GROUNDWATER HYDROLOGY FOR BEGINNERS

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Prologue

Groundwater hydrology is a vast subject that broadly comprises theories/ methodologies/ algorithms that facilitate essentially two kinds of activities. First: assessment and management of the groundwater resource at a regional level. This activity involves identification of the aquifer system and subsequent assessment of recharge and finally the utilizable resource at macro (say basin) level. This requires exposure to the governing differential equations in the cartesian coordinates, and their solutions analytically and numerically. Numerical solutions are usually termed as *modeling*. The second kind of activity essentially deals with the development of the estimated utilizable resource through wells, and estimation of the aquifer parameters. This requires exposure to *well hydraulics* that usually comprises analytical solutions of the governing differential equations in radial coordinated, and various strategies for analyses of pumping test data. This write-up is a rather preliminary text that gives only a broad over view and is inevitably devoid of details. This should be read in conjunction with another sister write-up title “groundwater modeling for beginners”.

Subsurface Water

All the water occurring in the pore space of a geological formation is termed as *Subsurface water*. Typical distribution of subsurface water over a vertical section extending from the ground surface to the lower impervious layer is shown in Fig 1.

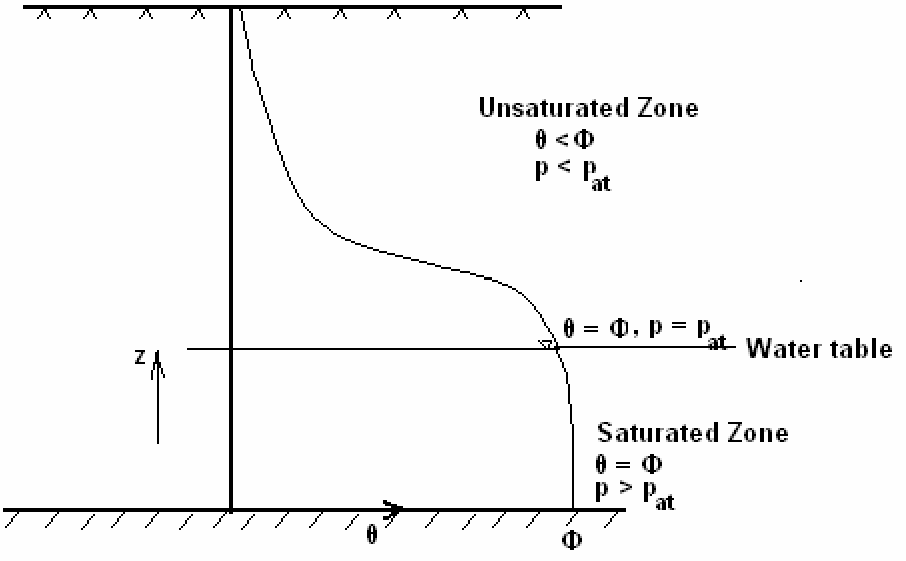

Figure 1 Subsurface Water

Unsaturated Zone

It can be seen that up to a certain depth below ground the moisture content is less than the porosity implying that only a part of the pore space is filled up by water and air occupies the remaining space. This zone termed as the unsaturated zone comprises water that sticks on to the individual grains (Fig 2) due to capillary forces and as such occurs at a sub-atmospheric pressure.

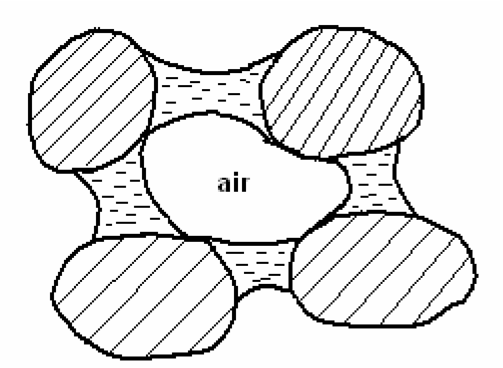

Figure 2 Occurrence of subsurface water in Unsaturated Zone

Saturated Zone

Fig 1 reveals that the subsurface water beyond certain depth occurs at a moisture content equal to the porosity implying that the entire pore space of the geological formation is filled up by water. This zone termed as saturated zone comprises water at a pressure greater than the atmospheric pressure.

Groundwater

Subsurface water occurring in the saturated zone can be tapped through boreholes since it is at a pressure greater than the atmospheric pressure and the pressure within the bore hole is atmospheric. On the other hand water in the unsaturated zone on account of being at a sub-atmospheric pressure would not enter into a bore hole. As such only the subsurface water occurring in the saturated zone can be developed through bore holes. Therefore, logically the subsurface water occurring in the saturated zone is termed as groundwater.

Aquifer

An aquifer is a geological formation satisfying the following pre- requisites.

