Using groundwater levels to estimate recharge

Richard W. Healy · Peter G. Cook

Abstract Accurate estimation of groundwater recharge is extremely important for proper management of groundwater systems. Many different approaches exist for estimating recharge. This paper presents a review of methods that are based on groundwater-level data. The water-table fluctuation method may be the most widely used technique for estimating recharge; it requires knowledge of specific yield and changes in water levels over time. Advantages of this approach include its simplicity and an insensitivity to the mechanism by which water moves through the unsaturated zone. Uncertainty in estimates generated by this method relate to the limited accuracy with which specific yield can be determined and to the extent to which assumptions inherent in the method are valid. Other methods that use water levels (mostly based on the Darcy equation) are also described. The theory underlying the methods is explained. Examples from the literature are used to illustrate applications of the different methods.

Résumé Une estimation précise de la recharge des nappes est extrêmement importante pour une gestion appropriée des systèmes aquifères. Il existe de nombreuses approches différentes pour estimer la recharge. Cet article passe une revue les méthodes basées sur les données piézométriques de nappes. La méthode d'analyse des fluctuations de nappe est probablement la méthode la plus largement utilisée pour estimer la recharge; elle nécessite la connaissance du rendement spécifique et des variations du niveau de la nappe au cours du temps. L'intérêt de cette approche tient en sa simplicité et dans son insensibilité au mécanisme qui fait s'écouler l'eau dans la zone non saturée. L'incertitude sur les estimations introduite par cette méthode est liée à la précision limitée avec laquelle le rendement spécifique peut être déterminé et à l'étendue de la validité des hypothèses inhérentes à la méthode. D'autres méthodes qui recourent aux niveaux piézométriques, pour la plupart basées sur l'équation de Darcy, sont également décrites, et la théorie supportant chacune de ces méthodes est expliquée. Des exemples tirés de la littérature sont utilisés pour illustrer des applications des différentes méthodes.

Resumen La estimación precisa de la recarga es de vital importancia para gestionar adecuadamente un acuífero. Existe numerosos enfoques para ello, si bien en este artículo se presenta una revisión de los métodos basados en datos de niveles piezométricos. El método de la fluctuación del nivel freático es quizás el más utilizado para estimar la recarga; requiere el conocimiento del coeficiente de almacenamiento específico y de las variaciones temporales del nivel. Entre sus ventajas, cabe citar su sencillez e independencia respecto al mecanismo de desplazamiento del agua en la zona no saturada. La incertidumbre en las estimaciones obtenidas con este método están relacionadas con la limitada precisión con que se puede determinar el coeficiente de almacenamiento específico y con la validez de las hipótesis de partida. Se describe también otros métodos basados en datos de nivel (la mayoría utiliza la ley de Darcy) y los conceptos teóricos de cada uno de ellos. Se ilustra la aplicabilidad de este grupo de métodos por medio de ejemplos recogidos en la bibliografía.

Keywords Groundwater recharge · Groundwater hydraulics · Groundwater flow · General hydrogeology

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Introduction

Recharge has been defined as "the entry into the saturated zone of water made available at the water-table surface, together with the associated flow away from the water table within the saturated zone" (Freeze and Cherry 1979). Groundwater recharge is a key component in any model of groundwater flow or contaminant transport. Accurate quantification of recharge rates is imperative to

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proper management and protection of valuable groundwater resources. A multitude of methods has been used to estimate recharge. These methods produce estimates over various time and space scales and encompass a wide range of complexity and expense. Information on different methods is contained in references such as Simmers (1988, 1997), Sharma (1989), Lerner et al. (1990), and Scanlon et al. (2002). Unfortunately, given the current state of the science, it is extremely difficult to assess the accuracy of any method. For this reason, it is highly beneficial to apply multiple methods of estimation and hope for some consistency in results – even though consistency, by itself, should not be taken as an indication of accuracy. Techniques based on groundwater levels are among the most widely-applied methods for estimating recharge rates. This is likely due to the abundance of available groundwater-level data and the simplicity of estimating recharge rates from temporal fluctuations or spatial patterns of groundwater levels.

This paper reviews methods for estimating groundwater recharge that are based on knowledge of groundwater levels. Most of the discussion is devoted to the use of fluctuations in groundwater levels over time to estimate recharge. This approach is termed the water-table fluctuation (WTF) method and is applicable only to unconfined aquifers. In addition to monitoring of water levels in one or more wells or piezometers, an estimate of specific yield is required. Other methods, addressed in less detail, include an approach developed by Theis (1937) that is based on Darcy's equation and requires knowledge of hydraulic conductivity and hydraulic gradient; Hantush's (1956) method for estimating interaquifer flow; a method derived from an analytical solution of the Boussinesq equation; and an approach that uses transform models and requires precipitation data.

The theory underlying the WTF and other methods is presented in detail. Mechanisms that cause water-level fluctuations in unconfined aquifers are closely examined. The concept of specific yield is scrutinized and techniques for its measurement (theoretical, laboratory, and field methods) are reviewed. A separate section is included on the WTF method as applied to fractured-rock systems. Detailed examples of all methods are presented to illustrate their applications.

Water-Table Fluctuation Method

As background, consider the groundwater budget for a basin. Changes in subsurface water storage can be attributed to recharge and groundwater flow into the basin minus baseflow (groundwater discharge to streams or springs), evapotranspiration from groundwater, and groundwater flow out of the basin (Schict and Walton 1961). The budget can be written as:

$$R = \Delta S^{\mathrm{gw}} + Q^{\mathrm{bf}} + ET^{\mathrm{gw}} + Q^{\mathrm{gw}}_{\mathrm{off}} - Q^{\mathrm{gw}}_{\mathrm{on}} \quad (1)$$

where R is recharge, ΔS^{gw} is change in subsurface storage, Q^{bf} is baseflow, ET^{gw} is evapotranspiration from groundwater, and $Q^{\mathrm{gw}}_{\mathrm{off}} - Q^{\mathrm{gw}}_{\mathrm{on}}$ is net subsurface flow from the study area and includes pumping; all terms are expressed as rates (e.g., mm/year).

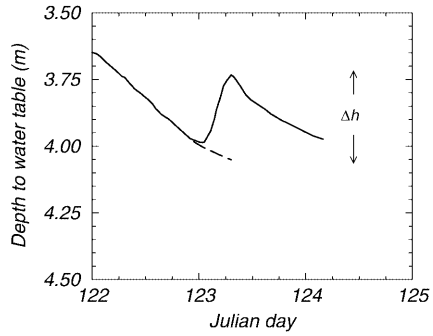

Fig. 1 Hypothetical water-level rise in well in response to rainfall. Δh is equal to the difference between the peak of the rise and low point of the extrapolated antecedent recession curve (*dashed line*) at the time of the peak

The WTF method is based on the premise that rises in groundwater levels in unconfined aquifers are due to recharge water arriving at the water table. Recharge is calculated as:

$$R = S_{\mathrm{y}} dh/dt = S_{\mathrm{y}} \Delta h/\Delta t \quad (2)$$

where S_{y} is specific yield, h is water-table height, and t is time. Derivation of Eq. (2) assumes that water arriving at the water table goes immediately into storage and that all other components of Eq. (1) are zero during the period of recharge. A time lag occurs between the arrival of water during a recharge event and the redistribution of that water to the other components of Eq. (1). If the method is applied during that time lag, all of the water going into recharge can be accounted for. This assumption is most valid over short periods of time (hours or a few days), and it is this time frame for which application of the method is most appropriate. The length of the time lag is critical in the success of this method. If water is transported away from the water table at a rate that is not significantly slower than the rate at which recharge water arrives at the water table, then the method is of little value.

For the WTF method to produce a value for total or "gross" recharge requires application of Eq. (2) for each individual water-level rise. Equation (2) can also be applied over longer time intervals (seasonal or annual) to produce an estimate of change in subsurface storage, ΔS^{gw}. This value is sometimes referred to as "net" recharge. To tabulate a total recharge estimate, Δh is set equal to the difference between the peak of the rise and low point of the extrapolated antecedent recession curve at the time of the peak. The antecedent recession curve is the trace that the well hydrograph would have followed in the absence of the rise-producing precipitation (Fig. 1). Drawing that trace is a somewhat subjective matter. If long-term hydrograph records are available, many recession curves should exist from which a trace can be patterned. Particularly difficult is the case where a

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water-level rise begins during the steepest part of the previous peak's recession. For an estimate of the net recharge, Δh is the difference in head between the second and first times of water-level measurement. The difference between total and net recharge is equal to the sum of evapotranspiration from groundwater, baseflow, and net subsurface flow from the site. With some additional assumptions, the WTF method can be used to estimate any of these parameters.

The WTF method for estimating groundwater recharge was applied as early as the 1920s (Meinzer 1923; Meinzer and Stearns 1929) and since then has been used in numerous studies (e.g., Rasmussen and Andreasen 1959; Gerhart 1986; and Hall and Risser 1993). White (1932) used a similar approach to estimate evapotranspiration from diurnal fluctuations in groundwater levels in the Escalante Valley of Utah. Weeks and Sorey (1973) also used data on groundwater fluctuations to estimate ET^{gw}.

The attractiveness of the WTF method lies in its simplicity and ease of use. No assumptions are made on the mechanisms by which water travels through the unsaturated zone; hence, the presence of preferential flow paths within the unsaturated zone in no way restricts its application. Because the water level measured in an observation well is representative of an area of at least several square meters, the WTF method can be viewed as an integrated approach and less a point measurement than those methods that are based on data in the unsaturated zone. The method, however, does have its limitations:

1. The method is best applied to shallow water tables that display sharp water-level rises and declines. Deep aquifers may not display sharp rises because wetting fronts tend to disperse over long distances. Nonetheless, the method has been applied to systems with thick unsaturated zones that display only seasonal water-level fluctuations.
2. Typically, recharge rates vary substantially within a basin, owing to differences in elevation, geology, land-surface slope, vegetation, and other factors. Wells should be located such that the monitored water levels are representative of the catchment as a whole.
3. The method cannot account for a steady rate of recharge. For example, if the rate of recharge were constant and equal to the rate of drainage away from the water table, water levels would not change and the WTF method would predict no recharge.
4. Other difficulties relate to identifying the cause of water-level fluctuations and calculating a value for specific yield. These two concerns are addressed in the following sections.

