Vadose-Zone Techniques for Estimating Groundwater Recharge in Arid and Semiarid Regions

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ABSTRACT

Both physical and chemical methods have been used to estimate recharge in arid and semiarid areas. Our review indicates that indirect, physical approaches, such as water balance and Darcy flux measurements, are the least successful, while methods using tracers (e.g., Cl, ^{3}H, and ^{36}Cl) have been the most successful in estimating groundwater recharge in dry regions. Lysimeters, which can directly measure root-zone drainage, have been useful in quantifying recharge, particularly for coarse soils, but are costly to construct and operate. Of the tracer techniques available, Cl balance techniques appear to be the simplest, least expensive, and most universal for recharge estimation. In Australian studies, under native vegetation in semiarid areas, Cl profiles were found to be remarkably uniform, indicating very low and relatively uniform rates of groundwater recharge. Following changes in land use, recharge appeared to become much more variable, increasing more than two orders of magnitude. Methods for scaling point estimates of recharge to large areas using indirect techniques (such as nondestructive electromagnetic induction) have also been developed. In deep unsaturated zones, the pressure response in the soil water may be recorded in the profile, and simple field measurements may be used to obtain semi-independent verification of recharge rates determined by using Cl balance techniques.

In arid and semiarid areas (where precipitation is less than ≈700 mm/yr), the native vegetation has often developed such extensive root systems that almost all of the precipitation is consumed by evapotranspiration. As a result, the local recharge flux under native vegetation is very low, but generally constant in time. Following clearing of native vegetation and establishment of comparatively shallow-rooted pastures and crops, local recharge increases. Variability of rainfall will result in recharge being highly variable with time.

In arid and semiarid regions, physical parameter methods, which rely on measurements of hydrological parameters, are problematic for several reasons. First, the fluxes are still low and changes in these parameters will be small and difficult to detect. Second, the temporal variability is such that measurements must be made for several years to obtain an estimate of mean values. Third, spatial variation brought about by local topography and soil texture changes requires a large number of sampling locations to assess recharge variability.

This review emphasizes chemical and isotopic methods for estimating local recharge in semiarid areas because they show more promise than do physical methods in such dry climates. Physical parameter methods assume more importance in wetter areas and for estimating artificial recharge where the fluxes are higher and the inputs more constant with time.

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METHODS

Recent reviews of a range of techniques for, and problems associated with, estimation of groundwater recharge have been given by Allison (1981, 1988) and Gee and Hillel (1988). In addition, Keith et al. (1982) have prepared a bibliography of recharge studies in semiarid areas. In their review on the use of environmental isotopes for modeling in hydrology, Dincer and Davis (1984) outlined some of the techniques available for recharge estimation using these isotopes.

To estimate local recharge, some techniques rely on measurements made in the unsaturated zone and, for others, in the saturated zone. One of the major problems in studying the unsaturated zone in semiarid areas is that measurements need to be made beneath the zone where uptake of water by roots is significant. For annual crops and pastures, this is not a great problem as rooting depths usually are <2 m and loss of water from depths greater than this are negligible. However, native vegetation and many trees and shrubs may have living roots to great depth. The significance of these deep roots in modifying annual water balance and recharge is uncertain.

Indirect Physical Methods

In principle, one of the simplest methods used for estimating local recharge, R, is empirical expressions of the type

$$R = k_1 (P - k_2) \qquad [1]$$

where P is precipitation and k_1 and k_2 are constants for a particular area. Such expressions have been used with varying degrees of success and are probably most useful for making "first-guess" estimates of recharge where annual recharge is fairly high, >50 mm/yr, and thus should seldom be used in arid or semiarid regions.

Methods relying on estimates of soil physical parameters fall into the following three classes: (i) soil water balance, (ii) zero-flux plane method, and (iii) estimation of water fluxes beneath the root zone using unsaturated hydraulic conductivity and the gradient in soil water potential.

These techniques will be discussed only briefly here because extensive literature is available on them.

Soil Water Balance

The soil water balance can be represented, in the absence of significant surface runoff, by

$$R = P - E_a + S \qquad [2]$$

where E_a is actual evapotranspiration and S is the change in soil water storage, which usually may be ignored if calculations are made on an annual basis. A number of techniques for estimating E_a, based on Penman-type equations (Howard and Lloyd, 1979; Rushton and Ward, 1979) and other methods (e.g. Alley, 1984), have been used. The data requirement of these methods is large and Howard and Lloyd (1979) suggested that large errors can occur unless the accounting period is <10 d. The soil water balance technique has been used extensively in temperate areas. However, it is unlikely to be successful in semiarid areas because of long periods of less than potential evaporation, when errors in E_a are greatest and P and E_a are nearly equal (Gee and Hillel, 1988).

Zero-Flux Plane Method

The zero-flux plane (ZFP) method relies on locating a plane of zero hydraulic gradient in the soil profile. Recharge during a time interval is obtained by summation of the changes in water content below the plane. Unfortunately, the method breaks down in periods of high infiltration when the hydraulic gradient becomes positive downward throughout the profile. This is when recharge fluxes are likely to be highest.

Use of this technique can give good estimates of recharge for periods during the year when the ZFP exists, as has been shown by studies in the United Kingdom (e.g., Wellings, 1984). However, it suffers from the same problems as the other two classes if annual values of recharge are required, because one of the other two methods must be used when the ZFP disappears.