I. It should be able to store groundwater in adequate measures.
II. It should be able to transmit groundwater in adequate measures.
III. It should receive adequate recharge.

Aquifer Types

Depending upon the hydrogeological setup an aquifer may be classified into the following types:

I. Unconfined aquifer
II. Confined aquifer
III. Leaky confined aquifer

Unconfined Aquifer

An unconfined aquifer bounded by a free surface at the top and an impervious boundary at the bottom (Fig 3). Free surface is essentially a surface at all points of which the pressure equals the atmospheric pressure. Groundwater professionals usually call it water table. It can be seen in Fig 1 that an unconfined aquifer essentially lies below the unsaturated zone and water table is the surface of demarcation between the unsaturated and the saturated zones.

Recharge to unconfined aquifers can occur vertically or laterally.

The vertical recharge emanates from the ground overlying the aquifer and flows through the unsaturated zone to the water table. It may comprise following components:

I. Recharge from rainfall
II. Recharge from applied irrigation
III. Recharge from canal seepage

The lateral recharge may occur when the stage of a stream hydraulically connected to the unconfined aquifer exceeds the water table elevation (Fig 4).

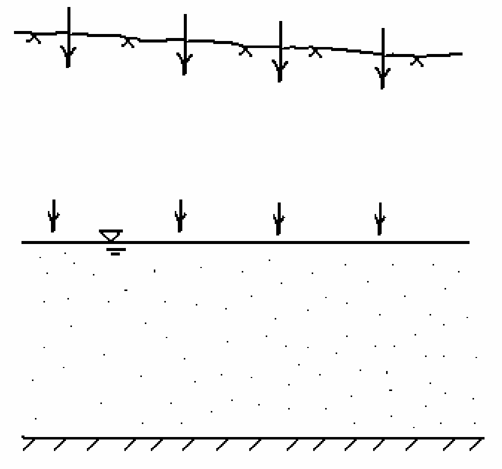

Figure 3 Unconfined Aquifer

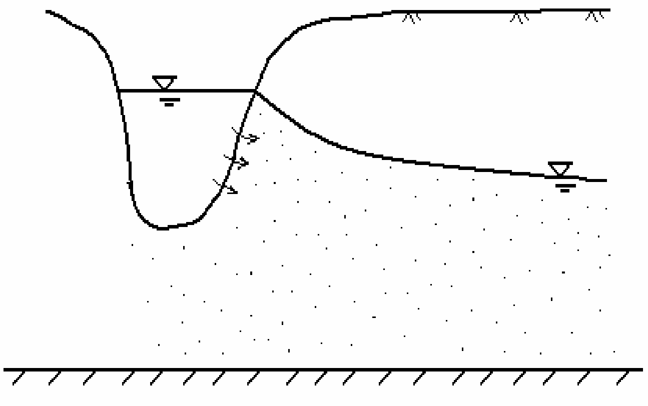

Figure 4 Lateral Recharge to an Unconfined Aquifer

Confined Aquifer

A confined aquifer is bounded by impervious boundaries at the top and bottom and the groundwater occurs at a pressure larger than the atmospheric pressure (Fig 5). Thus if a bore hole is drilled through a confined aquifer, water would rise into it through the pressure head. Thus the piezometric head of the aquifer equals the elevation of the lower face of the upper confining layer plus the pressure head. As such a confined aquifer has a piezometric surface designating its piezometric head.

A confined aquifer with its piezometric surface above the ground is termed as an artesian aquifer (Fig 6). If a bore hole is drilled in such aquifers, water would rise above the ground surface without any pumping.

Since a confined aquifer is bounded at its top by an impervious layer, it can not receive a vertical recharge. It may however be recharged vertically at its outcrop i.e., some region (usually mountainous) where the confined aquifer is exposed to the ground (Fig 7). It may also be recharged laterally from a hydraulically connected stream with its stage higher than the piezometric head (Fig 8). The lateral recharge is rather rare since most of the confined aquifers are deep-seated lying below the riverbeds.

