Journal of Hydrology (N.Z.) Vol. 25, No. 1, 1986

MEASUREMENT AND PREDICTION OF NATURAL GROUNDWATER RECHARGE — AN OVERVIEW[1]

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ABSTRACT

Management of groundwater is often carried out with little or no knowledge of recharge rates, resulting in misuse of the resource.

As direct measurement of recharge is extremely difficult, recharges commonly are estimated indirectly. This paper gives a brief overview of recharge processes, and the merits and limitations of estimation methods, considering spatial and temporal scales of aquifers and the specific objective in question. Estimation techniques should reflect key processes of the system, and, where possible, more than one method should be used. Aspects on which information is lacking and areas for fruitful research are identified.

INTRODUCTION

Groundwater recharge is the addition of surface water to an aquifer. Quantification of recharge is basic to the optimal use of groundwater resources, and for safeguarding their quality. Recharge may be natural (through infiltration of natural precipitation or from streambeds), induced (from water reservoirs, irrigation channels and bays) or artificial, with natural recharge by far the most prominent. This paper gives an overview of the processes involved, and techniques for estimating natural groundwater recharge.

RECHARGE PROCESSES

A schematic diagram (Fig. 1) summarizes the hydrological cycle, and illustrates how various processes are interrelated, and affect the flux of water being added to groundwater storage. The prominent processes are precipitation (intensity, duration and distribution), infiltration and redistribution of water in the unsaturated zone, and evapotranspiration. These processes are affected by weather, soil and vegetation. All must be considered in estimating recharge, although their relative effects will vary under different conditions. Consideration of the processes in detail is complicated and often unnecessary, more important is determining the relative significance of various factors, so that the approach can be simplified by approximations. For example, in a humid climate where soil water is not a limiting factor, evapotranspiration (E) from a perennial grassland catchment may be approximated by the estimates of potential evapotranspiration (E_p) computed from the Penman equation. For such conditions, if surface runoff (RO) could be measured

[1] Invited paper, presented at the New Zealand Hydrological Society 1984 Symposium, Hamilton, New Zealand

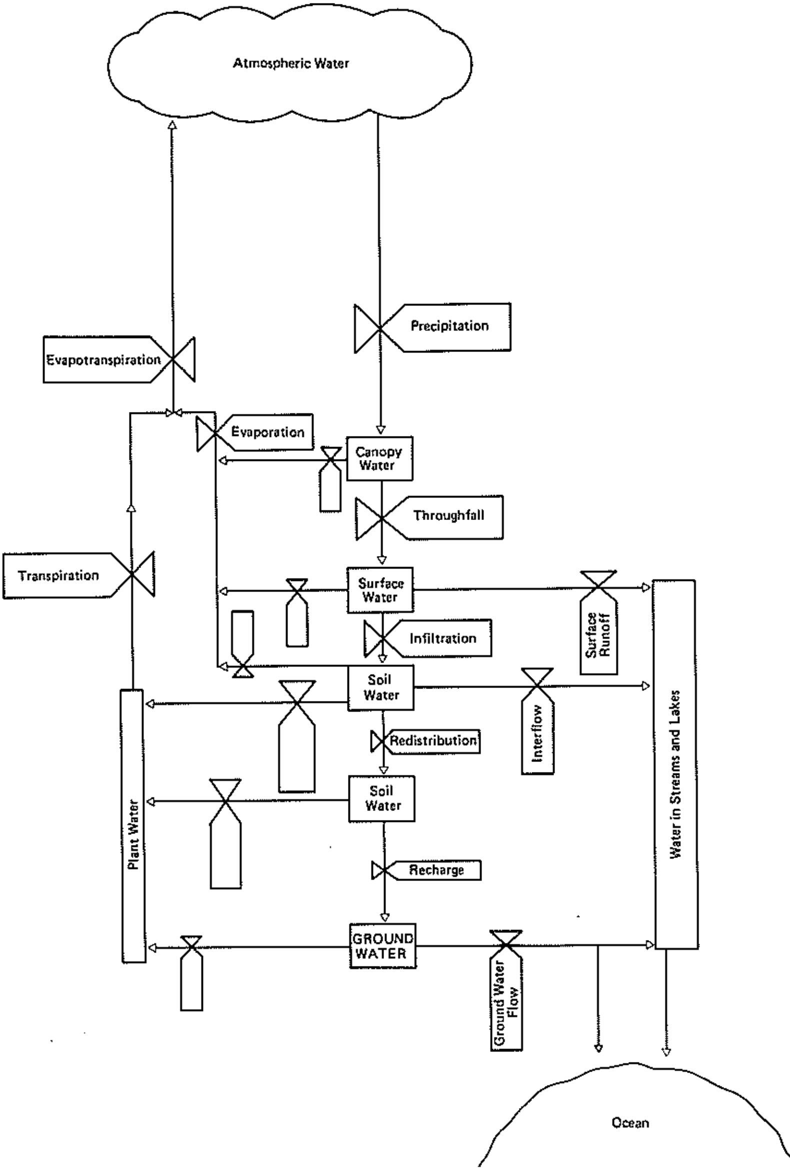

FIG. 1 — Schematic representation of the hydrological cycle showing processes involved in natural groundwater recharge. (Scheme adapted from Forrester, 1971.)