Causes of Water-Table Fluctuations in Unconfined Aquifers

Groundwater levels rise and fall in response to many different phenomena. Fluctuations are not always indicative of groundwater recharge or discharge. Changes in water levels occur over different time scales. Long-term fluctuations, over periods of decades, can be attributed to naturally occurring changes in climate and to anthropogenic activities (e.g., changes in land usage, pumpage, irrigation, and induced infiltration). Seasonal fluctuations in groundwater levels are common in many areas due to the seasonality of evapotranspiration, precipitation, and irrigation. Short-term water-table fluctuations occur in response to rainfall, pumping, barometric-pressure fluctuations, or other phenomena. The WTF method is best applied for short-term water-level rises that occur in response to individual storms. These conditions usually occur in regions with shallow depths to the water table. Long-duration, low-intensity precipitation events are not optimal because the slow, steady rate of water percolation to the water table may be of a similar magnitude as the rate of drainage away from the water table. Therefore, the height of the rise would be less than that expected from a short intense storm of the same total precipitation, and the recharge rate would be underestimated.

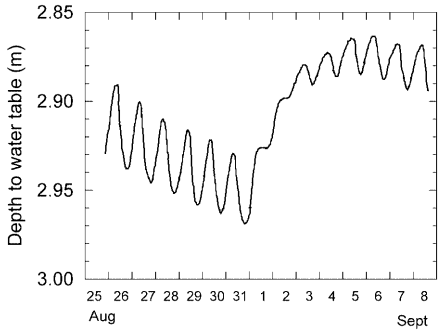

Fig. 2 Diurnal fluctuations in response to evapotranspiration by alfalfa in the Escalante Valley, Utah, USA, 25 August to 8 September 1926. Alfalfa was cut on 3 August (White 1932)

Application of the WTF method for estimating recharge requires identification of the water-level rises that are attributable to precipitation. This can be a formidable task. The following sections give some details on mechanisms other than precipitation that can induce short-term fluctuations in the water table.

Evapotranspiration

Water tables near the soil surface often exhibit diurnal fluctuations, declining during daylight hours in response to evapotranspiration and rising through the night when ET^{gw} is essentially zero. Figure 2 shows diurnal fluctuations in depth to the water table beneath a field of alfalfa in the Escalante Valley of Utah, USA, before and after cutting (White 1932). White developed a formula similar to Eq. (2) for estimating ET^{gw} based on such fluctuations. He assumed that ET^{gw} was zero between midnight and 4:00 AM and defined h' as the hourly rate of water-table rise during those hours. The total amount of groundwater discharged during one day, V_{ET}, was then calculated as:

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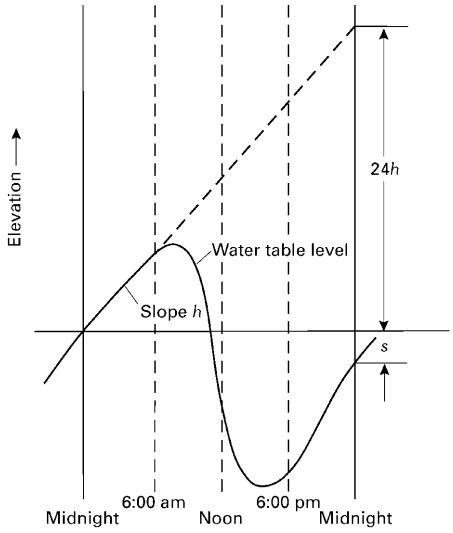

Fig. 3 Diagram of water-level fluctuations, showing variables needed for White's method for estimating *ET*; *s* is change in water level over 24 h. (After Troxell 1936)

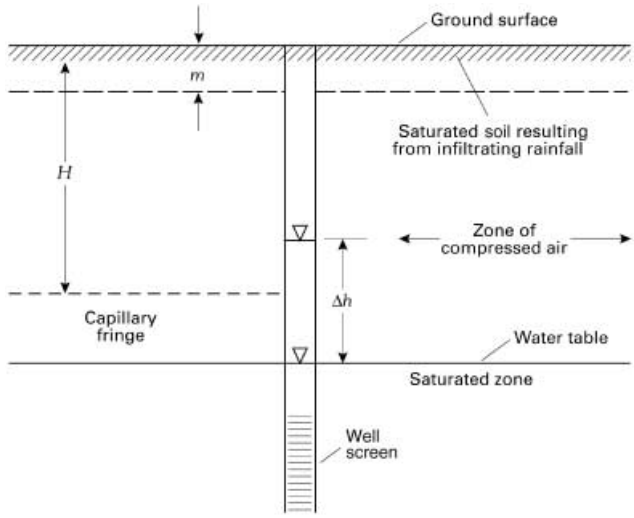

Fig. 4 Water-table rise in an observation well due to air entrapped between the water table and an advancing wetting front. *H* is initial depth to saturated zone, *m* is thickness of infiltrating saturated front, Δ*h* is water-level rise in well. (After Todd 1980)

$$V_{\mathrm{ET}} = S_{\mathrm{y}}\left(24h' + s\right) \tag{3}$$

where *s* is the water-level elevation at midnight at the beginning of the 24-h period minus the water-level elevation at the end of the period. Figure 3 illustrates the method. Using a specific yield of 0.073, White estimated seasonal ET^{gw} of the alfalfa to be 700 mm.

Atmospheric pressure

Changes in atmospheric pressure can cause fluctuations of several centimeters in water levels measured in observation wells. The fluctuations occur because pressure changes are transmitted much more rapidly through the open well than through the sediments overlying the aquifer. This is an observational effect rather than an effect on the aquifer itself. Although the phenomenon occurs in both confined and unconfined aquifers, different mechanisms are responsible for it in the two aquifer types. Consider a step increase in atmospheric pressure at land surface of ΔP. Because a well is open to the atmosphere, the pressure change on water in the well is equal to ΔP. For a confined aquifer, only a portion of that increase, $\alpha\Delta P$, is transmitted to the aquifer; this occurs instantaneously by way of grain-to-grain contact (Jacob 1940). The resulting pressure imbalance produces a decline in the water level within the well as water moves from the well to the aquifer. The factor $(1-\alpha)$ is called the barometric efficiency of the aquifer and generally takes on a constant value for confined aquifers.

For unconfined aquifers, a time lag occurs during which the pressure change at the land surface is propagated through the unsaturated zone to the water table. Air must move through the unsaturated zone to transmit a pressure change. Therefore, an imbalance exists between the pressure on the water in the well and the water in the aquifer until the pressure front arrives at the water table. This imbalance produces a change in the observed water level in the well. Weeks (1979) and Rojstaczer (1988) present in-depth analyses of this phenomenon. The length of the time lag increases with increasing depth to the water table and with decreasing vertical air diffusivity of the unsaturated-zone sediments. The barometric efficiency of an unconfined aquifer is not constant, because these two factors can change over time. Techniques for identifying and removing the effects of atmospheric-pressure changes from observed water levels are described in Weeks (1979) and Rasmussen and Crawford (1997).

Pressure transducers are often used to monitor water levels; these devices can be affected by atmospheric-pressure changes. Nonvented, or absolute, transducer readings may reflect atmospheric-pressure changes and could give the false impression that water-level fluctuations are much greater in magnitude than they are in reality. Rasmussen and Crawford (1997) present some guidelines on the use of transducers.

Entrapped air

Another phenomenon that affects water levels in unconfined aquifers is the presence of entrapped air between the water table and a wetting front advancing downward from the land surface. This phenomenon is particularly difficult to identify because it occurs in response to precipitation and, thus, is easily mistaken for recharge. The phenomenon takes place when surface soils become saturated and therefore impermeable to air. Figure 4 illustrates the phenomenon. As explained by Todd (1980), if *m* is the thickness of the saturated front infiltrating at land surface and *H* is the initial depth to the top of the capillary fringe (or saturated zone), then the rise in

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water level in a well tapping the water table, Δh, can be calculated as:

$$\Delta h = P_a m / (H - m) \rho g \quad (4)$$

where P_a is atmospheric pressure, ρ is density of water, and g is acceleration due to gravity. This phenomenon is most likely to occur where surface soils are fine textured and nearly saturated. Keys to identifying this effect include a very rapid water-level rise (because pressure is transmitted more rapidly than water is transported), an anomalously high rise in water level (greater than anticipated based on the given precipitation rate), and a quick dissipation of the water-level rise (usually in a matter of hours or days as the entrapped air is able to escape along the periphery of the recharge area).

This phenomenon has been termed the "Lisse" effect (Krul and Liefrinck 1946) after the village in Holland where it was first identified. It has been studied in detail with laboratory experiments (McWhorter 1971). The entrapped air, in effect, causes a decrease in the magnitude of the vertical hydraulic gradient. This attenuated gradient, in turn, restricts infiltration at land surface and increases runoff. Thus, entrapped air may not only give the false impression of recharge, it may also reduce the amount of recharge that would be expected in its absence. Meyboom (1967) attributed the immediate increased runoff in the Qu'Appelle River after a light rainfall to the Lisse effect. Even in the absence of the Lisse effect, air can become trapped in the largest pores behind a downward-moving wetting front, thus diminishing the cross-sectional area through which water can move. This phenomenon reduces the hydraulic conductivity of the sediments and therefore enhances runoff.

The occurrence of entrapped air does not entirely preclude use of the water-table fluctuation method, but it can greatly complicate it. To check on whether the Lisse effect is responsible for a rise in water levels, it would be necessary to determine if the subsurface soils were fully saturated or if the subsurface gas-phase pressure were greater than atmospheric pressure. Instruments such as a time domain reflectometry (TDR) or neutron probe can be used to measure the in situ moisture content; however, the accuracy of the measurements may not be sufficient to determine definitively whether or not the soils were fully saturated. Subsurface gas-phase pressures can be measured with a small-diameter gas piezometer installed in the unsaturated zone; in the presence of the Lisse effect, the piezometer would serve as a vent through which entrapped air could escape. Because of the difficulty in identifying occurrences of the Lisse effect, its prevalence in nature remains a matter of some debate.

