23	-
8	-	24	1.23
9	3.64	25	-
10	-	26	2.71
11	6.40		
12	1.49	28	-
13	2.24	29	2.06
14	-	ISA- 1	3.65
15	2.59	2	2.34
16	No measure	3	1.76

sensitivity analysis has not been done, but head values appear more sensitive to the change of recharge rate than that of hydraulic conductivity. Lower hydraulic conductivities bring higher water levels.

UNSTEADY STATE SIMULATION

An unsteady state model will simulate the aquifer response due to groundwater pumpage. The final steady state calibration is used as the starting point for the unsteady state simulation and calibration. Since the calibration of the model is based on comparing simulated water levels with the historical record, then the longer the period of data, the better the predictive capability of the model.

Pumpage/recharge for one-year cycle is shown as below :

Months

J	J	A	S	O	N	D	J	F	M	A	M
Net recharge Rainy season				Discharge for domestic use Dry season							

Table 3. Rising in groundwater level in galleries from May to September.

Well No.	Water level rising(m)	Well No.	Water level rising(m)
IGG- 1	2.31	IGG- 26	2.14
2	-	27	-
3	-	28	-
4	1.79	29	0.25
5	1.71	30	1.29
6	-	31	1.02
7	-	32	-
8	1.07	33	-
9	-	34	2.12
10	1.73	35	-
11	-	36	-
12	1.27	37	2.14
13	2.92	ISG- 1	-
14	-	2	Overflow
15	2.26	3	Almost no recharge
16	-	4	1.52
17	-	5	0.69
18	-	6	-
19	-	PDG- 1	1.92
20	-	2	2.59
21	Almost const.	4	-
22	Almost no recharge		
23	0.89		
24	Almost no recharge		
25	-		

As May is the last month of the dry season, the water leuel contour map of May, 1986(Fig. 9) was used as the criterion of the unsteady state simulation. Consequently, the period from October to May (8 months) was taken as the unsteady state calibration time.

The data taken from the steady state simulation, including the steady state water level data and other information on aquifer parameters and boundary conditions form the basic input data for the unsteady state runs. In addition,

Table 4. The aquifer parameters computed by different methods.

Well No.	Boulton's		Jacob's		Recovery	Theis best fit method		Radial model		
	T	Sy	T	S	T	T	S	K	T	Sy
IGK 1	16	0.002	67.97	0.0015	80.43	39.2	0.0044	0.7	34.32	0.0025
	16.6	0.01	34.81	0.0054						
IGK 2	30.3	0.002	53.11	0.0025	45.14	27.0	0.0043	0.7	33.83	0.0025
	14.2	0.02	27.06	0.005						
IGK 3	13.8	0.008	81.57	0.017	40.79	18.2	0.0337	0.37	18.11	0.02
	14.7	0.04	18.01	0.03						

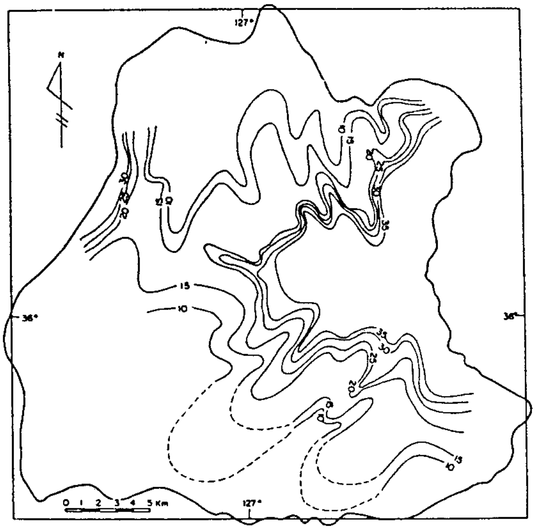

Fig. 4. Water table contour map of September, 1985.