or approximated, estimates of annual recharge (R) could be made by solving the traditional water balance equation, i.e.

$$R = P - [E_p + RO] \tag{1}$$

where P is annual precipitation. Here annual change in soil water storage is negligible. However, for a forested catchment in a semi-arid region, the use of the same approach will be fraught with danger. Forest evapotranspiration rates are reduced by soil water deficit during dry periods, but enhanced considerably during winter periods due to canopy interception and low aerodynamic resistance. Evapotranspiration for such a system can be estimated better using the Penman-Monteith (Monteith, 1980) equation, i.e.

$$\lambda E = \frac{s(R_n - G) + \rho c_p\,[e_s(T_a) - e_a]/r_a}{s + \gamma\,(1 + r_s/r_a)} \tag{2}$$

$$= \frac{s(R_n - G) + \rho\, c_p\,(\delta e)/r_a}{s + \gamma\,(1 + r_s/r_a)}$$

where λ is heat of vaporization, c_p is specific heat, R_n is net radiation, G is ground heat flux, $\gamma = (c_p/\lambda)$ is the psychrometric constant, e_s is saturation vapour pressure, T_a and e_a are ambient temperature and vapour pressure, r_s and r_a are surface and aerodynamic resistance of the vegetated surface, s is slope of the saturation vapour pressure — temperature curve at T_a, and δe is the vapour pressure deficit.

Similarly, the degree of complexity required in treating infiltration and redistribution of water and its interaction with water uptake by plants will depend on the soil-plant system in question.

RECHARGE ESTIMATION

The spatial and temporal scales used to estimate recharge will depend on the objective for which the information is required. The appropriateness of a method for a given set of conditions also will be affected by cost, convenience and availability of technical competence. For example, methods appropriate for determining localized recharge (i.e. plot size $<100\ m^2$) may not be appropriate for a large region ($>10^7 m^2$); and the accuracy required of recharge estimates will also differ.

Quantitative estimation of groundwater recharge has traditionally been the concern of hydrogeologists, who consider the flow of water only in the saturated zone; the unsaturated part of the profile has largely been ignored. More recently, soil physicists and geochemists have contributed to recharge estimation using observations of the unsaturated zone. Several methods of measuring and predicting recharge are being used, and others are being developed. Recharge estimation methods may be categorized broadly into hydrological and tracer methods.

Hydrological Methods

Methods in this group are based on the hydrological analysis of either the saturated zone or the unsaturated zone; rarely are both zones considered together.

Two methods developed by hydraulic engineers and hydrologists are: 'hydrograph separation' and 'flow net analysis', which are usually appropriate for a particular scale of catchment. The method based on hydrograph separation assumes equilibrium between recharge and discharge, and estimates that part of streamflow which could be ascribed to recharge. However, it cannot indicate where in the landscape recharge takes place. Despite considerable research, the method still requires subjectivity. The 'flow net analysis' method requires knowledge of the geometry of the system and aquifer properties, which may vary spatially. In an iterative empirical mode, the latter method gives reasonable estimates of recharge. Caution is needed in translating water table rises into recharge, since factors such as air entrapment, changes in atmospheric pressure, and hydrologic influence from surrounding areas may give rise to misleading conclusions.

The 'water balance method' is probably the most commonly used, and superficially the most attractive, method. Other components of water balance are estimated either directly or indirectly, and recharge is computed by the difference as a drainage term, defined at an arbitrary depth below the root zone. This lumped parameter approach can be used to measure annual recharge or recharge for individual events, over a spatial scale ranging from a plot to a region. The method gives acceptable estimates of recharge if precipitation exceeds potential evapotranspiration. However, when evapotranspiration is of a similar magnitude to precipitation, as under semi-arid conditions, recharge estimated by this method must be treated with caution. Despite several advances (Sharma, 1985), measurement and modelling of areal evapotranspiration for such conditions is uncertain. When recharge is only a small proportion of precipitation, the uncertainty of the recharge estimated by the water balance method is magnified.

