Assessment of Ground Water Recharge

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Introduction

Hydrologic phenomena are complex, and may not be fully understood. However, for solving water related problems, they may be represented in a simplified way by means of the systems concept. The hydrologic system may be conveniently decomposed into three subsystems: i) the atmosphere-land-water system involving the processes of radiation, precipitation, evaporation, interception, infiltration, and transpiration; ii) the land-surface-water system containing the processes of infiltration, overland flow, interflow, surface runoff, erosion and sedimentation process, surface water and groundwater interaction, and runoff to ocean; and iii) the subsurface water system containing the processes of groundwater recharge, stream aquifer well interaction and spring flow , and sea aquifer interaction.

Process level models i.e. models used for evaluation of (i) infiltration and its redistribution due to rainfall, (ii) surface runoff originating from rainfall and consequent stream stage rise, (iii) influent and effluent seepage from stream, (iv) seepage from a canal and tanks,(v) irrigation return flow, and (vi) groundwater withdrawal are integrated in a water balance study. The integrated macro level model can either be a lumped one or a distributed one. Using the water balance model groundwater recharge that takes place naturally can be ascertained when all other components are calculated with reasonable accuracy.

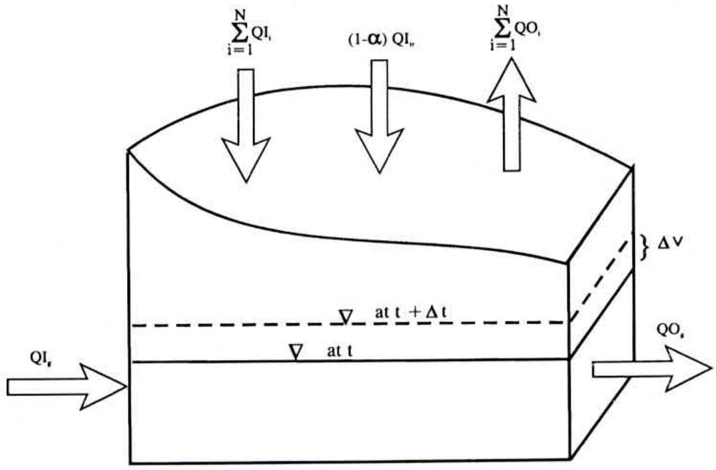

Fig. 1: A Saturated Control Volume of a Lumped Model

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Lumped Groundwater Balance Model

Considering the saturated flow control volume (Fig. 1), the following groundwater balance equation over a time period Δt as proposed by Khan (1980) is:

$$\sum_{i=1}^{N} Q_{\inf}(i) + (1-\alpha)Q_{iir} + Q_{ig} - \sum_{i=1}^{N} Q_{outf}(i) + Q_{og} = dS$$

in which,

$\sum_{i=1}^{N} Q_{\inf}(i)$ = total inflow into the saturated flow control volume comprised of recharge from precipitation, seepage from stream and canals, tank, and artificial recharge where water level lies at great depth,

Q_{ig} = the ground water inflow from adjacent sections, which includes influent seepage from stream;

$\sum_{i=1}^{N} Q_{outf}(i)$ = the total outflow from the section through pumping;

α = the irrigation efficiency over the section which is unknown;

Q_{og} = the ground water outflow to the adjacent sections, which includes effluent seepage from stream;

dS = change in storage in time period Δt.

The change in volume of water within the control volume during time Δt is equal to integration of change in water level multiplied by specific yield. ΔV can be computed either using the distributed groundwater flow model or by direct measurement. Groundwater levels data, stream cross section and stage data, water-spread area of the tank, aquifer parameters and water supplied for irrigation are required to carry out the water balance. From this equation only one unknown quantity can be estimated. The method requires reasonable estimate of specific yield in the zone of water table fluctuations, measurement of water table position at several observation wells evenly distributed in the study area, and transmissivity of the aquifer.

The same approach can be extended to the distributed ground water flow model. Deep percolation which represents a major component of irrigation return flow can be estimated using soil moisture modelling for each soil group and crop type. Using this estimation and other components of recharge and groundwater withdrawal, and applying the prevailing boundary conditions, the water level fluctuation can be simulated using a distributed groundwater flow model. Subsurface inflow and outflow components of water balance appearing in Khan's model and the change in storage ΔV can be known from groundwater flow modelling.