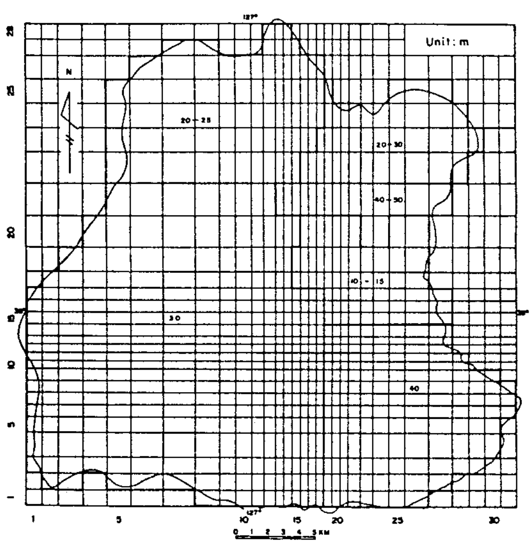

Fig. 5. Initial saturated thickness map.

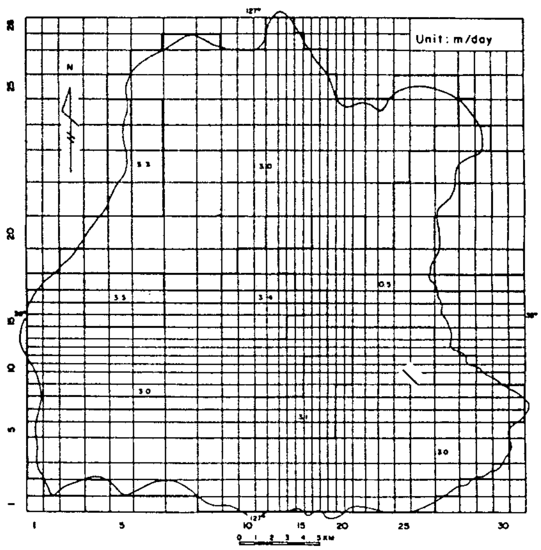

Fig. 6. Hydraulic conductivity distribution.

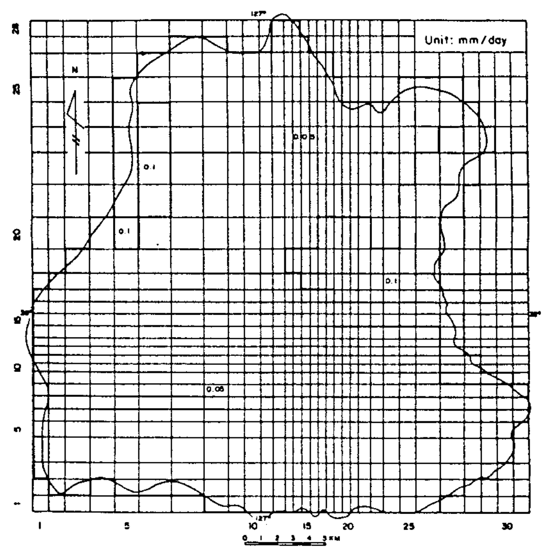

Fig. 7. Net recharge distribution during the rainy season.

specific yield and groundwater pumpage data are also required.

Specific Yield

As based on the pumping test result, uniform specific yield value equal to 0.02 was applied to all the nodes and this value was adjsted during later runs.

Groundwater Pumpage

In the dry season from October to May, discharge from the aquifer is only for domestic use. Pumpage for domestic purpose can be roughly estimated by observing drawdowns in galleries from September to May(Table 5). As shown in the table, more than half of the galleries do not show any change in water level during the dry season. The groundwater contours of May 1986 compared with September 1985 confirm this observation. Consequently a uniform 0.024mm/day was used at the first run as estimated in the steady state simulation.

Unsteady State Model Calibration

The aim of unsteady state calibration was the identification of specific yield and recharge/discharge rates as matching the May 1986 water levels.

In the first run, uniform specific yield 0.02 gave drawdowns of less than 1 m overall with a little up-lising in a portion of the central eastern part. In the second run, a specific yield of 0.01 gave more drawdowns in alomost all the area than in the first run. Lastly, a specific yield of 0.02 was assigned to the southern part, 0.01 assigned to eastern high land and northern tip part and 0.01 to 0.04 assigned to the central part(Fig. 10).