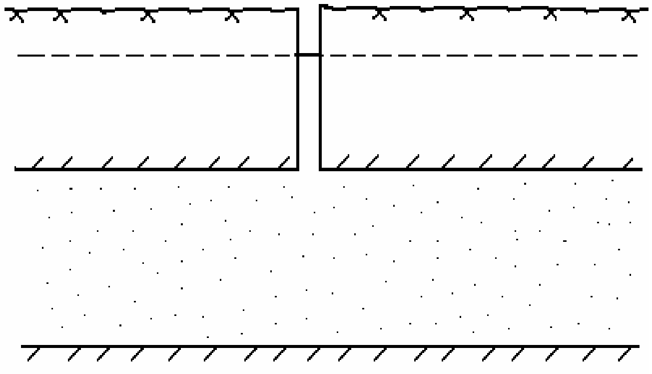

Figure 5 Confined Aquifer

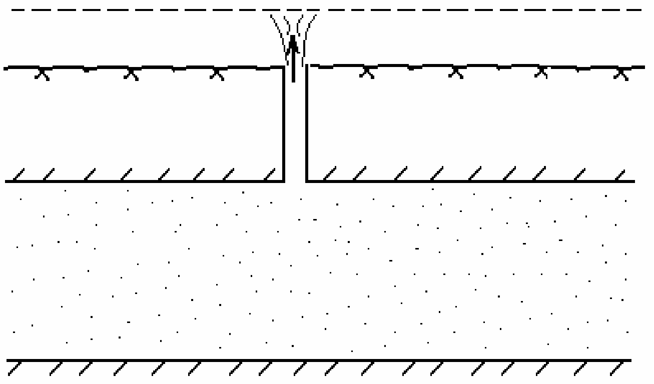

Figure 6 Artesian Aquifer

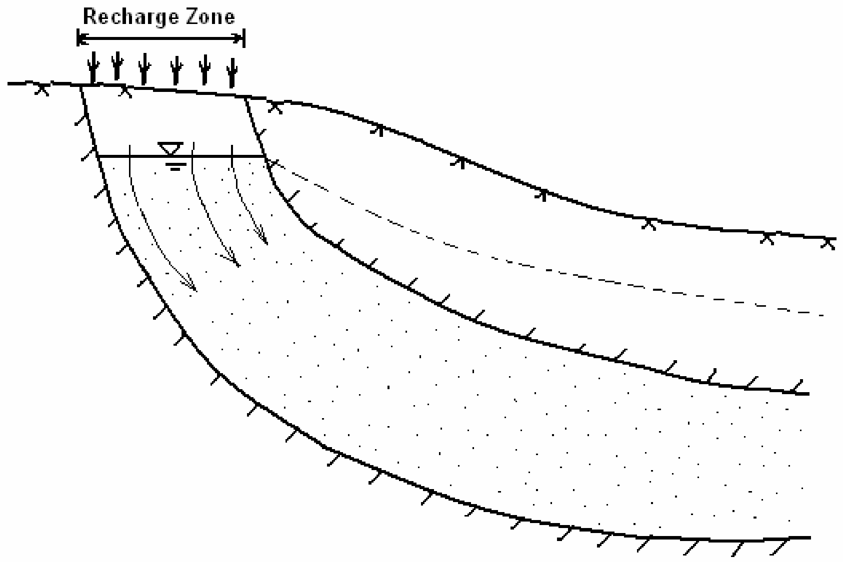

Figure 7 Recharge Zone of a Confined Aquifer

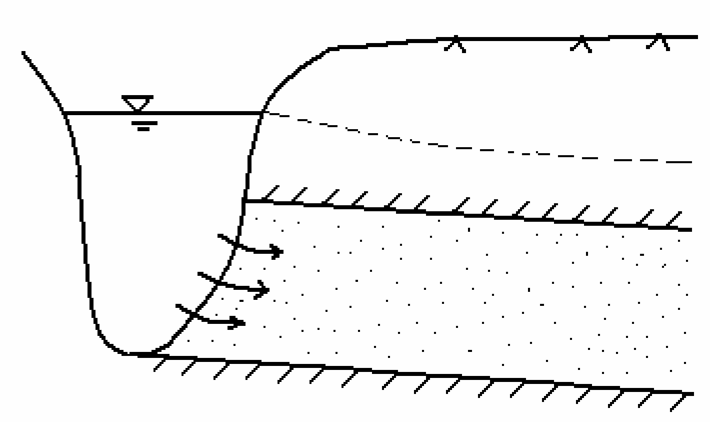

Figure 8 Lateral Recharge of a Confined Aquifer

Leaky confined aquifer

A leaky confined aquifer as the name implies is essentially a confined aquifer with its upper confining layer being leaky rather than impervious. Typically a leaky confined aquifer may be overlain by an unconfined aquifer (Fig 9). Thus, any difference between the water table elevation of the unconfined aquifer and the piezometric head of the underlying leaky confined aquifer would trigger off a vertical leakage through the leaky layer. The direction of the leakage could be downwards incase the water table is at a higher elevation (Fig 9) or vice versa.

