TRAINING COURSE

ON

SOFTWARE FOR GROUNDWATER DATA MANAGEMENT

UNDER

WORLD BANK FUNDED HYDROLOGY PROJECT

LECTURE NOTES

ON

BASICS OF GROUNDWATER

(UNIT-1)

BY

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BASICS OF GROUNDWATER

1.0 SYSTEM CONCEPT

Hydrologic phenomena are extremely complex, and may never be fully understood. However, in the absence of perfect knowledge, they may be represented in a simplified way by means of the systems concept. A system is a set of connected parts that form a whole. The hydrologic cycle may be treated as a system whose components are precipitation, evaporation, runoff, infiltration, evapotranspiration, groundwater and other phases of the hydrologic cycle. These components can be grouped into sub systems of the overall cycle; to analyze the total system, the simpler subsystems can be treated separately and the results combined according to the interactions between the subsystems (Chow et. al., 1988).

In Fig.1, the global hydrologic cycle is represented as a system. The dashed lines divide it into three subsystems: the atmospheric water system containing the processes of precipitation, evaporation, interception, and transpiration; the surface water system containing the processes of overland flow, surface runoff, subsurface and groundwater outflow, and runoff to streams and the ocean; and the subsurface water system containing the processes of infiltration, groundwater recharge, subsurface flow and groundwater flow. Subsurface flow takes place in the soil near the land surface; groundwater flow occurs deeper in the soil or rock strata.

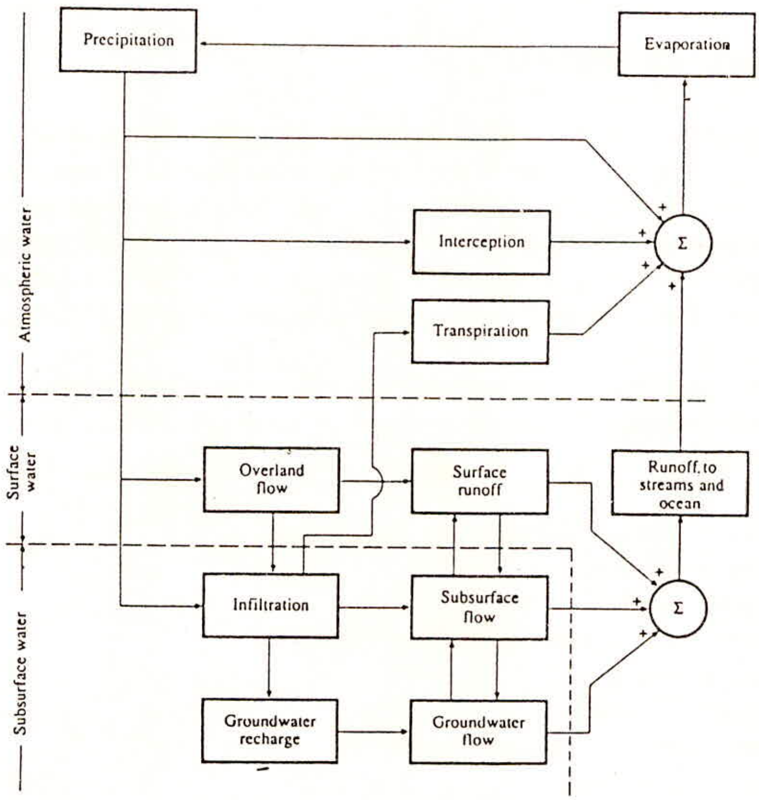


Fig.1 Block diagram representation of the global hydrologic system

In fluid mechanics, the application of the basic principles of mass, momentum, and energy to a fluid flow system is accomplished by using a control volume, a reference frame drawn in three dimensions through which the fluid flows. The control volume provides the framework for applying the laws of conservation of mass and energy and Newton's second law to obtain practical equations of motion.

By analogy, a hydrologic system is defined as a structure (for surface or subsurface flow) or volume in space (for atmospheric moisture flow), surrounded by a boundary, that accepts water and other inputs, operates on them internally, and produces them as outputs. The boundary is a continuous surface defined in three dimensions enclosing the volume or structure. A working medium enters the system as input, interacts with the structure and other media, and leaves as output. Physical, chemical, and biological processes operate on the working media within the system; the most common working media involved in hydrologic analysis are water, air, and heat energy.

The procedure of developing working equations and models of hydrologic phenomena is similar to that in fluid mechanics. In hydrology, however, there is generally a greater degree of approximation in applying physical laws because the systems are larger and more complex, and may involve several working media. Also, most hydrologic systems are inherently random because their major input is precipitation, a highly variable and unpredictable phenomenon. Consequently, statistical analysis plays a large role in hydrologic analysis.

