INTRODUCTION TO GROUND WATER BALANCE

BY
G.C. MISHRA

0.5 percent of the total water resources of the world is in the form of groundwater. According to Huisman (1972) not all this is available for exploitation since half is below 800m and therefore is too deep for economic utilization. The capacity of the underground resource should not be underestimated; about 98 percent of the usable fresh water of the Earth is stored under ground. The groundwater is estimated to be 7000 x $10^{12} m^3$.

The following is the approximate water inventory of the Earth: (Hamil and Bell,1985)

Storage Component	Volume of water $10^{12} m^3$	Total water %
Atmosphere	. 13	0.0009
Oceans	1350400	97.5868
Saline lakes and in land sea	105	0.0076
Ice cap and glaciers	26000	1.8789
Soil moisture	150	0.0108
Groundwater	7000	0.5059
Fresh water lakes	125	0.0090
Rivers	2	0.0001

Watersheds and groundwater systems perform identical functions in the hydrologic cycle. They convey water from higher elevation to lower ones and hold a certain volume of water in transient storage. An aquifer can be thought of as a pipe or conduit that transfers water from the recharge area to areas of discharge, and which has also a storage component. A simple equation representing flow through the aquifer is:

Recharge to the aquifer - Discharge from the aquifer = Change in groundwater storage.

Thus, if outflow exceeds inflow there is a reduction in the amount of water stored and the water level falls. Obviously this can only continue until such times - as the storage becomes exhausted, when the maximum yield available at particular any time will be equal to the inflow. For the purpose of a discussion on regional exploitation of groundwater it is convenient to distinguish several kinds of groundwater reserves (Mandel and Shiftan, 1981). Reserves held in live storage are depleted by natural drainage and can also be recovered by pumping. Reserves held in dead

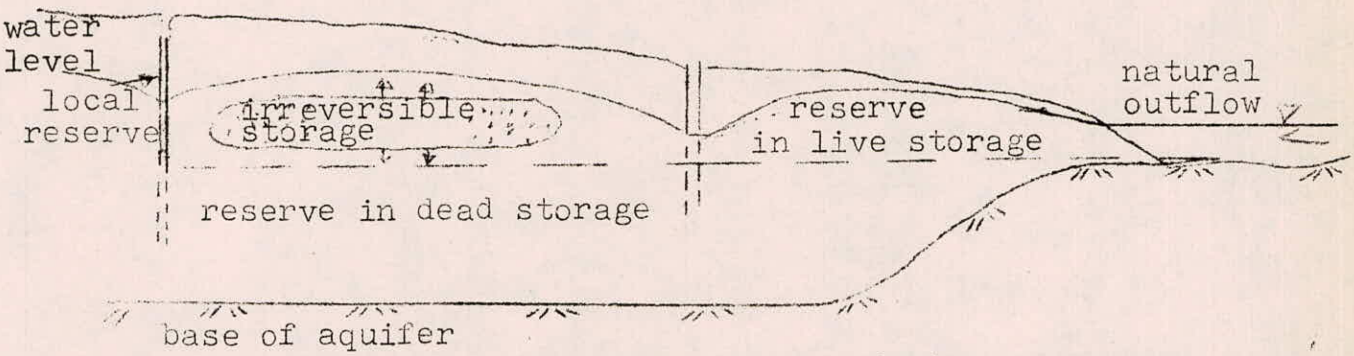
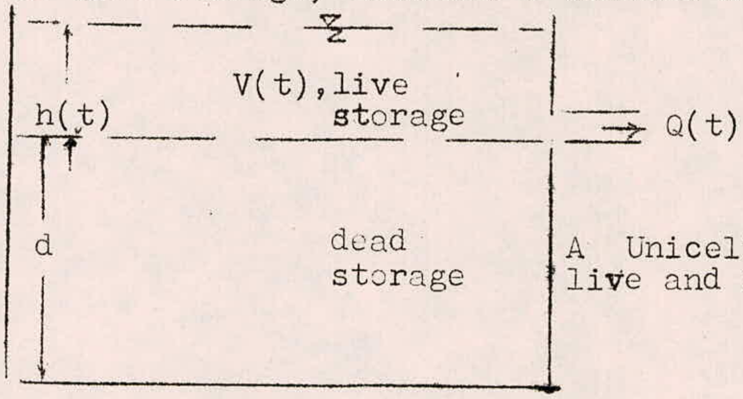


Fig.1 Classification of groundwater reserves

storage can be recovered only by pumping after the live reserves have been exhausted. However, water in dead storage is not stagnant. Cones of depression are formed by the abstraction of local reserves from the vicinity of pumped boreholes. Eventually all cones of depression spread outward over the aquifer, depleting the live storage or the dead storage. A fourth kind of reserves comprises those that are held in irreversible storage in semiconsolidated sediments. When the water level is lowered, the sediments undergo irreversible compaction, squeezing out water from their interstices. The processes leads to soil subsidence and frequently also to the deterioration of the water quality because of the influx of connate water from fine-grained strata. For the purpose of understanding natural discharge of live storage, consider a unicell model.