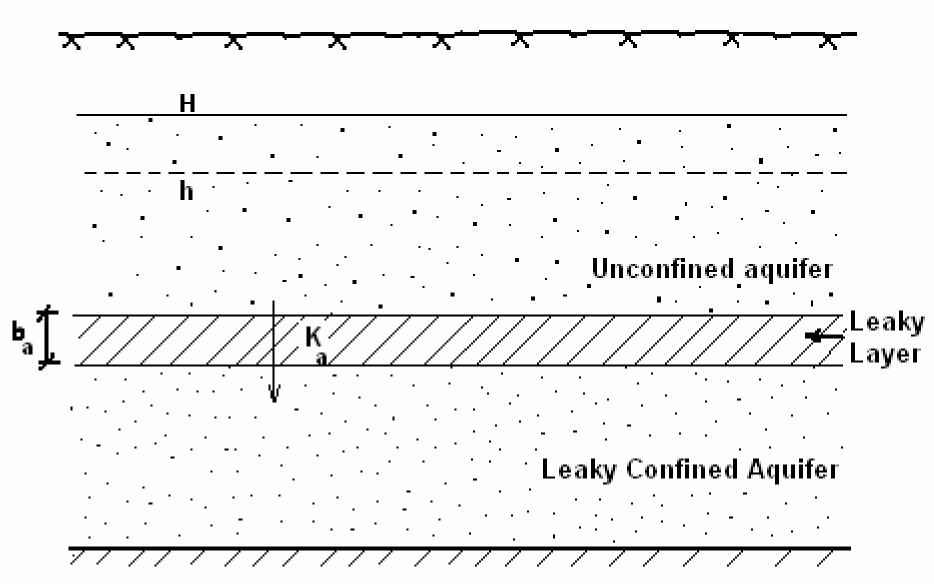

Figure 9 Leaky Confined Aquifer

Like a confined aquifer, a leaky confined aquifer may also be recharged laterally from its outcrop as well as from hydraulically connected streams. Further, it may receive additional recharge from overlying unconfined aquifer with the water table being higher than the piezometric surface.

Aquifer Parameters

Parameters of an aquifer essentially represent its characteristics quantitatively. Recalling that an aquifer is expected to store and transmit water in adequate measures, the aquifer parameters can be classified into following categories.

I. Flow parameters
II. Storage parameters
III. Derived parameters

Flow Parameters

Hydraulic conductivity: The fundamental flow parameter is the Hydraulic conductivity that correlates the Darcy's velocity (i.e., volume of water flowing per unit area per unit time) to the hydraulic gradient as per the famous Darcy's law. Darcy's law as proposed by Darcy can be written as follows:

$V = Ki$

Where V = Darcy's velocity, i is the hydraulic gradient and K is the permeability (now termed as hydraulic conductivity). It may be mentioned here that the presently accepted form of Darcy's law for anisotropic media is far more elaborate.

Hydraulic conductivity can be invoked to compute the flow rate in any direction corresponding to a given hydraulic gradient.

Transmissivity: For a confined/ leaky confined aquifer with horizontal or nearly horizontal confining layers, the horizontal flow rate (volume of water flowing horizontally per unit width per unit time) is related to hydraulic gradient by the parameter Transmissivity. It can be easily seen that Transmissivity equals the product of hydraulic conductivity (K) and the saturated thickness (b) (Fig 10).

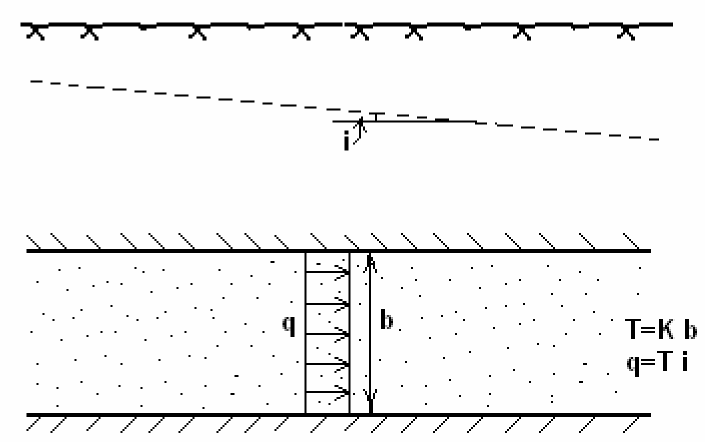

Figure 10 Transmissivity of an aquifer

Strictly speaking, the concept of transmissivity does not hold for an unconfined aquifer due to two reasons. Firstly, the saturated thickness in an unconfined aquifer changes with time as the water table fluctuates. Thus, the transmissivity at a given location may vary with time. Further, flow in an unconfined aquifer is caused by inclination of the water table. And, since water table forms the upper boundary a truly horizontal flow is not possible. However, we may define a time- averaged transmissivity of an unconfined aquifer provided the inclination of the water table is not large.

Storage Parameters

Storage parameters facilitate computation of release of water from groundwater storage. A brief description follows.

Release of water from groundwater storage: Any abstraction of water from the groundwater storage leads to a fall in its watertable/ piezometric head. Or inversely a fall in the water table/ piezometric head of an aquifer triggers off a release of water from its storage. There are two distinct mechanisms responsible for this water release viz., gravity drainage and Aquifer deformation.