2.0 GROUNDWATER SYSTEM

The groundwater system may be imagined as a huge natural reservoir or system of reservoirs in rocks whose capacity is the total volume of pores or openings that are filled with water. Groundwater may be found in one continuous body or in several distinct rock or sediment layers at any one location. Thickness of the groundwater zone is governed by the local geology, availability of pores or openings in the rock formation, recharge, and movement of water from areas of recharge toward points or areas of discharge.

It is nearly impossible to adequately summarize all types of geologic environments in which water can exit, but the list below presents some typical types of openings found in rocks (Driscoll, 1987).

1. Intergrain pores in unconsolidated sand and gravel
2. Intergrain pores in sandstone
3. Intergrain pores in shale
4. Systematic joints in metamorphic and igneous rocks
5. Cooling fractures in basalt
6. Solution cavities in limestone
7. Gas-bubble holes and lava tubes in basalt
8. Systematic joints in limestone
9. Openings in fault zones

Unfortunately, rock masses are rarely homogeneous and adjacent rock types may vary significantly in their ability to hold water. Nevertheless, intelligent groundwater assessment or use requires an understanding of how water exists in each type of rock or sediment medium.

A mathematical model describing the groundwater system must incorporate the following aspects (Keshari, 1994):

i. geometry of the aquifer boundary and its physical domain under investigation (large scale, smaller scale or individual well project)

ii. nature of the porous medium (heterogeneity, anisotrophy, porosity, hydraulic conductivity, storativity)

iii. mode of flow in the aquifer (3-dimensional, 2-dimensional or 1-dimensional)

iv. flow regime (laminar or turbulent)

v. relevant state variables and the area or volume over which the averages of such variables are taken (variables: hydraulic head, solute concentration, temperature)

vi. sources and sinks of water and of relevant pollutants within the aquifer domain and on its boundaries (point or distributed sources and sinks)

vii. boundary conditions of the aquifer domain under investigation (Dirichlet, Neumann, Cauchy)

viii. initial conditions for the flow and transport processes (in transient cases)

ix. nature of pollutant (conservative, radioactive, degradable or adsorbent)

x. dispersivities

The essential components of a mathematical model simulating flow and transport processes in an aquifer are:

i. definition of the geometry of the considered domain and its boundaries

ii. equations governing the flow and transport processes incorporating constitutive relationships (Darcy's law, Fick's law, decay law, adsorption isotherm, law of mass action)

iii. initial conditions for the flow and transport processes (in transient cases)

iv. boundary conditions of the aquifer domain under investigation for the flow and transport processes

3.0 AQUIFERS

An aquifer is a saturated bed, formation, or group of formations which yields water in sufficient quantity to be economically useful. Water-bearing formations and groundwater reservoirs are synonyms for the word aquifer. To be an aquifer, a geologic formation must contain pores or open spaces (both of these are often called interstices) that are filled with water. These interstices must be large enough to transmit water toward wells at a useful rate.

A non-dimensional parameter, K_c , designated as conductivity class has been defined (Bear and Verruijt, 1987) by the relation

$$K_c = -[2 + \log_{10} K]$$

in which K is the hydraulic conductivity in m/s. K_c is used as an index to classify the aquifers and the porous media as given in Table-1.

Table-1 : Classification of porous media/aquifer using conductivity class (K_c)

Porous media/Aquifer	K_c
Pervious	-2 to 2
Semipervious	2 to 6
Impervious	6 to 11
Good aquifer	-2 to 3
Poor aquifer	3 to 7
No aquifer	7 to 11

Both the size of pores and the total number of pores in a formation can vary remarkably, depending on the types of material and the geologic and chemical history. Individual pores in a fine-grained sediment such as clay are extremely small, but the combined volume of the pores can be unusually large. Subsequent compaction of clay reduces the pore space considerably. Although clay has a large water holding capacity, water cannot move readily through the tiny open spaces. This means that a clay formation under normal conditions will not yield water to wells, and therefore it is not an aquifer even though it may be water-saturated.

Ordinarily a clay or shale formation is nearly impermeable and is called an aquiclude, or a formation through which virtually no water moves. Formations which do yield some water, but usually not enough to meet even modest demands, are called aquitards. In reality, almost all formations will yield some water, and therefore are classified as either aquifers or aquitards. In water-poor areas, a formation producing small quantities of water may be called an aquifer, whereas the same formation in a water-rich area would be an aquitard.

Aquifers have two main functions in the underground phase of the water cycle. They store water for varying periods in the underground reservoir, and they act as pathways or conduits to pass water along through the reservoir. Although some are more efficient as pipelines (e.g., cavernous limestones) and some are more effective as storage reservoirs (e.g., sandstones), most aquifers perform both functions continuously.

