LECTURE-9

Modelling of Lake-Aquifer Interactions

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INTRODUCTION

Water bodies are the common features on the land surface and are important in respect of various water uses. Keeping in view their importance in many hydrologic and economic fields, their hydrologic studies are of much use. Field and theoretical studies on the interaction of large water bodies with aquifer are needed based on the variability of flow pattern near the boundaries of the water body. Meyboom (1966) and Freeze and Witherspoon (1967) are amongst the early investigators who stressed upon the groundwater flow pattern around a lake. Winter (1981) has discussed the uncertainties in the water balance of a lake. Some insight into the groundwater regime and discharge estimates from lakes were provided by Winter (1978), who used two- and three- dimensional steady state models for hypothetical groundwater-lake systems. McBride and Pfannkuch (1975) used a numerical model to evaluate the vertical component of groundwater flow of a lake for a number of hypothetical settings. In spite of the early recognition of the lake-water management studies, few attempts (Munter and Anderson, 1981; Winter, 1986; Cheng and Anderson, 1993, 1994) have been made towards the understanding of the interaction of water bodies with aquifer. Munter and Anderson (1981) showed that two- and three-dimensional groundwater flow models provide flexible and effective means of calculating flow rates around lakes. They observed that the anisotropy ratio of aquifer hydraulic conductivity has a significant effect on the simulated head distribution around the lake and magnitude and distribution of seepage from the lake. It is observed that no guidelines or method is available for estimating the recharge from water bodies.

In this paper, a diagnostic curve is developed for estimating the recharge from a water body, from the groundwater heads near the water body. The diagnostic curve is developed using the results of MODFLOW application. Additionally, the proposed diagnostic curve can also be used to estimate the aquifer parameters from measured values of the groundwater head and recharge.

DEVELOPMENT OF DIAGNOSTIC CURVE

MODFLOW (McDonald and Harbaugh, 1988) was used to generate the data for developing the diagnostic curve. The water body was assumed to be of square cross-section having uniform depth. The aquifer was assumed homogeneous and anisotropic. Different sizes of water body and different depths of water in the water body were considered. The anisotropic ratio (K_h/K_v, K_h = hydraulic conductivity in horizontal direction, and K_v = hydraulic conductivity in vertical direction) is considered as 1, 10, 100, 250, 500, and 750. Variable grid spacing with finer grids near the water body was considered. Coarser grids were taken away from the water body.

The difference between the head at a point and the initial head is designated as Δh and the depth of water in the water body is designated as ΔH. The analysis of data obtained using the MODFLOW suggests that the parameter X^2S/Tt (X = distance of observation point from the water body; S = storage coefficient of aquifer; T = transmissivity of aquifer; t = time measured since rise of water level in the water body) and $T\Delta h/Q_R$ (Q_R = rate of recharge from water body) are uniquely related and the relation is found to be the same for different values of ΔH but different for different values of K_h/K_v. Figs. 1, 2 and 3 show such relations (curves) for K_h/K_v = 1, 100 and 500, respectively. The curves for different values of K_h/K_v on double logarithmic paper are made to coincide to the curve for K_h/K_v = 1 with parallel shift of axes. Thus, a new parameter CX^2S/Tt is uniquely related to another new parameter $T\Delta h/(CQ_R)$ and this relation is the same for different values of K_h/K_v.

