GROUNDWATER RECHARGE

Sophocleous M.
University of Kansas, Lawrence, United States

Keywords: groundwater flow system, turnover time, physical methods, tracer methods, diffuse recharge, localized recharge, potential recharge, induced recharge, rejected recharge, capillary flow, macropore flow, fingered flow, available-water capacity, soil-moisture deficit, conceptual model, inverse model, water balance method, zero-flux plane method, hybrid water fluctuation method, lysimeters, seepage meters, uncertainties, hydrologic budget, regionalization, spatial variability, temporal variability, GIS.

Contents

1. Introduction and Terminology
2. Groundwater Flow Systems
3. Flow System Extensions
4. Sources and Mechanisms of Recharge
5. Conceptual Models of Recharge
6. Methodologies for Recharge Estimation
7. Factors Influencing Recharge, Predictive Relationships, and Recharge Regionalization
8. Difficulties and Challenges in Recharge Estimation
Glossary
Bibliography
Biographical Sketch

To cite this chapter

Summary

This work attempts to establish a hydrogeological framework for the understanding of natural groundwater recharge processes in relation to climate, landform, geology, and biotic factors. It begins with the concepts of groundwater flow systems, which form the basis for comprehending recharge processes. It then concentrates on the sources and mechanisms of groundwater recharge and stresses the importance of developing correct conceptualizations of recharge. Several recharge estimation methodologies are then outlined, with an emphasis on minimizing uncertainty. The article then discusses developing predictive relationships for recharge based on the major recharge-influencing factors, and regionalizing point recharge data. A discussion of difficulties that face the field of recharge assessment follows, with recommendations as to how to minimize them.

Although there are various well-established methods for the quantitative estimation of recharge, few can be applied successfully in the field. All are characterized by major uncertainties. When estimating groundwater recharge it is essential to proceed from a good conceptualization of different recharge mechanisms and their importance in the study area. Besides this conceptualization the objectives of the study, available data and
resources, and possibilities of obtaining supplementary data should guide the choice of recharge-estimation methods. A key to deciding on a methodology is related to the spatial and temporal scale of interest. If the major concern is obtaining good recharge estimates over a limited area, then the need for detailed information is evident. However, small-scale variability in local recharge ceases to be a major problem for regional studies. In addition, the inherent temporal variability of recharge has important implications for the measurement techniques adopted. Different measurement techniques provide recharge estimates with different temporal scales. For example in arid and semiarid areas, where deep drainage fluxes are low and water tables are deep, interpreting groundwater hydrographs and water table rises may be misleading for estimating rates of groundwater recharge; chemical and isotopic methods are likely to be more successful than physical methods in such cases.

1. Introduction and Terminology

The endless circulation of water as it moves in its various phases through the atmosphere, to the Earth, over and through the land, to the ocean, and back to the atmosphere is known as the hydrologic cycle. This cycle is powered by the Sun; through phase changes of water (i.e. evaporation and condensation) involving storage and release of latent heat, it affects the global circulation of both the atmosphere and oceans, and hence is instrumental in shaping weather and climate. The efficiency of water as a solvent makes geochemistry an intimate part of the hydrologic cycle; all water-soluble elements follow this cycle at least partially. Thus, the hydrologic cycle is the integrating process for the fluxes of water, energy, and the chemical elements. This cycle is the foundation of hydrological science and occurs over a wide range of space and time scales.

Figure 1 illustrates different parts of the land-based portion of the hydrologic cycle that affect an individual watershed or catchment.

![Figure 1. Schematic representation of the hydrologic cycle (from Freeze, 1974)](image)
Water enters the hydrologic system as *precipitation*, in the form of rainfall or snowmelt. It leaves the system as streamflow or *runoff*, and as *evapotranspiration*, a combination of evaporation from open bodies of water, evaporation from soil surfaces, and transpiration from the soil by plants. Precipitation is delivered to streams on the land surface as *overland flow* to tributary channels, and in the subsurface as *interflow* or lateral subsurface flow and *baseflow* following *infiltration* into the soil.

A portion of the infiltrated water enters the groundwater or aquifer system by passing through the *vadose or unsaturated zone*, and it exits to the atmosphere, surface water, or to plants. As Figure 1 shows, the flowlines deliver groundwater from the highlands towards the valleys, or from the recharge areas to the discharge areas. As the figure also shows, in a *recharge area* there is a component to the direction of groundwater flow that is downward. *Groundwater recharge* is the entry to the saturated zone of water made available at the water table surface. Conversely, in a *discharge area* there is a component to the direction of groundwater flow that is upward (Figure 1). *Groundwater discharge* is the removal of water from the saturated zone across the water table surface. The patterns of groundwater flow from the recharge to the discharge areas form *groundwater flow systems*, which constitute the framework for understanding recharge processes.

2. Groundwater Flow Systems

The route which groundwater takes to a discharge point is known as a *flow path*. A set of flow paths with common recharge and discharge areas is termed a *groundwater flow system*. The three-dimensional closed system that contains the entire flow paths followed by all water recharging the groundwater system has been termed a *groundwater basin*. Groundwater possesses energy mainly by virtue of its elevation (elevation or gravitational head) and its pressure (pressure head). Groundwater can also possess kinetic energy by virtue of its movement, but usually this energy is negligible because of groundwater’s low velocities. Groundwater moves from regions of higher energy to regions of lower energy. A measure of groundwater’s energy is the level at which the water stands in a borehole drilled into an aquifer and measured with reference to an (arbitrary) reference level or datum, such as sea level. This height at which water stands above a reference datum is called *hydraulic head*, or simply *head*. The hydraulic head, for most practical purposes, is composed of the sum of the pressure head and gravitational or elevation head. Both of these component forms of energy (in other words, elevation energy and pressure energy) are known as *potential energy*. The change in hydraulic head over a certain (arbitrary) distance along the groundwater flow path is called hydraulic gradient or head gradient and constitutes the driving force for groundwater movement. According to *Darcy’s law*, which describes the flow of groundwater through an aquifer, the groundwater flow rate is directly proportional to the cross-sectional area through which flow is occurring, and directly proportional to the hydraulic gradient. Gravity due to elevation differences is the predominant driving force in groundwater movement. Under natural conditions, groundwater moves “downhill” until it reaches the land surface, such as at a spring, or the root zone, where it is evapotranspired to the atmosphere.
Therefore, groundwater moves from interstream (higher) areas toward streams or the coast (lower areas). Except for minor surface irregularities, the general slope of the land surface is also toward streams or the coast. The depth to the water table is greater along the divide between streams than it is beneath a floodplain. In effect, the water table is usually a subdued replica of the land surface.

A groundwater flow pattern is controlled by the configuration of the water table, and by the distribution of hydraulic conductivity in the rocks. The water table, in turn, is affected by topography, and is controlled by the prevailing climate. The flow pattern is therefore a function of topography, geology, and climate. These three parameters have been collectively termed the hydrogeologic environment. In addition, biotic influences affect most aspects of the hydrologic cycle, including groundwater. Vegetation, for example, regulates the rate at which a land surface returns water vapor to the atmosphere, and humans alter nearly all aspects of water’s distribution and behavior on land.

Figure 2. Effects and manifestations of gravity-driven flow in a regionally unconfined drainage basin (adapted from Tóth, 1999)

Based on their relative position in space, three distinct types of flow systems have been recognized (right-hand side of Figure 2):
1. A local system, that has its recharge area at a topographic high and its discharge area at the immediately adjacent topographic low.

2. An intermediate system, characterized by one or more topographic highs and lows located between its recharge and discharge areas.

3. A regional system, that has its recharge area at the major topographic high and its discharge area at the bottom of the basin. Regional flow systems are at the top of this hierarchical organization; all other flow systems are nested within them.

On the basis of a comparative study of variations in selected geometric parameters—such as depth to impermeable basement, slope of the valley flanks, and local relief—the conditions under which local, intermediate, and regional systems may develop were elucidated. If local relief is negligible, and there is a general slope of topography, only regional systems will develop. Because no extensive unconfined regional system can span the valleys of large rivers or highly elevated watersheds, pronounced local relief generally is an indicator of a local system. The greater the relief, the deeper the local systems that develop. Under extended flat areas unmarked by local relief, neither regional nor local systems can develop. Waterlogged areas may develop, and the groundwater may be highly mineralized due to concentrations of salts.

The recognition that groundwater moves in systems of predictable pattern in topography-controlled flow regimes, and that various identifiable natural phenomena are regularly associated with different segments of the flow systems, was only made in the 1960s when the system-nature of groundwater flow was first understood. This recognition of the system-nature of subsurface water flow has provided a unifying theoretical background for the study and understanding of a wide range of natural processes and phenomena, and has thus shown flowing groundwater to be a general geological agent.