Other mechanisms

Other mechanisms cause fluctuations in water levels in unconfined aquifers. In general, these mechanisms are more easily identified or are less frequently encountered than those already discussed. Temperature variations affect groundwater levels due to freeze–thaw action and the temperature dependency of surface tension, air solubility, and air density. Pumping of wells and natural or induced changes in surface-water elevations greatly affect groundwater levels. Ocean tides influence water levels in aquifers near oceans. Confined aquifers are usually influenced to a greater degree than water-table aquifers, but analyses that have been developed for predicting confined-aquifer response as a function of time and distance from the ocean are useful for approximating behavior in unconfined aquifers (Todd 1980). Changes in groundwater flow into or out of the study area, due to processes occurring in adjacent areas, also generate water-level fluctuations. Additional information on these processes is in Todd (1980) and Freeze and Cherry (1979).

The Concept of Specific Yield

"The specific yield of a rock or soil, with respect to water, is the ratio of (1) the volume of water which, after being saturated, it will yield by gravity to (2) its own volume" (Meinzer 1923). The following formula is generally used:

$$S_y = \phi - S_r \quad (5)$$

where ϕ is porosity and S_r is specific retention (the volume of water retained by the rock per unit volume of rock). Specific yield is treated as a storage term, independent of time, that in theory accounts for the instantaneous release of water from storage. In reality, the release of water is not instantaneous. Rather, the release can take an exceptionally long time, especially for fine-grained sediments. King (1899) determined S_y to be 0.20 for a fine sand; however, it took two and a half years of drainage to obtain that value. The limitations of this definition are noted by Meinzer (1923), who also points out that it does not account for temperature and chemical effects. These problems have no doubt contributed to the wide range of values that are reported in the literature. For example, Table 1 displays average, coefficient of variation (CV), and range of S_y as a function of texture from 17 studies compiled by Johnson (1967), in which different techniques for determining S_y were used. As can be seen from the table, a great deal of variation exists within each textural class. The variability is attributed to natural heterogeneity in geologic materials, the different methods used for determining S_y, and, in large part, to the amount of time allotted to the determination (Prill et al. 1965). In general, the coarser sediments have lower CVs than the finer sediments. This behavior is expected because coarser sediments drain more quickly and, hence, S_y for these sediments shows less time dependency.

Soil scientists share a common concern with hydrologists: a need to define the amount of water that a soil can hold against gravity. This information is important in determining the ability of the soil to provide plants with water and in scheduling irrigation. The term "field capacity" was introduced by Veihmeyer and Hendrickson (1931) as "the amount of water held in the soil after the

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Table 1 Statistics on specific yield from 17 studies compiled by Johnson (1967)

Texture	Average specific yield	Coefficient of variation (%)	Minimum specific yield	Maximum specific yield	Number of determinations
Clay	0.02	59	0.0	0.05	15
Silt	0.08	60	0.03	0.19	16
Sandy clay	0.07	44	0.03	0.12	12
Find sand	0.21	32	0.10	0.28	17
Medium sand	0.26	18	0.15	0.32	17
Coarse sand	0.27	18	0.20	0.35	17
Gravelly sand	0.25	21	0.20	0.35	15
Fine gravel	0.25	18	0.21	0.35	17
Medium gravel	0.23	14	0.13	0.26	14
Coarse gravel	0.22	20	0.12	0.26	13

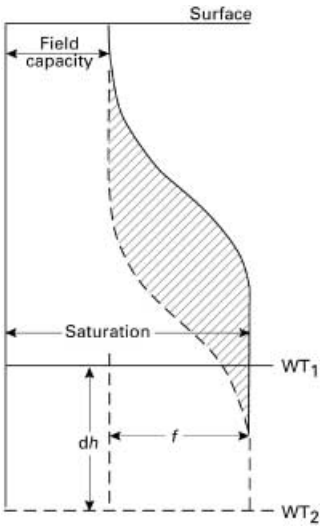

Fig. 5 Hypothetical moisture profiles above a declining deep water table. *Solid line* is initial profile for water table at WT1. *Dashed line* is final profile for water table at WT2. *Shaded area* represents water yield. f is $\phi-\theta_r$; dh is WT1–WT2

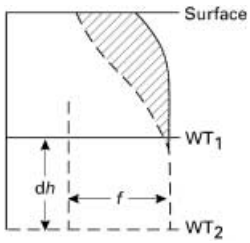

Fig. 6 Hypothetical moisture profiles above a declining shallow water table. *Solid line* is initial profile for water table at WT1. *Dashed line* is final profile for water table at WT2. *Shaded area* represents water yield. f is $\phi-\theta_r$; dh is WT1–WT2

excess gravitational water has drained away and after the rate of downward movement of water has materially decreased". Field capacity is equivalent to specific retention. Its definition is slightly more satisfying than that presented for S_r; it acknowledges that a time factor is involved, i.e., water does continue to drain from soil but a point is reached where the rate of drainage becomes insignificant. Determination of the significance of that rate must be made relative to the flow of water to or from the water table. This rate varies greatly among different sediments, as does the time required to reach that rate.

The fact that S_y is not constant but varies as a function of depth to the water table is described from the perspective of soil physics by Childs (1960). That analysis takes into account the moisture-retention curve and is briefly described here.

Consider the case of uniform porous media with an initial static equilibrium profile above a water table. Upon a change of the level of the water table, it is assumed that the static equilibrium profile is re-attained instantaneously. The moisture profile above the water table at both times is represented by the moisture-retention curve. Figure 5, after Childs (1960), displays two such profiles for an initial water level, WT1, and after the water level has declined by dh to a level of WT2. The shaded area represents the volume of water per unit area that is released from storage due to the water-level decline. A simple analysis shows this area to be equal to S_ydh.

Consider another case where the initial depth to water table is not great enough to allow moisture content at land surface to reach the value of residual moisture content (Fig. 6). Again, the shaded area represents the water yield. Now, however, that area is less than S_ydh. The discrepancy between actual yield and that calculated on knowledge of S_y and dh increases as depth to water table decreases. The extreme case occurs when the initial and final depths to water table are less than the height of the capillary fringe. For this case, no release of water would occur with changing water levels. In the field, this phenomenon, sometimes referred to as the reverse Wieringermeer effect, is reflected by a nearly instantaneous rise in water level in response to only a small amount of infiltration. Gillham (1984) used an innovative field and modeling study to illustrate this effect.

dos Santos and Youngs (1969) extended the work of Childs (1960) to account for time in addition to depth to water table in the definition of specific yield:

$$S_y = \mathrm{d}W'/\mathrm{d}t/dH/\mathrm{d}t + \phi - \theta(H) \qquad (6)$$

where dW'/dt is a term that accounts for the changing shape of the moisture profile above the water table over time, H is depth to water table, and θ is moisture content. $\theta(H)$ can be determined from measured moisture-retention curves or can be estimated on the basis of texture from published tables of soil properties (e.g., Carsel and Parrish 1988). In a series of two-dimensional drainage experiments in a 333-cm by 333-cm by 153-cm-deep sand tank, dos Santos and Youngs (1969) demonstrated the variability of specific yield as a function of depth to water table, time, and distance to a drain. In general, S_y decreased with increasing distance from a drain and in-

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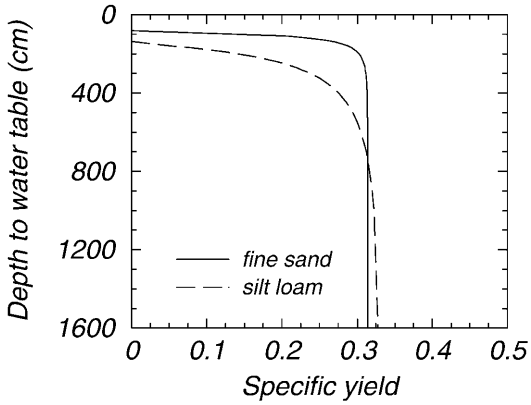

Fig. 7 Specific yield as a function of depth to water table for a fine sand and a silt loam. (After Duke 1972)

creased with time at any one location. In spite of these findings, the authors observed for the sand of their experiments that dW'/dt was often small and that neglecting the first term in Eq. (6) resulted in a fair approximation to the true specific yield. Duke (1972) likewise suggests setting S_y equal to $\phi-\theta(H)$. The effect of depth to water table on S_y is shown in Fig. 7 for two soils described by Duke (1972).

Although the foregoing discussion illustrates how the height of the capillary fringe and depth to water table affect S_y, the assumptions employed in the analysis are generally not representative of actual field conditions. Steady-state conditions probably never occur in the real world, and, as mentioned above, many years may be required for some soils to fully drain. Most water tables fluctuate on a seasonal or annual basis, so that complete drainage of the sediments never occurs. The hysteretic nature of the moisture-characteristic curve should also be considered. Soils are more easily wetted than drained. For example, if one were to wet a soil sample from an initially dry state in a laboratory experiment to a tension of h_1 and record its moisture content, and then fully saturate the sample and drain it to the same tension, h_1, one would find that the moisture content at h_1 after drainage was greater than that achieved during the wetting process. So the history of water-table movement prior to a recharge rise affects the true value of S_y. Ideally, specific retention would be defined as the moisture content at which hydraulic conductivity is negligible relative to the magnitude of the annual flux arriving at or leaving the water table. Of course, because that annual flux is unknown, that definition is not suitable. The following sections describe laboratory and field approaches for determining specific yield.

Laboratory methods

Specific yield is usually determined by measurement of porosity and specific retention and application of Eq. (5), because specific retention is more easily measured in the laboratory than is specific yield. Johnson et al. (1963) describe a column-drainage approach. A column is filled with undisturbed or repacked sediments and then saturated with water from the bottom. Water is allowed to drain from the bottom of the column, which is maintained as a constant boundary of zero-pressure head. The top of the column is vented to maintain atmospheric pressure. The test continues until drainage ceases. As discussed in the previous section, the time allowed for drainage has a large effect on the calculated values of specific retention and yield. Ideally, the test is carried out in a climate-controlled chamber where the relative humidity is kept close to 100%. Otherwise, drained water contained in a collection vessel as well as water from the core itself (via the vent) will evaporate; conceivably, the rate of evaporation could exceed the rate of drainage. If proper technique is not employed, it may be erroneously assumed that drainage has ceased. The height of the column also exerts a large influence on measured values. A column that is shorter than the height of the capillary fringe is likely to produce a value of S_y that is much less than that obtained from a column that is greater in height than that of the capillary fringe. Prill et al. (1965) estimated that it would take longer than a year for a 1.71-m-long column of medium sand to reach static equilibrium conditions after drainage commenced, although equilibrium conditions were attained in the lower 60 cm of the column after only 5 h. Cassel and Nielsen (1986) describe similar laboratory techniques for approximating field capacity. They note, however, that these techniques should only be used if it is not possible to determine field capacity using the field technique described in the following section.