Appendix1

Reach Transmissivity for a partially penetrating stream having large width

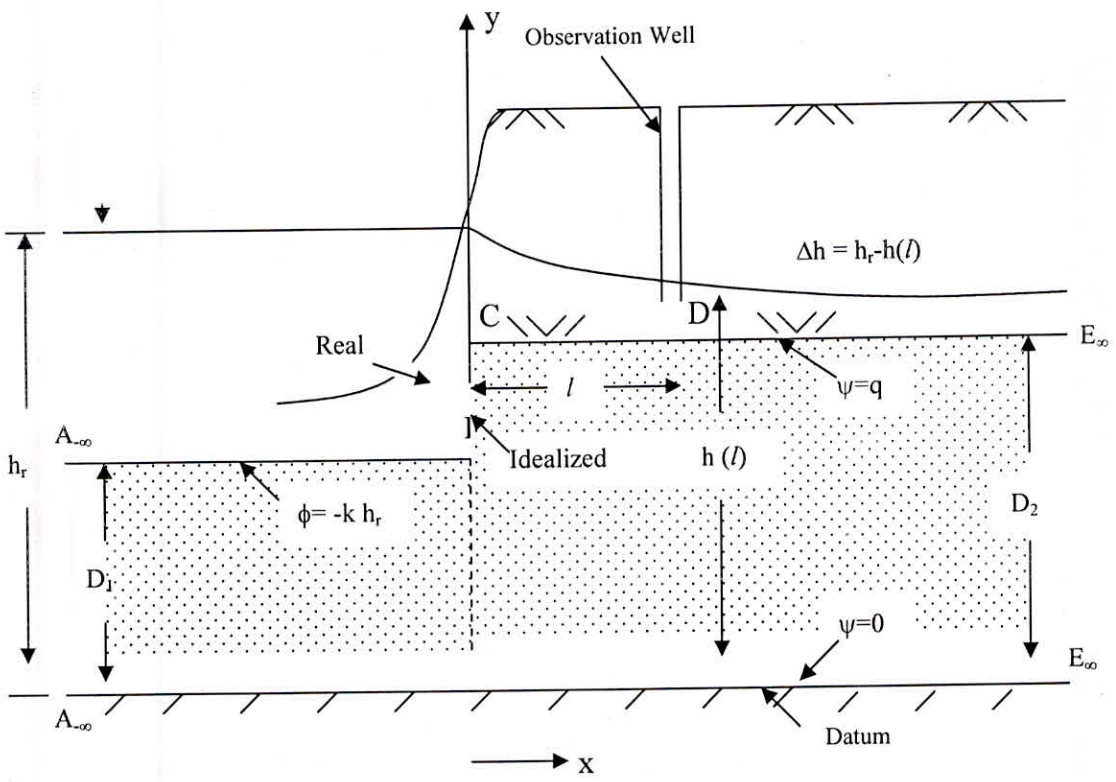

Fig. 2- A partially penetrating stream

Multiplying the difference in stream stage, and water level in an observation well located in the vicinity of a stream, $[h_r - h(l)]$, with a constant known as reach transmissivity constant Γ_r, the exchange flow between the stream and the aquifer can be estimated using the relation:

$$Q = \Gamma_r [h_r - h(l)].$$

The reach transmissivity constant has been determined using conformal mapping technique applicable for steady flow condition.

$$q = \frac{\pi k}{\ln \dfrac{\sqrt{1-c} + \sqrt{d-c}}{\sqrt{1-c} - \sqrt{d-c}}} \quad [h_r - h(l)] = \Gamma_r [h_r - h(l)]$$

$$c = 1 - (D_1 / D_2)^2$$

$$\frac{l}{D_1} = \frac{\sqrt{d-c}}{\pi} \int_{-1}^{1} \frac{\sqrt{c + \frac{(d-c)}{4}(1+u)^2}}{1 - c - \frac{(d-c)}{4}(1+u)^2} du$$

The ratio $q/[k(h_r - h(l))] = q/[k\Delta h]$ is a dimensionless reach transmissivity constant (Morel-Seytoux et al 1978) for a stream reach of unit length. As seen from figure the reach transmissivity constant is a function of l and D_1 / D_2.