A schematic overview of groundwater flow distribution, and some of the typical hydrogeologic conditions and natural phenomena associated with it in a gravity-flow environment, is presented in Figure 2. On the left side of the figure, a single flow system is shown in a region with insignificant local relief; on the right side, a hierarchical set of local, intermediate, and regional flow systems is depicted in a region of composite topography. Each flow system has an area of recharge, an area of throughflow, and an area of discharge. In the recharge areas, the hydraulic heads, representing the water’s potential energy, are relatively high and decrease with increasing depth; water flow is downward and divergent. In discharge areas, the energy and flow conditions are reversed: hydraulic heads are low and increase downward, resulting in ascending and converging water flow. In the areas of throughflow, the water’s potential energy is largely invariant with depth (the isolines of hydraulic heads are subvertical), and consequently flow is chiefly lateral. The flow systems operate as conveyor belts, with the flow serving as the mechanism for mobilization, transport (distribution), and accumulation of mass and energy, thus effectively interacting with their ambient environment.
3. Flow System Extensions

Studying flow systems in groundwater basins may help gain an understanding of the interrelations between the processes of infiltration and recharge in topographically high parts of the basin and of groundwater discharge through evapotranspiration and baseflow. For example, at least some of the water derived from precipitation that enters the ground in recharge areas will be transmitted to distant discharge points, thus causing a relative moisture deficiency in soils overlying recharge areas. Water that enters the ground in discharge areas may not overcome the upward potential gradient, and therefore becomes subject to evapotranspiration close to its point of entry. Water input to saturated discharge areas generates overland flow, but in unsaturated discharge areas infiltrating water and upflowing groundwater is diverted laterally through superficial layers of high hydraulic conductivity. Further, the ramifications of anthropogenic activities in discharge areas are immediately apparent. Some of these include:

- waterlogging problems associated with surface-water irrigation of lowlands,
- waterlogging problems associated with destruction of phreatophytes, or plants discharging shallow groundwater, and
- pollution of shallow groundwaters by gravity-operated sewage and waste-disposal systems located in valley bottoms in semiarid basins, where surface water is inadequate for dilution.

The spatial distribution of flow systems will also influence the intensity of natural groundwater discharge. In the example in Figure 2, the main stream of a basin may receive groundwater from the area immediately within the nearest topographic high, and possibly from more distant areas. If baseflow calculations are used as indicators of average recharge, significant error may be introduced in that baseflow may represent only a small part of the total discharge occurring downgradient from the line separating the areas of discharge from the recharge areas.

In groundwater hydrology today, the system concept is fundamental to thinking about a groundwater problem. System thinking is vital to the understanding of practical problems, such as groundwater contamination from point sources, or the impact of a structure such as a dam, waste disposal facility, or gravel pit. Many such studies suffer irreparably from the failure to place the local site in the context of the larger groundwater system of which the site is only a small part.

4. Sources and Mechanisms of Recharge

The sources of recharge to a groundwater system include both natural and human-induced phenomena. Natural sources include recharge from precipitation, lakes, ponds, and rivers (including perennial, seasonal, and ephemeral flows), and from other aquifers. Human-induced sources of recharge include irrigation losses from both canals and fields, leaking water mains, sewers, septic tanks, and over-irrigation of parks, gardens, and other public amenities.

Recharge from these sources has been classified as direct recharge from percolation of precipitation and indirect recharge from runoff ponding. Other classifications include
localized or focused recharge, preferential recharge, induced recharge, mountain front recharge, and others.

Direct or diffuse recharge is defined as water added to the groundwater reservoir in excess of soil-moisture deficits and evapotranspiration, by direct vertical percolation of precipitation through the unsaturated zone: that is, recharge below the point of impact of the precipitation. This mode of recharge is spatially distributed (diffuse), and results from widespread percolation through the entire vadose zone. It is typical of humid climates because frequent, regular precipitation maintains a high water content in the soil, so that there is little additional storage capacity in the vadose zone. Thus infiltration can be routed quickly through the vadose zone to the saturated zone. This recharge raises the water table, which leads to increased streamflow. Therefore, in humid climates flowing perennial streams are typically groundwater discharge areas, sustained by diffuse recharge in the basin.

Indirect recharge results from percolation to the water table following runoff and localization in joints, as ponding in low-lying areas and lakes, or through the beds of surface water courses. Two distinct categories of indirect recharge are evident:

- that associated with surface-water courses, and
- a localized or focused form resulting from horizontal surface concentration of water in the absence of well-defined channels, such as recharge through sloughs, potholes, and playas.

Recharge through such topographic depressions, which are common in the Canadian prairies and Great Plains of the United States, is also known as depression-focused recharge, and occurs where surface runoff or lateral flow of subsurface moisture accumulates within or beneath such depressions on the landscape. Thus, knowledge of lateral subsurface flow processes becomes important in understanding recharge processes. In arid and semi-arid regions, localized and indirect recharge are often the most important sources of natural recharge.

Percolation to the water table from streambeds takes two forms, depending on whether there is a saturated connection between the stream and the water table. Where no connection exists (Figure 3a), a situation typical of arid zones where water tables are generally deep, water moves downward from the streambed to the water table, forming a groundwater mound which then dissipates laterally away from the stream. As long as the mound is recharged by unsaturated flow, there is no hydraulic connection between the groundwater and the stream flow, in the sense that the recharge rate is almost unaffected by the groundwater levels. Yet, even when the unsaturated condition is present, the stream and aquifer may in fact be hydraulically connected in the sense that further lowering of the regional water table could increase channel losses. At some critical depth to the water table, however, further lowering has no influence on channel losses. At this depth, which depends mostly on soil properties and water stage in the channel, the aquifer becomes hydraulically disconnected from the stream. If the distance from the water table to the stream stage is greater than approximately twice the stream width, the seepage rapidly begins to approach the maximum seepage for an infinitely deep water table. The parameters determining the recharge process are the width, depth,
and duration of streamflow, and the hydraulic characteristics of the local material in and below the streambed.

Figure 3. Recharge from streambeds (a) with no hydraulic connection, and (b) with hydraulic connection

In less arid areas, water table levels tend to rise closer to the streambed. In these situations, a hydraulic connection will usually exist between the stream and the groundwater (Figure 3b), and the recharge rate will decrease as the water table rises. The recharge process will be dominated by horizontal rather than vertical flow, and will have a much shorter turnover or transit time than when there is no hydraulic connection. In these less arid environments, there is also likely to be recharge from general catchment percolation, and the mix between the two mechanisms may be hard to predict. Mountain front recharge typically involves complex processes of unsaturated and saturated flow in fractured rocks, as well as infiltration along channels flowing across alluvial fans. On a large scale, mountain front recharge through fractured bedrock is primarily a diffuse recharge process, whereas infiltration from mountain streams is considered a localized recharge process. Vertical leakage across low-permeability strata and underflow from adjacent aquifers (interaquifer flows) can be important sources of recharge, but typically they do not involve the vadose zone.

In areas where the potential recharge rate exceeds the rate at which water can flow laterally through the aquifer, the aquifer becomes overfull and available recharge is rejected, a condition known as rejected recharge. In this situation, groundwater pumping in recharge areas can increase the rate of underground flow from the area and more water can be drawn into the aquifer as induced recharge.

Two different flow mechanisms, termed capillary and viscous flow, drive potential groundwater recharge through the vadose zone. Capillary flow takes place in pores with a diameter less than approximately 3 mm in which capillary forces, together with gravity, determine the flow process. A porous medium in which capillary forces are dominant behaves like a sponge: in other words, no free drainage occurs even at high water contents, and capillary rise causes water to move upwards against the pull of gravity. The capillary flow process normally leads to stable wetting fronts, but unstable wetting fronts that are characterized by fingered flow sometimes form. (Fingered flow is unstable flow whereby the percolating water may concentrate at certain points to break into the sublayer in the form of finger-like or tongue-like protrusions.) Theoretical and experimental research results demonstrate that dry sandy soils are prime sites for the
occurrence of unstable wetting fronts (in other words, boundaries between the wetted and dry regions of soil during infiltration), which may be expected in dune fields that often provide a large portion of the recharge under semi-arid conditions. This type of flow also occurs in the transition of percolating water from a fine-textured top layer to a coarser-textured sublayer. Unfortunately, it is not yet possible to quantify the effects of fingered flow on recharge rates.

**Macropore flow** occurs in pores with a diameter or width larger than 3 mm, such as cracks in clay soils, rock fractures, fissures in sediments, solution channels, worm holes, and old root channels. In macropore flow, the effects of capillarity are no longer felt, and the flow process is dominated by viscous forces and gravity. Flow through macropores is also known as preferential or bypass flow. The resulting recharge is called **preferential recharge**, which preferentially takes place through such macropores, as opposed to diffuse recharge, which takes place through the entire vadose porous medium. The velocity with which water moves from the soil surface to the water table is often several orders of magnitude higher through macropores than through the soil matrix. Saturated flow through macropores can be quantified using Poiseuille’s equation, as opposed to Darcy’s equation for diffuse flow. However, capillary and macropore flow frequently occur simultaneously within the same soil mass without the presence of clearly defined macropores. The depth to which preferential flow is effective depends on the nature and connectivity of the macropores or preferred pathways, but they are rarely effective beyond the root zone depth of approximately 2 m.

![Diagram of macropore flow](image)

Figure 4. Schematic representation of the fluxes involved during infiltration into a macroporous soil: see text for explanation of symbols (from Germann and Beven, 1985)

The process of macropore flow, shown in Figure 4, is somewhat similar to localized recharge, albeit on a much smaller scale, because horizontal water movement is required. When the overall water input from precipitation or irrigation, $q^*(t)$, exceeds the infiltration capacity of the soil, $i(t)$, a horizontal overland flow, $o(t)$, is generated that
causes a water flux into the macropores, $q(0,t)$. This flux causes water content inside the macropore, $w(z,t)$, to increase. A fraction of the water, $r$, that occupies a macropore at a given depth will be absorbed by the soil matrix through the macropore walls, while the remainder will percolate downwards into the macropore, $q(z,t)$. When the infiltration rate, $i(t)$, decreases with time and with increasing antecedent soil-water content, the opportunity for overland flow, $o(t)$, and macropore flow, $q(0,t)$ increases.