Estimation of Water Fluxes

Several studies have reported attempts to use unsaturated zone hydraulic conductivity ($K(\theta)$ or $K(\psi)$, and hydraulic potential data, $\psi(\theta)$, to solve either Darcy's Law or Richards' equation in the unsaturated zone and to estimate soil water flux for periods of months to years (Sophocleous and Perry, 1985; Stephens and Knowlton, 1986). If the water flux is calculated at such a depth in the profile that no further extraction by roots occurs, then the flux will be equal to groundwater recharge:

$$R = K(\theta)\ \Delta H_t \qquad [3]$$

where ΔH_T is the total head gradient. For most soil systems, $H_T = H_g + H_m$, where H_g = gravity head and H_m = matric suction head.

Both $K(\theta)$ and $K(\psi)$ relationships are difficult and time consuming to determine, both in the field and in the laboratory, with difficulty and uncertainty increasing with soil dryness. Slight differences in measured water content translate into large differences in unsaturated hydraulic conductivity. The work of Stephens and Knowlton (1986) highlighted the problem of using these techniques for estimating groundwater recharge, especially where the fluxes are low. They found that the annual recharge flux varied by more than a factor of five (7–37 mm/yr), depending on how they computed mean hydraulic conductivity.

Sophocleous (1991, 1992) estimated recharge in the Great Bend Prairie of Kansas using a combination of physically based methods (measured precipitation, soil water tension, and water content profiles coupled with water table fluctuations). The most influential variables on recharge were total annual precipitation, average maximum soil-profile water storage in spring months, average shallowest depth to water table, and spring rainfall rate. Using geographic information systems (GIS) technology, Sophocleous (1992) developed an overlay mapping routine using available precipitation, soils, and water table data. From this, he was able to prepare area-wide maps of differing recharge regions that were in reasonable agreement with field measurements.

Direct Physical Methods

Lysimetry has long been used successfully to make direct estimates of drainage in temperate areas. Lysimeters are also widely used as a reference instrument from which evapotranspiration is measured for irrigated agriculture in arid and semiarid regions (Allen et al., 1992). Where appropriate, recharge can be estimated from lysimeter water balance. Gee et al. (1992, 1993) have successfully used lysimeters to estimate recharge at arid sites in the western USA. Drainage collected from lysimeters clearly shows that recharge can occur in desert climates, even where potential evapotranspiration exceeds precipitation by as much as a factor of 10 or more. At sites near Las Cruces, NM, and Richland, WA, lysimeters filled with coarse-textured soils, free of vegetation, were observed to drain (recharge) as much as one-half of the annual precipitation under strikingly different climatic conditions (Gee et al., 1993).

Lysimeters are expensive and permanent instruments, and typically are filled with disturbed soils, which generally have water content profiles that differ in some degree from those found in surrounding soils. Drainage can occur only when a water table develops at the base of the lysimeter, unless solution samplers (e.g., ceramic cup extractors) and a vacuum system are installed at the base of the lysimeter. This last factor, however, is unlikely to be a problem if the lysimeter is relatively deep and the vegetation is shallow rooted (Kitching and Shearer, 1982). While lysimeters have been useful in quantifying drainage at waste sites under arid conditions, they are limited in their ability to document the spatial variability produced by natural and human-induced changes in surface and subsurface flow pathways. Construction cost and logistics limit size and depth to generally no more than a few square meters of areal extent and seldom more than a 3-m depth, although lysimeters as deep as 18 m have been constructed (Allen et al., 1991; Gee et al., 1992, 1993).

All of the physical techniques (both direct and indirect methods) discussed above have limitations in arid and semiarid areas, principally because recharge fluxes are likely to be low, on the order of a few millimeters or less, necessitating long periods of measurement. Lysimetry, however, has the potential to overcome problems of low flux, if lysimeters are large enough and monitoring is long enough. Remaining problems for physical techniques include measurement uncertainties, expense, and inadvertent modification of the soil moisture regime.

Natural Tracers for Estimation of Local Recharge

Natural tracers have both advantages and disadvantages over physical methods. On the positive side, they represent a spatially uniform (at least to a first approximation) input to the soil water–groundwater system. In many cases a history of tracer fallout in precipitation is known, and, because the aquifer and the overlying unsaturated zone usually store sufficient water to represent many years of recharge, a historical record of recharge may be derived. The principal disadvantage of isotopes is that they may offer only an indirect measure of recharge and mechanisms of infiltration will affect the interpretation of results. However, the use of multiple tracers can often provide corroborative information needed for correct interpretation.

The natural tracers most commonly used in recharge studies are ^{3}H, ^{14}C, ^{36}Cl, ^{15}N, ^{18}O, ^{2}H, ^{13}C, and Cl. Of these, the first three are radioactive, with half lives of 12.3, 5700, and 301,000 yr, respectively. Their concentrations in the hydrologic cycle have been modified greatly by nuclear testing. Both ^{3}H and ^{36}Cl from atmospheric testing have been used for soil water tracing and recharge studies (Zimmerman et al., 1967; Gvirtzman and Margaritz, 1986; Phillips et al., 1988). Chlorine-36 has been used increasingly as more analytical facilities have become available. Input concentrations of the other isotopes mentioned above have also changed in time, but across a much longer time scale, due to changes in temperature and rainfall patterns. As far as we are aware, little is known of the temporal changes in the fallout of Cl.