3.1 TYPE OF AQUIFERS

Confined and Unconfined Aquifers

Aquifers may be classified as unconfined or confined depending on the presence or absence of a water table. For an unconfined aquifer a water table serves as the upper surface of the zone of saturation. The water table is defined as "... that surface in the groundwater body at which the water pressure is atmospheric." Unconfined groundwater, then, is water in an aquifer that has a water table in contact with the atmosphere through pores in the unsaturated soil above. Unconfined aquifers are sometimes called water table aquifers.

Confined groundwater, on the other hand, is water under pressure greater than atmospheric pressure. The upper boundary of a confined aquifer is an essentially impermeable formation that "traps" or "confines" water in the aquifer, sealing it off from the atmosphere (Fig. 2).

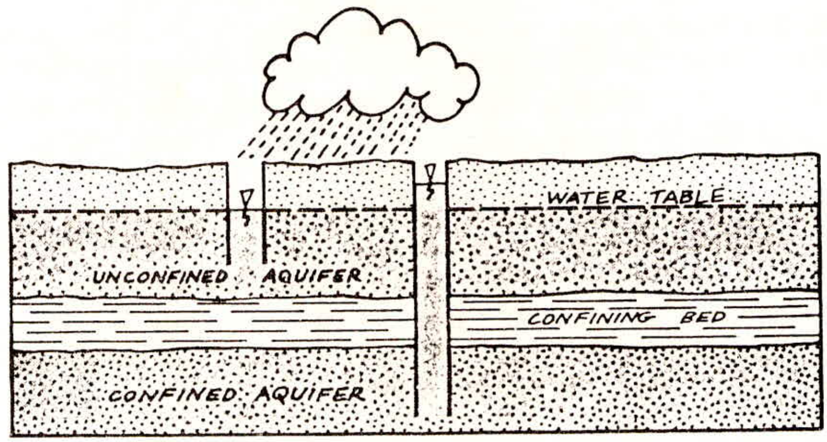


Fig.2 Typical occurrence of groundwater in confined and unconfined aquifers

When a well is drilled in an unconfined aquifer, water will remain approximately at the same level where it is first encountered. In wells tapping a confined aquifer, the water will rise in the well when it is first encountered during drilling, and will stand at a level above the top of the aquifer.

Water in a confined aquifer is under pressure because the aquifer is at a higher elevation in the recharge area than it is at the well location. Depending on local conditions, water from a confined aquifer will even rise up in the well until it flows out at the surface without the aid of a pump. Traditionally such a well is called an artesian well. Confined aquifers are often called artesian aquifers.

Storage in Unconfined Aquifers

An unconfined aquifer's pore space contains both live storage and dead storage. Gravity water makes up the live storage and capillary water the dead storage. When the water table rises, it saturates the overlying capillary fringe; when it falls, it leaves capillary water suspended above it. In either case only gravity water actually moves into or leaves storage in the aquifer, causing water levels in wells to rise or fall. For an unconfined aquifer, the ratio of aquifer volume holding capillary water is called specific retention; the ratio of aquifer volume containing gravity water that is free to move in or out of the pores is called specific yield. Specific retention plus specific yield equals porosity, which is the ratio of total open space to total volume of the aquifer.

Storage in Confined Aquifers

The implications for changes in storage as water levels fluctuate are very different for confined versus unconfined aquifers. In an unconfined aquifer, lowering the water table actually dewaters the upper part of the aquifer and reduces the volume of saturated material, but the aquifer and the water undergo essentially no physical changes. When water levels fall in wells tapping a confined aquifer, the pressure falls, but the aquifer remains completely saturated. The change in pressure in a confined aquifer affects both the water and the aquifer.

The aquifer responds to pressure change by expanding or contracting slightly as water levels rise or lower. This produces a small increase or decrease in porosity, and hence in storage space. Water, although usually thought of as being noncompressible, will also contract or expand slightly in response to pressure changes. Thus, when water is removed from storage, well levels fall, pressure in the aquifer falls, water expands slightly, and the aquifer contracts slightly. When water is added to storage, the reverse processes take place: well levels rise, pressure in the aquifer rises, water contracts slightly, and the aquifer expands slightly. Because the change in aquifer porosity or in water volume per unit change in pressure is very small, the numerical value of the storage coefficient for an artesian aquifer is much smaller than that for a water-table aquifer. The range in values for the storage coefficient in water-table aquifers is around 0.01-0.30, while the numbers for artesian aquifers are more like 0.00005-0.005 (Manning, 1989).