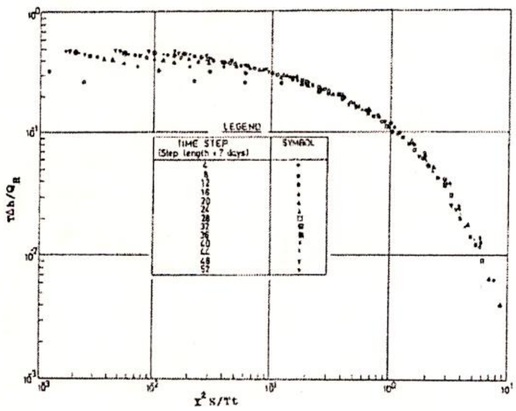

Fig. 1. Variation of X^2S/Tt with $T\Delta h/Q_R$ (K_h/K_v = 1)

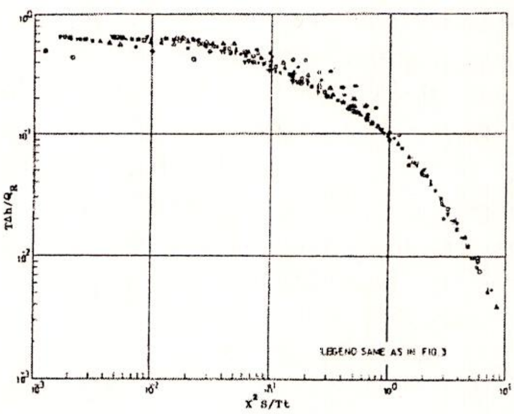

Fig. 2. Variation of X^2S/Tt with $T\Delta h/Q_R$ (K_h/K_v = 100)

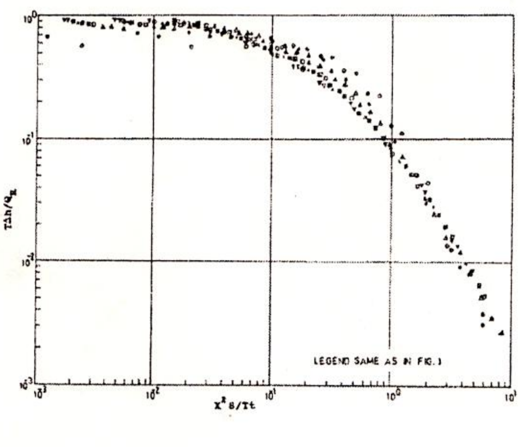

Fig. 3. Variation of X^2S/Tt with $T\Delta h/Q_R$ (K_h/K_v = 500)

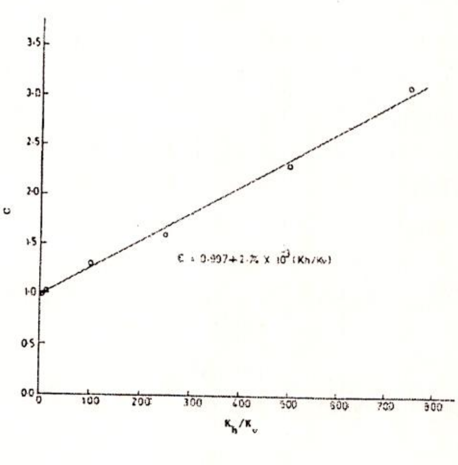

Fig. 4. Variation of C with K_h/K_v

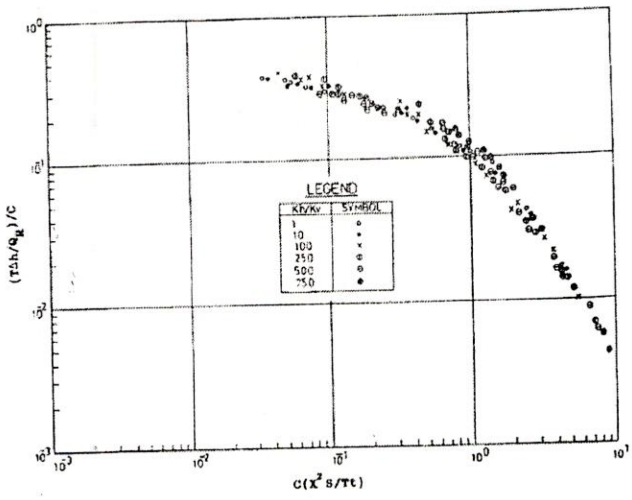

Fig. 5. Diagnostic curve

This unique relation between CX^2S/Tt and $T\Delta h/(CQ_R)$ is termed "diagnostic curve" and is shown in Fig. 5. The variation of C with K_h/K_v is shown in Fig. 4 and is expressed as