Of the tracers mentioned above, ^{3}H, ^{2}H, and ^{18}O most accurately simulate the movement of water because they form part of the water molecule. In most soils, ^{36}Cl and NO_3 move as the water does, but in some heavier textured soils anion exclusion may be a problem (Gvirtzman and Margaritz, 1986), and the tracer may move more rapidly than the water being traced.

Although piston flow is often able to explain the behavior of tracers in the field, there is mounting evidence (particularly at more humid sites) that water movement along preferred pathways is the rule rather than the exception (Gish and Shirmohammadi, 1991). Thus, non-piston-type flow must be dealt with in any comprehensive analysis of recharge. Using isotopic analysis, Allison and Hughes (1983) found ^3H much deeper than the recharge rate would imply in native forest, suggesting preferred flow of water along root channels.

Three techniques have been used for estimating recharge rates from tracer profiles in the unsaturated zone.

1. *From the position of the tracer peak.* The water in the profile above the peak in tracer concentration represents recharge since the time that peak occurred. Any bypass (preferential) flow will result in recharge being underestimated. It is recommended that the amount of water in the profile be estimated at the time of year that the soil moisture deficit is at its maximum.
2. *From the shape of the tracer profile in the soil.* This is generally more reliable than Method 1 since information about flow mechanisms can be obtained. In order to obtain estimates of mean annual recharge, $\bar{R}$, a weighting function taking into account year-to-year variations of recharge, is needed. A comparison of several possible weighting schemes was given by Allison and Hughes (1978).
3. *From the total amount of tracer, T_t, stored in the profile.* This is given by

$$T_t = \int_0^x T(z)\ \theta(z)\ dz \qquad [4]$$

where $T(z)$ is the tracer concentration of water in the unsaturated zone at a distance z beneath the surface and $\theta(z)$ is the volumetric water content. For evaporative tracers, such as ^3H, mean annual recharge can be estimated by

$$\bar{R} = T_t/\sum_{i=1}^{x} w_i\ T_{pi} \exp(-t\lambda) \qquad [5]$$

where T_{pi} is the tracer concentration of recharge water i years before the present; w_i is the annual recharge weighting factors, and λ is the tracer decay constant. In this analysis nonpiston flow can be handled since $\bar{R}$ is independent of the distribution of the tracer in the profile.

Tritium

Many studies on the estimation of recharge using natural ^3H in the unsaturated zone have been given in the literature (Schmalz and Polzer, 1969; Smith et al., 1970; Allison and Hughes, 1974; Foster and Smith-Carington, 1980; Tyler et al., 1992). Almost all studies reported have made use of the fact that the peak in ^3H in precipitation due to nuclear fallout has been preserved in the unsaturated zone.

To estimate recharge from ^3H profiles, the ^3H concentration of effective input to the soil water system must be known. Work by Gvirtzman and Margaritz (1986) and Gvirtzman et al. (1986) has demonstrated the usefulness of ^3H profiles for estimating recharge beneath an irrigated area. Their ^3H profiles show remarkable preservation with depth and enable annual recharge cycles to be seen. The profile results from differences in ^3H of irrigation water (summer) and rainfall (winter). Such discrimination indicates a very low effective diffusion coefficient for ^3H in this soil. The data also suggested that a very high percentage ($\approx$50%) of immobile water existed in the draining soil that was not readily exchanged with the mobile water. This led to ^3H profiles quite different from those predicted by piston flow.

The techniques for measuring ^3H are not valid in areas where the root zone is deep, and interpretation of data may become difficult if flow regimes are not found to be reasonably uniform. Also, it has been suggested that, in arid areas where soils are sandy and have a high air-filled porosity, it is possible for water vapor of high ^3H concentration to diffuse into soil water or groundwater systems. Estimates of recharge made using ^3H in such areas therefore would be anomalously high. However, this is significant only at $R \leq 1$ mm/yr. Careful consideration should be given to this possibility when interpreting data such as those given by Dincer et al. (1974).

Chlorine-36

In the last decade, a number of studies (Norris et al., 1987; Phillips et al., 1984, 1988; Scanlon, 1992; Cecil et al., 1992; Walker et al., 1992) have presented soil water profiles of bomb-pulse ^{36}Cl. In studies under arid or semiarid conditions (Phillips et al., 1988; Scanlon, 1992), the ^{36}Cl profile appeared to match that of the input signal; however, the pulse was still very near the soil surface. For example, Phillips et al. (1988) found that the ^{36}Cl fallout peak at one site was very closely reflected in the soil water ^{36}Cl and was still within the root zone at about 1-m depth. A Cl mass balance at their site gave a recharge flux of $\approx$0.03 mm/yr at a depth of 5 m beneath the surface. Using the depth to the ^{36}Cl peak resulted in an estimated recharge of 2.6 mm/yr. A cumulative Cl depth profile suggests that the bottom of the effective root zone at their site was $\approx$1.5 m and that it would take >125 yr for the bomb peak in ^{36}Cl to reach this depth. Scanlon (1992) reported a ^{36}Cl profile from an unvegetated arroyo. The ^{36}Cl was found at 0.5-m depth, suggesting a flux to that depth of 1.4 mm/yr. At the same site, the Cl mass balance, however, suggested a net recharge of 0.04 mm/yr at a depth of 2 m. Thus the bomb peak of ^{36}Cl, like ^3H, is not an ideal tracer for areas of low recharge or in areas of changing land use because it can still retain the imprint of the root zone velocity. It will probably be best suited for regions where the local recharge is expected to be higher than $\approx$30 to 50 mm/yr, but this will depend on the water-holding capacity of the soil and the rooting depth.