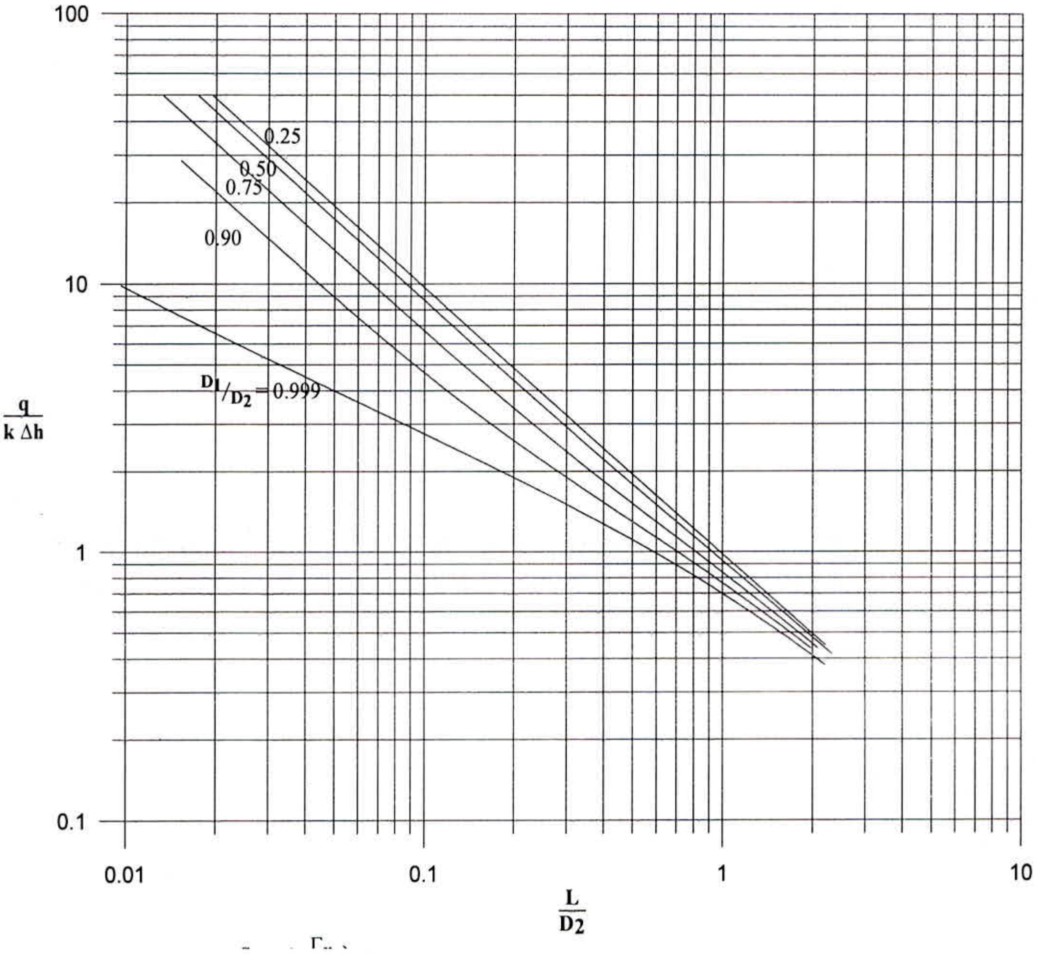

Fig.3. Dimensionless reach-transmissivity for different depth of penetration

Appendix2

Seepage from a Canal Reach

A canal having hydraulic connection with underlying aquifer is shown in Fig. 4. The recharge from unit length of the canal hydraulically connected to the aquifer can be assumed to have the following non-linear relationship with the potential difference between the canal and the aquifer (Rushton and Redshaw, 1979):

$$Q(t) = C_2 [1-\exp\{-C_3(h_r - h(0,t))\}] \quad (1)$$

where C_2 and C_3 are constants; h_r is the hydraulic head at the canal perimeter; and h(0, t) is the hydraulic head in the aquifer under the canal axis at time t. h_r and h(0, t) are measured upwards from the impervious bed of the aquifer which has been selected as the low datum. The hydraulic head, h(0, t), is governed by the recharge from the canal which occurs up to time t. The time should be measured from the instant, the seepage water from a canal joins the ground water. For convenience, it is reckoned since water is conveyed in the canal.