5. Conceptual Models of Recharge

The key to successful hydrological measurement and modeling is the appropriate conceptualization of the system of interest. The conceptual model includes the recognition of important hydrological processes, pathways, boundary conditions, spatial and temporal limits, inputs and outputs, and constraints. If the initial conceptual model is wrong, then recharge estimates based on this model will be unreliable.

For example, at the plot scale, important elements of a conceptual water balance model aimed at recharge prediction might be:

- the pattern and amount of evaporation with respect to land cover,
- the importance of overland flow,
- the existence of any lateral throughflow,
- the datum in the profile beyond which drainage will become groundwater recharge,
- the transience and frequency of recharge events, and
- the hydraulic pathway(s) that water may take through the profile.

At the catchment scale, the potential complexity of the correct conceptual model increases dramatically: it includes not only all of the considerations of the plot-scale recharge phenomena, but also the distribution of these phenomena in space, as well as the interaction of the water balance components of adjacent plots. For example, overland flow or shallow throughflow can become groundwater recharge downslope. The complexity of lateral flow systems, and their definition, becomes paramount at the catchment scale. Important considerations include:

- the presence of any confining bed(s), their depth, hydraulic conductivity, and distribution across the catchment;
- the hydraulic head surface of groundwater system(s) and the degree of confinement of aquifers across the catchment; and
- the geomorphic and geological features associated with the discharge of groundwater, which define, locate, and control saturated areas within the catchment.

Successful estimation of groundwater recharge depends on first identifying the probable flow mechanisms and important features influencing recharge for a given locality, since it cannot be assumed that a procedure developed successfully for one area will prove equally reliable for another. Thus, in each case involving recharge estimation modeling, conceptual models must be based on local data and experience.
In summary, the vital aspects of a conceptual model of catchment recharge processes must consider:

- what parts of the landscape contribute to groundwater recharge,
- how these areas change with time,
- if the topographic catchment is the same as the groundwater catchment,
- what controls recharge rates from place to place, and
- the importance of lateral redistribution of runoff and shallow throughflow to recharge downslope.

If appropriate, one can use a mean recharge rate over the entire catchment, or at least over that portion of the catchment subject to recharge. Expected rates or changes in the rate of recharge, however estimated or modeled, can be applied uniformly over this area. In other words, the recharge across the landscape can be treated one-dimensionally. The assumption here is that the lateral redistribution of water in the catchment takes place only after the recharge reaches the groundwater table, and that the subsequent discharge of this groundwater does not in turn change the area subject to recharge (that is, the discharge area does not grow significantly in size). The conditions where such an approach might be appropriate may be found in areas with deep, uniformly permeable soils, deep groundwater, and a very low topographic (hydraulic) gradient.

However, most catchments are heterogeneous in their topography, soil, geology, and land cover. To model catchment recharge in these systems, the spatial pattern of these influences on recharge must be taken into account. There are two basic approaches to this problem, depending on the nature of the recharge modeling to be undertaken. In the first general approach, the catchment is broken up into land units in which recharge can be expected to respond similarly according to climate inferences on recharge, and its relation to land use are then distributed spatially on this basis. In the second approach, the individual controls on recharge are distributed independently and serve as input into a spatially explicit water-balance model yielding recharge.

In either approach, an appreciation and understanding of scaling hydrologic parameters is essential. As one looks at larger areas of the landscape, and thus incorporates natural heterogeneity into the modeling, the parameter values used to represent hydrologic processes often change. For example, the saturated hydraulic conductivity of a soil profile will be different from the inferred conductivity of a hillslope: this in turn will be different from the inferred conductivity of an entire catchment, even if climate, geology, vegetation, and other variables are held constant. Thus, it is not reasonable to assume scale invariance in model parameters as one moves from point measurements to entire catchments. This even holds true for the mean of many point estimates of model parameters. The search for scale invariant model representations of hydrologic phenomena for catchments has yet to yield a solution. Indeed, it is unlikely that any general scaling theory can be developed because of the dependence of hydrological systems on historic and geological perturbations. In most watershed models, equations representing hydrologic processes across scales usually involve “effective” parameters: in other words, the parameter values change with scale.
6. Methodologies for Recharge Estimation

A number of methodologies are used to estimate recharge. These can be classified as:

- direct or indirect;
- physical, chemical, or isotopic;
- methods based on the analysis of inflow, outflow, or aquifer response;
- methods based on the unsaturated or saturated zones; and
- methods based on numerical modeling of groundwater flow, soil-water flow, both soil and groundwater flows, or modeling of the hydrologic balance at plot, field, or watershed scales.

Additional classifications also exist. Within each methodology, a number of estimation techniques are available. Here we combine these methodologies into two groups: physical methods and tracer methods.

6.1. Physical Methods for Recharge Estimation

Physical methods rely on direct measurements of hydrological parameters, or on estimates of soil and/or aquifer physical parameters. Physical methods are frequently used to estimate precipitation recharge because they are quick, inexpensive, or straightforward. However, these methods are often problematic in arid and semi-arid regions. There are several reasons for this:

- The low recharge fluxes largely depend on the vadose zone physical parameters, and significant variations in fluxes may occur with small changes in these physical parameters. Unfortunately, it is almost impossible to detect such small changes in physical parameters.
- The extreme temporal variability of arid climates means that long time series are needed to assess mean annual recharge rate.
- Spatial variability caused by changes in local topography, soil type, and vegetation requires a large number of measurement sites to assess the spatially averaged recharge rate.

Nevertheless, with prudent appreciation of their limitations, physical methods can be a helpful tool for evaluating precipitation recharge.

When characterizing groundwater recharge, a distinction between potential and actual recharge needs to be made. Potential recharge is soil-water that percolates below the root zone, whereas actual recharge is soil-water that reaches the aquifer. Most potential recharge water will be stored in the vadose zone at negative pressures (suctions) and is not available for exploitation. In addition, it may still be lost later by an increase in vegetation rooting depth, capillary rise, or upward vapor transport. Conversely, actual recharge is the amount of water that in fact reaches the water table, and can be pumped.

6.1.1. Indirect Physical Methods

Indirect physical methods for estimating groundwater recharge consist of:
• empirical methods,
• water balance methods based on estimates of soil physical properties, and
• numerical modeling methods.

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

$$R + k_1(P - k_2) \tag{1}$$

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. They are probably most useful for making first-guess estimates of recharge where annual recharge is fairly high (> 50 mm per year), and thus should seldom be used in arid or semiarid regions.

Methods relying on estimates of soil physical parameters generally fall into the following classes:

• soil-water balance,
• zero-flux plane method,
• estimation of water fluxes beneath the root zone using unsaturated hydraulic conductivity and the gradient in soil-water potential, and
• estimation of water fluxes in the saturated zone based on Darcy’s law and flow-net analysis.

Additional methods also exist that may not fit well into one of these classes, such as gravity surveys for measuring changes in aquifer storage resulting from recharge events. Increased accuracy in measuring temporal variations in the Earth’s gravity field has recently allowed the use of gravity observations to deduce subsurface water mass changes resulting from precipitation and consequent recharge events.

**Water Balances:** This group of methods estimates recharge as the residual of all other fluxes. The principle is that other fluxes can be measured or estimated more easily than recharge. Examples of water balance methods include:

• *Soil-moisture budgets:* rainfall and potential evapotranspiration are inputs to a soil-moisture accounting procedure, with actual evapotranspiration and recharge as the outputs.

• *River-channel water balances:* upstream and downstream flows are differenced to calculate recharge or—more accurately—*transmission losses*. (A related stream hydrograph analysis technique based on baseflow-separation techniques is founded on steady-state water-balance calculations, whereby the estimate of discharge based on baseflow-separation or baseflow-recession analysis of the stream hydrograph, must equal recharge. This technique is considered too empirical and approximate to give reliable quantitative estimates.)

• *Water-table rises:* the volume stored beneath a rising water table is equated to recharge, after allowing for other inflows and outflows such as pumping wells and aquifer throughflow.
The simplicity of the latter method made it a popular one. However, calculating the volume of water stored between lowest and highest water table positions over a study period interval involves reliable estimates of the aquifer’s specific yield values, which may be difficult to obtain.

The advantages of water-balance methods are that they use readily available data (rainfall, runoff, water levels), are easy to apply, and account for all water entering a system. The major disadvantage is that recharge is the residual or remainder of all other hydrologic components in the water balance equation, and constitutes only a small difference between large-number components, such as precipitation and evapotranspiration. Errors can be high, with the errors in all the other fluxes accumulating in the recharge estimate. For example, high river flow can often only be estimated to ±25%. If recharge is 25% of flow, the error in estimating it is ±100%. Other disadvantages include the difficulty of estimating other fluxes in the water balance equation. For example, evapotranspiration cannot be measured easily, yet it is often the largest outward flux. Physical properties like specific yield are central to some water-balance methods, such as those based on water-table rises, but are not easily defined or measured.