4.0 ANISOTROPY

If the properties of the soil that are responsible for the resistance to flow are independent of the direction such a material is said to be isotropic with regard to permeability. Not every soil possesses that property, however. In many soil deposits the resistance to flow in the vertical direction is considerably larger than the resistance to horizontal flow, due to the presence of layered structure in the soil, generated by its geological history. For such anisotropic porous media Darcy's law has to be generalised. The proper generalisation is, in terms of hydraulic conductivity (Verruijt, 1982)

$$q_x = -k_{xx}\frac{\partial\phi}{\partial x} - k_{xy}\frac{\partial\phi}{\partial y} - k_{xz}\frac{\partial\phi}{\partial z}$$
$$q_y = -k_{yx}\frac{\partial\phi}{\partial x} - k_{yy}\frac{\partial\phi}{\partial y} - k_{yz}\frac{\partial\phi}{\partial z} \qquad \ldots\ldots(1)$$
$$q_z = -k_{zx}\frac{\partial\phi}{\partial x} - k_{zy}\frac{\partial\phi}{\partial y} - k_{zz}\frac{\partial\phi}{\partial z}$$

The quantities q_x, q_y and q_z are the three components of the specific discharge vector q, where specific discharge vector denotes the discharge through a certain area of soil divided by that area. $\partial\phi/\partial x$, $\partial\phi/\partial y$ and $\partial\phi/\partial z$ are components of the hydraulic head gradient in x, y and z direction respectively.

These equations express the most general linear relationship between the specific discharge vector and the gradient of the groundwater head. The coefficients k_{xx} ,....,k_{zz} are said to be components of a second-order tensor. It is usually assumed, on the basis of thermodynamic considerations, that this is a symmetric tensor (that is, $k_{xy}=k_{yx}$, $k_{yz}=k_{zy}=k_{zy}$, $k_{zx}=k_{xz}$). It can be shown that this means that there exist three mutually orthogonal directions, the so-called principal directions of permeability, in which the cross-components vanish. Physically speaking this means that a gradient of the groundwater head in one of these directions leads to a flow in that same direction.

Consider the two-dimensional case of an anisotropic porous medium consisting of two orthogonal systems of channels of different cross-section, and hence of different resistance (Fig.3).

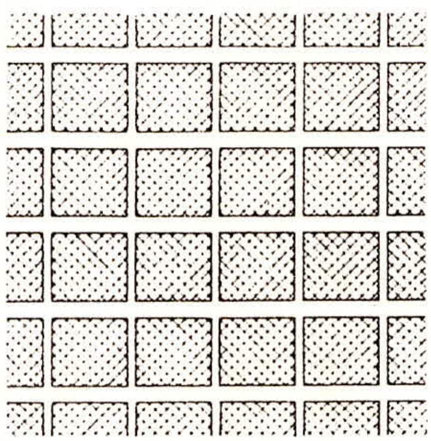


Fig. 3 Anisotropic porous medium

In this case the principal directions coincide with the directions of the channels, and one may write

$$q_x = -k_{xx}\frac{\partial\phi}{\partial x}$$
$$q_y = -k_{yy}\frac{\partial\phi}{\partial y} \quad \ldots.(2)$$

Where, $\phi = p/\gamma_w + z$ = the groundwater head
p = pressure
γ_w = unit weight of water
z = elevation head (the vertical z-axis is positive upwards).

If the channels in the x-direction are wider than those in the y-direction, the permeability k_{xx} will be greater than k_{yy}. Now consider the same situation being described with respect to coordinates ξ and η, which are obtained from x and y by rotation through an angle α.

$$\xi = x \cos\alpha + y \sin\alpha$$
$$\eta = y \cos\alpha - x \sin\alpha \quad \ldots.(3)$$

A vector q with components q_x and q_y can also be decomposed into components q_ξ and q_η , where

$$q_\xi = q_x \cos\alpha + q_y \sin\alpha$$
$$q_\eta = q_y \cos\alpha - q_x \sin\alpha \quad \ldots.(4)$$

incorporating equation (2) in (4)

$$q_\xi = -k_{xx}\cos\alpha\frac{\partial\phi}{\partial x} - k_{yy}\sin\alpha\frac{\partial\phi}{\partial y}$$
$$q_\eta = -k_{yy}\cos\alpha\frac{\partial\phi}{\partial y} + k_{xx}\sin\alpha\frac{\partial\phi}{\partial x} \quad \ldots.(5)$$

The partial derivatives $\partial\phi/\partial x$ and $\partial\phi/\partial y$ can be related to $\partial\phi/\partial\xi$ and $\partial\phi/\partial\eta$ with the aid of equation 3. This gives

$$\frac{\partial\phi}{\partial x} = \frac{\partial\phi}{\partial\xi}\frac{\partial\xi}{\partial x} + \frac{\partial\phi}{\partial\eta}\frac{\partial\eta}{\partial x} = \frac{\partial\phi}{\partial\xi}\cos\alpha - \frac{\partial\phi}{\partial\eta}\sin\alpha$$
$$\frac{\partial\phi}{\partial y} = \frac{\partial\phi}{\partial\xi}\frac{\partial\xi}{\partial y} + \frac{\partial\phi}{\partial\eta}\frac{\partial\eta}{\partial y} = \frac{\partial\phi}{\partial\xi}\sin\alpha + \frac{\partial\phi}{\partial\eta}\cos\alpha \quad \ldots.(6)$$