At first, discharge and recharge rates were inputted based on domestic use and natural infiltration. That is, the distributed discharge of 0.024mm/day was given to the majority of the area, the diffused recharge of 0.2mm/day to south-eastern part and various discharge values to the other part. Lastly, the highest rate 0.05mm/day discharge was given to the northern central part, then 0.04mm/day given to the western tip part, 0.001mm/day to the western central part, zero discharge to the eastern and western high lands and 0.02 to 0.03 mm/day given to majority of the other area and distributed recharge rate 0.2mm/day was given to the south-eastern part(Fig. 11).

The simulated water level contours of the unsteady state calibration(Fig. 12) compare well with the water table contours of May 1986, with small deviations in the south-western, northern and central parts. Actually, as many galleries show no drawdown from October 1985 to May 1986 on Table 6, the simulated water table contours in May 1986 have much similar trend to that in October 1985.

Prediction of Transient Aquifer Condition

Transient model simulation from 1 October, 1985 until 30 September, 1995 was conducted using steady state and non-steady state calibration data. Total 36 time steps were used by dividing one year into three time increments(one for the rainy season and two for the dry season). Net recharge during the rainy season and recharge and pumpage rates during the dry season and recharge and pumpage rates during the dry season were taken from the Fig. 7 and Fig. 11, respectively.

As result of the simulation, drawdowns ranging from 6m to 15m appear in the eastern part, whereas drawdowns less than 5m appear in most of the other part(Fig. 13). This is because the

Table 5. Drawdowns in galleries from September to May

Well No.	Grid No.	Drawdown(m)	Well No.	Grid No.	Drawdown(m)
IGG-1	17, 25	1.60	IGG-24		-
2		-	25		-
3		-	26	22, 7	1.85
4		-	27	23, 7	1.62
5		-	28		-
6	22, 13	1.53	29		-
7		-	30		-
8	20, 11	1.66	31	21, 7	1.35
9		-	32		-
10	20, 7	2.05	33		-
11		-	34		-
12	17, 7	1.08	35	25, 15	1.52
13	18, 6	2.62	36		-
14	19, 6	1.03	37		-
15	19, 10	2.28	ISG- 1		-
16		-	2		-
17		-	3		-
18		-	4		-
19		-	5		-
20		-	6		-
21		-	PDG- 1		-
22		-	2	7, 26	2.11
23	19, 5	1.81	4		-

high land has low specific yield 0.01 and low hydraulic conductivity 0.5m/day.

This means that regional water level variation will not be severe for 10 years if discharge and recharge conditions are maintained as 1985.

Yearly volumetric balance of groundwater at transient condition is presented in Table 6, where outflow from the aquifer is negative.

CONCLUSION

Table 6. Yearly volumetric balance at transient aquifer condition

Natural recharge	3,376,500 m^3
Withdrawal	-275,370 m^3
Inflow from interior constant head grids	-7,002,506 m^3
Boundary inflow	-1,140,215 m^3
Change in storage	5,026,975 m^3
Total mass balance	-14,616 m^3

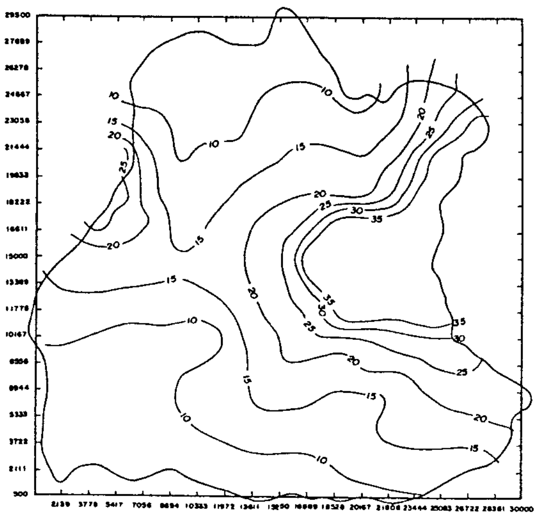

Fig. 8. Simulated water table of October 1985 by steady state calibration.

