Natural Recharge of Groundwater in Illinois

Bruce Hensel

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CONTENTS

FIGURES

TABLE

ABSTRACT

Recharge and infiltration are related hydrological processes. Infiltration will occur during any event when water is made available to an unsaturated ground surface. However, much of the water that infiltrates may not reach the water table as recharge. Some may be used to replenish soil-moisture deficits, and some may be returned to the atmosphere by evapotranspiration.

Groundwater recharge does not occur during every infiltration event; it will most likely occur when the soil is wet because less moisture is used to replenish moisture deficits. In Illinois, recharge most commonly occurs in late winter and early spring, when precipitation is abundant, evapotranspiration is low, and soil moisture is high. Recharge also frequently occurs in the late fall after plant transpiration has ceased. It occurs less frequently during the summer because evapotranspiration commonly uses more soil water than precipitation supplies.

Upland terranes in Illinois are typically areas of groundwater recharge. Hydraulic head in upland areas is generally high and hydraulic gradients are usually downward. The rate of recharge will be greatest in areas with surficial deposits of coarse sand and gravel, highly fractured bedrock, or karst topography. Groundwater discharge typically occurs where hydraulic head is low and hydraulic gradients are upwards, including lowland terranes characterized by marshes, lakes, and rivers. Recharge may occur in some lowlands, usually where there is surficial sand and gravel, especially if a water supply well is finished in the formation.

There are numerous methods of estimating recharge. Direct measurements of point recharge may be made with lysimeters. Indirect estimates may be made by calculating the Darcy flux in the saturated or unsaturated zone or by observing fluctuations of the water table in an observation well. Indirect measurements of recharge over a basin may be made by measurement of the water-table rise over a long time, calculation of a water budget, or by separation of base flow by stream hydrograph. Numerical models and analysis of groundwater chemistry can also be used to estimate recharge at point and regional scales.

INTRODUCTION

Recharge and infiltration are two separate but related processes that affect water within the groundwater portion of the hydrologic cycle. Both processes are important for quantifying the availability and the contamination potential of groundwater. The purpose of this report is to describe the factors that affect recharge and infiltration, list some methods that may be used to measure these two components of the hydrologic cycle, and describe the general recharge potential for several terranes commonly found in Illinois. This report takes a textbook approach, beginning with definitions and progressing into advanced concepts. It is primarily aimed at environmental professionals, engineers, regulators, and others who want to become more familiar with the factors that influence natural recharge of groundwater in Illinois.

Hydrologic Cycle

Within the relatively limited range of temperature and pressure conditions that occur near Earth's surface, water changes state from solid to liquid to gas and back. Because of its capacity to change state, water can move from the atmosphere to land and back again. This process, known as the hydrologic cycle, is vital to our planet's climate and water distribution system.

Figure 1 illustrates the basic components of the hydrologic cycle. Water is input to the atmosphere in a gaseous state through *evaporation* and *transpiration*. It returns to the surface as *precipitation*. Water at the land surface may flow to surface water bodies (lakes, streams, and rivers) as *runoff*, or it may *infiltrate* into the unsaturated zone. The *unsaturated zone* is the layer of soil and rock where pore spaces are partially filled with water and partially filled with air. Water that infiltrates may be returned directly to the atmosphere through evaporation and transpiration, or it may flow to a nearby surface water body as subsurface storm flow. It may also seep downward to the *saturated zone*, where all pores are filled with water, as *groundwater recharge*. Groundwater may complete the next step of the hydrologic cycle as transpiration or as *discharge* to a surface water body.

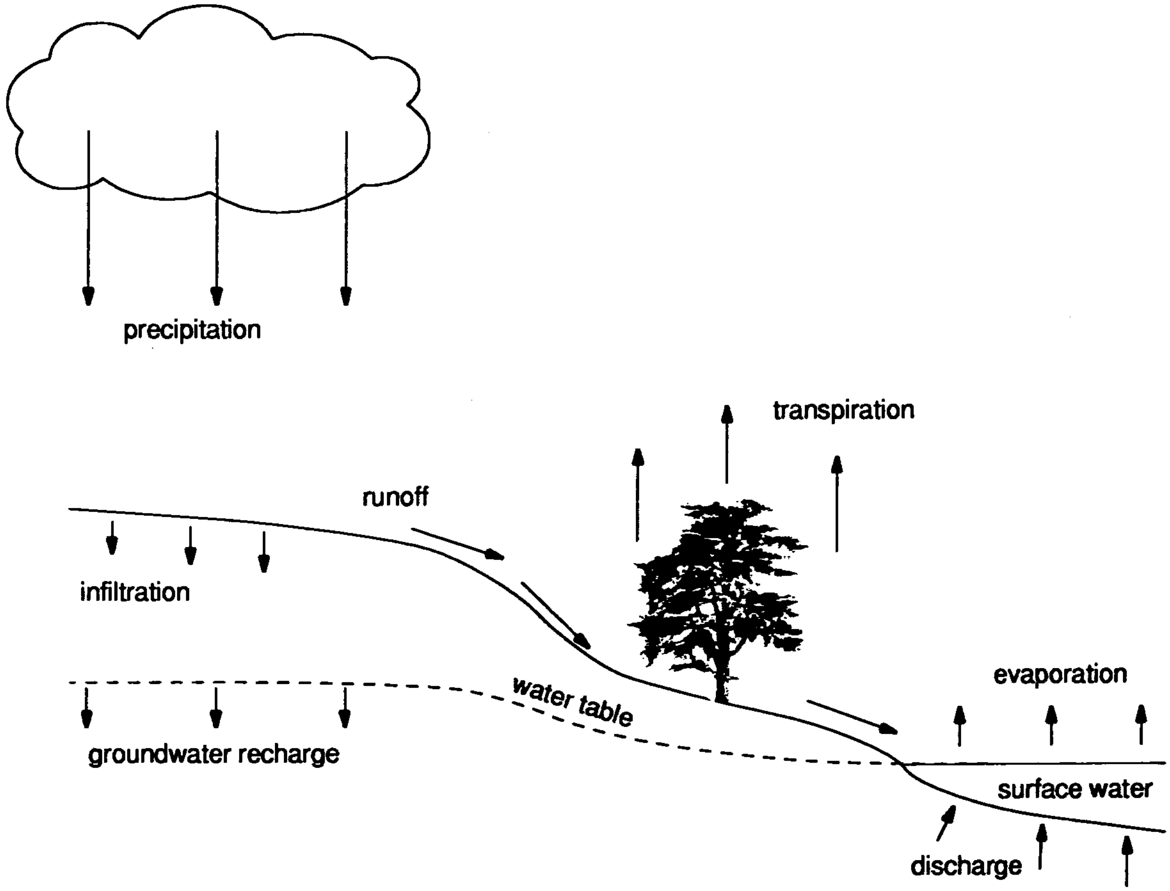


Figure 1 The hydrologic cycle.