$$C = 0.997 + 2.74x10^{-3}\frac{K_h}{K_v} \quad (1)$$

DETERMINING RECHARGE FROM A WATER BODY

The required data are: (1) values of the aquifer parameters T and S; (2) anisotropy factor K_h/K_v; (3) measurement of head changes in an observation well; (4) distance of observation point from the water body. The procedure to calculate the recharge from the water body is:

1. Calculate the value of C from Eq. (1) for known value of K_h/K_v .
2. Calculate the value of the parameter CX^2S/Tt .
3. Knowing the value of the parameter CX^2S/Tt , find the corresponding value of the parameter $T\Delta h/(CQ_R)$ from the diagnostic curve shown in Fig. 5.
4. Once the value of the parameter $T\Delta h/(CQ_R)$ is known, the rate of recharge from the water body, i.e., Q_R is determined.
5. Check for $CX^2S/Tt < 10^{-2}$.

DETERMINING AQUIFER PARAMETERS

The diagnostic curve can be expressed as

$$\frac{1}{C}\frac{T\Delta h}{Q_R} = f(u) \quad (2)$$

$$u = \frac{1}{C}\frac{Tt}{X^2S} \quad (3)$$

Equations (2) and (3) show that a double logarithmic plot of Δh vs t can be made to overlap the plot of $f(u)$ vs u (diagnostic curve) on the same scale with a parallel shift of axes. Once the graphs are matched, the aquifer parameters can be estimated from the dual coordinates of a selected point on the matched portion of the graphs:

$$T = CQ_R \left.\frac{f(u)}{\Delta h}\right|_m \tag{4}$$

$$S = \left.\frac{T}{CX^2}\frac{t}{u}\right|_m \tag{5}$$

where m = subscript denoting the values corresponding to selected point.

CONCLUSION

A diagnostic curve has been developed for estimating recharge from a water body using the groundwater heads near the water body. The proposed diagnostic curve can also be used to estimate the aquifer parameters from the measured values of recharge.

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APPENDIX

Modelling of Lake-Aquifer Interaction

DESCRIPTION OF MODFLOW

Mathematical Model

MODFLOW was originally developed by McDonald and Harbaugh (1988). The partial differential equation governing the three-dimensional unsteady (transient) movement of incompressible groundwater through heterogeneous and anisotropic medium is described by

$$\frac{\partial}{\partial x}\left(K_{xx}\frac{\partial h}{\partial x}\right)+\frac{\partial}{\partial y}\left(K_{yy}\frac{\partial h}{\partial y}\right)+\frac{\partial}{\partial z}\left(K_{zz}\frac{\partial h}{\partial z}\right)-W=S_s\frac{\partial h}{\partial t} \quad (1)$$

where x, y, z = cartesian coordinates aligned along the major axes of conductivities K_{xx}, K_{xx}, and K_{xx}; h = piezometric head [L]; W = volumetric flux per unit volume, which represents sources and/or sinks [T^{-1}]; S_S = specific storage of porous material [L^{-1}]; and t = time [T]. The variables appearing in Eq. (1) are defined in function forms as

$$S_S = S_S(x,y,z) \quad (2a)$$
$$K_{xx} = K_{xx}(x,y,z) \quad (2b)$$
$$K_{yy} = K_{yy}(x,y,z) \quad (2c)$$
$$K_{zz} = K_{zz}(x,y,z) \quad (2d)$$
$$h = h(x,y,z,t) \quad (2e)$$
$$W = W(x,y,z,t) \quad (2f)$$

Thus, in general, the specific storage and conductivities may be the functions of space and the piezometric head and sources/sinks may be the functions of space and time. Eq. (1) when combined with boundary conditions (flow and/or head conditions at the boundary of the aquifer system) and initial conditions (specified head condition s at $t = 0$) constitute a mathematical model of transient flow of groundwater.