Tyler and Walker (1993) presented a simplified model of water transport and root water uptake in the root zone. Their model illustrates that fallout tracers (such as ^{36}Cl) found in the root zone probably overpredict recharge rates unless the recharge rate is a significant fraction of the precipitation. This suggests that, for arid sites, where recharge is often <0.1 times the precipitation, ^{36}Cl would not be the tracer of choice since it will probably overpredict recharge. Other drawbacks to fallout ^{36}Cl are the difficulty in analyzing the near-background concentrations and the cost of analysis (only a few machines are available for the specialized analysis required and the cost-per-sample ranges from $500 to $1000 or more).

Chloride

Input of Cl occurs at the soil surface both in rainfall and as dry fallout. Samples collected from most stations include both wet and dry fallout, although the estimate of dry deposition may be modified by the collection vessel. The Cl may be of either atmospheric or terrestrial origin but, for the hydrologic studies described here, the net accession is required. Thus in areas that are laterally uniform it is possible that only Cl of oceanic origin need be considered. Hutton (1976) showed that Cl of atmospheric origin dominated up to 300 km from the coast in southern Australia. Blackburn and McLeod (1983) suggested that $\approx$50% of the Cl deposition at distances on the order of 700 km from the coast was of oceanic origin.

Most plant species do not take up significant quantities of Cl from soil water, thus Cl is concentrated by evapotranspiration in the root zone. In the simplest case of piston flow of water in the unsaturated zone, the Cl concentration in soil water should increase through the root zone to a constant value (Gardner, 1967). Provided the water table is deep or has the same concentration as the soil water, a depth–concentration profile such as shown in Fig. 1a should result. Under steady-state conditions, following Eriksson and Khunakasem (1969),

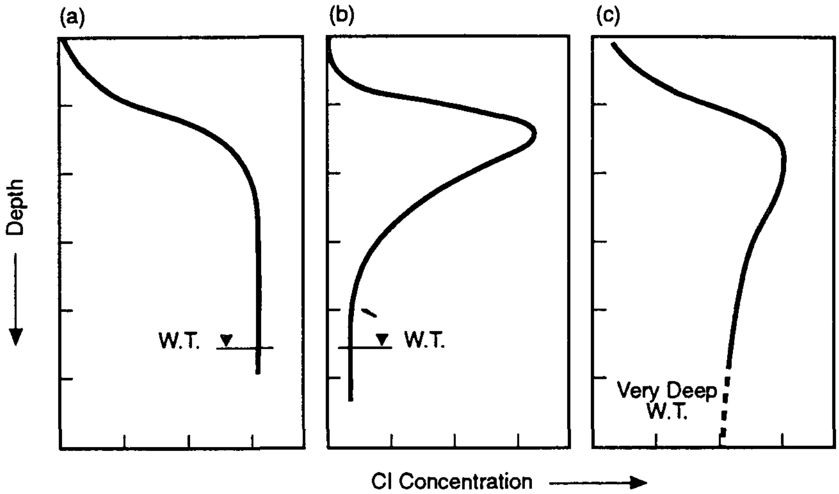

Fig. 1. Schematic depth profiles of the Cl concentration of soil water: (a) piston flow with extraction of water by roots; (b) extraction of water by roots, but with either preferred flow of water through and below the root zone, or diffusive loss of Cl to the water table (W.T.); and (c) a profile reflecting past (paleo-) recharge conditions (after Allison, 1988).

the flux of Cl at the surface is equal to the flux of Cl beneath the root zone, thus

$$\overline{P}\,\overline{C}_p = \overline{R}\,\overline{C}_s \qquad [6]$$

where $\overline{P}$ and $\overline{R}$ are mean annual precipitation and recharge, and $\overline{C}_p$ and $\overline{C}_s$ are the mean Cl concentrations in precipitation and soil water, respectively.

A more comprehensive model that describes water movement in terms of piston flow below a surface mixing layer has been used by Cook et al. (1992a) to estimate paleorecharge and paleoclimate from vadose-zone profile data. They reported that Cl profiles from Cyprus and northern Senegal appeared to show changes in recharge rates for periods up to 400 yr with variations in Cl well correlated with known variations in rainfall and fluctuations in lake levels.

The Cl mass balance technique in the unsaturated zone has been used successfully to evaluate recharge in a range of environments (Edmunds and Walton, 1980; Allison and Hughes, 1978; Sharma and Hughes, 1985). However, many of the depth profiles of Cl concentration in soil water show a more complex shape than that shown in Fig. 1a. Some idealized examples of these more complex shapes are given for comparison in Fig. 1b and 1c. Examples of a range of Cl profiles are given by Dimmock et al. (1974), Peck et al. (1981), and Phillips (1993). Several explanations have been offered for the more complex Cl profiles. Bulges in deep Cl profiles have been attributed to preferential flow (Peck et al., 1981), to diffusion of Cl to the water table (Cook et al., 1989a), and to paleoclimate-induced changes in recharge (Stone, 1992).