Numerous suggestions have been made for estimating field capacity or S_r on the basis of a laboratory-determined moisture-retention curve. Several authors have equated S_r with residual (or irreducible) moisture content, θ_r. McWhorter and Garcia (1990) suggest setting S_r equal to moisture content at a matric potential of –1.35 MPa. Johnson et al. (1963) prefer a matric potential of –50 kPa. Jamison and Kroth's (1958) proposal that volumetric moisture content at a matric potential of –33 kPa be used to estimate field capacity has gained widespread acceptance within the agricultural community. However, Cassel and Nielsen (1986) point out that this approach does not take account of soil texture. Field capacity for coarse-textured soils is usually better represented by moisture content at –10 kPa, whereas moisture content at a matric potential of much less than –33 kPa is appropriate for fine-textured soils.

Measurement of the moisture-retention curve in the laboratory is often time consuming and problematic. It is much easier to determine the particle-size distribution of a soil. Several attempts have been made to relate soil texture to the moisture-retention curve in general and to S_y in particular, using pedo transfer functions (e.g., Briggs and Shantz 1912; Arya and Paris 1981; Nimmo 1999). The simplest of these functions allows estimation of moisture content based on the percentages of sand, silt, and clay in the sample. Carsel and Parrish (1988) determined parameters for the van Genuchten (1980) equation that relates moisture content to matric potential

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Table 2 Values of S_y determined by type-curve matching for 18 aquifer tests. (After Prickett 1965)

Material	S_y	Material	S_y
Sand, medium to coarse	0.200	Sand, medium to coarse	0.250
Sand, medium	0.161	Sand, fine	0.113
Sand, medium	0.166	Sand, silty to medium	0.014
Sand, medium	0.181	Sand, fine to medium	0.192
Sand, fine to medium	0.032	Sand, fine to coarse	0.014
Sand, medium, silty	0.051	Sand, fine with clay	0.021
Sand, fine to medium	0.005	Sand, fine with clay	0.206
Sand, fine to medium	0.007	Sand, fine with silt	0.018
Sand, fine	0.09	Clay, silt, fine sand	0.039

Table 3 Values of specific yield from Nwankwor et al. (1984)

Method	S_y	Remarks
Neuman (1972)	0.07	–
Boulton (1963)	0.08	–
Volume balance	0.02	t=15 min
	0.05	40 min
	0.12	600 min
	0.20	1,560 min
	0.23	2,690 min
	0.25	3,870 min
Laboratory (ϕ–θ_r)	0.30	–

for a range of generic soil types. With those parameters, an estimate can be obtained of S_r as moisture content at any prescribed matric potential (e.g., –10 kPa, –33 kPa, –1.35 MPa).

Field methods

Aquifer tests

Values of S_y and transmissivity, T, for unconfined aquifers are commonly obtained from the analysis of aquifer tests conducted over a period of hours or days. Drawdown-versus-time data from observation wells are matched against theoretical type curves developed by Boulton (1963), Prickett (1965), Neuman (1972), and Moench (1995, 1996). A brief description of the process is presented here. The interested reader is referred to the above references as well as Walton (1970) for more detail. Ideally, an unconfined aquifer is pumped from a well screened through the entire thickness of the saturated zone. Observation wells are located at various distances from the pumping well. These may fully or partially penetrate the aquifer. When pumping begins, declining water levels (drawdowns) are recorded over time at each observation well. Early time recording is especially important. Drawdown versus time is plotted on a log–log scale, and type curves at the same scale are then overlaid on the plot. The curves are shifted (keeping coordinate axes parallel) until most of the data points lie on a curve. An arbitrary match point is then selected and the values of drawdown and time at that point are used to calculate T and S_y. Table 2 from Prickett (1965) shows S_y calculated for 18 aquifer tests for a variety of aquifer types. Similar to Table 1, Table 2 shows a wide range of values for S_y for materials of similar texture.

Aquifer tests provide in situ measurements of S_y and other aquifer properties that are integrated over fairly large areas, but the methods are not without some drawbacks. Interpretation of results is nonunique, as discussed by Freeze and Cherry (1979). The fact that a theoretical curve matches an experimental drawdown curve in no way implies that the aquifer adheres to the assumptions inherent to development of that theoretical curve. Expense is also a key concern; installation of pumping and observation wells may be cost prohibitive for many recharge studies. Aquifer tests require careful planning; useful guidelines are provided by Walton (1970) and Stallman (1971).

Volume-balance method

The volume-balance method for determining specific yield combines an aquifer test with a water budget of the cone of depression:

$$S_y = V_w / V_c \tag{7}$$

where V_w is the volume of water pumped out of the system at some point in time, and V_c is the volume of the cone of depression (the region between the initial and the final water table) at that same time (Clark 1917; Wenzel 1942; Remson and Lang 1955). Nwankwor et al. (1984) used this method to analyze results from an aquifer test at the Borden field site in Ontario, Canada. Their results (Table 3) show a trend of increasing S_y with increasing time: the longer they ran the aquifer test, the larger the calculated value of S_y became. Wenzel (1942) observed a similar trend in a sand–gravel aquifer. Nwankwor et al. (1984) attribute the trend to delayed drainage from the unsaturated zone and question the appropriateness of Neuman's (1972) assumption of instantaneous drainage from the unsaturated zone. An analysis by Neuman (1987) shows that these results could be explained by flow from outside the assumed radial extent of the cone of depression.

The topic of delayed yield from the unsaturated zone has been revisited several times in the literature. Nwankwor et al. (1992) repeated the Borden aquifer test, this time with tensiometers installed to monitor water movement above the water table. Results of that study show that analyses of aquifer tests that do not take into account unsaturated flow predict values of S_y that are unrealistically low. Moench (1994) shows that type-curve estimates can agree with those of volume-balance methods when composite drawdown plots from more than one well are used and when partial penetration is taken into account. Akindunni and Gillham (1992) used a numerical model to simulate the Borden pumping test. They show that the volume-balance approach is a viable method, and they demonstrate that delayed yield from above the water table can explain the time trend in S_y shown in Table 3.

Endres et al. (2000) used ground-penetrating radar (GPR) to measure the volume of the cone of depression.

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Results of the study were less than optimal. Using the GPR-determined value for V_c in Eq. (7) produced an unreasonably high value for S_y (in excess of 0.6). Although GPR has the great advantage of avoiding the need for observation wells, additional refinement is needed before this approach becomes practicable.

Water-budget methods

A simple water budget for a basin can be written as:

$$P + Q_{on} = ET + \Delta S + Q_{off} \quad (8)$$

where P is precipitation plus irrigation; Q_{on} and Q_{off} are surface and subsurface water flow into and out of the basin; ET is the sum of bare-soil and open-water evaporation and plant transpiration; ΔS is change in water storage; and all components are given as rates. Water storage takes place in several compartments: surface reservoirs, ΔS^{sw} (including water bodies as well as ice and snow packs); the unsaturated zone, ΔS^{uz}; and the saturated zone, ΔS^{gw}:

$$\Delta S = \Delta S^{sw} + \Delta S^{uz} + \Delta S^{gw} \quad (9)$$

Walton (1970) proposes using Eq. (8) in conjunction with Eq. (2) for estimating S_y during periods of water-level rise in winter months. ET is usually small during these months and the soil is nearly saturated, so that the change in water storage in the unsaturated zone is also small. The water-budget equation can be rewritten as:

$$S_y = (P + Q_{on} - Q_{off} - ET - \Delta S^{sw} - \Delta S^{uz}) / \Delta h / \Delta t \quad (10)$$

The value determined for S_y from Eq. (10) is then used with water-level rises throughout the year to generate an annual estimate of recharge. Gerhart (1986) and Hall and Risser (1993) applied this method during some winter periods (usually of week-long duration) with the assumption that ET, ΔS^{uz}, and net subsurface flow were zero. Best results should be expected in areas where surface and subsurface flow are easily determined. The method as given above, however, implicitly assumes that no difference exists in baseflow prior to and during the rise in groundwater levels. The validity of this assumption should be examined for each application.

Rasmussen and Andreasen (1959) used a water-budget approach to estimate S_y (they called the term *gravity yield*, Y_G). They measured or estimated all parameters on the right-hand side of Eq. (10). Water levels in 25 observation wells were used to determine Δh as the net change in water level over one and a half years. The method of convergent approximations (similar to simple linear regression) was then used to determine a basin-wide value of 0.11 for Y_G. The authors note that this value is less than the average value of 0.21 determined in the laboratory on core samples obtained from the basin. They attribute this difference to inadequate time for the sediments to fully drain between successive water-level rises.

A water-budget method was also employed by Gburek and Folmar (1999) to determine S_y for a fractured sandstone, siltstone, and shale system in east-central Pennsylvania, USA. Depth to the water table was approximately 7 m. Pan lysimeters were installed at depths of 1–2 m beneath undisturbed soil columns. These lysimeters were designed to capture all downward-moving water. The rate of water percolation measured by the lysimeters was assumed equal to the recharge rate. Water levels were measured in wells near the lysimeters, and S_y was calculated from Eq. (2). Good correlation was obtained between percolation rates and water-level fluctuations. An average value for S_y of 0.009 was obtained for eight events between 1993 and 1995.

Geophysical methods

Pool and Eychaner (1995) developed a method for using gravity measurements to estimate S_y. Micro-gravity measurements were made over transects that were kilometers in length. Changes in gravity over time were attributed to changes in subsurface water storage (combined unsaturated and saturated zones). Comparison of these changes in storage with changes in water levels in monitoring wells allowed Pool and Eychaner (1995) to determine values of S_y that range from 0.16–0.21 for an alluvial aquifer in the Pinal Creek Basin of central Arizona, USA. The authors claimed good agreement with values obtained from aquifer tests. Standard deviations in repeated gravity measurements during one day at a single station were equivalent to 0.04–0.14 m of water. Thus, application of the method should be restricted to regions where annual water-level fluctuations are substantially greater than 0.14 m.