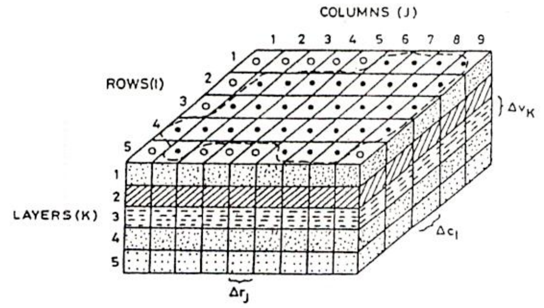
The analytical solution of Eq. (1) is not feasible for complex systems, therefore, numerical methods must be employed to obtain approximate solutions. The finite difference approach is one of such numerical methods, wherein, the continuous system described by Eq. (1) is replaced by a set of discrete points in space and time, and the partial derivatives appearing in Eq. (1) are replaced by finite differences between functional values at these points. Thus, this process leads to a system of simultaneous linear algebraic difference equations, the solution of which yields values of piezometric heads at specific points and time. These values are an approximation to the time varying head distribution that would be given by an analytical solution of the partial differential equation governing the flow process.

Discretization Convention

For the formulation of finite difference equations, the aquifer system needs to be discretized into a mesh of points termed nodes, forming rows, columns, and layers. Such spatial discretization of an aquifer system is shown in Fig. 1. An *i, j, k* coordinate system is used to conform to the computer array convention. If an aquifer system consists of "nrow" rows, "ncol" columns, and "nlay" layers, then

i = row index, i = 1, 2, ...nrow;
j = column index, j = 1, 2, ...ncol; and
k = layer index, k = 1, 2, ...nlay.

For example, Fig. 1 shows a system with nrow = 5, ncol = 9 and nlay = 5. With respect to the Cartesian coordinate system, points along a row are parallel to the *X*-axis, points along a column are parallel to the *Y*-axis, and points along vertical are parallel to the *Z*-axis. In this spatial discretization, nodes represent prisms of porous material termed cells in a conceptual sense. Within each cell, the hydraulic properties are considered constant. Thus, any value associated with or assigned to a node applies to or is distributed over the extent of the cell represented by that node. The width of cells along rows is designated as Δr_j for *j*th column; width of cells along columns is designated as Δc_i for *i*th row; and thickness of layers in vertical is designated as Δv_k for *k*th layer (see Fig. 1). Thus the cell with the coordinates (*i*, *j*, *k*) = (5, 3, 2) has a volume of $\Delta r_3 \Delta c_5 \Delta v_2$.



EXPLANATION

- - - - AQUIFER BOUNDARY

• ACTIVE CELL

o INACTIVE CELL

Δr_J DIMENSION OF CELL ALONG THE ROW DIRECTION (J) INDICATES THE NUMER OF THE COLUMN

Δc_I DIMENSION OF CELL ALONG THE COLUMN DIRECTION SUBSCRIPT (I) INDICATES THE NUMBER OF THE ROW

Δv_K DIMENSION OF CELL ALONG THE VERTICAL DIRECTION SUBSCRIPT(K) INDICATES THE NUMBER OF THE LAYER

Fig. 1. A discretized hypothetical aquifer system

Configuration of Cells

There exist two conventions for defining the configuration of cells with respect to the location of nodes, viz., block-centered and point-centered formulations. In both systems, the aquifer is divided with two sets of parallel lines, which are perpendicular to each other. In block-centered formulation, the blocks formed by the set of parallel lines are the cells and the nodes are at the center of the cells. In point-centered formulation, the nodes are assumed at the intersection points of the set of parallel lines and the cells are drawn around the nodes with faces half way between nodes. In either case of configuration, the spacing of nodes should be such that the hydraulic properties of the system are uniform over the extent of a cell. Both types of grid configurations have been shown in Fig. 2.