Gravity drainage: In unconfined aquifers a fall in the water table distorts the unsaturated zone in the following two ways:

1. A part of the saturated zone de-saturates, i.e., transforms into an unsaturated zone.
2. Elevation of all initially unsaturated points measured above the water table increases.

This in turn reduces the equilibrium soil moisture content (θ_e) of the modified unsaturated zone extending from new water table to the ground. The vertical distributions of θ_e before and after the water table fall are given in Fig 11. The area between the two curves represents the volume of water (per unit plan area) that can no longer be held in position against the action of gravity. This excess water starts draining down to the water table under the action of gravity at a rate which is highest at the beginning, and reduces as the time elapses. The drainage is thus a time-distributed phenomenon which continues until the entire excess water is drained down. Accordingly, this mode of water release is termed as “delayed yield” or “delayed gravity drainage”.

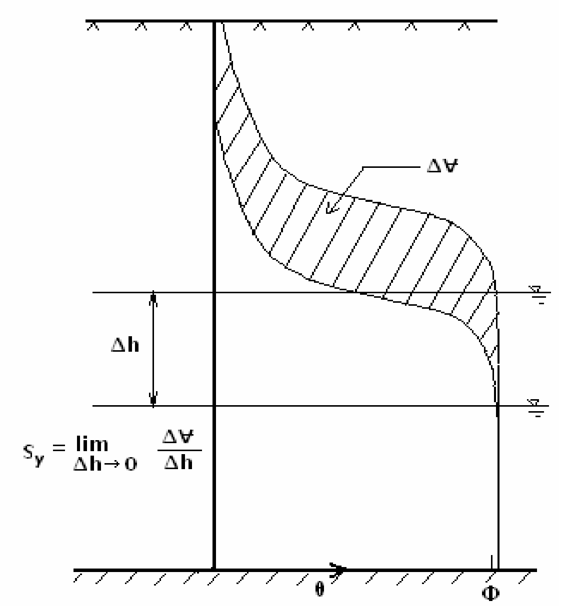

Figure 11 Vertically Drainable Water above Falling Water-table

Specific yield: The fundamental parameter related to the release of water under the action of gravity is the Specific yield. This represents the volume of water released from a unit area of aquifer under the action of gravity as the water table declines by one unit (Fig 12).

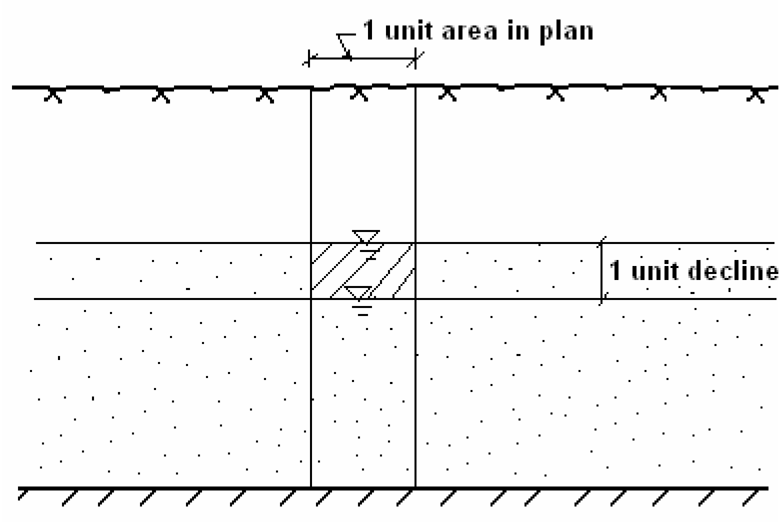

Figure 12 Specific Yield of an Unconfined Aquifer

Aquifer Deformation: An aquifer may be viewed as a medium comprising an inter-granular skeleton and water occurring in the pore space. The overburden pressure (w) is jointly borne by the two components leading to a water pressure (p) and inter-granular pressure (q).

$w = p + q$

A fall in the water table/ piezometric head in an aquifer causes a fall in the water pressure. Since w remains unchanged this pressure fall leads to an equal increase in the inter-granular pressure. The increased inter-granular pressure compacts the aquifer reducing its porosity. The fall in porosity reduces the water storing capacity of the saturated zone. The surplus water in the saturated zone is squeezed out. Further the volume of ejected water is marginally increased due to decrease of water pressure.

Storage coefficient: This represents the volume of water released from a unit area of aquifer through compaction of aquifer and expansion of water as the piezometric head/ water table declines by one unit (Fig 13).