Using these expressions the equations (5) can be written as

$$q_\xi = -k_{\xi\xi}\frac{\partial\Phi}{\partial\xi} - k_{\xi\eta}\frac{\partial\Phi}{\partial\eta}$$
$$q_\eta = -k_{\eta\xi}\frac{\partial\Phi}{\partial\xi} - k_{\eta\eta}\frac{\partial\Phi}{\partial\eta} \qquad ....(7)$$

where,

$$k_{\xi\xi} = k_{xx}\cos^2\alpha + k_{yy}\sin^2\alpha = \frac{1}{2}(k_{xx} + k_{yy}) - \frac{1}{2}(k_{xx} - k_{yy})\cos 2\alpha$$
$$k_{\eta\eta} = k_{yy}\cos^2\alpha + k_{xx}\sin^2\alpha = \frac{1}{2}(k_{xx} + k_{yy}) + \frac{1}{2}(k_{yy} - k_{xx})\cos 2\alpha \qquad ..(8)$$
$$k_{\eta\xi} = k_{\xi\eta} = (k_{yy} - k_{xx})\sin\alpha\cos\alpha = \frac{1}{2}(k_{yy} - k_{xx})\sin 2\alpha$$

Equations (7) are the description of a general flow, using the coordinates ξ and η. It is interesting to note the appearance of the cross-coefficients $k_{\xi\eta}$ and $k_{\eta\xi}$ vanish only if $k_{xx} = k_{yy}$ (when the soil is isotropic) or if $\alpha = 0$, π etc. (when the ξ, η -coordinates coincide with x and y). It now follows that a gradient of the groundwater head in ξ-direction not only leads to a flow in that direction, but also to a flow in η-direction. This can be realised physically by noting (see Fig.3) that in that case the channels in x-direction will transport much more water than the narrow channels in y-direction (because $k_{xx} > k_{yy}$). This means that the resultant flow will always have a tendency towards the most permeable direction. It also means that in general anisotropy cannot be formulated by simply using different coefficients in the three directions in which Darcy's law is formulated. In general the anisotropy law should be of the form of equation (1), with six independent coefficients. Fortunately in engineering practice it is usually acceptable to distinguish only between the permeability in vertical direction and one in horizontal direction, assuming that this difference has been created during the geological process of deposition of the soil. Then it may assumed that the x,y and z-directions are principal directions (if the z-axis is vertical), with $k_{xx} = k_{yy} = k_h$ and $k_{zz} = k_v$. Darcy's law can then be used in the form

$$q_x = -k_h\frac{\partial\Phi}{\partial x}$$
$$q_y = -k_h\frac{\partial\Phi}{\partial y} \qquad ....(9)$$
$$q_z = -k_v\frac{\partial\Phi}{\partial z}$$

which involves only two coefficients. They must be measured by doing two independent tests.

5.0 CONSTITUTIVE EQUATIONS FOR GROUND WATER FLOW

For the steady state flow of viscous, incompressible fluids (at small Reynolds numbers), the Navier-Stokes equations of motion, the most general equation governing fluid flow, in which conservation of momentum is embedded, reduce in form to generalised statement of Darcy's law. The mass balance equation (continuity equation) when combined with Darcy's law yields the equation of flow.

The equations of flow are partial differential equations with respect to time and space. They are of parabolic type. Both initial conditions in the whole modelled domain and boundary conditions on the boundary must be given for their solution. The initial conditions consist of the known head distribution at an initial time t_o from which the simulation is supposed to start. There are three possible types of boundary conditions which may apply to any part of the boundary of the modelled domain (Kinzelbach, 1986).

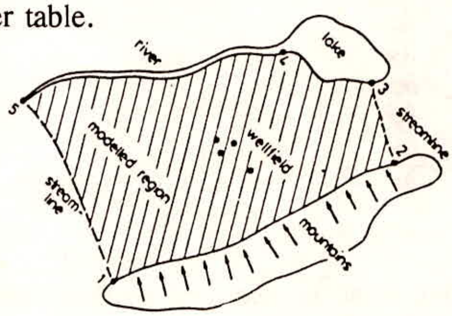
- Boundary conditions of the first kind (Dirichlet type) prescribe the head value. In a modelled domain there should be at least one point that constitutes a first-kind boundary. This is necessary to guarantee the uniqueness of the solution.

- Boundary conditions of the second kind (Neumann type) specify the boundary flux, which means the head gradient normal to the boundary. A special case of this type of boundary is the impervious boundary where the flux is zero. If streamlines form boundaries of the modelled domain they are treated as impervious boundaries. Wells can be viewed as inner boundaries of the second kind by cutting out a circle around the well and specifying the flow across the circle. In view of the discretization in mathematical models, wells are considered as distributed recharge and discharge sources and not as boundaries.