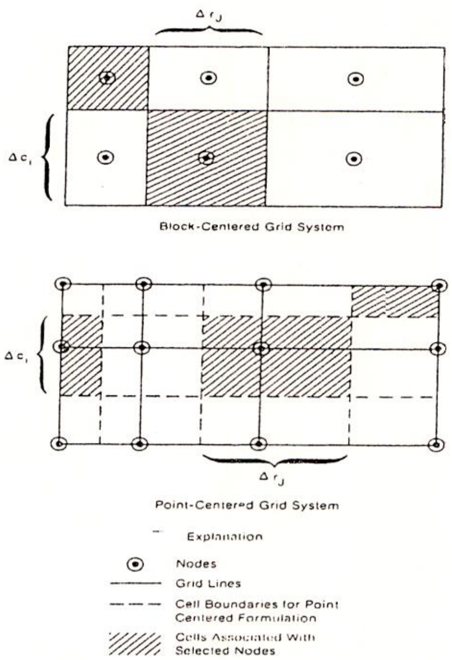


Fig. 2. Grids showing the difference between block-centered and point-centered grids

Finite Difference Equation

The following development of finite difference equation holds good for both type of cell-configurations. The groundwater flow equation may be written in a finite difference form by applying continuity equation. Thus, the algebraic sum of all flows into and out of cell must be equal to the rate of change of storage within the cell. Under the assumption that the groundwater is incompressible, the continuity equation for the flow to a cell can be written as

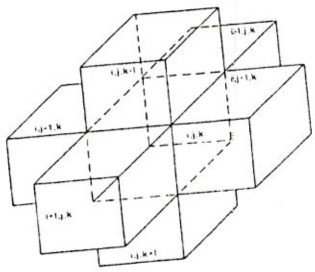
$$\sum Q_i = S_S \frac{\Delta h}{\Delta t} \Delta V \qquad (3)$$

where Q_i = inflow to cell [$L^3 T^{-1}$]; S_S = specific storage defined as the ratio of volume of water, which can be injected per unit volume of aquifer material per unit change in head [L^{-1}]; ΔV = volume of cell [L^3]; and Δh = change in head over a time interval Δt.

The right hand side of Eq. (3) represents the volume of water taken into cell-storage over a time Δt given a change of head of Δh. Outflow and loss in storage are represented by defining outflow as negative inflow and loss as negative gain.

For a three-dimensional problem, each cell is surrounded by six adjacent cells. Fig. 3 show a cell i, j, k along with the six adjacent cells, i -1, j, k; i +1, j, k; i, j -1, k; i, j

+1, k; $i, j, k-1$; and $i, j, k+1$. The net flow to the cell i, j, k is the algebraic summation of the flows into the cell from six adjacent cells.



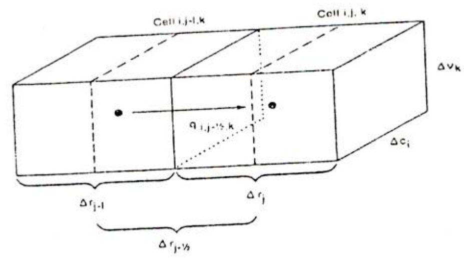


Fig. 3. Cell i,j,k and indices for the adjacent cells

Fig. 4. Flow into cell i,j,k from cell i,j-1,k

Using Darcy's law, the flow from each adjacent cell into the cell i, j, k can be obtained. The flow into the cell i, j, k in row direction from the cell i, j -1, k (see Fig. 4) is given by

$$q_{i,j-1/2,k} = KR_{i,j-1/2,k} \Delta c_i \Delta v_k \frac{h_{i,j-1,k} - h_{i,j,k}}{\Delta r_{j-1/2}} \qquad (4)$$