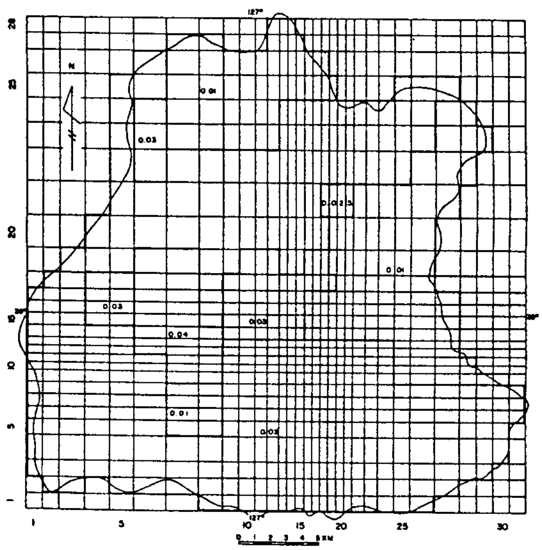

Fig. 10. Map of specific yield.

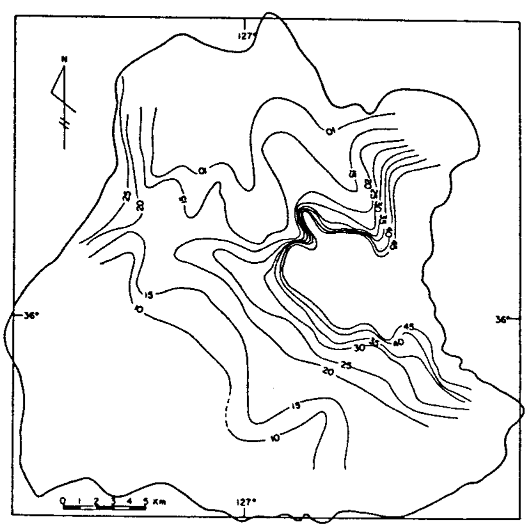

Fig. 9. Water table contour map of May, 1986.

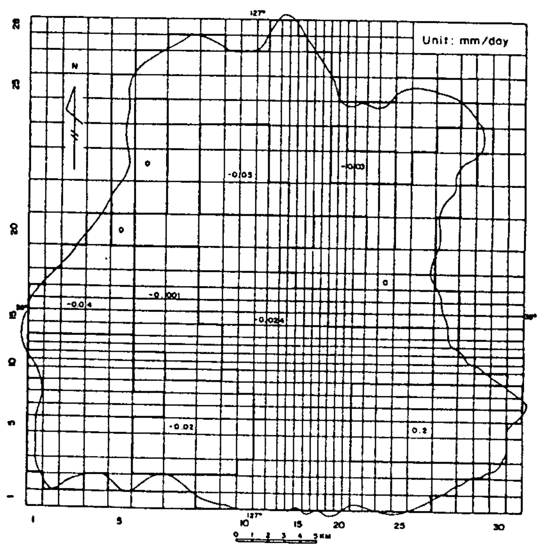

Fig. 11. Recharge and discharge distribution during the dry season.