The natural timescale for water-balance methods is the duration of a recharge event. Recharge processes are often nonlinear, so that estimates based on longer time intervals should be summed over the individual events rather than calculated for the whole interval at once. Long records are available for much of the data used for balances (rainfall, runoff), so that long time series of recharge can often be calculated.

A methodology consisting of a combination of soil-moisture budget and water-table rise analyses is known as hybrid water fluctuation method. This combination methodology, designed to minimize water balance errors, was successfully applied in the US central Kansas Plains region, which is characterized by semiarid to subhumid climate, relatively flat terrain, and a relatively shallow water table.

The hybrid water fluctuation methodology can be summarized as follows. Neutron-moisture profile readings are collected onsite at least once a week during the recharge season (usually spring and fall). The soil-water balance methodology for each recharge-producing storm period is applied and the resulting water table rise is noted, provided it is confirmed that the water-table rise is due to incoming soil-water from above, as checked with tensiometer readings and/or deeper water content measurements. The recharge estimate resulting from the soil-water balance is then divided by the corresponding water table rise to obtain an estimate of the effective storativity or fillable porosity of the region near the water table. Several such estimates are obtained and averaged. This average is, in effect, the site-calibrated effective storativity value. It can be used to translate each water table rise, tied to a specific storm period, into a corresponding amount of groundwater recharge. In the central Kansas prairies, which are characterized by mostly permeable sandy soils and shallow water table, the time lags between the occurrence of a recharge-causing rainstorm and the corresponding water-table rise usually range from less than a day to just a few days.
Recharge estimation errors in the hybrid water fluctuation method are reduced by running a storm-period-based soil-water balance throughout the year, in combination with the associated water level rise, thus avoiding masking short periods of recharge by the averaging effect of monthly or larger time-interval data. Furthermore, during the recharge-producing rainstorm periods under consideration, the evapotranspiration (ET) estimates are usually significantly smaller than the precipitation amounts. Therefore, even a large ET error on a relatively small quantity may not significantly affect the recharge estimate. Also, by employing the Complementary Relation Areal Evapotranspiration (CRAE) methodology, which permits areal ET to be estimated from its effects on the routinely-measured temperatures and humidities, the soil/plant system complexities can be avoided, as well as the need for locally-calibrated coefficients. (The CRAE concept takes into account interactions between the evaporating surfaces and the overpassing air, whereby a decrease in the availability of water for areal ET causes the overpassing air to become hotter and drier, which in turn causes the potential ET to increase.) In addition, the errors inherent in the soil-water balance approach can be appreciably reduced by corroborating the estimation of recharge with increases in soil-water content at depth, and with unequivocal fluctuations of the water table, provided it is relatively shallow. Furthermore, close monitoring of shallow and deeper hydraulic gradients using multi-level piezometers makes it possible to ascertain whether water table rises are due to lateral inflow at the site or to vertical accretion from rainfall percolation. This combined methodology results in better and more reliable recharge estimates than either the soil-water budgeting procedure or the water-table-rise analysis used singly, and does not require additional difficult-to-measure variables.

**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 this 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.

**Estimation of Unsaturated Water Fluxes:** Several studies have reported use of unsaturated zone hydraulic conductivity $K(\theta)$ or $K(\psi)$, and water retention 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. 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

$$R = K(\theta)\Delta H_T$$

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

Both $K(\theta)$ and $K(\psi)$ relationships are difficult and time consuming to determine, in the field or 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. As a result the annual recharge flux could vary significantly, depending on how the mean hydraulic conductivity is computed.

**Estimation of Saturated Water Fluxes:** An equivalent method for recharge estimation, based on saturated flow governed by Darcy’s law, is simpler, especially when assuming steady state conditions and employing flow-net analysis. The only measurements needed are values of hydraulic head and hydraulic conductivity to construct a quantitative flow net. This consists of a set of intersecting lines of equal hydraulic head values (known as equipotential lines) and flow lines representing two-dimensional steady flow through a porous medium (Figures 1 and 2). Two-dimensional, vertical flow nets constructed along the general groundwater flow direction from water table and hydraulic head field measurements provide an approximate but straightforward way of identifying areas of recharge and discharge, and thus of estimating recharge.

**Numerical Models for Estimating Recharge:** Different types of models are available for estimating groundwater recharge:

- numerical models that solve one-, two-, or three-dimensional forms of the water flow or Richards equation,
- parametric hydrologic models that use a numerical or analytical relationship between infiltration or precipitation and recharge,
- groundwater flow models, and
- combined or integrated watershed and groundwater models.

Numerical modeling methods take transient flows and storage changes into account and can include spatial variability of physical properties, of which hydraulic conductivity is one of the most important. However, data requirements and computing load are both high. Such models are used to estimate model parameters, in this case recharge, based on known values of hydraulic head. Such an approach is known as a solution of an inverse problem. This is in contrast to the forward or direct problem, where model parameters are considered known and hydraulic head is computed.

Should one possess the analytical expressions for hydraulic head and transmissivity in the groundwater flow equation, determination of recharge would be a trivial exercise of calculus in computing the derivatives of the groundwater flow equation. However, hydraulic heads are always measured with a degree of inaccuracy. Differentiating such noisy data leads to large errors in recharge estimation.

Integrated watershed and groundwater models allow a complete analysis of the land-based hydrologic cycle, thus providing the means for evaluating the impacts of land use, irrigation development, and climate change on both surface water and groundwater resources. Such models allow predictions of the impact of management changes on total water supplies, including recharge. The seasonal variation of water table levels and recharge can be more accurately predicted by the soil-moisture accounting system employed in the integrated model than by using only a groundwater model. This increased flexibility, however, comes at the expense of increased complexity and the expertise needed to use integrated watershed modeling effectively. Although integrated models require extensive data, such integrated modeling constrains the adjustment of
model parameters during calibration because overall water budgets must be observed. Whereas traditional methods used to calibrate groundwater models may include adjustments to recharge rates, in an integrated model recharge is completely constrained by the overall water budget for the surface-water system. In addition, stream–aquifer interactions, including stream-derived recharge, are constrained by the generated amount of surface runoff to streams. This, in turn, impacts on the stream stage and thus the driving forces for stream–aquifer interaction.

The principal advantages of the numerical methods are that they attempt to represent the actual physical processes of interest, and that they allow predictions of future recharge regimes resulting from different land uses and climatic changes. These advantages are often countered by the need to make simplifying assumptions in order to reduce the computational effort. For example, numerical models of the soil zone usually assume a single-porosity medium with no spatial variation in properties. In practice many soils may have dual porosity, with preferred pathways during high saturation: in other words, at times of recharge.

The correct timescale for such models depends on the rate of fluctuation of heads, varying from seconds for rainfall into soil to seasonal or longer spans for seepage between aquifers. Effectively addressing the multiple temporal (as well as spatial) scales involved in recharge estimation constitutes a major problem in modeling recharge processes. In addition to such obstacles and uncertainties, large data requirements often make application of numerical models difficult.

6.1.2. Direct Physical Methods

In contrast to the numerous indirect physical methods, there is only one direct method for estimating diffuse recharge. This involves the construction of a lysimeter. Lysimeters comprise enclosed blocks of disturbed or undisturbed soil, with or without vegetation, that are hydrologically isolated from the surrounding soil in order to assess or control various elements of the water balance. There is also only one direct method for estimating indirect recharge associated with surface water bodies in direct hydraulic connection with an underlying aquifer. This involves the use of seepage meters inserted in the streambed or lakebed that can provide direct point measurements of localized recharge.

Lysimeters are expensive and permanent instruments. They are typically filled with disturbed soils, which generally have water content profiles that differ to 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 (such as ceramic cup extractors) and a vacuum system are installed at the base. This last factor, however, is unlikely to be a problem if the lysimeter is relatively deep and the vegetation is shallow-rooted. While lysimeters have been useful in quantifying drainage at waste sites under arid conditions, they have limited 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 and a 3 m depth, although lysimeters as deep as 18 m have been constructed. Because lysimeters are effective for the study of recharge mechanisms, and
yield the high-quality data needed in computer model calibration for simulating the water balance, some specialists recommend that more lysimeter-recharge studies be undertaken worldwide in a variety of climatic and soil conditions. However, the initial construction costs and the long-term monitoring requirements represent a serious extended commitment.

Seepage meters were originally developed to measure canal seepage losses. They involve a seepage bell or cylinder that is pushed into the canal-bed sediment, the infiltration rate being measured by constant or falling head techniques. Their advantages include being:

- lightweight and easily transportable,
- relatively cheap,
- simple to operate,
- rapidly measurable, and
- producing observations that are directly convertible into a seepage value.

Difficulties are encountered in gravelly or stony sediments, or in sandy sediments, which may be washed from around the seepage cylinder by eddy currents and sediment disturbance, or because of an ineffective seal around the inserted seepage cylinder. The number of measurements per unit of area needed to arrive at a reasonable average depends on the degree of heterogeneity in the seepage loss at the specific site. In conclusion, the seepage meter gives a rapid and direct measurement at low cost, but the figures obtained are only point measurements.