where $q_{i,j-1/2,k}$ = volumetric flow discharge through the face between cells i, j, k and i, j - 1, k [L^3 T^{-1}]; $KR_{i,j-1/2,k}$ = hydraulic conductivity along the row between nodes i, j, k and i, j - 1, k [LT^{-1}]; $\Delta r_{j-1/2}$ = distance between nodes i, j, k and i, j -1, k [L]. The index $j-1/2$ indicates the space between nodes (Fig. 4). It does not indicate a point exactly half way between nodes. For example, $KR_{i,j-1/2,k}$ represents hydraulic conductivity in the entire region between nodes i, j, k and i, j -1, k. Since the grid dimensions and hydraulic conductivity remain constant throughout the solution process, Eq. (4) may be rewritten by combining the constants into a single constant that is termed as "hydraulic conductance" or simply "conductance" of the cell.

$$q_{i,j-1/2,k} = CR_{i,j-1/2,k}\left(h_{i,j-1,k} - h_{i,j,k}\right) \qquad (5)$$

where

$$CR_{i,j-1/2,k} = \frac{KR_{i,j-1/2,k}\Delta c_i \Delta v_k}{\Delta r_{j-1/2}} \tag{6}$$

The $CR_{i,j-1/2,k}$ is the conductance in *i*th row and *k*th layer between nodes i, j, k and $i, j-1, k$ [$L^2\ T^{-1}$]. Thus, the conductance is defined as the product of hydraulic conductivity and cross-sectional area of flow divided by the length of flow path. Here, *C* refers for the conductance and *R* refers for the row direction.

Equations similar to Eq. (5) can be written approximating the flows into or out of cell i, j, k through the remaining five faces. Such equations are written below.

$$q_{i,j+1/2,k} = CR_{i,j+1/2,k}\left(h_{i,j+1,k} - h_{i,j,k}\right) \tag{7}$$

$$q_{i-1/2,j,k} = CC_{i-1/2,j,k}\left(h_{i-1,j,k} - h_{i,j,k}\right) \tag{8}$$

$$q_{i+1/2,j,k} = CC_{i+1/2,j,k}\left(h_{i+1,j,k} - h_{i,j,k}\right) \tag{9}$$

$$q_{i,j,k-1/2} = CV_{i,j,k-1/2}\left(h_{i,j,k-1} - h_{i,j,k}\right) \tag{10}$$

$$q_{i,j,k+1/2} = CV_{i,j,k+1/2}\left(h_{i,j,k+1} - h_{i,j,k}\right) \tag{11}$$

Eqs. (5) and (7)-(11) represent the flow into the cell i, j, k from all six adjacent cells. There may be some flow to the cell i, j, k from external sources. Seepage from the streambeds, drains, areal recharge, evapotranspiration and flow from wells are sources of external flows. These can be taken care of by additional terms that account for the flows into the cell from outside the aquifer. These flows may depend on the head in the receiving cell or may be independent of the head in receiving cell but are independent of the heads in other cells of the aquifer. The flow from outside the aquifer, which is represented by *W* in Eq. (1), may be represented, in general, as

$$a_{i,j,k,n} = p_{i,j,k}h_{i,j,k} + q_{i,j,k,n} \tag{12}$$

where $a_{i,j,k,n}$ = flow from the *n*th external source into cell i, j, k [$L^3\ T^{-1}$]; $p_{i,j,k,n}$ = a constant [$L^2\ T^{-1}$]; $q_{i,j,k,n}$ = a constant [$L^3\ T^{-1}$]. For example, consider a well in the cell i, j, k with its discharge as $q_{i,j,k,1}$. In this case, the discharge from the well is assumed to be independent of head in the cell i, j, k, hence $p_{i,j,k,1} = 0$ and

$$a_{i,j,k,1} = -q_{i,j,k,1} \tag{13}$$

If a second external source ($n = 2$) is taken to be seepage from a riverbed, which is proportional to the head difference between the river stage (i.e., $R_{i,j,k}$) and receiving cell i, j, k (i.e., $h_{i,j,k}$). Therefore, Eq. (12) takes the form