Meyer (1962) proposes use of a neutron meter for determining change in water storage within both the unsaturated and saturated zones due to changes in groundwater levels. The method requires that a neutron-probe access tube be installed adjacent to an observation well. The total amount of stored water can be determined at any time by measuring moisture contents between the water table and the zero-flux plane (the horizontal plane in the unsaturated zone that separates downward-moving water from water that moves upward in response to ET). The difference in stored water between any two measurement times, ΔS^{gw}, is the amount of water that has been added or subtracted from storage. Invoking Eq. (2), specific yield can be determined as:

$$S_y = \Delta S^{gw} / \Delta h / \Delta t \quad (11)$$

Weeks and Sorey (1973) used this approach to calculate S_y. Then, assuming that water-level declines were due only to evapotranspiration, they used an equation similar to Eq. (2) to estimate ET^{gw} in the Arkansas River valley of Colorado, USA. Sophocleous (1991) used a similar technique for estimating S_y and recharge in the Great Bend Prairie region in Kansas, USA.

Field-capacity tests

Field experiments for determining field capacity are sometimes useful for estimating S_y. Field-capacity tests are conducted in the upper 2 m of soil. Field-capacity estimates should be representative of specific retention

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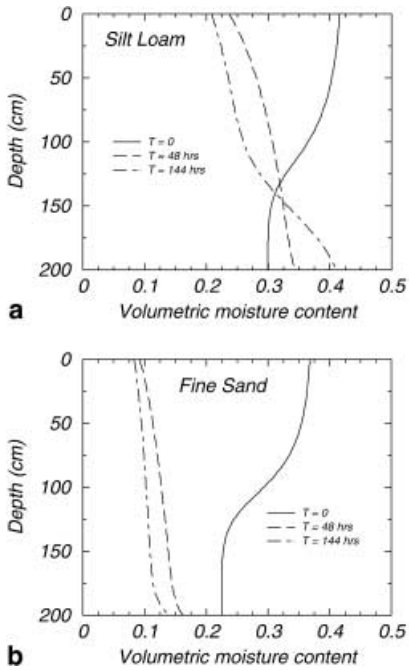

Fig. 8 Simulated moisture profiles after drainage of 0, 48, and 144 h for **a** a silt loam and **b** a fine sand

of the sediments at the water table, if the water table is not much deeper than this and if the sediments are fairly uniform. The procedure for determining in situ field capacity is described in detail by Cassel and Nielsen (1986). In brief, the procedure is to saturate a small plot, let it drain, and then determine the soil-moisture content. The area selected should be at least 3–4 m on a side. An impermeable dike is constructed around the perimeter. Water is then applied to the surface in sufficient quantities to allow it to infiltrate to a depth of at least 75 cm. The amount of water required is calculated based on the initial moisture content of the plot. The plot is then covered with an evaporation barrier, such as polyethylene sheeting. After 48 h, the barrier is removed and volumetric moisture content is determined throughout the depth of wetting near the center of the plot. Measurements can be made with a TDR probe or neutron or gamma-attenuation moisture gage, but more commonly moisture contents are determined gravimetrically by obtaining cores of known volume. The cores are then oven dried and field capacity *FC* is calculated as:

$$FC = V_w / V_s \tag{12}$$

where V_w is volume of water and V_s is total volume of the core. Cassel and Nielsen (1986) recommend obtaining at least five samples at each depth in order to obtain a representative value.

The 48-h waiting period for redistribution is somewhat arbitrary. A numerical experiment was carried out to simulate the field-capacity test for the fine sand and silt loam described by Duke (1972) and used in Fig. 7 with the VS2DT variably saturated flow model (Healy 1990). Initial moisture content was set at the value that represented an effective saturation of 0.5. An amount of water sufficient to saturate the top 100 cm was applied to the top of the 350-cm-tall column. The bottom boundary was treated as a zero constant-pressure head. Figure 8 shows the initial moisture pattern in the upper 200 cm of the column after application of the water, as well as the patterns that develop after 48 and 144 h of drainage. Drainage continues after 48 h for both soils. The amount of additional drainage for the sand could be considered insignificant and the 48-h period appears to be sufficient for estimation of field capacity. For the silt loam, a substantial amount of additional drainage occurs between 48 and 144 h. Therefore, 48 h is probably an inadequate time frame for determining S_y for this material.

Further thoughts on specific yield

The preceding discussion indicates that considerable uncertainty remains as to what value of S_y to use in a particular study. It is disconcerting that so little progress has been made in addressing the limitations of S_y in the century that has passed since the pioneering work of Hazen (1892) and King (1899). Although a large variability exists in both laboratory and field values of S_y, laboratory values are generally greater than those obtained from field tests. Laboratory measurements of S_r are largely dependent on the amount of time allowed for drainage as well as the length of the test column. They are usually run for longer periods of time than are field tests. S_r determined in the laboratory is often quite close to the value of θ_r. For some applications, such as determining the effect of long-term pumping of an aquifer, laboratory values of S_y are appropriate, but for estimation of groundwater recharge by the WTF method, laboratory values of S_y are probably too large. Meyboom (1967) was well aware of this discrepancy between laboratory and field estimates of S_y. He multiplied laboratory-determined values of S_y by 0.5 to arrive at a field value that he then used with Eq. (2) to estimate recharge in a prairie setting. Such a simplistic approach is not expected to have widespread applicability.

A reasonable course of action is to start with an estimate of $S_y = \phi - \theta(H)$, as suggested by dos Santos and Youngs (1969) and Duke (1972). This definition requires knowledge of the moisture-characteristic curve. In the absence of that information, an estimate of $\theta(H)$ can be obtained on the basis of texture from compilations of generic soil properties (e.g., Carsel and Parrish 1988; Leij et al. 1996). Adjustments to that estimate may be warranted for various reasons. In general, the more coarse-grained the soil, the more easily it drains; hence, the more likely the laboratory determined S_y will resemble the true S_y. Hysteresis and the timing between water-level rises also influence S_y. If multiple rises occur in close proximity in time, the sediments probably would not fully drain between rises. Therefore, it would be appropriate to use different values of S_y for the different rises. The shorter the time between water-level rises, the smaller the value of S_y should be.

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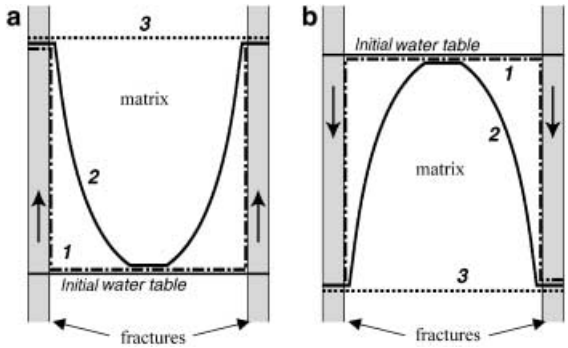

Fig. 9 Diagrams of saturation of a fractured-rock matrix under **a** rising and **b** declining water tables. Where the rate of water-table rise is rapid relative to the matrix permeability *1*, the matrix remains unsaturated as the water level in the fracture rises. The specific yield is equal to the fracture porosity. Where the rate of water-level rise is slow, and the matrix permeability is high *3*, the water table rises evenly in both the fracture and matrix. The specific yield is given by Eq. (13). More usually, the matrix partially fills as the water table rises, and the specific yield is between these two limiting values *2*. Congruent behavior occurs for declining water tables

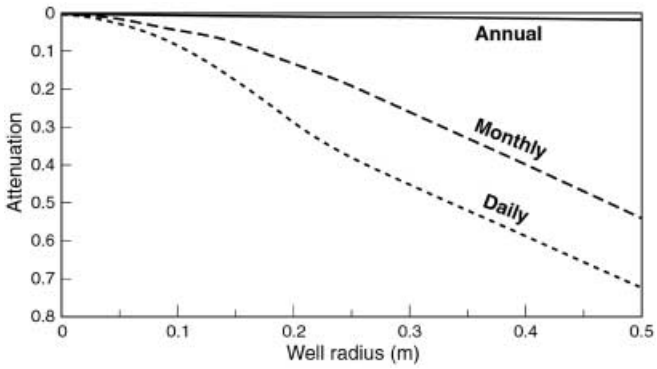

Fig. 10 Amplitude of water-level fluctuation observed in a piezometer relative to that in the aquifer as a function of well radius for varying signal frequency. Values less than one represent attenuation of the signal within the piezometer. Simulations are for an aquifer hydraulic conductivity of 10 m/year, and storativity of 0.005. (After Simmons et al. 1999)

Fractured-Rock Systems

Much of the previous discussion in this paper focuses on interpretation of water-level variations and measurement of specific yield in sedimentary (porous-media) aquifers. In fact, relatively little research has been done on interpretation of hydrograph response in fractured-rock aquifers. Porosities of igneous and metamorphic rocks are often less than 1%, and even some sandstones and limestones have total porosities of less than a few percent (de Marsily 1986). Where porosities are low, very large variations in groundwater level occur in response to recharge. For example, Bidaux and Drouge (1993) measured water-level rises of approximately 15 m in response to rainfall events of approximately 50 mm over 24 h in fractured Cretaceous limestones and marly limestones in southern France.

Zuber and Motyka (1998) define the specific yield, n_s, of a fractured, cavernous rock as:

$$n_s = \beta_c n_c + \beta_f (l - n_c) n_f + \beta_p (1 - n_c - n_f) n_p \qquad (13)$$

where n_c, n_f, and n_p are the cavern, fissure, and matrix porosities, respectively; and β_c, β_f, and β_p denote the fractions of these pore spaces that can be drained under the force of gravity. They note that β_c and β_f are usually close to unity, which arises because fracture apertures of greater than about 4 µm are empty at field capacity (matric potential of –33 kPa). $\beta_p n_p$ is simply the specific yield of the matrix blocks. Using Eq. (13), the authors estimated the specific yield of karstic aquifers in southern Poland from measurements of the cavern volumes intersecting boreholes, fracture porosity from rock exposures, and matrix specific yield measured in the laboratory.