6.2. Tracers for Recharge Estimation

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 years, respectively. Their current concentrations in the hydrologic cycle have been affected greatly by nuclear testing. Both tritium, \(^3\)H and chlorine-36, \(^{36}\)Cl from atmospheric testing have been used for soil-water tracing and recharge studies. Chlorine-36 has been used increasingly as more analytical facilities have become available. Input concentrations of the other isotopes mentioned above have also changed over time, but across a much longer time scale, due to changes in temperature and rainfall patterns. However, little is known about the temporal changes in the fallout of Cl.

Of the tracers mentioned above, tritium (\(^3\)H), deuterium (\(^2\)H), and oxygen-18 (\(^{18}\)O) most accurately simulate the movement of water because they form part of the water molecule. In most soils, chlorine-36 and nitrate (NO\(_3\)) move as the water does, but in some soils with heavier textures anion exclusion may be a problem, and the tracer may move more rapidly than the water being traced.

Most of the recently developed isotope techniques are aimed at determining the age of water, which in turn permits calculation of groundwater travel time. The recharge rate \(R\) can then be calculated by \(R = L \phi_e / \eta_a\), where \(\phi_e\) is the effective porosity, \(L\) is the...
distance along the flow path, and \( t_a \) is the travel time or age of the groundwater at the distance \( L \). There are three basic types of groundwater dating methods:

- methods that rely on input concentrations that have changed over time and whose properties are well known, such as the radioactive noble gas krypton-85, and the synthetic organic compounds chlorofluorocarbons (CFCs), used for dating young waters (less than about 40 years old);
- tracers for which input concentrations have been constant, and decreases in concentration with time occur due to radioactive decay, such as \(^{14}\)C, used for dating waters over the time scales of 200 to 20 000 years; or
- methods where the input concentrations may have changed with time but can be determined because both parent and daughter isotopes are measured, such as \(^{3}\)H/\(^{3}\)He (tritium/tritiogenic helium), which ratio is also used to date young waters (0 to about 50 years).

Tracer infiltration mechanisms will affect the interpretation of results. Although piston (or plug) flow can often explain the behavior of tracers in the field, there is convincing evidence, particularly from humid regions, that water movement along preferred pathways is the rule rather than the exception. Thus, preferential or non-piston-type flow must be dealt with in any comprehensive analysis of recharge. For example, \(^{3}\)H was found 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 the recharge since the time that peak occurred. Any bypass (preferential) flow will result in recharge being underestimated.

2. *From the shape of the tracer profile in the soil*: generally more reliable than Method 1 above because information about flow mechanisms can be obtained. In order to obtain estimates of mean annual recharge, \( \bar{R} \), a weighting function that takes into account year-to-year variations of recharge is needed.

3. *From the total amount of tracer, \( T_i \), stored in the profile*: this is given by

\[
T_i = \int_0^z T(z)\theta(z)dz
\]

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 \(^{3}\)H, mean annual recharge can be estimated by

\[
\bar{R} = T_i / \sum_{i=1}^{\tilde{z}} w_i T_{p_i} \exp(-t\lambda)
\]
where $T_{pi}$ is the tracer concentration of recharge water $i$ years before the present, $w_i$ is the annual recharge weighting factors, and $\lambda$ is the tracer decay constant. In this analysis, non-piston flow can be handled because $\bar{R}$ is independent of the distribution of the tracer in the profile.

Tracer methods have a number of attractive attributes. Their movement is governed mainly by the long-term mean soil-water fluxes that lead to recharge. (Many water-balance or soil-water-pressure-based techniques measure fluxes on a much smaller time scale than is needed for recharge estimates.) In addition, the use of tracers does not necessitate frequent visits to the field. With tracers, it is possible to estimate smaller fluxes than with other methods. Finally, they are often the only viable alternative.

The choice of tracer depends on the situation. In most cases, the tracer is used to follow water movement and hence should move with the water. It therefore needs to be mobile and soluble; it should not be strongly retarded by the soil or aquifer matrix. Ideally, the tracer should be non-reactive and should not transform during transport. Of course, the tracer needs to be easily measured and extracted from the soil. If artificial tracers are used additional constraints need to be satisfied, such as low natural levels of the tracer in the environment, low toxicity, and low radioactivity. For environmental tracers, it is desirable to have large natural variations of tracer concentrations in the landscape. These constraints usually mean that only anions (Cl, Br, $^{36}$Cl) or isotopically-labeled water molecules ($^2$H, $^{18}$O, and $^3$H) can be used.

The choice of tracer is mainly determined by the timescale of the recharge process. Use of artificial tracers requires that the bulk of the applied tracer has passed through the root zone. The timescale associated with leaching through the root zone is $Z_r \theta / R$, where $Z_r$ is the root zone depth, $\theta$ the volumetric water content, and $R$ the recharge rate. For example, in a humid climate (with $\theta = 0.1$, $Z_r = 1$ m, and $R = 100$ mm year$^{-1}$), the timescale is one year. However, in an arid climate with a recharge rate of 10 mm year$^{-1}$, the timescale is 10 years. While the former timescale is short enough for the tracer to be applied and the soil sampled in succeeding years, the latter is probably not. However, it would be suitable to use a bomb tracer (i.e. a tracer resulting from nuclear testing).

Bromide is the most widely used artificial tracer and $^3$H and $^{36}$Cl are the most suitable bomb tracers. However, $^3$H and $^{36}$Cl are both too expensive for spatial and temporal variability studies that require many samples. The use of tritium as an artificial tracer is not generally recommended because of concerns about radioactivity and difficulty in application. For such investigations, the use of chloride as an environmental tracer is generally recommended. Employment of multiple tracers can often provide corroborative information needed for correct interpretation.

Because tracers do not measure water flow directly, a number of problems can lead to over- or underestimation of recharge. These problems include secondary (unknown) tracer inputs, mixing and dual flow mechanisms; such problems only arise if the sources, sinks, and pathways of the tracer are not fully understood. Part of the recharge going through preferred pathways (such as root channels or fissures) may invalidate the results of a tracer study.
6.3. Accuracy of Recharge Estimates

Because recharge is not easy to measure directly, estimates of its value are prone to large errors. Four common types of error are discussed below. The most serious and common type of error is an error in the conceptual model. It arises when the recharge process is not fully understood, or when too many simplifying assumptions are made. For example, in a given study it may be assumed that excess irrigation water applied in parks becomes recharge, whereas in reality a low-conductivity layer causes perching and horizontal flow to a surface drain. Alternatively, a monthly time unit might be used for a soil-moisture budgeting model in a semi-arid area, resulting in zero recharge being estimated, whereas occasional short wet spells overcome soil-moisture deficits to cause some recharge.

Another common error relates to temporal and spatial variability. Most recharge processes are non-linear in relation to time. For example, a low-intensity rainfall might cause no recharge because of a high rate of evapotranspiration, whereas the same amount over a shorter time period might be sufficient to saturate the soil and cause recharge. Thus, errors will arise if temporal variations are ignored: for example by using monthly, annual, or long-term average data. Recharge is also non-linear with respect to spatial variations of inputs and physical properties of soils and aquifers.

Measurement error is also a significant issue, and is connected with the equipment used to make measurements. This kind of error is generally not overlooked. The final type of error—calculation errors—can be avoided by care and checking, especially of units. A particularly difficult type of calculation error can occur with numerical computer models unless they are rigorously tested under a wide range of conditions.

Error analysis, or sensitivity analysis, can show which variables in an equation lead to the highest errors, and special effort can be concentrated on obtaining the most accurate estimates for these. However, this approach will not help if the conceptual model is wrong. More than one method of estimation using other data should be used to provide an independent check. Table 1 summarizes six different methods for estimating natural groundwater recharge from precipitation. These methods were tested and compared in a sandy till area in southeastern Sweden. As mentioned earlier, the desired resolution in time is an important criterion in method selection. The interest may vary from estimation of instantaneous recharge to long-time averages. Table 2 classifies the methods indicated in Table 1 according to the time and areal resolution. Clearly there is a need for comparative studies, in which several methods are applied to minimize the uncertainty in estimations of groundwater recharge.
<table>
<thead>
<tr>
<th>Method/model</th>
<th>Category</th>
<th>Required input data</th>
<th>Calibration</th>
</tr>
</thead>
<tbody>
<tr>
<td>One-dimensional soil water flow model (SOIL)</td>
<td>inflow</td>
<td>precipitation, temperature, wind speed, relative humidity, soil water retention properties, hydraulic conductivity, groundwater outflow</td>
<td>measured ground-water levels</td>
</tr>
<tr>
<td>Soil moisture budget model</td>
<td>inflow</td>
<td>precipitation, temperature, wind speed, relative humidity, size of soil moisture reservoir, soil moisture-recharge relation</td>
<td>soil water flow model</td>
</tr>
<tr>
<td>Groundwater level fluctuations</td>
<td>aquifer response</td>
<td>groundwater levels, specific yield</td>
<td></td>
</tr>
<tr>
<td>Chloride concentration</td>
<td>aquifer response</td>
<td>precipitation, wet and dry deposition of chloride, concentration of chloride in groundwater</td>
<td></td>
</tr>
<tr>
<td>Spring discharge</td>
<td>outflow</td>
<td>spring discharge, size of catchment area</td>
<td></td>
</tr>
<tr>
<td>------------------</td>
<td>---------</td>
<td>------------------------------------------</td>
<td></td>
</tr>
<tr>
<td>Catchment area model (PULSE)</td>
<td>outflow</td>
<td>precipitation, temperature, potential evapotranspiration</td>
<td>size of soil moisture reservoir, soil moisture-recharge relation, outflow from the groundwater reservoir</td>
</tr>
</tbody>
</table>