As mentioned above, perennial native vegetation usually develops much deeper root systems than the annual vegetation that often replaces it following agricultural development. Allison and Hughes (1983) found Cl profiles of the type shown in Fig. 1c beneath native eucalyptus vegetation where maximum Cl concentrations in the soil solution were ≈14 000 g/m^3. Very similar profiles were found at several locations beneath the native vegetation. Chloride profiles taken beneath land cleared for agriculture (≈80 yr previously) also showed very similar profiles, but with the very high concentrations displaced deeper in the profile, suggesting a piston flow process had occurred at this site. Allison and Hughes (1983) estimated that recharge at increased from <0.1 mm/yr (under native vegetation) to ≈3 mm/yr following land clearing. A possible explanation of the shape of the chloride profile shown as Fig. 1c is that it represents a record of the change in recharge with time (Allison et al., 1985). Much higher recharge in the past could give rise to lower Cl concentrations deeper in the profile. Stone (1992) suggested that paleoclimate records support the hypothesis that ancient recharge differed significantly from present in the Murray Basin, South Australia, and that observed Cl profiles with distinct bulges corroborate the paleorecord. In contrast, Cook et al. (1989a) attributed the shape of Cl profiles (similar to Fig. 1c) simply to slow diffusion of Cl from below a deep (>20 m) root zone to the underlying water table.

Oxygen-18 and Deuterium

These stable isotopes have proved particularly valuable for determining the origin of groundwater; consequently, hundreds of studies have been reported on this topic. The variation of the isotopic composition of rainfall with elevation has been used (Arnason, 1977). Variations of isotopic composition with rainfall intensity and the changes that occur following evaporation have led to a determination of the possible sources of groundwater in arid areas (Gat, 1984; Gat and Naor, 1979; Issar et al., 1984).

Relatively few studies using stable isotopes have been carried out in the unsaturated zone. Thoma et al. (1979) found that the seasonal variations of the ^2H composition of rainfall were preserved in a sand dune in France. A knowledge of water content would then enable estimates of recharge to be made. Bath et al. (1982) and Saxena and Dressie (1984) found that some of their profiles of ^{18}O and ^2H in soil water showed cyclical variations in depth that corresponded to seasonal variations in rainfall. Knowledge of water content and an assumption of piston flow enabled estimates of local recharge to be made.

The three studies referred to above were all carried out in temperate areas where recharge is ≈200 mm/yr or more. In drier climates, strongly positive values of the displacement of either ^2H or ^{18}O concentration from the concentration found in meteoric (rain) water (expressed as δ_2 or δ_{18}) may occur near the surface due to evaporation through the soil surface (Munnich et al., 1980), leading to the possibility of identifying an annual marker in the soil water. However, depth profiles of δ_{18} or ^2H in soil water, shown in Fig. 2, where the mean annual recharge, R, is estimated from Cl profiles to be ≈50 and ≈3 mm/yr, do not show cyclic variations in either of these isotopes. The most likely reason for the absence of such peaks in the soil water profile is that, if piston flow occurs, the low recharge will ensure that peaks and troughs in isotopic values

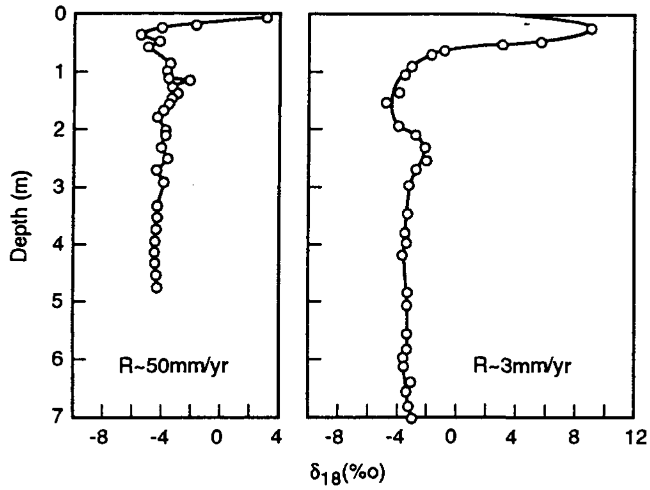

Fig. 2. Depth profiles of ^{18}O δ-values (δ_{18}) in soil water at two sites in southern Australia where the mean annual recharge is 50 and 3 mm/yr, respectively (after Allison, 1988).

will be separated by only short distances in the soil profile, and diffusional redistribution in both liquid and vapor phases will occur. Allison et al. (1984) developed a technique based on the displacement of the δ_{18} and δ_2 values of soil water from the local meteoric line, which may enable estimates of local recharge to be made under conditions of low recharge. They showed that the displacement of δ_2 from the meteoric line should be linearly related to $R^{-1/2}$ and found this relationship to hold for the four sites they studied. The nature of this suggests more precise estimates as recharge becomes lower, but the technique requires further testing. Barnes and Allison (1988) developed detailed models for the movement of stable isotopes in both the liquid and vapor phases in soils.