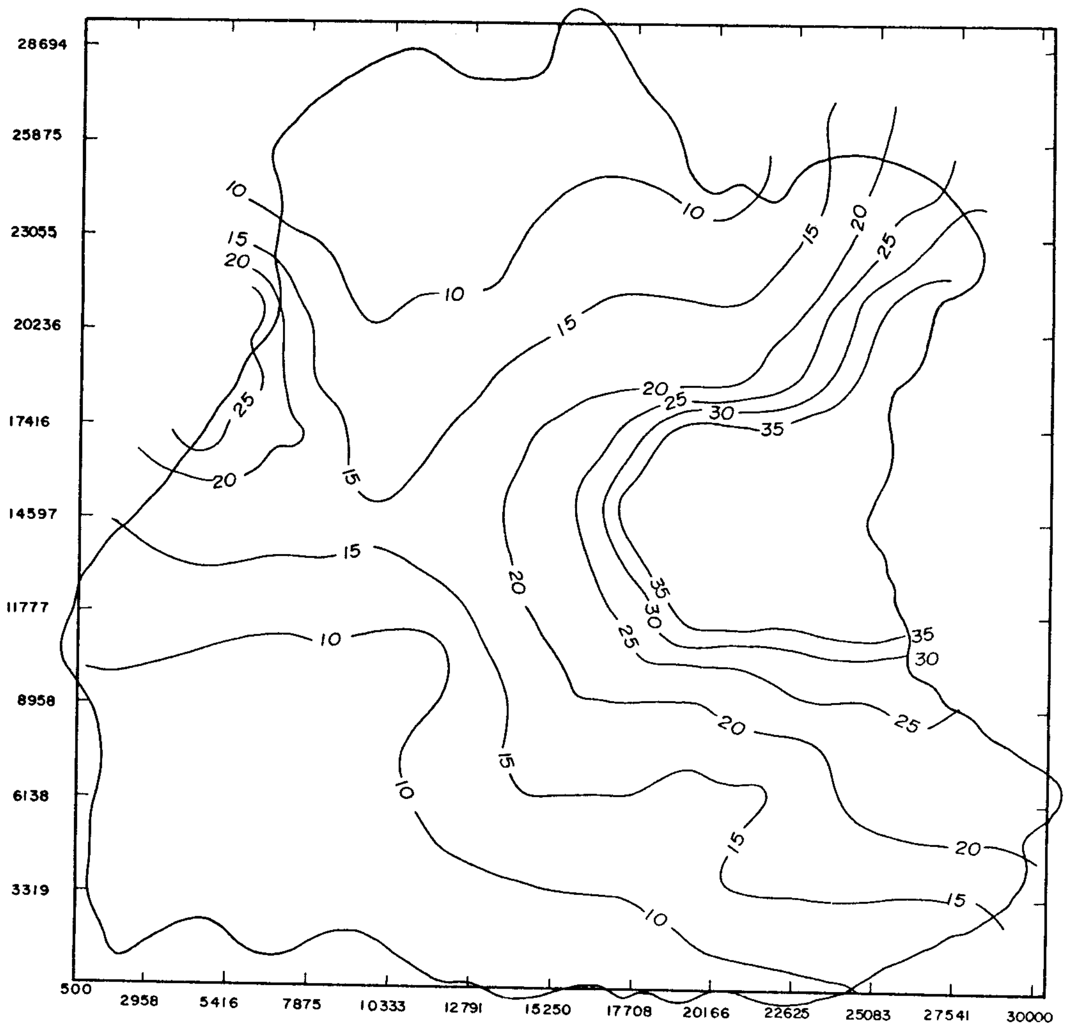

Fig. 12. Simulated water table of May 1986 by unsteady state calibration.

(1) A finite difference modelling was performed to simulate groundwater flow system in Igsan Area.

(2) Recharge distributin shows higher recharge values 0.1mm/day in the eastern high land and north-western boundary portion and 0.05mm/day in the remaining part during the rainy season from June to September.

(3) Hydraulic conductivity distribution was determined as 0.5m/day to higher eastern portion, values between 3.1 and 3.5m/day to lineaments, and 3.0m/day to the other part.

(4) More than half of the study area was given a specific yield of 0.02, higher eastern land, northern tip portion and lower central portion given lower value 0.01 and middle part given values between 0.025 and 0.04.

most of the other part.

(5) During the dry season, the distributed discharge of 0.05mm/day was assigned to the northern central part, 0.04mm/day to middle western part, 0.001 to western central part, zero to north-estern and north-western parts, and 0.02mm/day to 0.3mm/day to remaining part excluding south-eastern part where distributed recharge of 0.2mm/day was given.