Recalling that the release of water following a fall of piezometric head occurs on account of compaction of aquifer and expansion of water, Storage coefficient (S) can be quantified in terms of the elastic properties of aquifer skeleton and water as follows:

$$S = \gamma\theta\beta b\left(1 + \frac{\alpha}{\theta\beta}\right)$$

Where θ = porosity of aquifer, β = inverse of bulk modulus of elasticity of water, b = thickness of aquifer, γ = unit weight of water and α = inverse of bulk modulus of elasticity of aquifer skeleton.

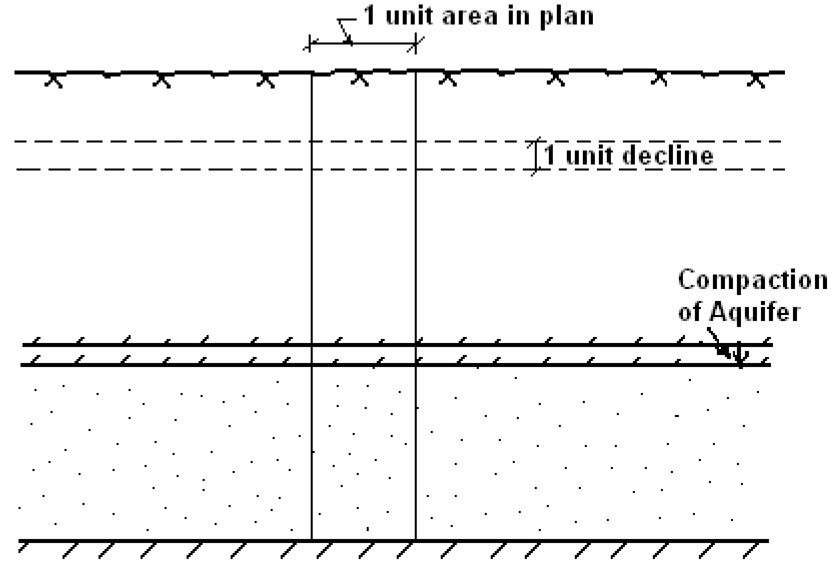

Figure 13 Storage Coefficient

Specific storage: Another parameter related to the release of water by compaction of aquifer and expansion of water is the Specific storage. This represents the volume of water released from a unit area and unit thickness of aquifer, through compaction of aquifer and expansion of water as the piezometric head/ water table declines by one unit. It can be seen that the Specific storage equals the Storage coefficient divided by the saturated thickness.

Storage parameters of an unconfined aquifer: A fall in the water table in an unconfined aquifer leads to desaturation as well as a fall of pore water pressure. Thus, water is released from storage due to gravity drainage, and also due to compaction of aquifer and expansion of water. As such an unconfined aquifer shall be characterized by the Specific yield and the Storage coefficient/ Specific storage.

Storage parameters of an confined/ leaky confined aquifer: A fall in the piezometric head of a confined/ leaky confined aquifer shall lead to release of water due to compaction of aquifer and expansion of water only. Thus, such aquifers shall have only the Storage coefficient/ Specific storage.

Derived Parameters

A few parameters derived from the fundamental flow and storage parameters are as follows:
Delay index: In accordance with the definition of Specific yield (S_y), the release of water from a unit area under the action of gravity as the water table declines by Δh is: Sy. Δh. However this release shall not be instantaneous but shall be distributed over time. As per Boulton's concept the rate of release (Y(t)) declines exponentially in accordance with the following equation.

$$Y(t) = \alpha.\ S_y \Delta h\ e^{-\alpha t}$$

Where $1/\alpha$ is the aquifer parameter termed as delay index.

Hydraulic Resistance: Consider a system of aquifers comprising an unconfined aquifer underlain by a leaky layer and a leaky confined aquifer (Fig 9). As discussed earlier any difference between the water table elevation (H) of the unconfined aquifer and the piezometric head (h) of the leaky confined aquifer shall trigger of a vertical leakage through the intervening leaky layer. Designating the hydraulic conductivity and the thickness of the leaky layer as K_a and b_a respectively, and invoking Darcy's law the leakage discharge per unit area (L) can be written as follows.

$$L = C.(H - h)$$

Where $C = (K_a/b_a)$, termed as the Hydraulic resistance of the leaky layer.

Leakage Factor: The leakage factor (B) of a leaky confined aquifer is defined interms of the Hydraulic resistance (C) of the leaky layer and the Transmissivity (T) of the leaky confined aquifer as follows:

$$B = (C.\ T)^{0.5}$$

Estimation of Parameters

Various parameters described in the preceding paragraphs can be estimated by conducting appropriate pumping tests on the aquifers. A pumping test essentially involves pumping water from a well usually at a constant rate. Simultaneously the drawdown of water table elevation/ piezometric head in a few observation wells/ piezometers in the vicinity of the pumping well is monitored at advancing times as the pumping progresses. The pumping and the observation wells tap the aquifer to be tested. The time- drawdown data emanating from various observation wells/ piezometers are analyzed to estimate the relevant parameters invoking Well hydraulics described in the following section.