Aquifers and the Water Table

Any definition of an aquifer is subjective because an *aquifer* is defined as any unit of soil or rock capable of supplying adequate water to meet the demands of a given user. Thus, a unit that yields an adequate water supply for an individual household may be not be considered an aquifer by a municipality that would need to supply water to hundreds or thousands of users. A unit not considered to be an aquifer is an *aquitard* or *confining zone*. Generally, aquifers in Illinois are composed of continuous, coarse-grained sands and gravels, porous sandstone, or highly fractured or weathered limestone and dolomite. Aquitards are usually composed of continuous layers of fine-grained silty to clayey till and lacustrine deposits, shale, and massive to slightly fractured limestone and dolomite.

The *water table* is the surface that divides the saturated from the unsaturated zone; as such, it has no thickness. In many lowland situations, the water table is shallow or at the ground surface; in the latter case, there would be no unsaturated zone. Beneath uplands, the water table may be very deep. In Illinois, this may be as much as 30 meters (100 ft) deep, although depths of 2 to 4 meters (6 to 13 feet) are more typical. The water table does not necessarily occur within an aquifer. Throughout the central plains portion of Illinois, the water table is commonly at a depth of 1 to 4 meters, within surficial till confining units that cover much of this area.

Infiltration and Recharge

Definitions As illustrated in figure 1, water must infiltrate the ground before it can recharge groundwater. *Infiltration* may be defined as the entry and associated downward flow of water from the ground surface into the unsaturated zone (Freeze and Cherry 1979).

Groundwater *recharge* is defined as the advance of water into the saturated zone that is associated with groundwater flow away from the water table and within the saturated zone (Freeze and

Cherry 1979). Therefore, recharge generally does not occur unless the hydraulic gradient has a downward component. This definition identifies groundwater recharge as occurring when infiltrating water passes the water table, regardless of the characteristics of geologic and aquifer materials.

Recharge to an aquifer occurs when water enters a hydrostratigraphic unit defined as an aquifer, rather than when water passes the water table. Groundwater recharge and recharge to an aquifer are the same only when the water table occurs within the aquifer. Recharge to a confined aquifer is also termed *leakage.*

In this text, the term recharge will only be used to refer to groundwater recharge. Water that enters an aquifer will be referred to as recharge to an aquifer or leakage.

Relationship Not all water made available at the ground surface infiltrates the soil and not all water that infiltrates soil will recharge groundwater. The source of water for infiltration may be rain, melting snow or ice, or irrigation water. When this water comes in contact with unsaturated soil or rock, some part of it may infiltrate that material. Water that infiltrates may take several courses: it may be drawn back to the surface by evaporation or through plant roots as transpiration; it may be redistributed through the unsaturated soil column and replenish soil-moisture deficits without recharging groundwater; it may flow to a surface water body as subsurface storm flow; or it may seep to the saturated zone and recharge groundwater.

Infiltration will occur when water contacts an unsaturated, porous ground surface, but only water that is not returned to the atmosphere as evapotranspiration and is not needed to replenish soil-moisture deficits is available for recharge. The timing, volume, and rate of groundwater recharge are functions of climatic patterns and flow processes in the saturated and unsaturated zones (Rehm et al. 1982).

INFILTRATION

Infiltration Rate and Capacity

The *infiltration rate* is the volume flux of water that flows into a soil or rock per unit of ground surface area (Hillel 1982). The *infiltration capacity* is the maximum possible infiltration rate in soil and rock for a given soil-moisture condition; infiltration capacity is reached when the rate of water available at the ground surface exceeds the ability of the material to adsorb water (Horton 1940). Sandy soils typically have higher infiltration capacities than clayey soils. Infiltration capacity is not constant (fig. 2). It decreases as the moisture content of the soil increases (wetting) and increases as the moisture content decreases (drying). When a soil/rock is saturated, its infiltration capacity is a function of its capacity to transmit water when saturated (Freeze and Cherry 1979). Water applied to the surface at a rate exceeding the infiltration capacity may accumulate (pond) at the surface or it may runoff to temporary ponds or permanent bodies of surface water.

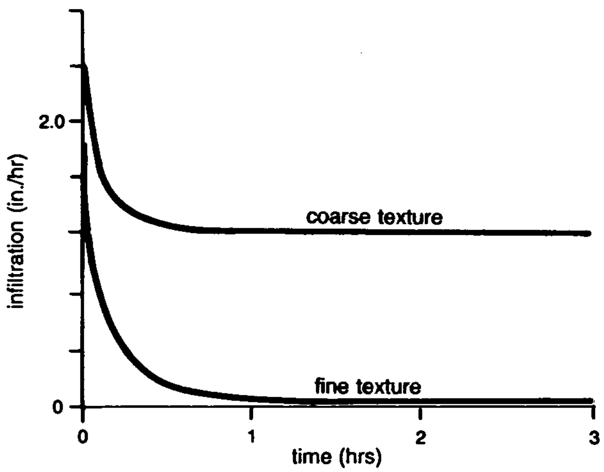


Figure 2 Graph showing decrease in infiltration capacity for two soil textures during a preciptation event (from USEPA 1990).

Suction Forces and Hydraulic Conductivity in the Unsaturated Zone

In unsaturated soils, pore pressure is negative, relative to atmospheric pressure. This negative pressure is expressed in positive notation as *tension.* Tension is a result of suction forces that have two primary components: *capillarity,* the tendency for water to form a meniscus between particles; and *adsorption,* the attraction of water to the surface of soil/rock particles (Hillel 1982). Suction forces are caused by hydrogen bonding between the molecules of water and other molecules of water, soil, and rock (Lehr 1988). The bonds are generally stronger in fine-grained soils and

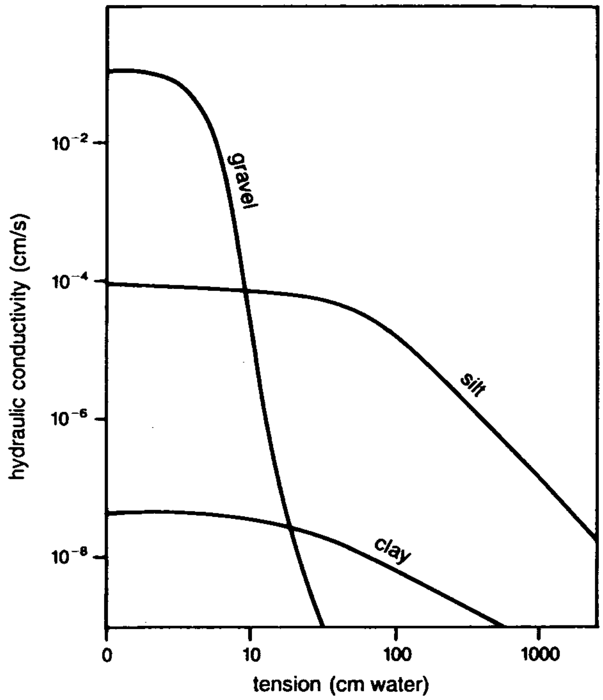


Figure 3 Relationship of hydraulic conductivity to soil tension for three types of unsaturated soil materials.

weaker in coarse-grained soils because fine-grained soils have more surface area where hydrogen bonding can occur.