Although this definition is consistent with the definition of specific yield for porous media, it does not necessarily facilitate routine use for interpretation of water-level variations. The problem arises because in most fractured-rock systems, the permeability of the matrix is usually very low, so that the time required for it to fill and drain is very long. The fracturing of low-porosity formations creates a total porosity distribution that is essentially bimodal. Most recharge through the unsaturated zone occurs rapidly along discrete, permeable fractures, which may become saturated during rain events, even though surrounding micropores remain unsaturated. Thus, water levels in fractures may rise while most of the formation remains unsaturated (Fig. 9a). In this case, the specific yield would be equal to the fracture porosity. This situation is most likely to occur in response to large rainfall events where the matrix permeability is very low, so the rate of water-level rise would be very rapid, as would the rate of subsequent water-level decline (Fig. 9b).

Where the rate of water-level rise is slower, such as in the case of deeper water tables displaying only seasonal variations in water level, then the water level in fractures and matrix rise together. In this case, specific yield becomes equal to that determined in Eq. (13). In the case of a declining water table, the time for the matrix to drain may be extremely slow. The aquifer matrix may supply water as baseflow to streams for many months after the water table (as reflected in fractures and piezometers) has declined (Price et al. 2000).

Hydrograph variations for piezometers installed in fractured rocks often provide a poor record of water-level variations within the aquifer itself. This counterintuitive scenario occurs where the permeability of the aquifer is low and the storativity of the aquifer is very low relative to the storativity of the piezometer or well. Simmons et al. (1999) note that short-term variations in aquifer water level are significantly attenuated within the well, particularly where the well radius is large (Fig. 10). The degree of attenuation increases as the storativity of the aquifer decreases. Longer-term variations in aquifer water level (such as annual cycles) are less attenuated. This effect sometimes manifests itself as very smooth hydrographs in low-porosity aquifers, which do not show

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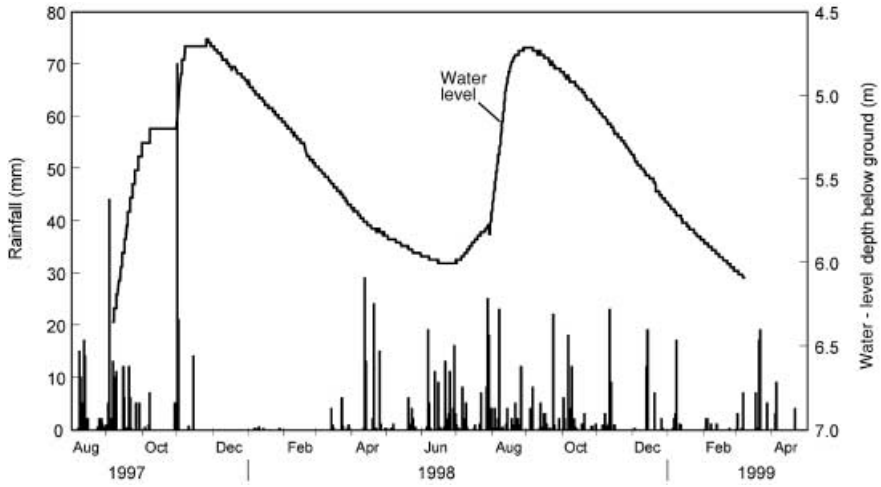

Fig. 11 Water-level fluctuations in a piezometer screened in Mintaro Shale, and daily rainfall at Clare, South Australia. The magnitude of the seasonal water-level fluctuations and the independently estimated recharge rate are consistent with a value of specific yield close to the total porosity. Short-term fluctuations, in response to daily rainfall events are generally absent, due to attenuation of short-wavelength variations by the large storage capacity of the well

responses to daily rainfall events except in extreme cases.

Figure 11 shows a well hydrograph from a piezometer completed in the Mintaro Shale near Clare, South Australia. At this site, the fracture porosity is estimated to be about 10^{-3}, based on outcrop mapping, and the matrix porosity is estimated to be 0.01–0.05, from helium porosimetry. The permeability of the matrix is extremely low ($<10^{-12}$ m/s). The water table at this site varies smoothly throughout the year; most rainfall events do not produce measurable changes in water level in the piezometer. This is not attributed to a lack of recharge, but rather to attenuation of these short-term signals by the large storage capacity of the piezometer (50 mm PVC). The only notable short-term rise in the water table occurred after 30 October 1997, when the water level rose approximately 200 mm in response to 70 mm of rainfall. Although a large fraction of the rain that fell on this date probably reached the aquifer, the relatively small water-level rise (relative to the aquifer porosity) is attributed to this attenuation. However, the annual cycle in water level is approximately 2 m, which is consistent with a recharge rate of approximately 40 mm/year and a specific yield (0.02) much closer to the matrix porosity than to the fracture porosity.

Most methods for estimating specific yield in fractured rocks are not sufficiently accurate to permit estimation of recharge using derived values. Interpretation of aquifer tests in fractured-rock systems is often difficult because of problems of nonuniqueness. Specific-yield values determined from these tests are usually unreliable (Bardenhagen 2000). The water-budget method is the most widely used technique for estimating specific yield in fractured-rock systems, probably because it does not require any assumptions concerning flow processes. Gburek et al. (1999) compared the recession of well hydrographs in shale and interbedded shales, siltstones, and sandstones from Pennsylvania, USA, with the base flow recession curve over a 40-day period for a stream draining the aquifer. Through calibration of a groundwater flow model, specific yield was estimated to be 1×10^{-2} in the overburden, 5×10^{-3} in the highly fractured rocks at shallow depths, and 1×10^{-4} in poorly fractured material below 22-m depth. As discussed previously, Gburek and Folmar (1999) used a water-budget method and estimated specific yield to range from 7×10^{-3} to 1×10^{-2} for the highly fractured zone at the same site.

Moore (1992) compared stream-flow hydrographs with groundwater hydrographs from shale and limestone aquifers on the Oak Ridge Reservation, Tennessee, USA, and estimated a specific yield of approximately 2×10^{-3} from slopes of the recession curves. Using an approach analogous to hydrograph separation, Shevenell (1996) estimated specific yields of 1×10^{-4}, 1×10^{-3}, and 3×10^{-3} for conduits, fractures, and matrix elements, respectively, of the limestone and dolomite Knox Aquifer at Oak Ridge, by apportioning segments of well recession curves to these different flow regimes.

Considerable effort has been devoted to evaluating S_y in fractured-rock systems. It remains unclear, however, whether estimates of S_y generated for these systems are of sufficient accuracy to permit their use in estimating recharge or if, because of the complexities of such systems, the application of the WTF method is at all valid.

Examples

Beaverdam Creek basin

Rasmussen and Andreasen (1959) studied the water budget of the Beaverdam Creek basin on the Delmarva Peninsula of Maryland, USA. The 51-km^2 drainage basin ranges in elevation from 4–26 m above sea level and receives on average 109 cm of precipitation annually. Beneath the basin, which is located on the Atlantic Coastal Plain, the unconsolidated sedimentary rocks consist mostly of sand, silt, clay, greensand, and shell marl. The water table is generally within a few meters of land surface. Quaternary surficial sands and silts of up to 22 m in thickness overlie Tertiary sand aquifers. Water levels were measured in 25 observation wells on a weekly basis. Stream discharge was monitored at the outlet of the basin. Precipitation was measured weekly at 12 sites within the basin.

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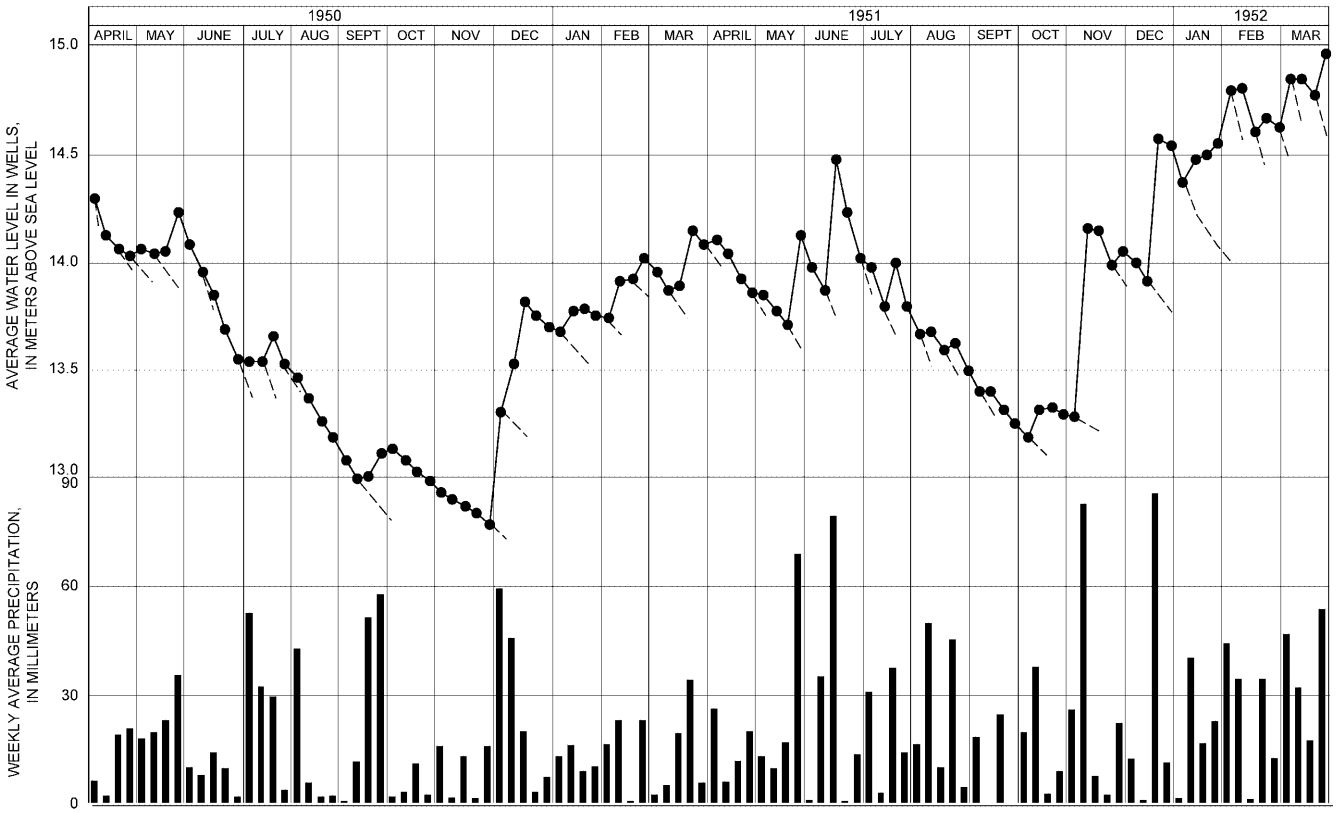

Fig. 12 Hydrograph of average water level in wells and bar graph of weekly average precipitation for Beaverdam Creek basin, Maryland, USA. (After Rasmussen and Andreasen 1959)

Figure 12 shows the average water level in the observation wells and total precipitation on a weekly basis for April 1950 through March 1952. Water levels are highest in late winter and early spring. Precipitation was fairly evenly distributed throughout the year. Recharge was calculated using Eq. (2) on a monthly basis. As previously described, S_y was estimated by a water-budget method to be 0.11; Δh was taken as the cumulative rise in water level for the month (i.e., the sum of all rises that occurred). To account for drainage from the water table that takes place during rises in water levels, water levels prior to rises were extrapolated to their expected positions had there been no precipitation. The rise was then estimated as the difference between the peak level and the extrapolated antecedent level at the time of the peak. The extrapolations are represented by the dashed lines in Fig. 12. Table 4 shows monthly estimates of recharge for April 1950 through March 1952.