(6) The simulated water level contours of the steady state and the unsteady state simulations compare quite well with the water table contours of September 1985 and May 1986, respectively.

(7) As a result of transient model simulation from 1 October, 1985 to 30 September, 1995, most of the other part.

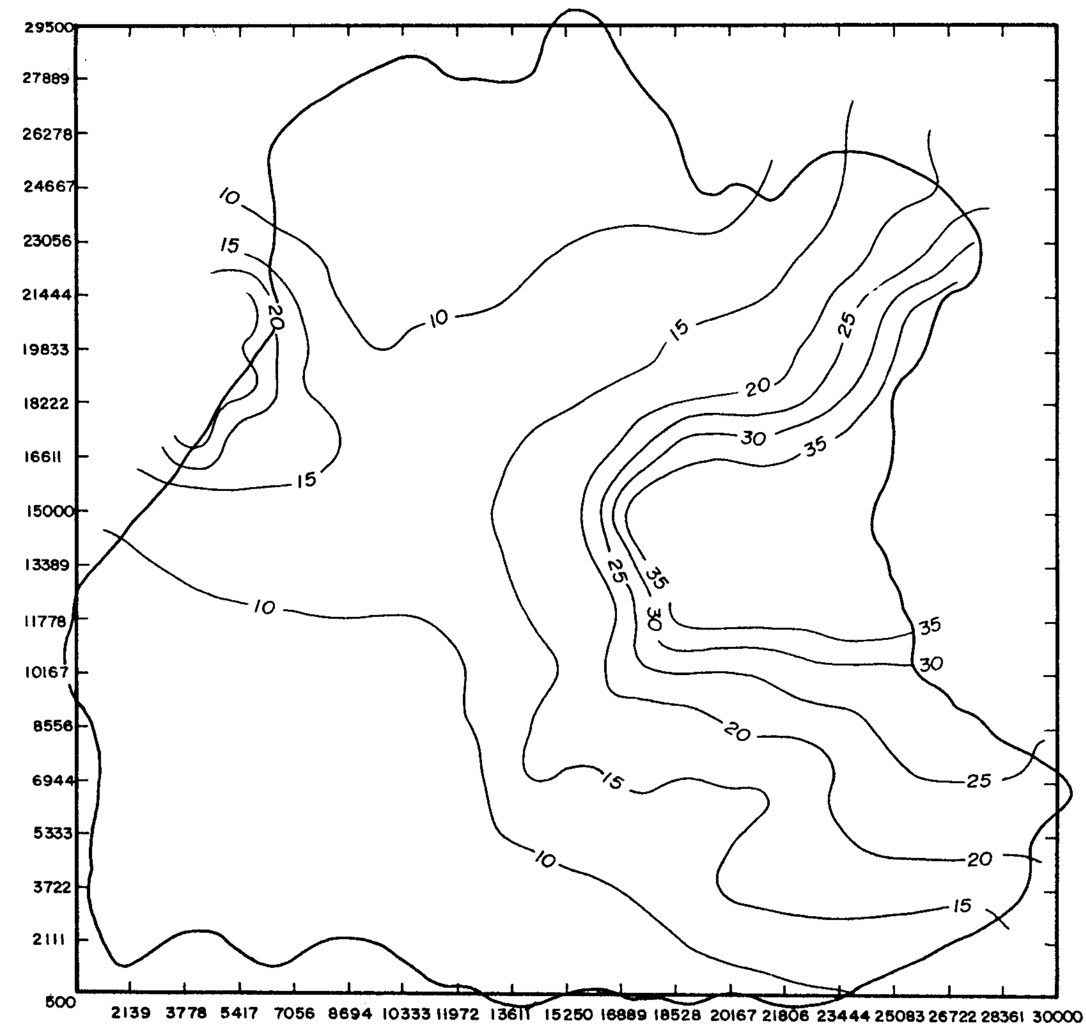

Fig. 13. Simulated water level of September 1995.