Table 1. Comparison of six different methods for estimation for groundwater recharge tested in southeastern Sweden (from Johansson, 1988)
<table>
<thead>
<tr>
<th>Timescale Method/Model</th>
<th>Instantaneous</th>
<th>Events</th>
<th>Monthly</th>
<th>Seasonal</th>
<th>Annual</th>
<th>Long-time average</th>
</tr>
</thead>
<tbody>
<tr>
<td>One dimensional soil water flow model (SOIL)</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Soil moisture budget model</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Groundwater level fluctuations</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Chloride concentration</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Spring Discharge</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Catchment area model</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
Table 2. Classification of the applied methods for estimating groundwater recharge, according to the resolution in time of their results. A dashed line indicates point values of groundwater recharge and a solid line indicates an areally integrated value (from Johansson, 1988)

7. Factors Influencing Recharge, Predictive Relationships, and Recharge Regionalization

7.1. Factors Influencing Recharge

The key environmental factors controlling recharge are climate, soils and geology, vegetation and land use, topography, and depth to water table. The water-balance equation is commonly used to quantify the components of the hydrologic cycle

\[ P + I = RO + D + ET + S \]  

where \( P \) is precipitation, \( I \) is irrigation, \( RO \) is surface runoff, \( D \) is deep drainage and recharge, \( ET \) is evapotranspiration, and \( S \) is water stored in the soil. Under non-irrigated conditions, where \( I = 0 \), the left-hand side of Eq. (5) is fixed in the sense that it is outside human control. Hence, a decrease in any of the variables on the right-hand side forces an equal increase in the other terms to maintain the equality (i.e. the water balance). For example, a decrease in surface runoff \( RO \) (for example, as a result of increased infiltration through better tillage practices) may increase the amount stored in the soil profile \( S \); an increase in \( S \) would tend to increase deep drainage (and recharge) \( D \), and evapotranspiration \( ET \). Clearly, a basic understanding of this water cycling is needed in order to understand and estimate aquifer recharge.

Most direct measurements of hydrologic variables related to recharge assessments provide only point measurements or estimates and do not integrate such variables (shown in Eq. (5)) in relation to space and time. Recharge varies across the landscape because the aforementioned controlling factors vary, but finding ways to estimate and predict this spatial and temporal variability, and to regionalize point measurements, remains a major problem in recharge assessments.

A daily water balance modeling analysis (based on Eq. (5)) for the Rattlesnake Creek basin in south-central Kansas, an approximately 3500 km\(^2\) semiarid to subhumid agricultural basin, demonstrated that soil factors, plant cover, and land use practice are important controls on groundwater recharge. The importance of each of these factors is detailed below. Although such results are highlighted for the case study of this particular basin, they are general enough to be valid for any agricultural plain region of similar climate in the world.

Soil factors, such as the available water capacity (AWC) of soil profiles, exert a dominating influence. AWC is the volume of water available to plants if the soil were at field capacity: in other words, the moisture content held by a soil against the pull of
gravity after the excess water has drained out of a saturated or nearly-saturated soil. The AWC of each soil determines the maximum limit of actual evapotranspiration (ET) that can be extracted without additional infiltration and the maximum soil moisture deficit possible. (Soil-moisture deficit is an estimate of the degree to which soil-moisture content has dropped below field capacity.) Thus, given the same hydroclimatic conditions and crop cover, a soil with a relatively low AWC will exhibit a relatively small soil-water deficit, and smaller amounts of water will be lost through ET compared to losses from a soil with higher AWC (Figures 5 and 6). The AWC also determines the amount of water that can infiltrate into the soil before deep drainage occurs, since it acts as a buffer for infiltrating water. Given the same initial-moisture conditions, a soil with higher AWC can absorb more infiltrating water than low-AWC soils. Thus, deep drainage decreases with increasing AWC (Figure 7).

Figure 5. Soil-moisture deficit versus available water capacity for grassland, dryland, and irrigated cropland for the upper two-thirds of the Rattlesnake Creek basin in south-central Kansas (from Sophocleous and McAllister, 1987)
Figure 6. Actual evapotranspiration versus available-water capacity for grassland, dryland, and irrigated cropland for the upper two-thirds of the Rattlesnake Creek basin in south-central Kansas (from Sophocleous and McAllister, 1987)

The water balance is also greatly influenced by plant cover and land-use practices. The largest element of the water balance in Eq. (5) in the south-central Kansas basin under study is the ET component, as can be seen for native grassland in Figure 8. The impact of vegetation on the hydrologic balance is complex and depends on factors such as crop coefficients (i.e. empirically determined coefficients relating potential ET to crop ET), growth stages, rooting depths, soil, water, and climatic conditions as used in the soil-water balance simulation model.

The crop coefficients vary with the stage of crop growth. Mature plants have greater ability to extract soil-moisture from all soil horizons, and thus have larger crop coefficients than young plants. The crop with the largest crop coefficients employed in the soil-water balance model in south-central Kansas is alfalfa. In addition, alfalfa is continuously grown from one year to the next with multiple harvests without replanting or land fallowing. Prairie grasses have the next highest overall crop coefficients in the study region, with a long growing season. All other crops have lower crop coefficients and are grown only part of the year.
The greatest deep drainage occurred in irrigated wheat fields, mainly because of the shallow rooting depth of wheat, while the lowest values occurred in alfalfa and grassland acreages (Figure 9). Decreased amounts of deep drainage in the northeastern portion of the Kansas study basin (Figure 8), where precipitation is lower, are recorded from grasslands, indicating the dominant effect that precipitation and vegetation exert on deep drainage.
Figure 8. Grassland water-balance components for (b) the lower one-third and (a) the rest of the Rattlesnake Creek basin in south-central Kansas (from Sophocleous and McAllister, 1987)
In general, the effect of irrigation is to increase both evapotranspiration and deep drainage significantly, as can be seen in Figures 6 and 7, respectively. For summer crops such as sorghum and soybeans, as well as alfalfa, most of the irrigation amounts are spent in evapotranspiration activities, with negligible amounts for deep drainage. Grasslands feature reduced deep drainage and runoff, and increased soil-moisture deficits compared to cropland acreages (Figures 5, 6, and 7). From the areal distribution of the various components of the water balance, it was concluded that single average values of hydrologic variables used in management practices are not realistic, and that a spatial discrimination approach to managing water resources is needed.

A computerized water-balance procedure such as the one used in the Rattlesnake Creek basin study can be used to demonstrate and predict human and natural impacts on the hydrologic cycle. The hydrologic effects of vegetation changes, weather modification, extreme weather conditions, and so on can be readily estimated during the planning process using the methodology of combining classification techniques (to identify hydrologically “homogeneous” unit areas within the heterogeneous basin) and water balance modeling employed in the Rattlesnake Creek basin study. Thus, had the basin been entirely covered by prairie grasses (as it probably was during predevelopment time) and had the 1982/1983 precipitation regime prevailed (which is about 10% below average), the overall basin deep drainage is model-predicted to have been 29 mm year\(^{-1}\), compared to less than 4 mm year\(^{-1}\) if alfalfa were planted exclusively in the basin. If the entire basin were planted with dryland wheat under 1982/1983 precipitation conditions, the overall basin deep drainage is model-predicted to have been 130 mm year\(^{-1}\). Such figures can be arrived at by multiplying the deep-drainage amounts for the
corresponding crop and soil complex by the planted area, summing these figures, and then dividing by the area of interest. Similarly, the hydrologic effects of manipulating the proportion of various crops and the amounts of irrigation within any soil-association area can thus be assessed.

Provided that future precipitation patterns can be established, then, under known vegetation and land-use practices, various components of the water balance, such as deep drainage and surface runoff within the basin, can be predicted using the presented methodology. An example of the relative effects of an approximately 19% precipitation difference on the components of the water balance, keeping the precipitation time-pattern constant, is shown in Figure 8. This figure represents actual grassland data from the northeastern portion of the Kansas basin study area, which received 480 mm of annual precipitation (Figure 8b), and the rest of the basin, which received an annual precipitation average of 592 mm (Figure 8a). Note the large increase in deep drainage in the higher precipitation region, especially in low-AWC soils, compared to the deep drainage in the lower precipitation region.

7.2. Predictive Relations and Recharge Regionalization

Although there have been numerous studies to estimate recharge in specific areas, there has been no systematic attempt to develop generic, predictive relationships for quantifying recharge based on the aforementioned controlling environmental factors. This is important for groundwater management and protection. Recharge studies in the agricultural plains of central Kansas resulted in the development of such an approach based on classification and statistical analyses, and taking advantage of Geographic Information Systems (GIS) capabilities for “mapping” recharge and its controlling factors. Because of the generality and applicability of the methodology to most semiarid to humid plain regions of the world with relatively shallow water tables, the Kansas case study will be outlined below.