A Unicell Model depicting live and dead storages

The live storage at any time is given by

$$V(t) = ASh(t) \qquad ..(1)$$

Where V(t) is volume of live storage, S is the storage coefficient of the aquifer, h(t) is the elevation of the water level above the outlet. If the aquifer thickness 'd' below the out let is very large in comparison to h(t), h+d can be assumed to be approximately a constant equal to 'b'. Also

$$Q(t) = kbc\ h(t) \qquad ..(2)$$

in which c is a dimensionless parameter representing the flow pattern. Further the change of volume will be equal to negative of the flow rate. Hence,

$$Q(t) = -\frac{dV}{dt} \qquad ..(3)$$

Replacing h(t) in equation (2) by making use of equation (1)

$$Q(t) = \frac{Kbc}{As} V(t) \qquad ..(4)$$

Differentiating,

$$\frac{dv(t)}{dt} = \frac{AS}{Kbc} \frac{dQ(t)}{dt}$$

Therefore,

$$Q(t) = -\frac{As}{Kbc} \frac{dQ(t)}{dt}$$

or

$$\frac{dQ(t)}{dt} + \frac{Kbc}{AS} Q(t) = 0$$

Solving the above equation with the initial condition that at $t = 0$, $Q = Q_o$, the natural discharge rate is found to be

$$Q(t) = Q_o \, e^{-\frac{Kbc}{As} t}$$

The total volume of water which is present in the storage at a time t at which the flow rate is Q_o and which will be discharged subsequently is given by

$$\int_0^\infty Q(t)dt = \int_0^\infty Q_o e^{-\frac{Kbc}{AS} t} dt$$

$$= Q_o \left. \frac{e^{-\frac{Kbc}{AS} t}}{-\frac{Kbc}{AS}} \right|_0^\infty$$

$$= Q_o \frac{AS}{Kbc}$$

If the discharge takes place at the rate Q_o, the time in which all the storage will be depleted is $\frac{AS}{Kbc}$. This time is known as the time of depletion. Let it be designated as t_o.

A final expression for discharge is

$$Q(t) = Q_o e^{-\frac{t}{t_o}}$$

t is measured from the instant Q_o is measured. Also the depletion time t_o multiplied by the current flow rate is the volume of water that remains in the live storage.

In the natural steady state prior to exploitation, the discharge rate is equal to natural replenishment i.e.,

$$Q_o = R$$

Assuming that pumping and replenishment are constant and pumping is smaller than replenishment, the final rate of drainage $Q_s = R - P_s$. During the transition period, the natural discharge rate is calculated as follows:

$$-\frac{dv}{dt} = Q(t) - (R - P_s) = -t_o \frac{dQ}{dt}$$

or

$$-\frac{dt}{t_o} = \frac{dQ}{Q(t)-(R-P_s)}$$

or

$$Q(t) = (R - P_s) + P_s e^{-t/t_o}$$

The last term on the right-hand side is outflow in excess of the final equilibrium state. The volume of water that will be lost through natural drainage until the final equilibrium state is established as

$$\int_o^{\infty} P_s e^{-t/t_o} dt = t_o P_s$$

Groundwater balance is carried out for a quantitative evaluation of groundwater resources and their change under the influence of man's activities (Sokolov and Chapman, 1974). The study of water balance in a groundwater basin forms a basis for the rational use, control and redistribution of groundwater resources in time and space (i.e., interbasin transfers, stream flow control, etc.). Knowledge of the water balance assists the prediction of the consequences of artificial changes in the regime of groundwater basins. With water balance data it is possible to compare individual sources of water in a system, over different periods of time and to establish the degree of their effect on variations in the water regime. Further, the initial analysis used to compute individual water balance components, and the co-ordination of these components in the balance equation make it possible to identify deficiencies in the distribution of observational stations, and to discover systematic errors of measurement. Finally, water balance studies provide an indirect evaluation of an unknown water balance component from the difference between known components.

REFERENCES

1. Huisman, L. (1972), 'Ground Water Recovery', Mac Millan, London.

2. Sokolov, A.A. and T.G. Chapman (1974), 'Methods for Water Balance Computations',An international guide for research and practice, The UNESCO Press.

3. Mandal, S. and Z.L. Shiftan (1981), 'Ground Water Resources Investigation and Development', Academic Press,

4. Hamill, L. and F.G. Bell (1985), 'Ground Water Resources Development' Butterworths.