(8) This model can be utilized to evaluate groundwater potential of Igsan Area and to allow management of this resources but should be improved as new data becomes available.

ACKNOWLEDGEMENTS

The authors would like to express their thanks to Drs. Daniel K. Sunada and James W. Warner of Colorado State University for providing of their valuable Colorado State University Finite Difference Model and helpful comments on this work. Special thanks are also due to Dr. R. K. Prasad, Central Ground Water Board, India for his helpful advice in many respects.

REFERENCES

ADC(1981) Groundwater Report on Igsan Area.

Boonstra, J. and de Ridder, N. A.(1981) Numerical modelling of groundwater basins, International Institute for Land Reclamation and Improvement (ILRI), The Netherlands, p.177.

Kim, Y. K.(1987) Hydrogeological Study of Igsan Area, U. N. D. P final report.

Lee, D. W. et al. (1987) Hydrogeological Study of Igsan Area, Research on Applied Geology of Korea, KIER, p.51–59.

McWhorter, D. B. and Sunada, D. K.(1977) Groundwater Hydrology and Hydraulics, Water Resources Publications, U. S. A., p.252.

Morris, D. A. and Johnson, A. I. (1967) Summary of hydrologic and physical properties of rock and soil materials, as analyzed by the Hydrologic Laboratory of U. S. G. S, 1948–60, U. S. G. S. Water Supply Paper 1839–D, p.42.

Prickett, T. A. and Lonnquist, D. G.(1971) Selected Digital Computer Techniques for Groundwater Resource Evaluation, Illinois State Water Survey Bulletin 55, p.62.

Rushton, K. R. and Redshow, S. C.(1979) Seepage and Groundwater Flow, John Wiley & Sons, p. 231–266.

Sunada, D. K. (1985) Instruction for the use of the Colorado State University Finite Difference Model.

Sunada, D. K. (1988) Instructions for the use of the Finite Difference Model, Version 2.0.

Todd, D. K.(1980) Groundwater Hydrology, John Wiley & Sons, Inc., p.71–72.

Trescott, P. C. and Larson, S. P. (1976a) Supplement to Open File Report 75–438 (U. S. G. S. Open File Report 76–591).

Trescott, P. C., Pinder, G. F. and Larson, S. P.(1976b) Finite difference model for aquifer simulation in two dimensions with results of numerical experiments, U. S. G. S., Techniques of Water Resources Investigations, Book 7, Automated Data Processing and Computations.

익산 지역의 지하수 유동 모델

함 세 영 · 김 연 기

요약 : 익산지역의 지하수유동계를 평가하기위한 수리지질학적 모델링이 수행되었다.

연구지역의 범위는 790km²이다. 본지역의 지질은 쥬라기의 대보화강암과 편마상화강암 그리고 선캠브리아기의 변성암으로 구성되어 있다. 채수량은 편마상화강암지역에서 가장 높으며, 이는 10m내지 25m두께의 비교적 두껍게 발달된 풍화대때문이다.

본 모델 시뮬레이션에는 콜로라도주립대학의 유한차분법모델이 사용되었다. 본 모델은 분균일한 격자간격을 가지며, 28행과 31열로 이루어져 있다.

본 모델은 정류상태와 부정류상태에서 보정되었다. 정류상태 시뮬레이션의 결과를 1985년 9월의 수위등고선과 비교하여 투수율과 우기의 순 충진율을 결정하였다.

부정류상태 시뮬레이션은 지하수 채수에 따른 대수층의 반응을 알기위한 것이다.

부정류상태 보정에서는 1986년 5월의 수위등고선과 맞춤으로써 비산출율의 크기와 분포 그리고 건기의 채수량과 충진량을 결정하였다.

보정된 모델을 이용하여 1985년 10월부터 1995년 9월까지의 지하수채수 및 자연적인 충진에 의한 지하수위변동을 예측하여 보았다. 보정된 모델은 광역적인 지하수개발계획에 이용될 수 있다. 그러나 국부적인 지하수위변동을 예측하는데는 이용될 수 없다.