Water moves through unsaturated soils in the form of continuous films of water that coat the soil particles and fill small pores. In very dry soil, there are few continuous films and interconnected water-filled pores and, accordingly, there will be little water movement. As the soil wets, the films become thicker and more continuous, larger pores fill with water, and the interconnectivity of the pores increases. More and larger pathways are thus available for water movement.

The capacity of a soil/rock unit to transmit water is termed *hydraulic conductivity*. The hydraulic conductivity of a saturated material is fairly constant in any given direction over time; however, the hydraulic conductivity of an unsaturated soil is a function of its moisture content. Hydraulic conductivity has distinct directional components and will vary from place to place within a formation. In very dry soils with few connected water-filled pores, the hydraulic conductivity may be as low as 1/100,000 of its saturated value (Hillel 1982). As the soil-moisture content and the number of available pathways for water movement increase, hydraulic conductivity increases.

The hydraulic conductivity of saturated coarse-grained materials is typically orders of magnitude greater than it is in saturated fine-grained materials. The opposite is usually true in unsaturated soils. As a soil dries, the largest pores drain first, with progressively smaller pores draining as suction increases. The drainage of soil pores decreases the pathways available for fluid flow, causing an exponential decrease in hydraulic conductivity. In fine-grained soils, the pores are small and retain more water under a given suction than coarse soils with larger pores. Therefore, hydraulic conductivity decreases more slowly and has a greater value in fine-grained, small-pored soils than in a coarse-grained soil at a given tension value (fig. 3; Hillel 1982).

Factors that Affect Infiltration

Hydraulic head *Hydraulic head*, also known as *potential head*, is a measure of the potential energy of groundwater at a given point beneath the surface. Groundwater in both the saturated and unsaturated zones moves from points of high head to points of low head. Infiltration into a soil or rock occurs when the hydraulic head in the soil/rock is lower than the hydraulic head at the ground surface. The change in hydraulic head over a given distance is the *hydraulic gradient*.

Hydraulic head has three components: pressure head, elevation head, and velocity head (fig. 4). The *velocity head* results from the inertia of water. Because the movement of water is relatively slow, velocity head is negligible compared with the other heads and is usually ignored. In unsaturated materials, suction forces create a negative *pressure head* that is usually expressed as *tension* or *suction*. In saturated materials, there are no net suction forces that influence flow. Pressure head in the saturated zone is positive because the water at any point is subjected to pressure from the weight of the water above. *Elevation head* is a measure of the potential for water to move downward in response to gravity; it is sometimes referred to as *gravity head*.

Infiltration will always occur in a porous unsaturated soil or rock because the negative pressure head in the soil is lower than the zero-pressure head of free water of the ground surface. However, infiltration will only occur in saturated soils if the hydraulic head in the soil is lower than the

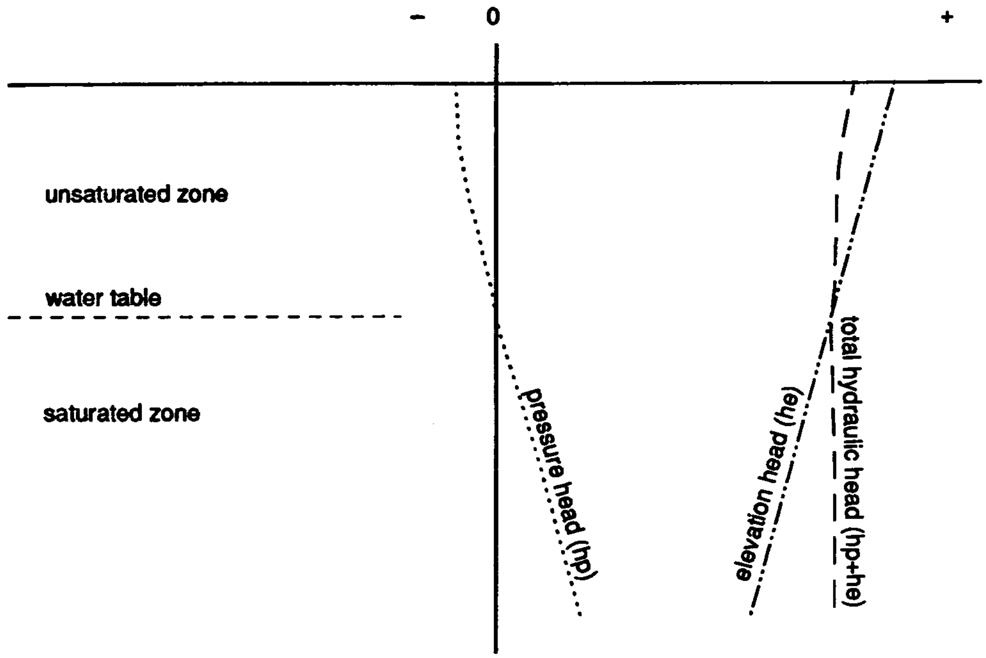


Figure 4 Vertical graph showing typical change in pressure, elevation, and total head with depth in the soil profile. Pressure head is negative in the unsaturated zone and positive in the saturated zone.

hydraulic head of the free water on the ground surface, a condition that may occur if water is ponded on the surface.

Degree of initial soil moisture The infiltration capacity of a soil will be greatest when it is dry and least when it is wet. During a heavy rainfall, there may be more infiltration in a dry soil than in a similar wet soil because the precipitation rate is more likely to exceed the infiltration capacity of the wet soil and cause some of the precipitation to run off rather than infiltrate. Infiltration capacity decreases with increasing soil wetness because the pressure head in the soil increases in response to decreasing suction forces, thereby reducing the hydraulic gradient. When the soil becomes saturated, there is no appreciable suction force and water will only infiltrate in response to elevation head (gravity forces).

Texture The effects of texture on infiltration are most pronounced when a soil is wet. Table 1 shows minimum infiltration rates (infiltration capacity) for wet soils. In general, the infiltration capacity under wet conditions will be greatest in coarse-grained soils and least in fine-grained soils (Musgrave 1955).

Table 1 Typical miinimum infiltration rates in wet soils (from Hillel 1980).