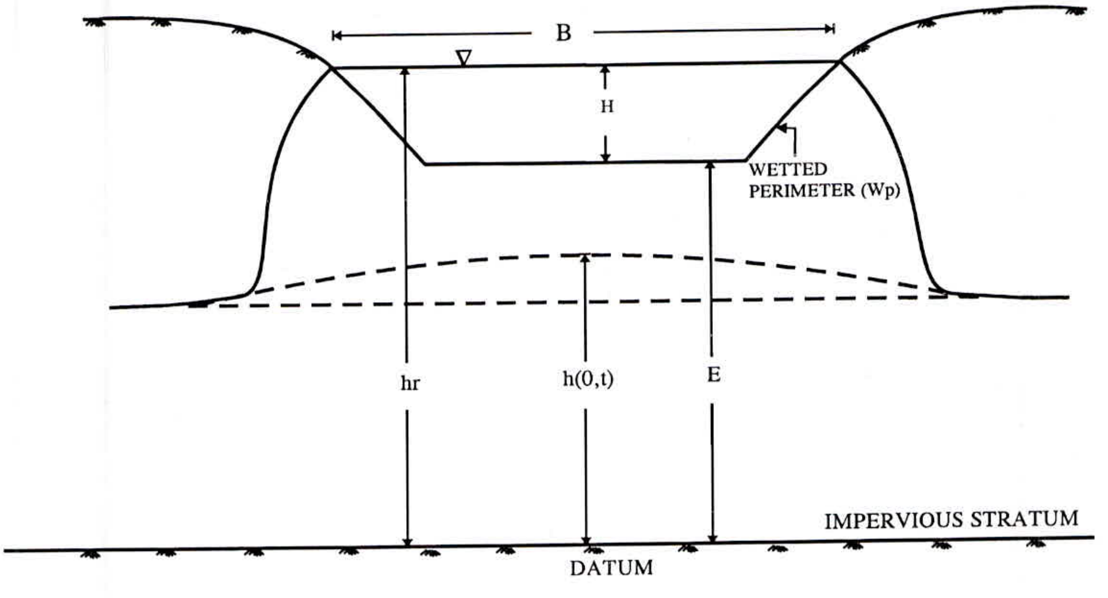

Fig. 4: A Canal Hydraulically Connected with Aquifer

The seepage from unit length of a stream, when the water table is at very large depth, is given by (Kozeny, 1931, vide Harr 1962):

$$Q = K\,(B+AH) \quad (2)$$

where B=width of the stream at the water surface, and H= the maximum depth of water in the stream. For a stream with a curved perimeter, the parameter 'A' is equal to 2 (Kozeny, 1931). For trapezoidal section, the constant *A* can be obtained from Vederinkov's graph (vide Harr, 1962).

Applying the condition that the exponential term in Eq. (1) tends to zero for very large value of $[h_r - h(0, t)]$, Eqs. (1) and (2) yield

$$C_2 = K(B+AH) \quad (3)$$

Substituting C_2 in Eq. (1)

$$Q(t) = K\,(B+AH)\,[1-\exp\{-\,C_3\,(h_r - h(0,t)\}] \quad (4)$$

For small difference between h_r and $h(0, t)$, the higher order terms of the polynomial expansion of the exponential term appearing in Eq. (4) can be neglected and the seepage for small potential difference can be approximated to be:

$$Q(t) = K(B+AH)\,C_3[h_r - h(0,t)] \quad (5)$$

For small potential difference between the stream and aquifer, the exchange flow rate has a linear relationship with the potential difference. The linear relationship proposed by several investigators is of the form:

$$Q(t) = \Gamma_r \Delta h \quad (6)$$

Γ_r is the reach transmissivity. Equating equations (5) and (6),

$$C_3 = \Gamma_r / [K\,(B+AH)\,] \quad (7)$$

Following Herbert (1970), an expression for reach transmissivity for unit length of a stream is given by:

$$\Gamma_r = \pi K / \log_e\,[0.5(E+H)/\,R] \quad (8)$$

where, E = saturated thickness of the aquifer below the bed of the stream; H = maximum depth of water in the stream; R = radius of the equivalent semi-circular section of the stream equal to w_p/π; w_p = wetted perimeter of the stream. Herbert's formula is applicable for $0.5(E+H) > R$.

References

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