$$\begin{aligned} a_{i,j,k,2} &= CRIV_{i,j,k,2}\left(R_{i,j,k} - h_{i,j,k}\right) \\ &= -CRIV_{i,j,k,2}h_{i,j,k} + CRIV_{i,j,k,2}R_{i,j,k} \end{aligned} \tag{14}$$

where $CRIV_{i,j,k,2}$ = conductance of riverbed in cell *i, j, k* [$L^2 T^{-1}$]. The term $-CRIV_{i,j,k,2}$ corresponds to $p_{i,j,k,2}$ and the term $CRIV_{i,j,k,2}R_{i,j,k}$ corresponds to $q_{i,j,k,2}$. Similarly, all other external sources or stresses can be represented by Eq. (12). If there are *N* external sources or stresses affecting a single cell, the combined flow is expressed by

$$\begin{aligned} QS_{i,j,k} &= \sum_{n=1}^{N} a_{i,j,k,n} \\ &= \sum_{n-1}^{N} P_{i,j,k,n}h_{i,j,k} + \sum_{n=1}^{N} q_{i,j,k,n} \\ &= P_{i,j,k}h_{i,j,k} + Q_{i,j,k} \end{aligned} \quad (15)$$

where

$$P_{i,j,k} = \sum_{n=1}^{N} p_{i,j,k,n} \quad (16)$$

$$Q_{i,j,k} = \sum_{n=1}^{N} q_{i,j,k,n} \quad (17)$$

While writing the continuity equation of the form given by Eq. (3), for the cell *i, j, k*, the tem $\sum Q_i$ consists of flows to the cell from six adjacent cells, and all external flow rates to the cell. The flows from six adjacent cells into the cell *i, j, k* are given by Eqs. Eqs. (5) and (7)-(11). The flow from external sources into the cell *i, j, k* is given by Eq. (15). Substituting these equations into Eq. (3), we get

$$\begin{aligned} &CR_{i,j-1/2,k}\left(h_{i,j-1,k} - h_{i,j,k}\right) + CR_{i,j+1/2,k}\left(h_{i,j+1,k} - h_{i,j,k}\right) + \\ &CC_{i-1/2,j,k}\left(h_{i-1,j,k} - h_{i,j,k}\right) + CC_{i+1/2,j,k}\left(h_{i+1,j,k} - h_{i,j,k}\right) + \\ &CV_{i,j,k-1/2}\left(h_{i,j,k-1} - h_{i,j,k}\right) + CV_{i,j,k+1/2}\left(h_{i,j,k+1} - h_{i,j,k}\right) + \\ &P_{i,j,k}h_{i,j,k} + Q_{i,j,k} = S_{S\,i,j,k}\Delta r_j \Delta c_i \Delta v_k \frac{\Delta h_{i,j,k}}{\Delta t} \end{aligned} \quad (18)$$

where $S_{S_{i,j,k}}$ = specific storage of cell *i, j, k* [L^{-1}]; and $\Delta r_j \Delta c_i \Delta v_k$ = volume of cell *i, j, k* [L^3]. The Eq. (18) can be written in the backward difference form by specifying the flow terms at time t_m, the end of the time interval, and approximating the temporal derivative of head over the time interval t_{m-1} to t_m, i.e.,

$$\begin{aligned} &CR_{i,j-1/2,k}\left(h^m_{i,j-1,k} - h^m_{i,j,k}\right) + CR_{i,j+1/2,k}\left(h^m_{i,j+1,k} - h^m_{i,j,k}\right) + \\ &CC_{i-1/2,j,k}\left(h^m_{i-1,j,k} - h^m_{i,j,k}\right) + CC_{i+1/2,j,k}\left(h^m_{i+1,j,k} - h^m_{i,j,k}\right) + \\ &CV_{i,j,k-1/2}\left(h^m_{i,j,k-1} - h^m_{i,j,k}\right) + CV_{i,j,k+1/2}\left(h^m_{i,j,k+1} - h^m_{i,j,k}\right) + \end{aligned}$$