- Boundary conditions of the third kind (semipervious boundary, mixed boundary conditions) specify a linear combination of head and flux at a boundary. They are used at semipervious (leakage) boundaries. If a leaky river forms the boundary the appropriate condition is of the form:

$$Q = \Gamma_r (h_r - h) \qquad \text{....(10)}$$

An example for the specification of boundary conditions in a aquifer is shown in Fig.4. At the edge of a shallow phreatic aquifer the boundary may be moving with the rising or falling water table.



boundary 1-2: prescribed flux (nonzero)
boundary 2-3: zero flux
boundary 3-4: prescribed head
boundary 4-5: semipervious
boundary 5-1: zero flux

Fig.4 Example of boundary condition in a flow model

While the equation of the confined aquifer is linear, the equation of the phreatic aquifer is nonlinear. This does not present major problems for the numerical solution. In cases where the spatial and temporal variations of head are small compared to the thickness of saturated flow, the equation can be linearized. The phreatic aquifer may in that case be described by the confined aquifer equation.

6.0 GROUNDWATER FLOW MODELS

Groundwater flow in aquifers is very often described by equations obtained by combining Darcy's law, as a dynamic equation with the continuity equation, adapted to the specific hydrogeological conditions of the system being investigated. The most frequently employed forms of the combined equations for two-dimensional flow modelling in aquifers are as given below. Fig.5 shows a definition sketch of semi-confined aquifer.

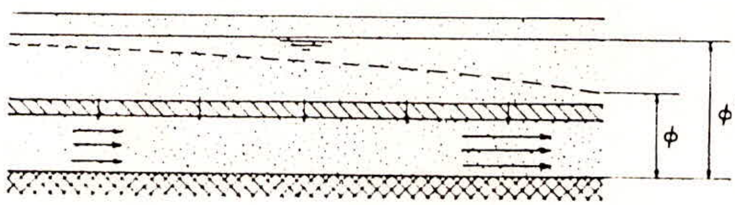


Fig.5 Semi-confined aquifer

For steady flow:

$$\frac{\partial}{\partial x}\left(T\frac{\partial \phi}{\partial x}\right) + \frac{\partial}{\partial y}\left(T\frac{\partial \phi}{\partial y}\right) - \frac{\phi' - \phi}{c} = C \qquad \ldots.(11)$$

For unsteady flow:

$$\frac{\partial}{\partial x}\left(T\frac{\partial \phi}{\partial x}\right) + \frac{\partial}{\partial y}\left(T\frac{\partial \phi}{\partial y}\right) + q = S\frac{\partial \phi}{\partial t} \qquad \ldots.(12)$$

where,

T = transmissivity;
ϕ = groundwater head;
S = specific storage coefficient;
q = discharge (sources and sinks);
x,y = horizontal co-ordinates.
c = resistance of confining layer (Fig. 5)

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CURRENT METHODS OF GROUND WATER ASSESSMENT IN THE COUNTRY

1.0 GROUNDWATER ESTIMATION COMMITTEE NORMS

Presently the ground water potential is assessed using the Groundwater Estimation Committee norms (GEC 1984). The estimation should be made using a river basin as a unit. For the block level water balance, it has been suggested to consider the subsurface inflow and outflow making use of the ground water level contour maps of the region of which the block is a part.

For calculating the annual recharge during monsoon, the following formula is used:

$$\text{Monsoon Recharge} = (\Delta S + DW - Rs - Rigw - Ris) \times \frac{Normal\ monsoon\ RF}{Annual\ monsoon\ RF} + Rs + Ris$$

where,

ΔS = change in ground water storage volume during pre- and post-monsoon periods (April/ May to November).

DW =Gross ground water draft during monsoon.

Rs = Recharge from canal seepage during monsoon.

Rigw = Recharge from recycled water from groundwater irrigation during monsoon.

Ris = Recharge from recycled water from surface water irrigation during monsoon.

RF = Rainfall.

The change in ground water storage volume during pre- and post-monsoon periods is computed as:

ΔS = [Area] x [water level fluctuation] x [specific yield]

The area not suitable for recharge like hilly and saline areas are excluded.

The annual replenishable ground water recharge is computed as follows:

Total annual recharge = Recharge during monsoon + Non-monsoon rainfall recharge + Seepage from canals + Return flow from irrigation + Inflow from influent rivers + Recharge from submerged lands, lakes etc.

The following points should be taken care of while using the norms:

(1) The change in groundwater storage over an interval of time is governed by the subsurface inflows and outflows to/from the domain besides recharge from different sources. If the control volume is not a watershed, but a block, the subsurface inflow and outflow should be considered as suggested in the norms.