Soil Type	Infiltration Rate (cm/hr)
Sand	>2.0
Sandy-silty soils	1.0-2.0
Loam	0.5-1.0
Clayey soils	0.1-0.5

Soil layering will also affect infiltration (fig. 5). When a coarse-grained soil is underlain at a shallow depth by a fine-grained soil, infiltration may be impeded at the fine-grained layer. A perched water table may form in the coarse-grained soil because the volume of water that infiltrates into the coarse-grained soil may be greater than the volume that can flow through the fine-grained soil. In a flat-lying area, the coarse-grained soil may become saturated, in which case water will

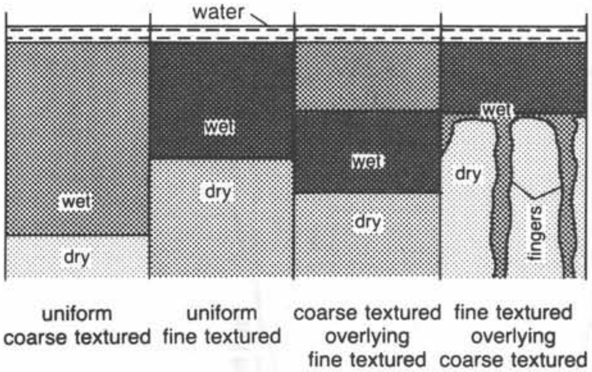


Figure 5 Effect of soil layering on infiltration (Hillel 1980).

pond and infiltrate at the minimum infiltration rate (saturated infiltration capacity) of the fine-grained soil. If this soil profile occurs on a slope, water in the coarse-grained layer may drain to a surface water body, with little flow into the fine-grained layer; such drainage is termed subsurface runoff. In either case, the rate and volume of infiltration may be decreased, relative to a uniform coarse-grained soil.

When a fine-grained surficial soil is underlain at shallow depth by a coarse-grained soil, the surficial infiltration rate will be dictated by the properties of the surficial soil layer. Because suction forces are stronger in fine-grained soils than in coarse-grained soils, however, water may accumulate in the fine-grained soil and not exit that layer until the tension at its base is lower than the tension in the underlying coarse-grained layer. When that tension is reached, water may move through the coarse-grained layer, often in saturated columns, or *fingers*, separated by dry soil (Hillel 1980).

Infiltration into the fine-grained surficial layer of this soil profile would not differ much from infiltration into a uniform, fine-grained soil profile, except for two conditions. First, the moisture content in the upper layer will increase faster than if the soil were uniform, because flow into the underlying coarse-grained layer may be retarded until tension in the upper layer is reduced to nearly zero (Frind et al. 1977). This will cause the infiltration capacity of the surficial layer of this profile to decrease faster than it would for a uniform, fine-grained profile, although the minimum infiltration capacity should be the same. Second, the possibility for deep infiltration may be reduced in some instances because the amount of infiltrated water may not be great enough to cause a tension decrease sufficient for flow into the lower, coarse-grained layer. This water will then be held in the upper layer where it may be returned to the atmosphere through evapotranspiration (Cartwright et al. 1988).

General soil structure Highly aggregated soils generally have greater infiltration rates than non-aggregated soils because the aggregates act as large grains, allowing infiltration rates similar to coarse sands (Musgrave 1955). Organic matter promotes formation of aggregates, so fine-grained soils with a high degree of organic matter generally have higher infiltration rates than soils with little organic matter. Rain drops falling on unprotected soil or intense tillage can break up soil aggregates, causing a dense soil layer to form (*rain packing*). This dense layer will have low hydraulic conductivity and, subsequently, a low infiltration rate (Musgrave 1955).

In most humid regions, vegetation protects soil from rain packing and dispersal, and the supply of humus and microfauna activity can help keep soil structure open (Dunne 1978). The efficiency of vegetation in maintaining soil structure, however, depends on the type of vegetation. There is generally less infiltration under intertilled crops such as corn, peanuts, and soybeans than under grass, trees, and mulch because the latter three preserve soil structure (Musgrave 1955).

Precipitation rate and duration There can be no infiltration unless there is a source of water at the ground surface. In Illinois, this source of water is commonly rain or melting snow.

As long as the rate of precipitation is less than the infiltration capacity of a soil, most of the water will infiltrate at a rate equal to the precipitation rate. As the moisture content of the soil increases, the infiltration capacity will decrease. Eventually the infiltration capacity may become lower than the precipitation rate. When this occurs, the infiltration rate will be equal to the infiltration capacity. At this time, some water will pond in surface depressions, where it will continue to infiltrate or evaporate, and some may run downslope as overland flow (Dunne 1978). The surface infiltration rate will increase if the depth of the pond increases because the elevation head of the pond increases, causing the gravity gradient to increase (Rehm et al. 1982).

As the duration of rainfall increases, the volume of water available for infiltration increases. As an example to illustrate the effects of precipitation rate and duration on infiltration, consider a soil profile that is subjected to two rain events where the initial moisture content of the soil is the same at the beginning of both events. The first event is a storm where 2 cm of rain falls in 1 hour. The second event is a slow, steady rain where 2 cm of rain falls in 24 hours. More water will infiltrate in the second than in the first event because the precipitation rate in the first event will probably exceed the soil's infiltration capacity, and a large part of the rain water will run off rather than infiltrate. The slow, steady rain of the second event falls at a rate that does not exceed the soil's infiltration capacity; therefore, most of that water will infiltrate.

Land use As mentioned previously, trees and grasses generally promote infiltration because soil structure is maintained by these types of vegetation, and heavy tillage for row crops generally destroys soil structure, thus reducing soil permeability and subsequent infiltration. In urban areas, there is little infiltration beneath paved surfaces. Lawns in suburban areas are areas of relatively high infiltration, although soil compaction caused by human activities on grasses may cause lower infiltration than in unused natural areas. Therefore, relative infiltration rates and capacities will be highest in wooded areas, meadows, and lawns, lower on lands planted with small grains and row crops, and lowest in paved urban areas.

Infiltration through Macropores

The preceding discussion assumes that infiltration is through intergranular pore openings. Infiltration also occurs through secondary pore openings known as *macropores*, which include fractures, animal burrows, and root channels. Infiltration through intergranular pore openings is a function of capillary forces and gravity; infiltration through macropores is mainly a function of gravity (Germann and Beven 1985).

Numerous papers have described the importance of infiltration through macropores. Williams and Farvolden (1967) found that joints in clay-rich till controlled the movement of groundwater through the till. Ehlers (1975) observed that macropores in tilled and untilled loess soils, with openings at the soil surface, can significantly influence the rate and pattern of water entry into field soils, especially under ponded conditions. Glass et al. (1988) showed that macropores in a soil cause fingers through which water may move to the saturated zone, and that these fingers may have a higher moisture content than the surrounding soil when overall soil-moisture conditions are dry. Hendry (1983) measured tritium concentrations in a soil profile and concluded that fractures in till allowed rapid downward movement of water during large storms events. Using a ring infiltrometer, Smettem and Collis-George (1985) concluded that infiltration rates in a well-drained fine sandy-silty loam were predominantly controlled by the number and size of macropores.

Germann and Beven (1985) stated that macropore flow only lasts for short periods and macropores may account for a small percentage of total porosity; yet the flow capacity in macropores may exceed that of the soil matrix by several orders of magnitude. They also concluded that the ratio of macropore flow volume to soil matrix flow volume was greatest for fine-grained soils and least in very coarse soils.

The presence of macropores in a soil or rock does not necessarily mean that infiltration will occur through macropores, particularly if they are not open to the atmosphere. Water in the unsaturated zone does not readily move from fine-grained to coarse-grained materials or from porous materials to open spaces such as macropores. Macropores must therefore be in contact with free water or a saturated material to effectively conduct water (Ehlers 1975). Thus, infiltration through macropores will most likely occur when the upper surface of the soil or rock is saturated or when water has ponded at the ground surface.