Localized recharge rates (spatial scale $<15\ m^2$) can be measured very accurately by using weighing and even non-weighing lysimeters. Using a representative soil-vegetation sample, however, is of prime concern since replications may be very expensive.

Another group of methods requiring knowledge of the hydraulic properties of soil is based on analysis of water flow, primarily in the unsaturated zone. These methods can be used on their own, or in conjunction with methods such as the water balance method. A generalized approach would be to solve Richard's three-dimensional transient water-flow equation with a source/sink term. However, because of a lack of appropriate data on hydraulic properties of soil, the flow equation is usually considered in one-dimensional form, i.e.

$$\frac{\partial \theta}{\partial t} = \frac{\partial}{\partial z}\left(K(\psi)\frac{\partial \psi}{\partial z}\right) - \frac{\partial K(\psi)}{\partial z} + S(\psi) \tag{3}$$

where θ, t and z are volumetric water content, time and profile depth respectively, K is soil hydraulic conductivity, ψ is soil water potential and S is the source/sink term. Methods to compute water uptake by plant roots (S term) require information on the distribution of root density, and soil water status as a function of time. Provided soil hydraulic properties ($K(\theta)$, $\psi(\theta)$) are characterised adequately and S can be modelled satisfactorily, water flux below the root zone can be simulated by this method. A simplified version of this approach (Rose and Stern, 1965) is to compute drainage flux below the root zone (Z) using the steady state, one-dimensional water flow equation:

$$q_w = K_Z \frac{d\psi}{dz} - K_Z \tag{4}$$

where q_w is instantaneous water flux below depth Z, $d\psi/dz$ and K_Z are the soil water potential gradient and average hydraulic conductivity at Z respectively. Under certain conditions, if a zone of zero-flux gradient (total water potential gradient being zero) can be identified within a profile, then the depletion in water storage above the zero-flux zone could be considered as evapotranspiration, and the reduction below the zone as groundwater recharge (Cooper, 1979).

The above methods assume one-dimensional water flow, which may be a reasonable approximation in flat topography with vertically uniform profiles. In strongly layered systems, two- or three-dimensional flow may need to be considered. The major weaknesses of this method are the awkwardness of measuring soil water potential gradients in deep profiles, a lack of field data on hydraulic properties of soils, and inadequate quantitative description of field variability. More research is needed on these aspects.

Tracer Methods

Estimation of recharge using tracers is based on the conservation of mass of the tracer, and the assumption that the tracer moves freely with water. Ideally a tracer should be highly soluble in water, relatively inert and foreign to the system, with negligible uptake by vegetation.

Two kinds of tracers are in use, those which occur naturally in the environment, and those which are applied. Among the natural environmental tracers are various isotopes (e.g. ^{13}C, ^{14}C, ^{3}H, ^{2}H, ^{18}O) and chemical ions (e.g. Cl^-, NO_3^-, $SO_4^=$). Examples of applied tracers are: Br^-, Cl^-, I^-, ^{3}H, rhodamine dye etc. For many purposes ^{2}H, ^{3}H and ^{18}O are preferable since they are incorporated in water molecules.

Tracers commonly are used in hydrologic studies, but most of these studies yield indirect information. The information derived may be qualitative or quantitative. Hydrological interpretation of tracer results depends, among other things, on the validity of the physical model of water flow for the system in question. Thus, the accuracy of recharge estimated using a tracer technique would depend on how realistic was a particular physical model used for interpreting the results and how realistically were the required assumptions for the model being met for the system.

Tracer studies may involve the saturated or the unsaturated zone, or both. Recharge estimates based on the use of environmental tracers (such as Cl^-, and ^{3}H) in the saturated zone, give long-term averages ($>$ year) and an areally-integrated value, which is difficult to obtain using soil physical methods. Regional recharge rates are required for resource planning. Studies based on the monitoring of tracers in unsaturated profiles usually yield point ($<0.1\ m^2$) recharge values. Thus the technique could be employed to measure spatial variability, and to delineate the effects of various factors on recharge.

The most successful non-isotopic environmental tracer in hydrologic studies is chloride, especially in coastal areas where aeolian Cl^- is deposited from winds carrying precipitation from the ocean. Chloride has been used for estimating areal recharge, based on its concentrations in the saturated zone (e.g. Eriksson and Khunakasen, 1969; Kitching *et al.*, 1980), and localized recharge by considering its depth distribution in soil water (Allison and Hughes, 1978; Kitching *et al.*, 1980; Peck *et al.*, 1981; Sharma *et al.*, 1983). As an example, average vertical water fluxes computed from Cl^- data are shown in Fig. 2. From the depth distributions of chloride concentration of soil water (C) and volumetric water content (θ) observed beneath a native woodland site in a sandy coastal region of Western Australia, time-averaged vertical water flux (q_w) was computed using a steady state water and solute flow model, i.e.