Rasmussen and Andreasen (1959) also attempted to identify the different components of recharge, R: baseflow, Q^{bf}; change in storage in the saturated zone, ΔS^{gw}; and evapotranspiration from groundwater, ET^{gw}. They use a modified form of Eq. (1):

$$R = \Delta S^{gw} + Q^{bf} + ET^{gw} \tag{14}$$

It was assumed that net surface-water and groundwater flow were zero. ΔS^{gw} was calculated as $S_y \Delta h_n / \Delta t$, where

Table 4 Monthly change in water level and groundwater recharge for Beaverdam Creek basin, Maryland, USA. (After Rasmussen and Andreasen 1959)

Month	Change in water level, ΔH (cm)	Groundwater recharge, R (cm)
1950		
April	21.3	2.3
May	45.7	5.0
June	6.1	0.7
July	33.5	3.7
August	6.1	0.7
September	27.4	3.0
October	0.0	0.0
November	45.7	5.0
December	65.5	7.2
1951		
January	24.4	2.7
February	38.1	4.2
March	39.6	4.4
April	13.7	1.5
May	50.3	5.5
June	67.1	7.4
July	44.2	4.9
August	24.4	2.7
September	9.1	1.0
October	21.3	2.3
November	108.2	11.9
December	71.6	7.9
1952		
January	74.7	8.2
February	48.8	5.4
March	97.5	10.7

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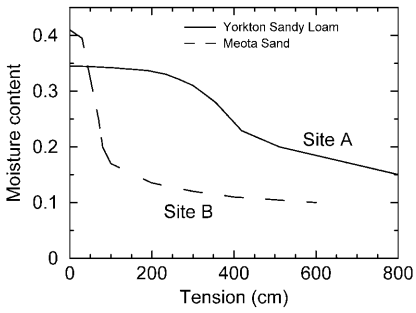

Fig. 13 Moisture-retention curves for Yorkton loamy sand (site A) and Meota sand (site B) (Freeze and Banner 1970)

Δh_n is the net change in head over each month (i.e., the difference in head between the end and the beginning of the month). Baseflow was determined by hydrograph separation of the runoff data. ET^{gw} was then calculated as the residual of Eq. (14). For a 2-year period, annual precipitation averaged 105.2 cm, of which groundwater recharge was 54.1 cm. Recharge was partitioned into 27.3 cm of base flow, 2.2 cm of increase in groundwater storage, and 24.6 cm of evapotranspiration of groundwater.

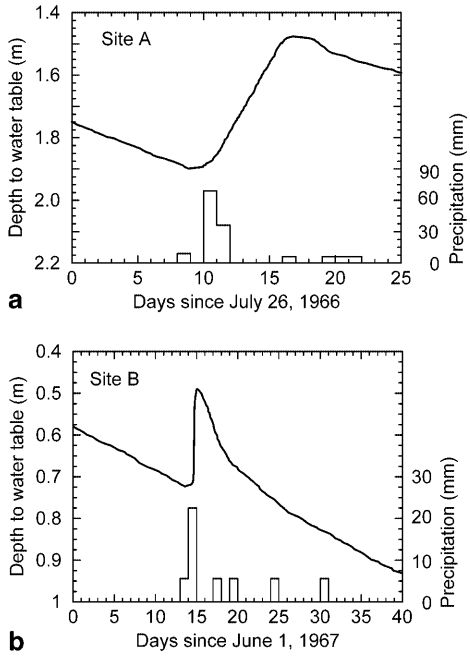

Fig. 14 Precipitation and depth to water table for **a** site A, and **b** site B (Freeze and Banner 1970)

Spirit Lake drainage

Freeze and Banner (1970) instrumented the unsaturated and saturated zones at three sites in the Spirit Lake basin of southeastern Saskatchewan, Canada. Instruments, installed at several depths, included soil-water tensiometers, electrical-resistance blocks for monitoring soil-moisture content, and piezometers in the saturated zone. A rain gage was also installed at each site. Laboratory measurements were conducted to determine moisture-retention curves, saturated hydraulic conductivity, bulk density, and other physical properties. Although Freeze and Banner (1970) did not use the water-table fluctuation method to estimate recharge at their sites, they present sufficient information to allow application of the method for a couple of discrete rainfall events.

The moisture-retention curves for Yorkton sandy loam (site A) and Meota sand (site B) are displayed in Fig. 13. The curve for the sandy loam was determined with field data, whereas that for the sand is based on laboratory data. Figure 14a shows rainfall and depth to the water table for site A for 25 days during July and August 1966. The groundwater level was receding, most likely due to evapotranspiration and lateral drainage, prior to the rainfall that began on 4 August. The level then began to rise gradually, reaching a peak on 10 August. Based on extending the recession line beneath the peak, the rise in water level is approximately 49 cm. Specific yield is estimated as $\phi-\theta(H)$, with H equal to 1.8 m. Using the moisture-retention curve, $\phi=0.35$ and $\theta(1.8\text{ m})=0.32$. A value of 0.03 is obtained for S_y. Therefore, a total recharge of 14.7 mm is calculated for the 6-day period. Precipitation during this period was slightly more than 76 mm. After the water table reached a peak, its recession was not as steep as it was prior to the rise. This difference in recession rates is due to the slow drainage of the soils in response to the large storm and to the additional low-intensity rainfall that continued almost daily between 10 and 20 August. During this period, recharge to the water table was probably still occurring. As presented in this paper, the WTF method is incapable of estimating recharge that occurs during hydrograph recession. If the slope of the recession curve in the absence of additional recharge (the theoretical recession slope dh_r/dt) were known, then the additional recharge could be estimated as the product of S_y and the difference between the actual and the theoretical recession slopes ($d(h-h_r)/dt$). However, the theoretical recession curve cannot be predicted, because it is a function of location, season, water-table height, current and antecedent climate, and many other parameters.

Figure 14b shows precipitation and water-level rise at site B for parts of June and July 1967. The water level is in decline for the most of the period, with the exception of a sharp rise that occurred on 15 June in response to less than 25 mm of rainfall. The difference between the peak and the trace of the recession curve is about 0.26 m. S_y is calculated to be 0.08, using the same approach as described for site A. Recharge for the 1-day period is then estimated to be 21 mm, which is very close to the total amount of rainfall. This result is not surprising, given the permeable soil and the shallow water table. Site B appears to be a more appropriate location than site A for application of the water-table fluctuation method; at site B, the water table is shallower and the sediments are more permeable. These factors contributed to the more rapid water-level rise at that site. All of the water arriving

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at the water table at site B probably went into storage, at least for a short period of time. Because the rise at site A was so long and gradual, some water arriving at the water table was likely lost to evapotranspiration or baseflow prior to the time of peak water level. These losses would not be included in the estimated recharge rate.

Other Methods

Description

Although the water-table fluctuation method is the most widely used method for estimating recharge, other methods that are based on measurement of water levels merit some attention. Theis (1937) used Darcy's equation to estimate flow through a cross section of the Southern High Plains aquifer in New Mexico, USA. A simple water budget was constructed assuming steady conditions and no water extraction. The calculated flow rate was divided by the contributing up-gradient area to arrive at an estimate of recharge. The approach requires estimates of hydraulic conductivity (K), hydraulic gradient, and the area up-gradient from the cross section. Using a range of values of K determined from laboratory and aquifer tests and water levels from a network of observation wells, Theis (1937) estimated the annual recharge rate to be 3–7 mm, values that are approximately an order of magnitude less than previous estimates. The method is easy to apply if information on K and water levels is available. However, determination of K and the hydraulic gradient can be costly. It is widely accepted that K is highly variable in space, so the method is limited by the ability to determine representative values of K.

Leakage through aquitards is an important source of recharge for regionally extensive deep, confined aquifers. In a strict sense, this leakage does not conform with the previously stated definition of recharge. Lerner et al. (1990) refer to leakage as interaquifer flow. The subject is briefly addressed here, because attempts to quantify leakage usually rely on knowledge of groundwater levels. As pumping lowers water levels in the confined aquifer, water is drawn out of the confining units or adjoining aquifers, in accordance with Darcy's law. To study the flow of water through confining beds, Hantush (1956) expressed Darcy's law as:

$$R = Q_c/A_c = \left(k'/m'\right)\Delta h \tag{15}$$

where Q_c is the volumetric rate of leakage through the confining bed; A_c is the area over which leakage is occurring; (k'/m') is the leakage coefficient or leakance, where k' is the vertical hydraulic conductivity and m' is the thickness of the confining bed; and Δh is the difference in total head between the aquifer and the top of the confining bed. The method requires measurements of water levels (head) in the aquifer and at the top of or above the confining bed as well as measurements or estimates of k'. Hydraulic conductivity can be measured in the laboratory, if core samples of the confining bed are available. Piezometers in the confining bed can also be used to conduct single-borehole slug tests for estimating k' (Neuzil 1986). Alternatively, k'/m' can be determined through analysis of a constant-rate pumping test on the confined aquifer or from drawdown vs. distance or drawdown vs. time data in a manner similar to that described above for type-curve matching (Walton 1970; Neuman and Witherspoon 1972).