$$P_{i,j,k} h^{m}_{i,j,k} + Q_{i,j,k} = S_{S\,i,j,k} \Delta r_j \Delta c_i \Delta v_k \frac{h^{m}_{i,j,k} - h^{m-1}_{i,j,k}}{t_m - t_{m-1}} \quad (19)$$

An equation similar to Eq. (19) can be written for each of the n cells in the system. Since there is only one unknown head for each cell, we are left with a system of n equations and n unknowns. Such a system of equations can be solved simultaneously to get the values of head at all cells.

Boundary Condition and Initial Condition

The type of boundaries that may be imposed in the model include constant head, no-flow, constant flow, and head dependent flow. Different types of boundaries are represented by the different types of cells. Cells may be designated as "inactive cell," "constant head cell," and "variable head cell." The variable head cells are those in which the head varies with time. Therefore, an equation similar to Eq. (19) is required for each variable head cell. The head remains constant with time in the constant head cells and these cells do not require an equation similar to Eq. (19), however, the adjacent variable head cells will contain non-zero conductance terms representing flow from the constant head cell. The "no flow cells" are those to which there is no flow from adjacent cells. Neither an equation is formulated for a no flow cell nor the equations for the adjacent cells contain a term representing flow from the no flow cell. The use of no-flow and constant-head cells to simulate boundary conditions is given in Fig. 5. The constant-flow and head-dependent flow boundaries can be represented by a combination of no-flow cells and external sources.

Solution

In most cases, the actual number of equations of the form of Eq. (19) would be less than the total number of model cells. This is because the number of equations is only equal to the number of variable head cells. The objective of the transient simulation is to predict the head distribution at successive times for the given initial head distribution and boundary conditions. The initial head distribution consists of a value of $h^{1}_{i,j,k}$ at each point in the mesh at the beginning of the first time step. The time is discretized into a number of discrete time steps for the finite difference process. The first step of the solution process is to calculate values of $h^{2}_{i,j,k}$, i.e., head at the end of first time step. Therefore, in Eq. (19), the subscript m is taken as 2 and the subscript m -1 is taken as 1. Once such equations are formed for each variable head cells, an iterative method is used to obtain the values of $h^{2}_{i,j,k}$. An iterative method starts with an initial trial solution. This trial solution is used to calculate through a procedure of calculations, an interim solution that more nearly satisfies the system of equations. The interim solution then becomes a new trial solution and this procedure is repeated. Each repetition is called iteration. The process is repeated until the trial and interim solutions are nearly equal, i.e., for each node, the difference between the trial head values and the interim head values is smaller than some arbitrary established value, usually termed as "closure criterion." The interim solution is then regarded as a good approximation to the solution of the system of equations under given initial and boundary conditions.

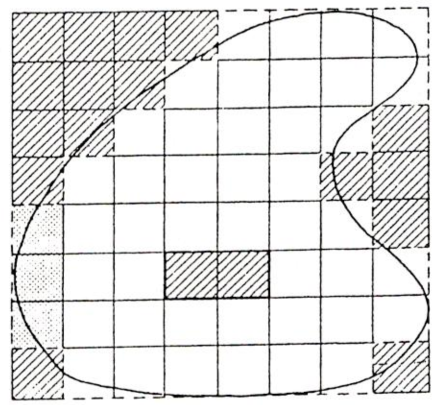


Fig. 5. Discretized aquifer showing boundaries and constant head cells

Other Applications

The MODFLOW can also be used to solve the isolated problems of groundwater flow, the analytical solutions for which are not available. Singh (1990, 1997) has made efforts to solve a couple of such problems.

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