Nitrate

Another involuntary tracer that may be used to give information on the rate of water movement in the unsaturated zone is NO_3, which has come into increasing agricultural use in many areas since about 1950. Because some of this added NO_3 is leached below the root zone, the change from higher to lower concentrations of NO_3 in soil water at depth is an indication of the position in the profile of recharge originating at the time of increased use of NO_3. A knowledge of the amount of water stored in the profile should then enable an estimate of R (see also Barnes et al., 1993). In some situations the reverse occurs where, following clearing, NO_3 associated with the native vegetation may possibly be a marker associated with the time of clearing.

Applied Tracers

In contrast with the tracers discussed above, which are input each year to the unsaturated zone in precipitation, use of an applied tracer necessitates a one-time application, followed by sampling of the tracer pulse with time. Ideally, the tracer should be applied beneath the root zone and a sufficient time allowed to elapse between injection and sampling to allow the depth interval traversed by the tracer peak to be measured accurately. These conditions make the technique valuable in temperate regions where the rooting depth is shallow and recharge rates are high. In semiarid regions where roots are usually deeper and recharge fluxes lower, this technique will be less useful and will not be discussed in detail here. Artificially added tracers that have been used with success are ^{18}O, ^{2}H, ^{3}H, and Br (Blume et al., 1967; Zimmerman et al., 1967; Saxena and Dressie, 1984; Athavale et al., 1980; Sharma et al., 1985). Sharma et al. (1985) applied Br at the soil surface, and their data show strong evidence of nonpiston flow for several meters depth, through the root zone of native vegetation in a sandy soil. McCord and Stephens (1987) applied Br at shallow depths on the crown (top), side slope, and toe (bottom) of a sand dune in New Mexico and observed significant differences in downward displacement (recharge) of the Br tracer. Little lateral spreading occurred at the crown or toe location, while significant lateral displacement was observed on the side slope. Differences in shape and extent of the Br plume were attributed to topographic (slope) features and moisture-dependent anisotropy of the nearly uniform, but slightly layered sand.

Observed Solute and Pressure Fronts in Fields with Altered Land Use

Clearing of native eucalyptus vegetation in the Murray River Basin in South Australia has resulted in large increases in soil-water flux. This is due to the lower water requirements and shallower depths of rooting of the crop and pasture species grown following clearing. It is estimated that fluxes beneath the root zones (i.e., groundwater recharge) have increased from ≈0.05 to 0.20 mm/yr (Allison and Hughes, 1983; Leaney and Allison, 1986) to ≈3 to 40 mm/yr (Allison and Hughes, 1983; Allison et al., 1985).

Jolly et al. (1989) and Walker et al. (1991) used a combination of tracer and physical methods to estimate recharge. Matric suction profiles (obtained using the filter-paper method of Greacen et al., 1989) were combined with Cl profile data and used to interpret the changes in recharge rates due to land clearing. Cores were taken from two holes, one drilled in a site with native vegetation and the other from a cleared field nearby. Figures 3a and 3b show the profiles of water content, Cl, and matric suction. The Fig. 3a profile is typical of those found under native eucalyptus vegetation in this region. The mean annual recharge at this site, calculated using the Cl technique of Allison and Hughes (1983), is 0.08 mm/yr. In contrast, Fig. 3b shows water content, Cl, and matric suction profiles for an adjacent field, which was cleared for 9 yr. A pressure front has developed to a depth of ≈7.5 m and the solute front is lower in the profile (now between 4 and 5 m) than that under native vegetation. Using the technique of Allison and Hughes (1983), the postclearing recharge rate is estimated to be approximately 45 mm/yr. Assuming that the pressure and solute fronts were at the surface and 2 m, respectively, prior to clearing, then the ratio of the rate of movement of the solute front to that of the pressure front is between 0.27 and 0.40.

Evidence of preferred pathway flow, such as reported by Johnston et al. (1983), was not found in any of the cores taken at this site and a ratio of 0.33 for solute to pressure is in reasonable agreement with the theory developed by Warrick et al. (1971) for the transfer of solute in unsaturated soils that do not exhibit preferential flow. The solute front represents the boundary between water stored in the soil before clearing and water that has entered since clearing. The water that has entered the profile postclearing has moved downward and displaced the preclearing water ahead of it. This has resulted in the downward propagation of a pressure front ahead of the solute front. Below the pressure front, the soil water is still moving at the preclearing rate, while the soil water above the pressure front is moving at the new rate. When the pressure front reaches the water table, water of high salt concentration will be added to the aquifer at the postclearing recharge rate. Such observations provide clear evidence for the deleterious effects of land clearing on the salinization of groundwater in the Murray River Basin (Allison et al., 1990).