Well Hydraulics

Well hydraulics refers to a variety of equations describing the spatial variation of the drawdown of water table/ piezometric head in the vicinity of a pumping well. A few prominent equations are described in the following paragraphs.

Confined aquifer:

The steady state drawdown in a confined aquifer at any radial distance r from a pumping well is given by the following equation (usually termed as Theim's equation).

$s(r) = Q/(2\pi T). \ln(r/R)$

Where Q is the pumping discharge, T is the Ttransmissivity and R is the distance of a constant head boundary from the well.

The unsteady state drawdown in a confined aquifer at any radial distance r from a pumping well and at time t since the beginning of the pumpage is given by the following equation (termed as Theis' equation).

$$s(r,t) = \frac{Q}{4\pi T}\int_u^\infty \frac{e^{-x}}{x}dx;\ u = \frac{r^2 S}{4Tt}$$

$\int_u^\infty \frac{e^{-x}}{x}dx = W(u)$; known as well function

W (u) is well tabulated in all the standard text books of groundwater.

Leaky confined aquifer

The steady state drawdown in a leaky- confined aquifer at any radial distance r from a pumping well is given by the following equation (usually termed as Hantush's equation).

$s(r) = Q/(2\pi T). K_0(r/B)$

Where B is the Leakage factor of the leaky confined aquifer and K_0 is the modified Bessel's function of second kind and zero order.

The unsteady state drawdown in a leaky confined aquifer at any radial distance r from a pumping well and at time t since the beginning of the pumpage is given by the following equation (termed as Theis' equation).

$$s(r,t) = \frac{Q}{4\pi T} \int_u^\infty \frac{1}{x} \exp\left[-x - \frac{\left(r/B\right)^2}{4x}\right] dx$$

Or $s(r,t) = \frac{Q}{4\pi T} W(u, r/B)$

$W(u, r/B)$ is well tabulated for a wide range of u and r/B values.

Unconfined aquifer:

The steady state drawdown in an unconfined aquifer at any radial distance r from a pumping well is given by the following equation (usually termed as Theim's equation).

$H^2 - h^2 = Q/(\pi K). \ln(r/R)$

Where K is the hydraulic conductivity.

The unsteady state drawdown in an unconfined aquifer at any radial distance r from a pumping well and at time t since the beginning of the pumpage is given by the following equation (termed as Boulton's equation).

$$s = \frac{Q}{4\pi T}\left[W(U_a, U_y, r/B, N)\right]$$

$$U_a = \frac{r^2 S}{4Tt};\ U_y = \frac{r^2 S_y}{4Tt};$$

$$B^2 = \frac{T}{S_y \alpha}\ \ and\ \ N = \frac{S + S_y}{S_y}$$

Thus, strictly speaking the well function, in accordance with Boulton's theory, should be tabulated or plotted for the four dimensionless parameters viz., U_a, U_y, r/B, N. However such a tabulation or plotting would have been prohibitive. As a practical solution, the no. of parameters is cut down to 2 as follows:

(i). It has been revealed that the earlier drawdowns are almost independent of U_y, whereas the late drawdowns are almost independent of U_a. Thus, out of the two parameters U_a and U_y only one is active at a time.

(ii). The parameter N, which any way is bound to assume very large values, is assumed to be infinity. This assumption is nearly satisfied if N exceeds 30.

Thus, the well function has been tabulated for the two dimensionless variables i.e. U_a or Uy and r/B.

Estimation of Vertical Recharge

Vertical recharge (say from rainfall) forms a predominant component of groundwater resource of an unconfined aquifer. The recharge can be estimated by one of the following approaches.

I. Modeling of flow through unsaturated zone
II. Studying the impact of recharge on water table

Modeling of flow through unsaturated zone: As has been discusses earlier the vertical recharge to the water table essentially emanates from the infiltration/ seepage as it routs through the unsaturated zone extending from ground/ bed to the water table. Consequently the most direct method of its estimation is modeling of flow through unsaturated zone. However this approach apart from requiring a high level of modeling expertise is prone to a large number of uncertainties arising out of unknown or poorly known soil and evapotranspiration characteristics. Therefore this approach is has generally not found acceptance among practicing professionals.

Impact of recharge on water table: This approach is usually based upon a lumped water balance study of the area of concern wherein all inflows into and outflows from the aquifer and the consequent change of storage during the period of concern are plugged into the continuity equation to estimate one unknown (say vertical recharge). The continuity equation can be written as follows:

$$V_{VR} + V_{LR} - V_O = \Delta S$$

Where V_{VR} = volume of unknown vertical recharge, V_{LR} = volume of lateral recharge, V_O= volume of outflow (comprising pumpage, evapotranspiration, lateral outflow to hydraulically connected streams), and ΔS = change of storage. The vertical recharge can be estimated provided the other terms appearing in this equation can be estimated from independent data.