Infiltration through Frozen Soils

Storey (1955), Startz (1969), and Rehm et al. (1982) found that infiltration and subsequent groundwater recharge is possible even through concrete frost. The rate of infiltration may be reduced because the permeability of the frozen soils decreases as winter progresses and more ice forms in the soil voids. They also noted that the effect of freezing on permeability is greater in soils with fine texture than in soils with coarse texture. Rehm et al. (1982) noted that the major effect of seasonally cold temperatures is to tie up potential infiltration as snow or ice. As this snow or ice melts, the water will be free to move down through the soil to the saturated zone with little or no loss to the atmosphere as evapotranspiration.

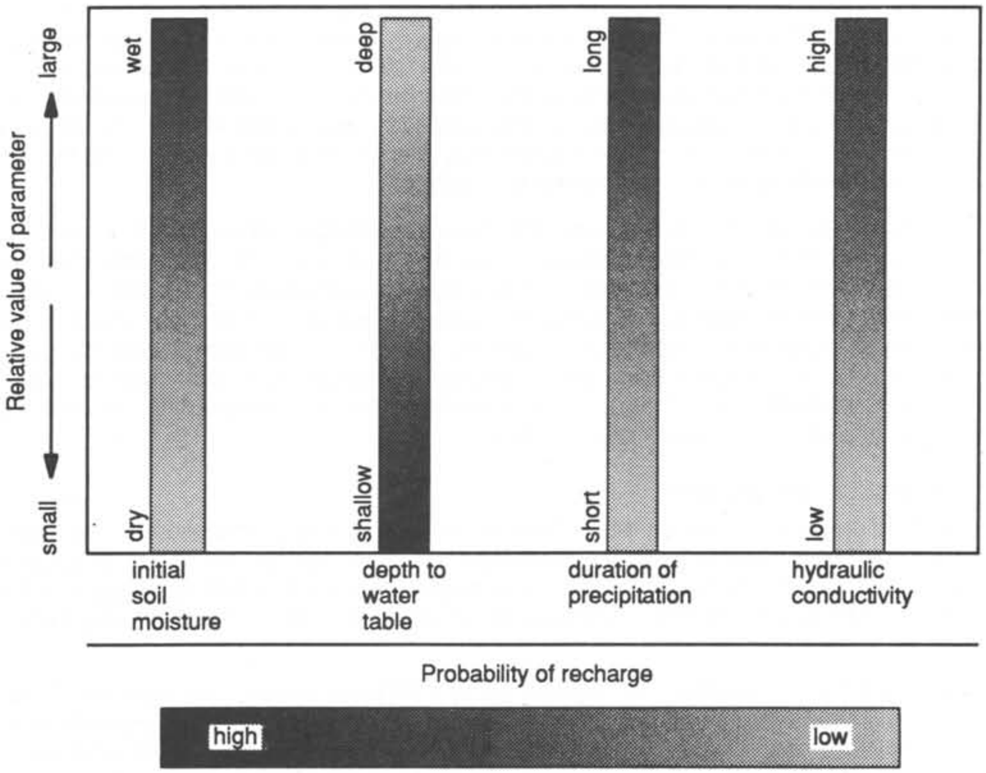


Figure 6 Relative effect of various hydrogeologic parameters on the probability of recharge occurring for a single precipitation event.

Generalized Model of Infiltration

This discussion of infiltration provides a background of the environmental factors that affect infiltration, from which a generalized model of infiltration can be developed. Infiltration occurs when water is supplied to the ground surface by rainfall, snow melt, or irrigation. If the rate of water supply is lower than the infiltration capacity of the upper soil layer, water will infiltrate without ponding, otherwise some water will infiltrate and some will pond or run off. The moisture content of the soil will increase and the infiltration capacity will decrease as infiltration occurs. If the soil becomes saturated, the infiltration capacity will reach a minimum value, which is a function of the saturated hydraulic conductivity of the soil or rock (Freeze and Cherry 1979). The infiltration capacity at saturation is generally higher for coarse-grained soils than for fine-grained soils. Thus, rates of infiltration generally are greater in sandy soils than in clayey soils.

In fine-grained soils, infiltration also occurs through macropores (Germann and Beven 1985). Infiltration through macropores will usually occur during heavy rain or snow melt, when the macropores are in contact with free-standing water or when the soil is saturated.

Once water has infiltrated, it may be drawn back to the atmosphere as evapotranspiration, used to replenish moisture deficits in an unsaturated soil/rock formation, diverted toward a nearby surface water body as subsurface storm flow, or continue downward to recharge groundwater.

RECHARGE

Factors that Affect the Occurrence and Rate of Recharge

As stated earlier, groundwater recharge is defined as the advance of water into the saturated zone that is associated with groundwater flow away from the water table and within the saturated zone (Freeze and Cherry 1979). The occurrence and rate of recharge are functions of the (1) character (texture, moisture content, specific moisture capacity, and tension) and thickness of the overlying unsaturated materials, (2) topography, (3) vegetation, (4) land use, (5) depth to the water table, (6) intensity, duration, and distribution of water application, and (7) air temperature (Walton 1965). Layering of deposits, macropores, and other soil nonuniformities may also affect recharge (Steenhuis et al. 1985).

The water table may rise during many recharge events. Recharge and water-table rise are not the same, however, nor does recharge always cause a water-table rise (Freeze 1969). When groundwater flow has a downward component — a necessary condition for recharge — the water table slowly falls as water drains downward. Therefore, recharge is necessary to maintain a given elevation of the water table.

Freeze (1969) concluded that, all other conditions being constant, the possibility of recharge is greater under the following circumstances:

- initially wet soil conditions rather than dry;
- rainfalls of long rather than short duration;
- shallow rather than deep water-table conditions;
- soils with high hydraulic conductivity, low specific moisture capacity, or high moisture content over a wide range of tension values.

Freeze (1969) states that the initial soil moisture regime is probably the most important of these parameters. If a soil profile has a relatively high moisture content prior to an infiltration event, the water will be more likely to reach the water table because less water will be needed to make up moisture deficits in the unsaturated zone (Rehm et al. 1982). Figure 6 graphically illustrates the importance of these factors in determining the probability of a precipitation event causing groundwater recharge.

Rehm et al. (1982) observed three general environmental factors that affect recharge: the spatial and temporal distribution of precipitation, the saturated and unsaturated hydraulic properties of a heterogenous hydrogeologic environment, and the relationships between topography and runoff.

Precipitation The potential for water movement through the unsaturated zone is more dependent on the duration than on the intensity of the rainfall event (Freeze 1969). Because runoff will be less and infiltration will be greater, a low-intensity event of long duration will more effectively recharge groundwater than a high-intensity event of short duration.