Estimates of groundwater recharge can also be obtained by graphical analysis of flow nets for both confined and unconfined aquifers. Although flow-net analyses have been replaced by groundwater flow models in most applications, they nonetheless deserve some discussion because they have enjoyed widespread use in the past. The technique was first proposed by Forchheimer (1930). A brief description is given here; details are provided in Cedergren (1977). The approach assumes steady flow in a two-dimensional section (either vertical or horizontal). A flow net for a homogeneous, anisotropic system consists of two sets of orthogonal lines: flow lines and equipotential lines. A flow line represents a path over which a water molecule travels from a recharge zone to a discharge zone. The region between two adjacent flow lines is termed a stream tube. An equipotential line represents a line of equal piezometric head or water level. The two sets of lines are drawn to form a series of rectangles (actually, quasi-rectangular elements) that represent equal amounts of flow. Using Darcy's law as in Eq. (15), flow through a rectangle, Q, can be written as:

$$Q = AK\Delta h/\Delta l \tag{16}$$

where A is the cross sectional area of one of the rectangles, K is hydraulic conductivity, and $\Delta h/\Delta l$ is the hydraulic gradient across the rectangle. Recharge for any rectangle is then calculated as Q/A.

Su (1994) developed a method for estimating recharge from knowledge of the time history of water levels in an observation well. His approach assumes an unconfined aquifer overlying a sloping impervious base. An analytical solution was developed for the Boussinesq equation and from this an equation was derived for estimating recharge as a function of head, the derivative of head with respect to time, and seven aquifer or soil parameters. For the watershed that was used to test the model (Uriarra Forest, Australian Capital Territory), the author concludes that terms involving most of these parameters are insignificant and that the following equation is sufficient for estimating recharge:

$$R = S_y h\left(\mathrm{d}h/\mathrm{d}t\right)/b \tag{17}$$

where b is average aquifer thickness. If little spatial variability exists in water levels, then h is essentially equivalent to b and Eq. (17) reverts to Eq. (2). Whether the more complex formula presented by Su always degenerates to Eq. (17) is difficult to determine. Besbes and de Marsily (1984), Morel-Seytoux (1984), and Wu et al. (1997) developed transfer-function methods for relating infiltration patterns to patterns of recharge. These approaches treat water movement through the unsaturated

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Table 5 Estimates of groundwater recharge rates, in m/year, for an upland area of North Dakota, USA, by the water-table fluctuation method (WTF) and the Hantush method for 1979 and 1980 (Rehm et al. 1982)

Material/location	WTF method		Hantush method	
	1979	1980	1979	1980
Sandy material	0.017	0.08	0.71	0.73
Fine-textured material	0.0018	0.0011	0.091	0.081
Sloughs	–	–	0.60	0.70

zone in a manner less rigorous than the Richards equation. Their ease of application could potentially be beneficial in the analysis of spatial variability of recharge. The models require data on precipitation as well as water levels. These methods have not yet gained widespread usage.

Example

Rehm et al. (1982) conducted a recharge study for an upland area of central North Dakota, USA. They used three different methods for estimating recharge: the water-table fluctuation method, the Hantush method, and a flow-net analysis. The 150-km^2 study area contained 175 piezometers and water-table wells. The water table is in glacial deposits of sand, gravel, and loam. These deposits overlie bedrock, silt, and clay units that confine the Hagel Lignite Bed aquifer. Thirty-eight observation wells were used to estimate recharge by the WTF method. S_y was determined from soil cores using Eq. (5) with S_r assumed to be equal to moisture content at –30 kPa. An average value for S_y of 0.16 was obtained. Estimates of recharge to the water table were made for sand and fine-textured materials. Average values are shown for 1979 and 1980 in Table 5.

Ten groups of nested piezometers were used to determine vertical hydraulic gradients for the Hantush method. Hydraulic conductivity was measured in the field at each piezometer using single-hole slug tests. The measured hydraulic conductivity was assumed to be equal to the vertical hydraulic conductivity. Sites were located in the sandy and fine-textured bedrock mentioned above, as well as below two of the many sloughs that are present in the study area. Measured values of K range from 3×10^{-9} m/s for clayey bedrock to 6×10^{-6} m/s for sands. Vertical gradients range from 0.006 in the sands to 1.2 in the fine-textured material. Average rates of recharge are listed in Table 5. They are considerably greater than those estimated from the WTF method. The differences are likely caused by errors inherent in the methods and natural heterogeneities within the system. Rehm et al. (1982) calculated an areal average recharge rate for the study area of 0.025–0.115 m/year by weighting the estimates in Table 5 on the basis of areal coverage of the different hydrogeological settings.

Assuming steady-state conditions, a horizontal flow net for the Hagel Bed was constructed with water-level data for April 1977. The potentiometric surface was contoured at 2-m intervals and a series of flow lines was drawn perpendicular to the contours, thereby forming rectangles of approximately equal area. The flow through each rectangle is described by Eq. (16). The calculated areally averaged recharge rate was 0.010–0.033 m/year, the difference in these two numbers owing to the uncertainty in the hydraulic conductivity of the confining bed underlying the aquifer. These numbers are within the range of estimates obtained from the WTF and Hantush methods.

Summary and Discussion

The water-table fluctuation method is based on the premise that rises in groundwater levels in unconfined aquifers are due to recharge arriving at the water table. Recharge is calculated as the change in water level over time multiplied by specific yield. This approach is a gross simplification of a very complex phenomenon, namely, movement of water to and from the water table. Favorable aspects of the WTF method include its simplicity and ease of use: it can be applied for any well that taps the water table, and an abundance of available water-level data exists. The method requires no assumptions on the mechanisms for water movement through the unsaturated zone; hence, the presence of preferential flow paths does not restrict its use. Recharge rates calculated with the WTF method are values that are integrated over areas of several square meters to hundreds or thousands of square meters. This is a distinct advantage relative to point-measurement approaches, such as methods that rely on measurements within the unsaturated zone. Wells should be located so that the water levels they monitor are representative of the aquifer as a whole.

The WTF method is best applied to systems with shallow water tables that display sharp rises and declines. Analysis of water-level fluctuations can, however, be useful for determining the magnitude of long-term changes in recharge caused, perhaps, by changes in climate or land use. Allison et al. (1990) observed groundwater levels that were steadily increasing at 0.1 m/year following the clearing of native vegetation in southern Australia. Assuming a specific yield of 0.2, this corresponds to an increase in recharge of 20 mm/year, a value consistent with recharge rates determined by other independent methods. Changes in climate or land use are not reflected immediately at the water table. Time is required for the pressure front from increased deep drainage to move downward through the unsaturated zone. This time delay is related to the recharge rate, the soil-water content, and the depth to the water table (Jolly et al. 1989). Allison et al. (1990) observed that 70 years after the change in land use at their site, groundwater levels were rising only where the depth to the water table was less than 40 m. Presumably the pressure front was still being propagated through the unsaturated zone in areas with a deeper water table.

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Limitations of the WTF method include ensuring that water-level fluctuations are indeed due to recharge arriving at the water table and are not in response to the presence of entrapped air, changes in barometric pressure, evapotranspiration, or other complicating phenomena. Determining a proper value for specific yield is a difficult endeavor. Values of specific yield determined from laboratory or field tests are usually dependent on the amount of time allowed for the test, with longer times corresponding to greater values. The WTF method is only capable of estimating recharge when water is arriving at the water table at a greater rate than it is leaving, a condition that produces a water-level rise. Recharge can still be occurring even when a well hydrograph shows that water levels are declining. Such an occurrence simply indicates that the rate of recharge is less than the rate of water movement away from the water table. If water movement away from the water table were equal to the steady recharge rate, no change in water level would occur, and the WTF method would predict no recharge.

Application of the WTF method to fractured-rock aquifers offers some peculiar challenges. Fractures usually serve as the primary conduits for water movement, but they account for a small percentage of the total storage available in the aquifer. Therefore, care must be exercised in selecting a value for specific yield and in analyzing water-level fluctuations.

Accurate quantification of recharge rates is imperative to proper management and protection of valuable groundwater resources. A major concern in application of the methods described in this paper, as well as most other methods for estimating recharge, is the difficulty in assessing the uncertainty associated with any given estimate. Ideally, recharge estimates should be presented along with statistical confidence levels or as a range of likely values. Unfortunately, the hydrogeology discipline has not yet arrived at a point where this is practical. Uncertainties and inaccuracies arise from several sources: spatial and temporal variability in processes and parameter values, measurement errors, and the validity of assumptions upon which different methods are based. Although uncertainty can be quantified for some of these parameters (e.g., measurement error), formal uncertainty analyses cannot be used to evaluate the validity of assumptions. In the face of this dilemma, prudence dictates application of multiple techniques for estimating recharge. Although such an approach does not necessarily guarantee improved accuracy, resolution of inconsistencies among different estimates aids in identifying errors, invalid assumptions, or limitations of specific methods. The simplicity of estimating recharge rates from information on temporal fluctuations of groundwater levels is attractive. In spite of the potential errors inherent in application of the WTF method, this simplicity, as well as the need to apply multiple estimation methods, suggests that this method should be used wherever appropriate.

Future research related to use of groundwater levels to estimate recharge would be beneficial in several areas. In addition to improved approaches for assessing uncertainty, a refined definition of specific yield is highly desirable. The current definition of S_y and its use in the WTF method do not properly account for flow and storage processes within the unsaturated zone. A revised definition would account for depth to water table, the moisture-characteristic curve of the sediments (including hysteresis), and the timing between water-table rises and declines. A valuable first step in this direction would be an explicit formulation of the term dW'/dt suggested by dos Santos and Youngs (1969). This term describes the temporal change in the shape of the moisture profile above the water table in response to changes in water-table height. The WTF method would be more accurate and useful if an expression could be developed to predict the slope of the hydrograph recession curve in the absence of recharge. This would permit estimation of recharge based on changes in that slope. The approaches based on transfer-function models should also be further explored. Although they have not been widely applied, they have the potential benefit of relating recharge to precipitation.

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