Lateral flow volumes (comprising the lateral recharge and the lateral outflows) can be estimated by drawing a flow network comprising water table contours and the flow lines. The water table contours are drawn by interpolating the available discrete point water table elevation data from observation wells/ piezometers. Subsequently the flow lines are superposed over the water table contours ensuring the requirement that the flow lines intersect the contours normally. The flow rate in the annular space between two stream lines (termed as a stream tube) is computed by invoking Darcy's law (Fig 14). The flow rate (q_k) in kth stream tube is thus written as follows.

$$q_k = b_k. T. (H_2 - H_1)/L$$

Where b_k is the average width of k^{th} stream tube, T is the Transmissivity, and $(H_2 - H_1)$ is the fall of water table in a distance L along the stream tube.

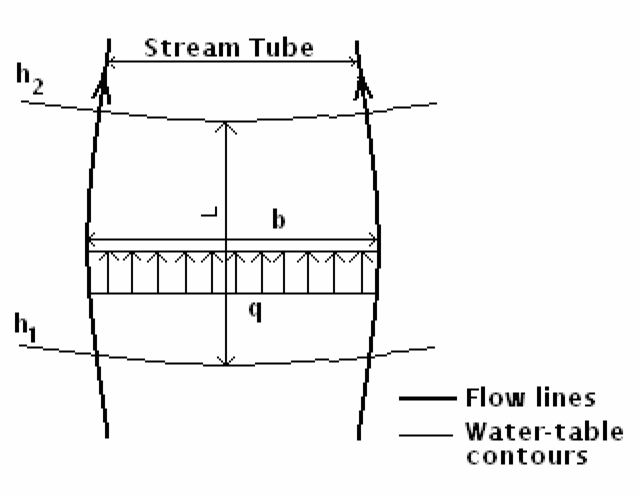

Figure 14 Discharge through a Steam tube

The total stream- aquifer interflow can be computed by summing up the flow rates in the stream tubes.

The change of groundwater storage in a given time period can be computed from the water table contours at the beginning and the end of the stipulated period. The corresponding storages can be computed by summing up the storages available between the adjacent water table contours (Fig 15). The storage (S) available in the area (a) falling between two water table contours (H_1 and H_2) can be computed as follows.

$S = [0.5*(H_1 + H_2) - H_b].a.Sy$

Where H_b = RL of the lower impervious layer.

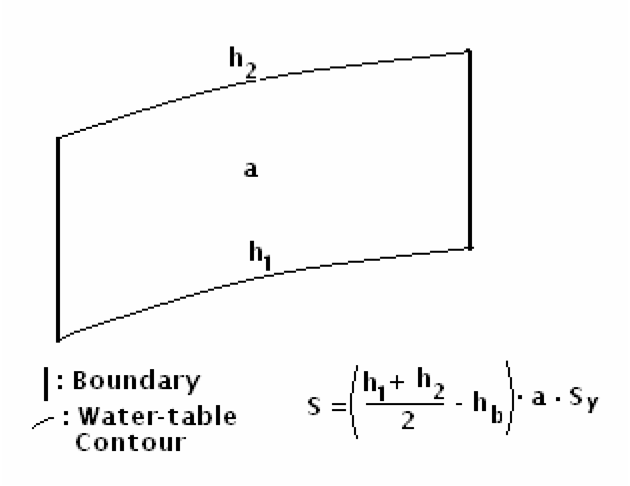

Figure 15 Computation of Groundwater Storage

Estimation of Groundwater Resource

The entire vertical recharge is not available for groundwater development because of the inevitable losses comprising primarily the evapotranspiration and the baseflow contributions to the hydraulically connected rivers. Groundwater resource can be deemed to be the utilizable part of the vertical recharge (i.e., the vertical recharge – losses). The allowance for the losses while stipulating a groundwater resource is linked to the projected water table elevations. Larger allowance for the losses would ensure a higher water table elevation following the development of the groundwater resource.

Thus following the estimation of vertical recharge there could be two different approaches for assessing the groundwater resource. First approach is to assume the losses arbitrarily or based upon past experiences. This approach though quite simple is not the most desirable one since it does not permit the assessment of the impact of the groundwater development on the water table/environment.

The second approach which can essentially be considered as an extension of the first approach may involve the following steps.

I. Estimate the vertical recharge
II. Stipulate losses and hence arrive at a tentative figure for the groundwater resource.
III. Project the response of the aquifer to the development of the tentative groundwater resource through modeling. The response could comprise among others, the depths to water table, lateral flows, sea water intrusion etc.
IV. Incase the response is acceptable, the tentative resource may be accepted. Otherwise modify the resource and go to step three.

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