Numerous field studies have shown that recharge in temperate and humid regions occurs most commonly in the spring when snow melts, precipitation is high, and evapotranspiration is low. Recharge may also occur during late fall after vegetation becomes dormant, so that evapotranspiration is greatly reduced. Under these conditions, the water table is usually high and soil moisture is often near saturation (Rehm et al. 1982, Steenhuis et al. 1985, Walton 1965). For example, Steenhuis et al. (1985) measured recharge on Long Island and estimated that 75% to 90% of precipitation from mid-October to mid-May becomes recharge. Little summer and early fall precipitation percolates all the way down to the saturated zone to recharge groundwater. During the summer, high evapotranspiration rates following an infiltration event can remove much of the infiltrated water before it reaches the water table (Rehm et al. 1982).

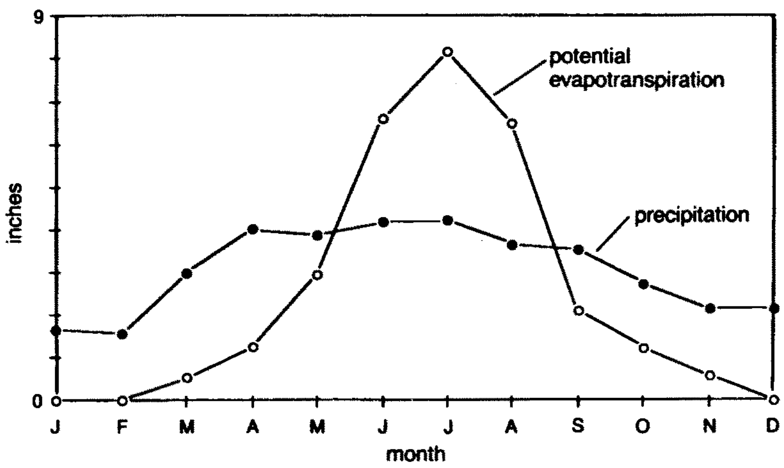


Figure 7 Monthly precipitation and potential evapotranspiration rates in Illinois (Bowman and Collins 1987).

Figure 7 shows potential evaporation and precipitation rates for an east-west band across north-central Illinois. As can be seen in the figure, potential evapotranspiration is greater than precipitation in the summer months. Given this relationship, it can be inferred that evapotranspiration and resultant soil-moisture deficits use most summer precipitation in Illinois. Summer recharge only occurs during periods of excessive rainfall (Walton 1965).

The preceding discussion applies to cases of infiltration through granular soil pores. If recharge occurs primarily through macropores, then recharge may be more likely during high-intensity precipitation events when water will pond, allowing flow into the macropores. An extreme example of macropore recharge would be in an area of karst topography. During precipitation events of low intensity, water will soak into the soil, and recharge, if any, will be by infiltration through intergranular pores. During high-intensity precipitation events, water will pond on the ground surface. Ponded water will run off to the lowest areas on the landscape — sinkholes. In a recharge sense, sinkholes are very large macropores, and water that runs off to these sinkholes will rapidly recharge the karst aquifer.

Hydrogeologic environment Rehm et al. (1982) observed that recharge rates in North Dakota were about one to two orders of magnitude higher in areas where coarse-grained, sandy soil was at the surface, compared with areas where fine-grained soil occurred at the surface. Infiltration through fine-grained soils is generally slower because the infiltration capacity and saturated hydraulic conductivity of these soils are generally orders of magnitude lower than in coarse-grained soils.

As stated previously, the depth to the water table may also affect the potential for and rate of recharge. Similar geologic profiles may not have similar recharge characteristics if the water table of one is deeper than the other. A profile with a deep water table will require more water to replenish and maintain the moisture requirements in the unsaturated zone than one with a high water table. In other words, given equivalent infiltration events, more water may be needed to replenish moisture deficits in the profile with the deeper water table; thus, less water will be available for recharge than in the profile with the shallow water table.

Topography The simplest indicator of a recharge area is a topographic high (Freeze and Cherry 1979). Because groundwater flows from areas of high hydraulic head to areas of low hydraulic head, it is not uncommon to find downward flow associated with a groundwater recharge area in the portions of a groundwater basin where the topographic elevation and the elevation head are high, and upward flow associated with a groundwater discharge area in the lowest portion of the basin where the elevation head is low (fig. 8).

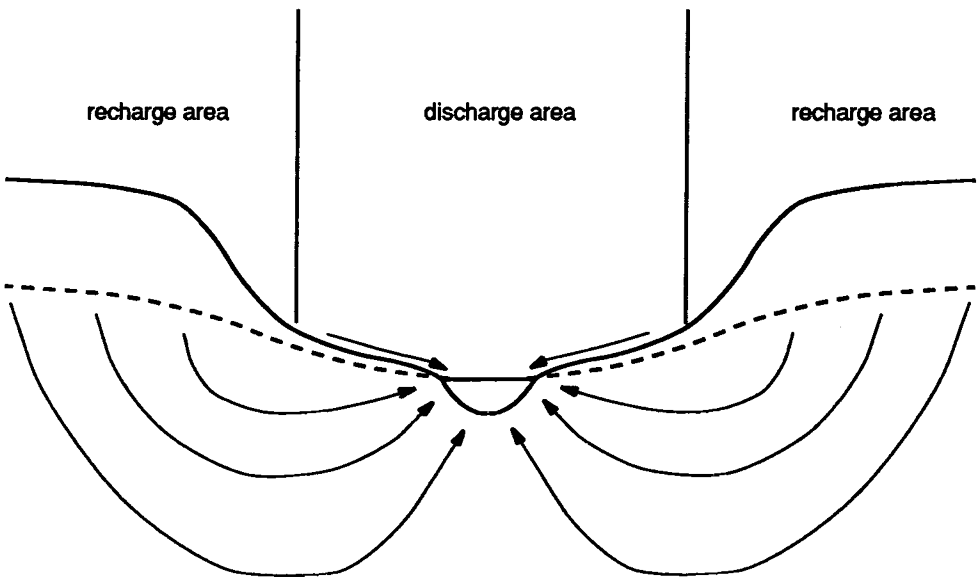


Figure 8 Schematic representation of topographic influences on recharge and discharge areas. Recharge in Illinois generally occurs on uplands and discharge generally occurs on lowlands.

Topography may also affect the spatial distribution of recharge. Rehm et al. (1982) found that recharge often occurred beneath small surface depressions where ponding occurred. Closed depressions, however, will not always be recharge areas. Meyboom (1966) and Williams (1968) showed that closed depressions may be recharge or discharge areas, or may alternate, depending on the geometry of the depression, volume of precipitation, and the transpiration needs of vegetation in the depression.

Freeze and Cherry (1979) noted that small lakes (or marshes) are usually discharge areas, but they can be areas of depression-focused recharge. Because regional discharge generally occurs in lowland areas and regional recharge usually occurs in uplands, the probability of depression-focused recharge is greatest when the depression is located on an upland and least when on a lowland.