A comprehensive analysis has been developed by Walker et al. (1991) to estimate increases in recharge following land clearing. The method, which accounts for transient movement of Cl, does not rely on piston-flow assumptions. It uses mea-

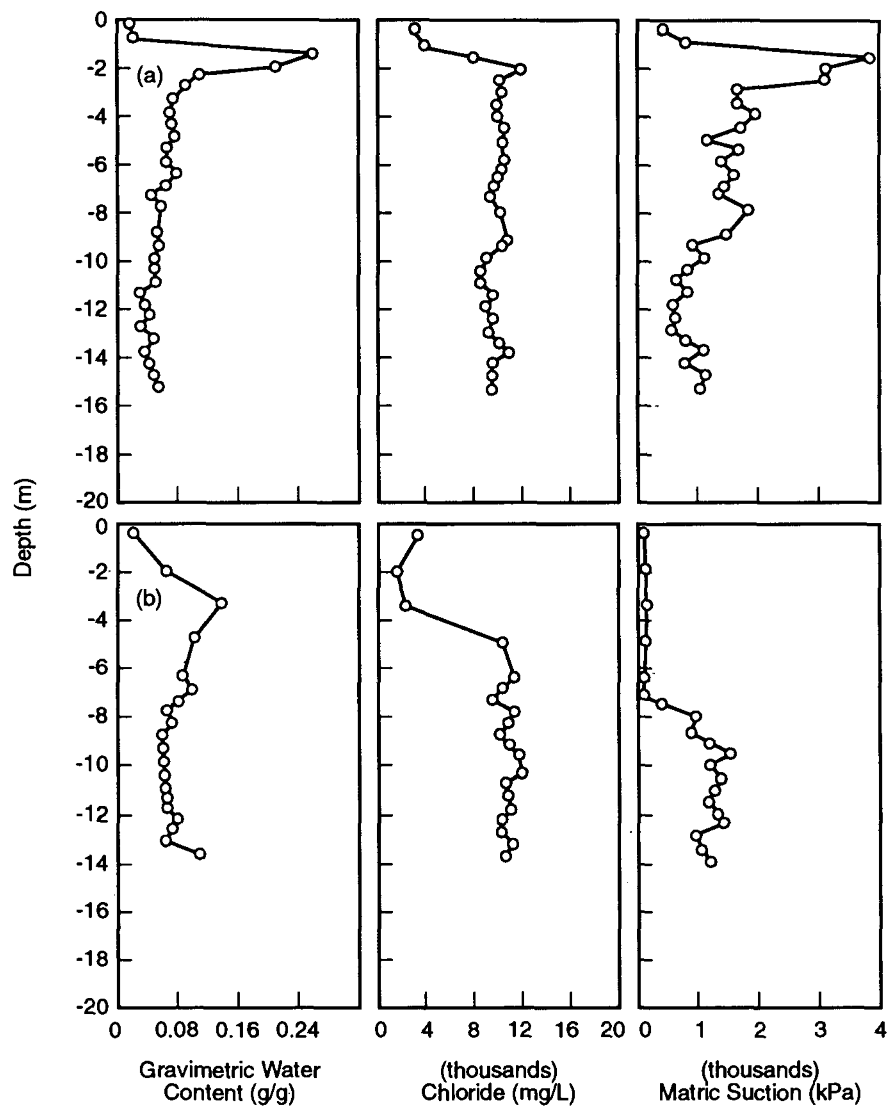

Fig. 3. Gravimetric water content, Cl in soil water, and matric suction profiles of a core hole drilled (a) beneath eucalyptus vegetation and (b) in a field cleared of native eucalyptus vegetation for 9 yr (after Jolly et al., 1989).

sured Cl and matric suction profiles to quantify drainage below the root zone. Test cases from a semiarid region of South Australia, where land clearing had influenced recharge, demonstrated the utility of the method.

In an extreme case of altered land use, Tyler et al. (1992) observed significant soil water flux beneath a nuclear subsidence crater on the Nevada Test Site. Subsidence craters, some in excess of 200 m in diameter, have formed as a result of belowground nuclear detonations. While the region is generally arid (annual precipitation is approximately 150 mm), periodic ponding and the presence of riparian species at the bottoms of the craters suggest that the craters may be localized sources of recharge. Borehole data from beneath a crater formed in 1971 revealed post-1952 soil water ^3H to a depth of 36 m and wetting front in excess of 50 m. No ^3H was observed below a depth of 3.4 m in the undisturbed soils adjacent to the crater, suggesting that the soil water flux under natural conditions was small. By applying several transport models, Tyler et al. (1992) calculated the local recharge flux below the crater to be approximately 0.6 m/yr, or four times in excess of the local precipitation. It was postulated that the altered surface topography, coupled with the coarse-textured alluvial soils beneath the crater, allowed deep infiltration of ephemeral runoff concentrated at the bottom of the crater.

While nuclear subsidence craters may be a somewhat anomalous alteration of the soil surface, analogies may be drawn to other more natural arid land forms. Any features that tend to concentrate ephemeral runoff into areas of permeable soils are likely sites for recharge. It is therefore likely that landforms such as arroyos, washes, and some closed basin features (desiccated playas) may act as local sources of recharge in spite of the regional aridity.

Variability of Local Recharge

It is now realized that there can be considerable variation in rates of local recharge over the scale of a few meters in many soil types (Johnston, 1987; Sharma and Hughes, 1985). Figure 4 shows four profiles of Cl concentration in soil water from what appears to be a uniform field, where the surface soils were sands and sandy loams. All cores were taken within a circle of 100-m radius, and it is believed that the large variations in Cl concentration reflect correspondingly large changes in recharge rate (Cook et al., 1989a). The native vegetation was removed from the field about 50 yr ago and information from this field suggest that Cl profiles, and hence recharge, are quite spatially variable. Other sites are equally nonuniform (Allison and Hughes, 1983).