Although geostatistical methodologies and multivariate statistical techniques such as cluster analyses are useful tools for regionalizing point measurements, the usually small number of experimental sites for recharge estimation precludes usage of such techniques. To develop practical relationships between annual recharge and easily measured, independent recharge-controlling factors for the south-central Kansas plains (Great Bend Prairie region), advantage was taken of recently-completed multi-year (1985–1992) field-based recharge assessment studies at ten sites in that region. As a result, a number of multiple regression analysis models were developed depending on the number of controlling factors considered. Most of the ten recharge-assessment field sites were located in grassland and adjacent to irrigated cropland fields. This analysis showed that, given the vegetation cover considered, the most influential variables in recharge estimation were, in order of decreasing importance, annual precipitation (PCP: major climatic variable), average maximum soil-profile water storage (AWC: major soil variable), average shallowest depth to water table (DTW: major groundwater condition variable), and average springtime precipitation rate (RATE: secondary climatic variable).
Each of these factors then was used to zone the region for recharge estimation, and was mapped as a separate GIS layer or coverage. Thus, four GIS (ARC-INFO) data layers were constructed for the region, based on the results of the multiple regression analysis as shown in Figure 10. Each data layer was classified into the same number of data classes (six in all) and assigned a class rank. Then the overlays were combined to produce a master map of “homogeneous” zones. GIS technology is ideally suited to such overlay analysis.

Figure 10. Four recharge-related GIS coverages. Solid circles indicate recharge-assessment sites. Crosses indicate climatic stations (adapted from Sophocleous, 1992).

An ARC-INFO overlay analysis procedure was conducted to identify areas of differing recharge in the study region. The regression coefficients of the developed multiple regression models, normalized to 1, were used to weigh the class rankings of each recharge-affecting variable. Based on this classification scheme, an area-wide recharge map (Figure 11) indicating five differing recharge regions was derived. The recharge zonation agreed well with the field-estimated recharge values at the sites.

In addition, the fact that the GIS-based regression estimates for recharge were of the same magnitude as other independent estimates, even during an extreme flooding period in the summer of 1993, attests to the robustness of the methodology, although additional tests are desirable. It was concluded that the combination of multiple regression and GIS
overlay analyses is a powerful, robust (even under extreme conditions), and practical approach to regionalizing small samples of recharge estimates.

Figure 11. GIS coverage showing recharge zonations. Solid circles indicate recharge-assessment sites (adapted from Sophocleous, 1992).

8. Difficulties and Challenges in Recharge Estimation

Quantification of the rate of natural and human-induced groundwater recharge is a basic prerequisite for efficient groundwater resource management, and is particularly vital in arid and semiarid regions where such resources are often the key to economic development. However, the rate of aquifer replenishment is one of the most difficult factors to measure in the evaluation of groundwater resources.

Although there are various well-established methods for the quantitative estimation of recharge, few can be applied successfully in the field. A 1988 international recharge-estimation workshop concluded that “no single comprehensive estimation technique can yet be identified from the spectrum of methods available; all are reported to give suspect results.”

Difficulties in reliably quantifying groundwater recharge stem from a variety of factors. These include the limited capability to identify and quantify the probable recharge mechanisms and important features influencing recharge for a given locality, the nonlinear recharge response with time, the highly variable areal distribution of groundwater recharge, the scarcity of hydrogeological data, and the complexities of the hydrologic balance in general.
Because of these uncertainties, project designs and management strategies need to be flexible enough not to require radical change if initial predictions prove wrong, due to incorrect assumptions about recharge rates or other hydrogeological factors. Groundwater recharge estimation must be treated as an iterative process that allows progressive collection of aquifer-response data and resource evaluation. In addition, more than one technique needs to be used so that results can be verified.

When estimating groundwater recharge, one must start with a good conceptualization of different recharge mechanisms and their importance in the study area. To identify the probable flow mechanisms, the field evidence must be examined carefully. The recharge mechanism at a particular location can depend on a variety of factors, which may be different from the influencing factors elsewhere. Therefore, just because a method has successfully estimated the recharge in one locality one should not assume the same method could be used in another, even if the situation appears to be similar. Once the recharge mechanisms have been defined, calculations can be carried out to estimate the recharge. In addition to being based on a good conceptualization, the choice of methods should be guided by the objectives of the study, available data and resources, and possibilities of obtaining supplementary data.

Consideration of the following questions may facilitate recharge estimation:

- How much recharge can the aquifer accept? A full aquifer will reject further water, which must then find another destination.
- How much water can the unsaturated zone transmit? High potential recharge rates (for example, from rivers or irrigation canals) may not be able to pass through low conductivity layers.
- What other destinations are there for potential recharge, and how large are they?
- How much potential recharge is there?
- What is the actual recharge? This step considers the balance and destinations of all water from the source, based upon the first four questions.
- How do other estimates compare? As previously mentioned, more than one method should be used whenever possible.

The key to deciding on a recharge estimation methodology is the spatial and temporal scale of interest. If the major concern is obtaining good recharge estimates over a limited area (for example, for waste disposal or local water supply purposes), then the need for detailed information is evident. In this situation, multiple site investigations, which also require identification of preferential flow contributions, are needed. Conversely, for projects on a regional scale, or those requiring only preliminary recharge estimates, groundwater-based methods (such as those involving interpretation of fluctuations in groundwater levels) are relevant and small-scale variability in local recharge ceases to be a problem.

The inherent temporal variability of recharge has important implications for the measurement techniques adopted. Different measurement techniques provide recharge estimates with different temporal scales. For example, applied tracers and lysimeters are only able to provide information on recharge over the period of measurement: usually no more than a few years. Meteorological water-balance techniques, and those involving
interpretation of fluctuations in groundwater levels, likewise can only provide information on recharge over the period of record. Chloride displacement techniques provide a mean recharge rate since a change in land use, while bomb tracers indicate a mean recharge rate since peak fallout (a period of about forty years). Chloride mass balance methods have a much longer temporal scale, typically in the order of hundreds to thousands of years (dependent on the recharge rate and the thickness of the unsaturated zone). The techniques adopted will depend upon the purpose of the study. Where interest is in estimating long-term recharge rates, a long temporal scale of measurement is desirable. On the other hand, if interest is on the effect of land management on recharge, those techniques with smaller temporal scales are required.

In areas where the annual variability of recharge is very high, measurement techniques with long timescales will be required to estimate the long-term mean annual recharge rates with any accuracy. Where the annual variability of recharge is lower, measurement techniques with shorter timescales will be suitable.

The temporal variability of the soil-water flux should decrease with depth. If the recharge rate is sufficiently low, and the water table sufficiently deep, then below a certain depth the temporal variability of drainage will approach zero. At these depths, even the measurement of the soil-water flux over a short time scale should be sufficient to infer the long-term drainage rate. This decrease in the temporal variability of soil-water flux with depth may cause problems in estimating recharge from hydrograph records. If much of the temporal variability is lost during passage through the unsaturated zone, then these methods may underestimate the recharge rate. For example, the coefficient of variation (CV) of annual drainage below the root zone of Banksia woodland (10 m deep) was found to be 64%, whereas the CV for drainage at 20 m (water table recharge level) was 11%.

Thus, groundwater-based methods may be inappropriate for estimating rates of groundwater recharge in areas where deep drainage fluxes are low and water tables are deep. In particular, methods which involve interpretation of hydrograph records will underestimate recharge. Individual recharge events will not usually be seen as rises in water tables if the time lag is greater than a few days. Seasonal variations in drainage will likewise not be reflected in water table variations if the time lag is much greater than a few months.

Tracer methods seem the most reliable for point recharge measurements in arid areas. The best method will depend on the magnitude of recharge flux. Chloride appears most reliable over drainage rates from less than 1 mm to 100 mm year\(^{-1}\). At deep drainage rates of more than 100 mm year\(^{-1}\), measurement errors and anion exclusion may become important. Bomb tracers \(^{3}H\) and \(^{36}Cl\) are suitable for recharge rates greater than 20 mm year\(^{-1}\). The accuracy of drainage estimates obtained with natural tracers should generally not be assumed to be better than ±50%.

Temporal variability of recharge is related to temporal variability of precipitation. The variability of annual recharge increases rapidly as the mean annual recharge decreases. For mean annual recharge of 30 mm year\(^{-1}\), measurements over a span of at least 15–20 years have been suggested.
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 one order of magnitude or more, especially when measuring physical parameters in the drier ranges. An advantage of tracers is that they integrate all 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 (for example, 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 rates. Combining Cl (tracer) and suction profile (physical) information, at sites where changing land use has significantly altered recharge rates is useful in quantifying both past and present recharge rates at such locations.

In estimating groundwater recharge in arid regions, indirect, physical approaches, such as water balance and Darcy flux measurements, were found the least successful, whereas methods using tracers (such as Cl, $^3$H, and $^{36}$Cl) have been found to be the most successful. Of the tracer techniques available, Cl-balance techniques appear to be the simplest, least expensive, and most widely used for recharge estimation.

Nevertheless, advances in recent years show that the value of water-balance and Darcian methods should not be underestimated. Reliability of water-balance methods for recharge estimation depends on the precision with which the water balance components have been determined. In arid and semiarid regions, application of this method is more difficult than in humid regions because precipitation is frequently only slightly different from actual evapotranspiration; small errors in these two components thus cause large errors in recharge estimates. To minimize such errors, one can use a combination methodology such as the hybrid water fluctuation methodology, which uses a storm-by-storm water balance analysis in combination with analyses of vadose zone moisture and water table fluctuations. However, the hybrid water fluctuation methodology is applicable predominantly to relatively flat, semiarid-to-humid regions with relatively shallow water table.