The general rule in humid regions is that discharge occurs in lowlands where groundwater may maintain lake and stream levels, and recharge occurs in uplands where flow is away from the water table and toward discharge areas (fig. 8). Slopes that lie between uplands and lowlands are more difficult to characterize. Philip (1991) theorized that infiltration on slopes may be slightly less than on horizontal surfaces because the effect of capillary forces on infiltration is decreased, although gravity remains constant. He showed that even on slopes of 30°, the reduction in infiltration rate and volume is usually less than 5%, as compared with a similar soil profile on a horizontal surface. In addition to a slight reduction in infiltration rates, other factors will also affect recharge on sloped surfaces. When the slope is saturated, the rate of infiltration is reduced because the soil is at its infiltration capacity, and excess water will probably run off rather than pond. Therefore, if all other soil factors are equal, the total potential for recharge on slopes may be less than on flat uplands because there is less ponded water. However, all other factors are generally not equal. The water table beneath a sloped surface is often deeper at the top than at the base. If the sloped surface is in an upland setting, there may be more recharge at the base, where the water table is often shallow, than at the top, where the water table is often deep, because more water is needed to replenish moisture deficits in the thicker unsaturated zone at the top. In a lowland setting, the base of the slope may be in an area of groundwater discharge. In this case, recharge can only occur near the top of the slope.

Distribution of Recharge

The rate of groundwater recharge and discharge varies from place to place throughout the areal extent of a groundwater basin (Freeze and Witherspoon 1968, Kempton and Cartwright 1984). Recharge water can follow several routes to reach the saturated zone, including intergranular seepage, macropore systems that allow flow to the saturated zone and have little effect on soil-moisture conditions, and overland flow to depressions where the water table may be at the ground surface and groundwater flow is downward (direct recharge). As discussed previously, recharge through sinkholes in karst terrains is a special case (Gerhart 1986).

When the infiltration capacity of the surficial soil is low, as may be the case for clayey soils, recharge may be concentrated in closed depressions where runoff water will generally pond (Rehm et al. 1982). In such cases, an inverted water table may develop. When the infiltration capacity of the surficial soil is high, as is common for sandy soils, recharge may not be affected by minor topographic fluctuations.

Land use is another factor that may affect the distribution of recharge. Recharge is less likely to occur in urban areas than in rural areas because a significant part of any urban area is paved and storm sewers direct runoff from the paved areas directly to surface water bodies. In suburban areas, recharge may be greater than in rural areas because there is no evapotranspiration from paved areas. In addition, runoff from the paved areas is often directed to ditches, yards, and stormwater retention cells where the water may infiltrate, thereby increasing the amount of water potentially available for recharge.

Vegetation may also have an effect on the distribution of recharge because it can influence the infiltration rate in the soil, and thus the amount of water potentially available for recharge. Also, some types of vegetation transpire a greater amount of water than others, affecting the amount of soil water available for recharge. Vegetation also provides shade, which may reduce evaporation

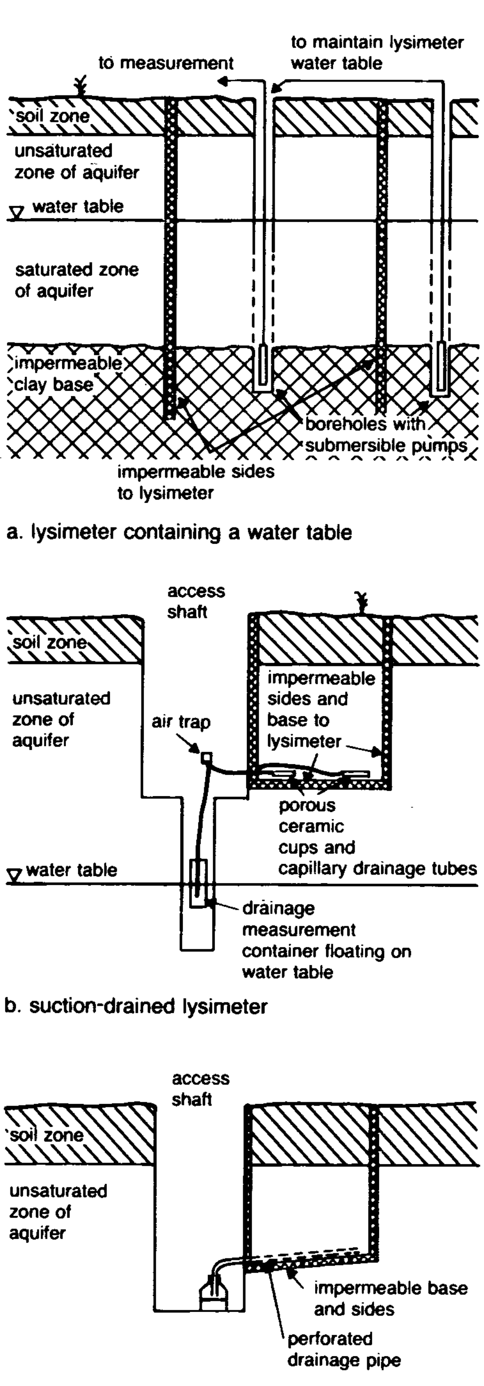


Figure 9 Schematic drawings of three types of lysimeters that can be used to measure recharge (Lerner 1990).

at the ground surface, thereby increasing the amount of water available for recharge. Nevertheless, increased transpiration generally exceeds decreased evaporation.

METHODS TO ESTIMATE RECHARGE

Introduction

No single, easy and accurate method can be used to directly measure the rate or amount of recharge entering the groundwater flow system. Variations in geologic and soil conditions, and precipitation patterns reduce even the most exacting methods to mere estimates of recharge.

Estimates of recharge are made at two scales: point measurements for areas as small as a square meter, and basinwide measurements for regional recharge. Point measurements are more exacting than basin measurements and thus generally have a higher degree of accuracy. Many point measurements, however, cannot be used to estimate basinwide recharge unless the basin is relatively homogeneous or measurements are made in all types of terranes found within the basin. Basin measurements may be more prone to error, and they do not reflect differences in recharge characteristics that may occur from location to location within the basin, but they are still the most practical way to estimate regional recharge rates and volumes.

Point Measurements of Recharge

Direct measurement A lysimeter is an in-situ device, used to measure quantities of water used by plants, evaporated from soil, and lost by deep percolation. It is constructed with impermeable sides and, in some cases, an impermeable bottom, so that the flow of water through the undisturbed block of soil within the device can be measured. The lysimeter must be large enough to minimize edge effects and deep enough to enclose entire root systems. It is therefore expensive and difficult to construct (Lerner 1990).

Three types of lysimeters have been used to estimate groundwater recharge rate and volume. One is a water-table lysimeter (fig. 9a), in which water is pumped in or out so that the water table inside is kept at the same level as the water table outside. A tightly controlled water budget can then be maintained with this device by monitoring pump volumes and water levels inside the lysimeter. A second type is a suction-drained lysimeter (fig. 9b). This device has a drainage layer installed just above the water table. Recharge is equal to the volume of water drained. A third type of lysimeter is gravity drained (fig. 9c). It is similar to but simpler than the suction-drained lysimeter. These devices are typically used in coarse-grained soils where there is little capillary rise and where conditions at depth have little influence on vertical water movement (Lerner 1990). Examples of lysimeter applications are given by Kitching and Bridge (1974) and Kitching et al. (1977).