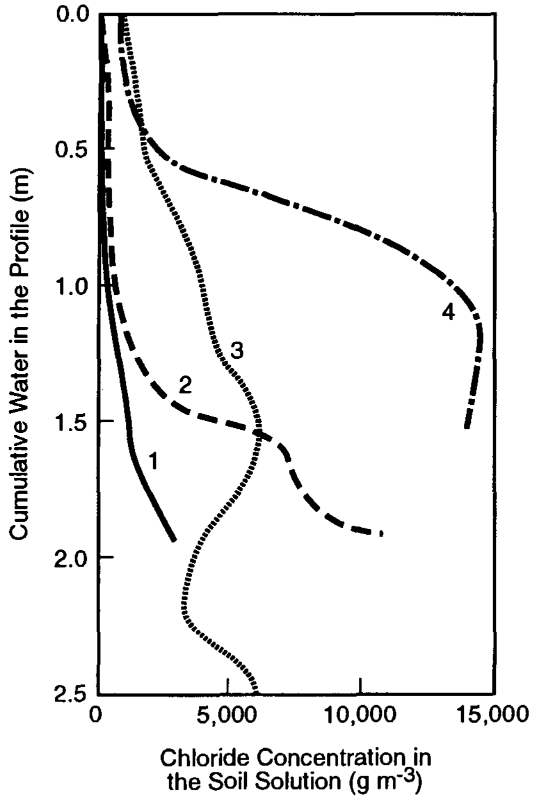

Fig. 4. The relationship between the Cl concentration of soil water and cumulative water stored in the profile in a field, cleared about 50 yr ago, under a pasture–crop rotation. Rainfall is ≈300 mm/yr. All sites are within a circle of 100-m radius. Curve numbers reflect different cores (after Allison, 1988).

The reason for this variability is unclear. It may be due to changes in the water-holding capacity of the soil, preferential flow (from deep cracks and root channels), or, alternatively, the infiltration characteristics of the surface may be such that runoff followed by localized recharge may occur. Whatever the reason, identification, quantification, and development of statistical and probably soil and geomorphological techniques for estimating recharge across a sizable area remain seminal problems in recharge evaluation.

Small-scale variability of local recharge is not always a problem. Athavale et al. (1980) were able to make what appears to be a reliable estimate of local recharge for an area of ≈1600 km^2 in India from estimates at 28 sites. Allison and Hughes (1978) found that natural ^{3}H profiles and, hence, estimates of local recharge, varied from one soil type to another, but did not vary greatly within a soil type. They were able to use ^{3}H profiles together with a soil map to estimate total local recharge to an area of similar size as the Indian work. Foster and Smith-Carington (1980) found that ^{3}H profiles up to 10 km apart in Norfolk, England, showed "remarkable lateral uniformity."

At present it appears impossible to predict areas where spatial variability will be high. Therefore, there seems to be no alternative but to make estimates of local recharge at many sites within an area of interest and to then assess the degree of spatial variability.

Assessments of Spatial Variations in Recharge

Frequency-domain electromagnetic (EM) or transient electromagnetic (TEM) methods have been used with success to assess soil salinity (Williams and Baker, 1982; Buselli et al., 1986). However, because these techniques respond primarily to soil electrical conductivity, the best correlation between soil conductivity, as measured by the instrument, and some field characteristic, is expected to be a function of both water content and salt concentration. Unfortunately, the magnitude of the recharge flux is inversely proportional to the Cl (or salt) concentration in soil water and not directly related to soil water content. Cook et al. (1989b, 1992b) using a portable EM unit, developed a calibrated model of soil conductivity using the EM unit in a field with widely varying recharge and found a good fit between modeled and measured values. Their results suggest that EM methods are promising for obtaining spatial variations in recharge when coupled with a Cl mass balance method.

DISCUSSION AND CONCLUSIONS

This review has concentrated on methods to estimate recharge in areas of reasonably low rainfall (<700 mm/yr). In general, the greater the aridity, the smaller and potentially more variable the recharge flux. When recharge rates are only a few millimeters per year or less, chemical and isotopic methods are likely to be more successful than physical methods, such as water balance methods, which rely on measured or estimated values of water flux. Water flux estimates are often in error by as much as an order of magnitude or more. An advantage of tracers is that they integrate all of the processes that combine to affect water flow in the unsaturated zone. Tracer behavior is generally a much more robust indicator of water movement in a porous medium than is the solution of the equations of water flow, especially when soils are relatively dry (e.g., for arid sites).

In estimating recharge, a method such as ^{3}H profiling relies on estimating the amount of the tracer beneath the soil surface, thus the precision of the estimate of recharge will increase with recharge rate. In contrast, for a tracer such as Cl, the concentration of which is inversely proportional to recharge rate, the precision of the estimate will increase with decreasing recharge rate. Combining Cl (tracer) and suction profile (physical) information, at sites where changing land use has significantly altered recharge rates has been useful in quantifying both past and present recharge rates at such locations.

Probably the most difficult and important problem to be overcome in estimating recharge is the prediction and assessment of its variability. Across some quite large areas it appears to show little lateral variability, while in other, apparently similar, areas it can range across at least an order of magnitude. Use of rapid and nondestructive techniques such as EM methods may offer some hope here.

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