Despite their importance for groundwater management and protection, no generic predictive relationships for quantifying recharge based on the major controlling factors of climate, soils, vegetation, and land use have been usefully developed. In addition the problem of regionalizing point measurements, given the spatial and temporal variability of recharge and aquifer heterogeneity, remains a serious one. Although a methodology to address such problems has been developed (Section 7.2), additional studies and approaches are needed to tackle these challenges.
Glossary

Arid: Said of a climate characterized by dryness, variously defined as rainfall insufficient for plant life or less than 250 mm of annual rainfall.

Baseflow: Stream flow derived mainly from groundwater seepage into the stream.

Bomb tracer: A tracer resulting from nuclear testing, such as tritium or chlorine-36.

Capillary flow: The flow that takes place in pores with a diameter less than approximately 3 mm in which capillary forces, together with gravity, determine the flow process.

Catchment: See watershed.

Conceptual model: An interpretation or working description of the characteristics and dynamics of a physical system.

Confining bed: A geological unit of significantly lower hydraulic conductivity than an aquifer stratigraphically adjacent to one or more aquifers.

Deep drainage: Drainage of water below the root zone.

Diffuse recharge: Water added to the water table by vertical percolation of precipitation through the unsaturated zone. Also known as direct recharge.

Discharge area: An area in which water is lost naturally from the saturated zone.

Fallow: The period during which land is left to recover its productivity after cropping, mainly through accumulation of water and nutrients, attrition of pathogens, or a combination of these factors.

Fingered flow: Unstable flow whereby the percolating water may concentrate at certain points to break into the sublayer in the form of finger-like or tongue-like protrusions.

Flow path: The route groundwater takes to a distant point.

Flow net: The set of intersecting lines of equal hydraulic head values and flow lines representing two-dimensional steady flow through a porous medium.


Groundwater flow system: A set of groundwater flow paths with common recharge and discharge areas. Flow systems are dependent on both the hydrogeologic characteristics of the soil/rock material and landscape position. Areas of steep or undulating relief tend to have dominant local flow systems (discharging in nearby topographic lows such as ponds or streams). Areas of gently sloping or nearly flat relief tend to have dominant regional flow systems (discharging at much greater distances than local systems in major basin topographic lows or oceans).

Hydrogeologic environment: The physical and chemical conditions resulting from the combination of topography, geology, and climate.
**Hydrologic budget or water balance:** An accounting of the inflow to, outflow from, and storage in a hydrologic unit such as a drainage basin, aquifer, soil zone, lake, or reservoir.

**Indirect recharge:** Recharge that results from percolation to the water table following runoff and localization in joints, as ponding in low-lying areas and lakes, or through the beds of surface water courses.

**Induced recharge:** Recharge to groundwater by infiltration, either natural or anthropogenic, from a body of surface water as a result of the lowering of the groundwater level below the surface-water level.

**Interflow:** Subsurface lateral flow that can enter streams quickly enough to contribute to the rising streamflow hydrograph response to a storm.

**Localized recharge:** Recharge that results from horizontal surface concentration of water in the absence of well-defined channels, such as sloughs, potholes, and playas. Also called focused or depression-focused recharge.

**Macropore flow:** The flow that takes place in a wide range of large pores such as cracks in clay soils, rock fractures, fissures in sediments, worm holes, and old root channels. Preferential and by-pass flow are alternative names for macropore flow.

**Matric suction:** Soil-water potential (energy) resulting from the capillary and absorptive forces due to the soil matrix.

**Mountain front recharge:** Recharge that involves complex processes of unsaturated and saturated flow in fractured rocks, as well as infiltration along channels flowing across alluvial fans.

**Natural recharge:** Naturally occurring water added to an aquifer. Natural recharge generally results from snowmelt and precipitation or storm runoff.

**Perched groundwater; perching:** A superficial body of groundwater separated from an underlying main body of groundwater by an unsaturated zone due to a sufficiently low hydraulic conductivity layer that supports this body of perched groundwater; the act of causing a body of groundwater to form above a low-permeability layer in an unsaturated zone.

**Piston flow or plug flow:** Purely advective flow without dispersion or diffusion of the dissolved components.

**Potential energy:** The energy deriving from elevation and/or pressure.

**Potential recharge:** Soil-water that percolates below the root zone and has the potential of reaching the aquifer, whereas *actual recharge* is soil-water that actually reaches the aquifer.

**Preferential recharge:** Recharge that takes place preferentially through macropores, as opposed to *diffuse recharge*, which takes place through the entire vadose porous medium.

**Recharge area:** The area that contributes water to an aquifer. Normally considered to be the natural area of recharge, as contrasted with a constructed recharge basin.

**Rejected recharge:** Potential recharge that exceeds the rate of flow through an
Residence time: The length of time between the input of water as infiltration or recharge and its output as runoff or discharge. Also known as transit time or turnover time.

Residual: In the case of recharge, the remainder of all other hydrologic components in the water balance equation.

Semiarid: Said of a type of climate in which there is slightly more precipitation (250 to 500 mm) than in an arid climate, and in which sparse grasses are the characteristic vegetation.

Sensitivity (analysis): In model application, the degree to which the model result is affected by changes in a selected model input representing hydrogeologic framework, hydraulic properties, and boundary conditions.

Transmission losses: Streamflow losses through seepage in ephemeral streams.

Unconfined (or water table) aquifer: An aquifer in which the water table is at the upper boundary of the groundwater flow system that is at atmospheric pressure.

Unsaturated or vadose zone: The unsaturated (i.e. not completely filled with water) zone lying between the Earth’s surface and the top of the groundwater.

Water table: The upper boundary of an unconfined aquifer at atmospheric pressure.

Watershed: That surface area which drains to a specified point on a watercourse, usually a confluence of streams or rivers.

Bibliography

Allison G.B., Gee G.W. and Tyler S.W. (1994). Vadose zone techniques for estimating groundwater recharge in arid and semiarid regions. Soil Science Society of America Journal 58, 6–14. [This presents a useful discussion of recharge estimation methods, with special emphasis on tracer techniques. Much of the material in section 6.2 was taken from this reference.]

Cook P.G. (1993). The Spatial and Temporal Variability of Groundwater Recharge: A Case Study in Semi-Arid Areas of the Western Murray Basin. Doctoral thesis, School of Earth Sciences, Flinders University of South Australia, Adelaide. [This dissertation presents, among other things, a discussion on how the temporal and spatial variability of recharge impacts upon the estimation methodology. Much of the material in section 8 was taken from this reference.]


Sophocleous M.A. (1992). Groundwater recharge estimation and regionalization: the Great Bend Prairie of central Kansas and its recharge statistics. *Journal of Hydrology* 137, 113–140. [This paper presents a recharge regionalization methodology based on field data analysis, statistical analysis, and GIS-based overlay analysis. Much of the material in section 7.2 was taken from this reference.]

Sophocleous M.A. and McAllister J.A. (1987). Basinwide water-balance modeling with emphasis on spatial distribution of groundwater recharge. *Water Resources Bulletin* 23, 997–1010. [This paper employs a simple soil-water agro-hydrologic balance model and the hydrologic response unit concept to analyze the impact of soil, vegetation, and land-use factors on recharge over an entire hydrologic basin. Much of the material in section 7.1 was taken from this reference.]

Sophocleous M.A. and Perkins S.P. (2000). Methodology and application of combined watershed and groundwater models in Kansas. *Journal of Hydrology* 236, 185–201. [This paper comprehensively outlines the methodology for integrating watershed and groundwater models for recharge estimation and other purposes, and presents three real-world applications of integrated modeling.]


Tóth J. (1999). Groundwater as a geologic agent: An overview of the causes, processes, and manifestations. *Hydrogeology Journal* 7, 1–14. [This is the most recent exposition of Tóth’s classic work on regional flow systems.]

**Biographical sketch**

**Marios Sophocleous** received his B.Sc. in Natural Sciences and Geology from the School of Physics and Mathematics of the University of Athens, Greece, in 1971, his M.Sc. in Water Resources from the Civil Engineering Department of the University of Kansas in 1973, and his Ph.D. in Geology, with specialization in Hydrogeology, from the University of Alberta, Canada, in 1978. Since 1978 he has been employed at the Kansas Geological Survey, where he became Senior Scientist in 1987. He is editor of the *Journal of Hydrology*, Associate Editor of the *Hydrogeology Journal*, member of the editorial boards of *Computers and Geosciences, Natural Resources Research*, and *Current Research in Earth Sciences* (Kansas Geological Survey’s peer-reviewed electronic journal), and Adjunct Professor of Geology at The University of Kansas. He was the 1997 recipient of the Best Practice Paper Award by the Irrigation and Drainage Engineering Division of the American Society of Civil Engineers. His areas of research include experimental investigations and numerical modeling of soil-water and groundwater flow and pollutant transport, aquifer-recharge processes, stream–aquifer interactions, regional groundwater flow and watershed hydrology, soil–vegetation–atmosphere hydrologic interactions, integrated watershed/groundwater modeling, and water-resources evaluation and management.
To cite this chapter