$$q_w = \frac{1}{C} [J_s + D_s \theta \frac{\partial C}{\partial z}] \quad (5)$$

where J_s is the average chloride input to the system (= $\bar{P}$ $\bar{C}_p$, where $\bar{C}_p$ is the time-averaged chloride concentration of the long-term average annual rainfall $\bar{P}$), D_s is the diffusion/dispersion coefficient of the solute, and $\partial C/\partial z$ is the slope of the observed chloride concentration with respect to depth. The computed vertical water flux below the rooting depth (~10m) of the native vegetation is interpreted as the average recharge rate to the groundwater. Complications may arise in interpreting chloride data, for example, when water movement through the profile deviates significantly from vertical, or when water flow through the profile is not steady and uniform. Even in relatively uniform, sandy profiles, some 50% of annual recharge may occur via movement of water through preferred pathways, bypassing the soil matrix (Sharma and Hughes, 1985). In some forested lateritic profiles with clayey subsoil, recharge through macropores amounts to more than 95% (Peck et al., 1981). Several aspects of the interpretation of chloride profiles still remain unresolved.

Tritium, generated by atmospheric nuclear tests, is the most common environmental isotope used in recharge studies (e.g. Smith *et al.*, 1970; Allison and Hughes, 1972). Tritium (^3H) concentrations of the saturated zone have been used to estimate the ratio of annual recharge to total groundwater storage. Localized recharge rates have been estimated from the distribution of tritium in the unsaturated part of the profile, and found to agree with other estimates, such as from the chloride method. Both ^{14}C and ^3H have been used to infer the age of groundwaters. The fact that enrichment of ^{18}O and ^2H occurs with evaporation, has stimulated hydrochemists to investigate whether depth-distributions of these isotopes could be used to estimate recharge (e.g. Allison *et al.*, 1984; Sharma and Hughes, 1985). Although more research is needed, it appears that use of $^2H/^{18}O$ is of value only in areas where direct evaporation is a major component of total evapotranspiration.

CONCLUDING REMARKS

Since it is usually difficult to assess the accuracy of recharge estimates, it is advantageous to use more than one method if possible. Recharge rates

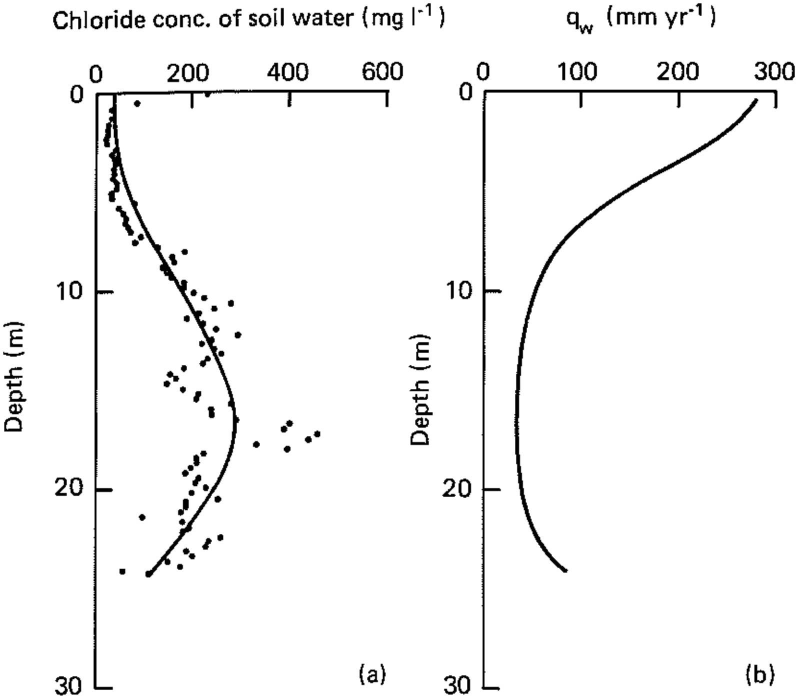

FIG. 2 — (a) Depth distribution of chloride concentration in soil water, and the cubic spline fitted to these data points. Data are for a sandy profile sampled beneath a native woodland site on the north Swan Coastal Plain, Western Australia.

(b) Depth distribution of average vertical water flux q_w (mm yr^{-1}), computed by solving Equation (5) with the data from 2 (a).

are spatially variable on a relatively small scale (1–10m), and appropriate methods are required to assess the response of aquifers at varying scales. For many systems, existence of preferred pathways of waterflow has been identified. Research is needed to quantify the spatial distribution of such pathways and their contribution to groundwater recharge.

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