(2) The term ΔS is governed by all the inflows to the domain as well as all the outflows from the flow domain. Therefore, the recharge from influent rivers and submerged lands during monsoon are already considered while computing ΔS. Therefore, influent seepage from rivers during non-monsoon and recharge from submerged lands during non-monsoon should be accounted while computing annual recharge.

(3) The ground water balance should be cross checked with the base flow from the sub basin/ basin.

(4) A watershed should be considered as a unit for proper evaluation of the components of water balance.

(5) According to Chaturvedi formula applicable for alluvial areas in Indo Gangetic plain

$$\text{Recharge from rainfall(mm)} = 13.93[RF(mm)-381]^{2/5}$$

This shows that the recharge is a nonlinear function of rainfall.

However the normalisation procedure assumes the relation to be linear. The normalisation procedure overestimates ground water potential for rainfall less than the normal.

(6) The specific yield values recommended are:

Sandy alluvial area	12 to 18 %
Silty clayey alluvial areas	5 to 12%
Valley fills	10 to 14%
Granites	2 to 4%
Basalts	1 to 3%
Laterite	2 to 4%
Sand stone	1 to 8%
Weathered phyllites,shales, Schist & associated rocks	1 to 3%

The crux of the estimation of groundwater potential is the specific yield used for the computation. If the above values are used without field verification, the computed values will be either very much under estimated or over estimated. For example, for silty clayey material, one may over estimate upto 140% or under estimate upto 58% depending upon the specific yield value adopted. Therefore the specific yield values to be adopted should be cross checked by carrying out water balance for nonmonsoon season.

(7) The seepage losses from canal should be estimated considering the prevalent potential causing seepage.

(8) It has been assumed that the effluent flow to a stream from an aquifer is a part of the groundwater potential and this can be exploited. This may not be feasible because of minimum flow requirements from environmental consideration. Part of the groundwater potential should be left in coastal aquifer to prevent saline intrusion.

(9) Transpiration from forested area needs to be considered in the water balance.

Since the specific yield varies from 1 to 30%, it is advisable to use an alternate method which does not make use of water table fluctuation and specific yield for the computation of the recharge. An alternate method is suggested in the following paragraphs.

2.0 ESTIMATION OF GROUNDWATER RECHARGE FROM RAINFALL BY SOIL-WATER-BALANCE METHOD

Based on the conceptual model of Ben Zvi and Goldstoff (vide Bear, 1979), a soil-water-balance method is suggested to compute recharge from rainfall. SCS method has been used to compute initial abstraction (Hawkins, 1978). Surface runoff, which is contributed only from the saturated part of a watershed, has been estimated by the method proposed by Manely(1977) which accounts for spatial variability of infiltration in a watershed. The area bound by a horizontal line representing the excess daily rainfall over initial abstraction, and the graph of saturated hydraulic conductivity with cumulative area fraction, defines the area contributing to the runoff for a particular rainfall event. Rainfall minus initial abstraction minus surface runoff is the quantity which infiltrates to the root zone. The soil moisture in excess to field capacity for a rainfall event is taken as deep percolation and regarded as the recharge to groundwater from rainfall. Evaporation which controls the initial abstraction and evapotranspiration which depletes soil moisture in the root zone have been accounted in the model.

Data required:

(i) Daily rainfall,
(ii) Daily potential evaporation,
(iii) Total area of the watershed,
(iv) Build-up area (roads, buildings etc.),
(v) Land use map and treatment/practice,
(vi) Maximum root zone depth, D_0
(vii) Field capacity, θ_f
(viii) Initial root zone soil moisture, θ (t - Δt)
(ix) Infiltration capacity at a number of places in the watershed.

Step-by-Step Procedure

1. Find hydrologic soil groups in the watershed.

A: Infiltration capacity = 7.5 - 11.5 cm/hr,
B: Infiltration capacity = 4.0 - 7.5 cm/hr;
C: Infiltration capacity = 0.13- 4.0 cm/hr,
D: Infiltration capacity = 0 - 0.13 cm/hr,,

2. Find antecedent moisture conditions (AMC) from 5-day antecedent rainfall

AMC Group	Dormant Season (cm)	Growing Season, (cm)
I	<1.3	<3.6
II	1.3 to 2.8	3.6 to 5.4
III	>2.8	>5.4

3. Plot Thiessen polygon areas of hydrologic soil groups and superimpose on land use map.

4. For the demarcated areas obtained in step 3, find the runoff curve numbers from tables (U.S Conservation Service, 1964) depending upon the land use, treatment practice, hydrologic soil group and antecedent moisture conditions. Estimate the weighted average curve number, CN^0 .

5. Find the storage index

$$S(t - \Delta t) = 25400/ CN^0 -254$$

6. Compare daily rainfall (P) with (0.2 x storage index) and estimate the initial abstraction (I_a) as follows:

$$I_a(t) = 0.2S(t - \Delta t) \quad \text{if } P(t) > 0.2S(t - \Delta t)$$

$$I_a(t) = P(t) \quad \text{if } P(t) < 0.2S(t - \Delta t)$$

7. Prepare the plot as given in figure 1 and obtain the runoff (Q) as shaded area. The shaded area is

$$Q = [P(t) - I_a(t)]\, b + 0.5\, [P(t) - I_a(t)]^2 (1 - b) / K_{sat}$$

$$\text{if } [P(t) - I_a(t)] < K_{sat}$$

$$Q = P(t) - I_a(t) - 0.5\, (1-b)\, K_{sat}$$

$$\text{if } [P(t) - I_a(t)] > K_{sat}$$

where b is the fraction of the build-up area.

8. Estimate infiltrated water (Q_i) as follow:

$$Q_i = P(t) - I_a(t) - Q \qquad \text{if } P(t) > 0.2\, S(t - \Delta t)$$
$$= 0 \qquad \text{if } P(t) < 0.2\, S(t - \Delta t)$$

9. Estimate the root zone soil moisture, $\theta(t)$ and ground water recharge, Q_r (t) on t^{th} day as follows:

If $[\theta_f - \theta(t - \Delta t)]\, D_0 < Q_i(t)$,
$\theta(t) = \theta_f$;

and

$$Q_r(t) = Q_i(t) - [\theta_f - \theta(t - \Delta t)]\, D_0 .$$

If $[\theta_f - \theta(t - \Delta t)]\, D_0 > Q_i(t)$,
$\theta(t) = [\theta(t - \Delta t) + Q_i(t) / D_0] / [1 + E_p(t) / \theta_f D_0)]$;

and $Q_r(t) = 0$.

(D_0 can be taken as 1.5 m for areas without vegetation).

10. Estimate the evaporation losses from upper reservoir, $E_{ua}(t)$, and lower reservoir, $E_{la}(t)$ as follows:

For Rainy Day:

$$E_{ua}(t) = E_p(t) \qquad \text{if } E_p(t) < I_a(t)$$
$$= I_a(t) \qquad \text{if } E_p(t) > I_a(t)$$

$$E_{la}(t) = E_p(t)\, \theta(t)/\theta_f$$

For Dry Day:

$$E_{ua}(t) = E_p(t) \qquad \text{if } E_p(t) < [\, I_a(t - \Delta t) - E_{ua}(t - \Delta t],$$

$$E_{ua}(t) = I_a(t - \Delta t) - E_{ua}(t - \Delta t) \qquad \text{if } E_p(t) > [\, I_a(t - \Delta t) - E_{ua}(t - \Delta t)],$$

$$E_{la}(t) = E_p(t)\, \theta(t)/\theta_f$$

At any time, $E_{ua}(t)$ is not less than zero.

11. Update the storage index as follows:

Rainy Day:

$S(t) = S(t - \Delta t) + E_{ua}(t) + E_{la}(t) - [Q_i(t) - Q_r(t)]$,

Dry Day:

$S(t) = S(t - \Delta t) + E_{ua}(t) + E_{la}(t)$,

12. GO TO STEP 6 for next day.

(According to SCS method, $\Delta t = 1$ day. Rushton and Word (1979) have found that Δt should not exceed 1 day. If SCS method is used for computation of runoff, Δt can be taken as one day. If Manely method is used Δt should be one hour for computation of runoff and infiltration).

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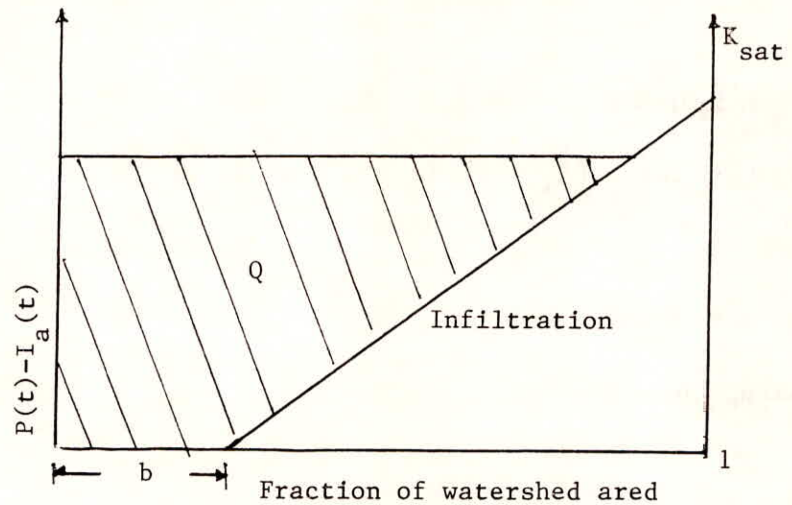

Fig.1- Estimation of overland flow