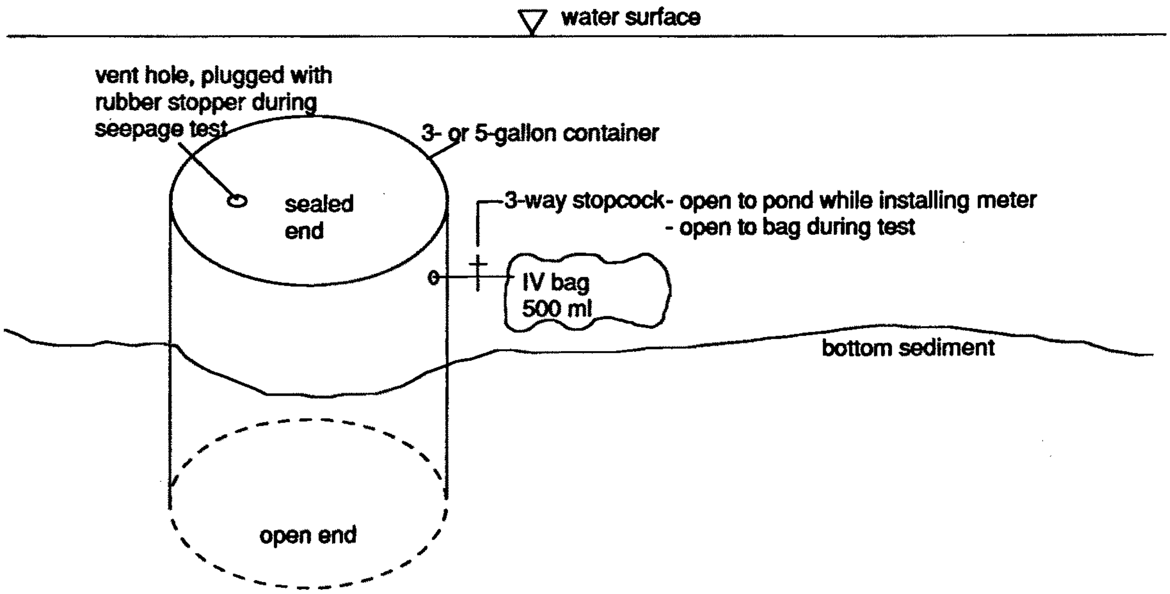


Figure 10 Schematic drawing of a seepage meter that can be used to measure recharge or discharge in a surface water body. Seepage is measured by change in water volume in IV bag.

Seepage meters may be used to measure recharge from (or discharge to) surface water bodies. Seepage meters generally consist of cylindrical containers with sealed tops and open bottoms of known area (fig. 10). The seepage meter is placed in the bottom sediment of the lake or stream for a given period of time, and a flexible bag containing a known volume of water is attached. The seepage flux is then calculated by

$$\text{Flux} = \frac{(V_o - V_f)}{A(t_f - t_o)} \tag{1}$$

where A = the area of the meter,
V_o = the volume of water in the bag at the beginning of the test (t_o),
V_f = the volume of water in the bag at the end of the test (t_f).

For this equation, groundwater recharge is indicated by a positive flux. Seepage meters need to be calibrated because they can disrupt groundwater flow, often resulting in erroneously low seepage rates. This calibration can be performed by setting the seepage meter in a sand tank having a known flux (Cherkauer and McBride 1988).

Darcy flux The Darcy flux is calculated from head measurements at nested piezometers. If flow is downward and the water table is stable over a given period of time, the Darcy flux will equal the rate of groundwater recharge. The Darcy flux is calculated by multiplying the vertical hydraulic conductivity of the material between two adjacent piezometers by the vertical gradient (fig. 11)

$$v = K_Z \frac{(H_1 - H_2)}{Z} \tag{2}$$

where v = the Darcy flux,
K_Z = the vertical hydraulic conductivity,
H_1 = the head in the shallow piezometer,
H_2 = head in the deep piezometer,
Z = the vertical distance between the two piezometers.

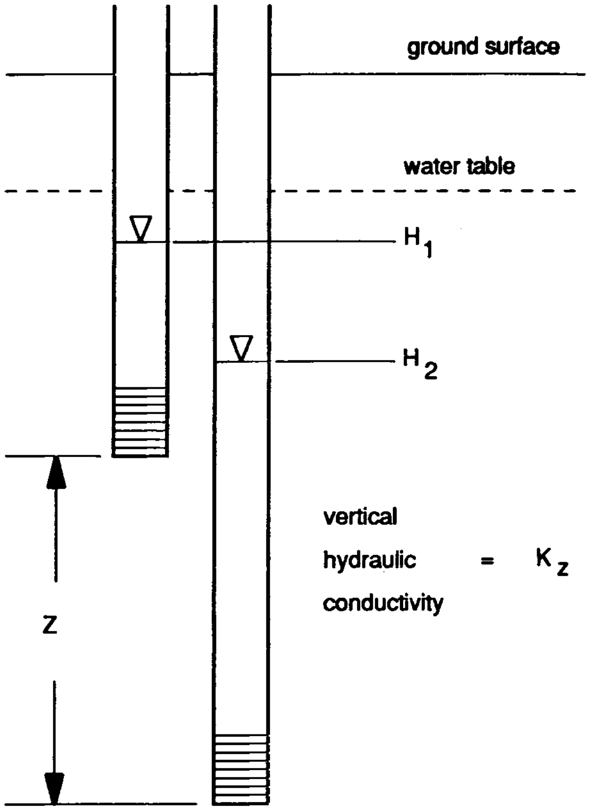


Figure 11 Hydrogeologic parameters needed to calculate recharge by the Darcy flux method.

Given this head notation, a positive value indicates downward flow (recharge) and a negative value indicates upward flow (discharge). An example of the Darcy flux calculation is given by Zebarth et al. (1989). Conversion of the Darcy flux to a recharge value requires that (1) the water table remains relatively constant so that it can be assumed that the amount of water passing the water table equals the amount of downward flow between the two piezometers, and (2) no change in storage takes place.

This method may also be used in the unsaturated zone, by using head values obtained from nested tensiometers. The relationship between hydraulic conductivity, moisture content, and soil tension must be known because vertical hydraulic conductivity is not constant in the unsaturated zone. For this calculation, the assumption is made that all downward flow below the tensiometers will eventually pass the water table as recharge. Steenhuis et al. (1985) gives an example of this calculation.

Water-table fluctuation Changes in the elevation of the water table at an individual observation well can be used to estimate recharge if the specific yield of the formation is known. The height of water-table rise at an observation well is multiplied by the specific yield of the formation to obtain a recharge value. Hughes et al. (1976) used the cumulative rise of the water table to calculate annual recharge near

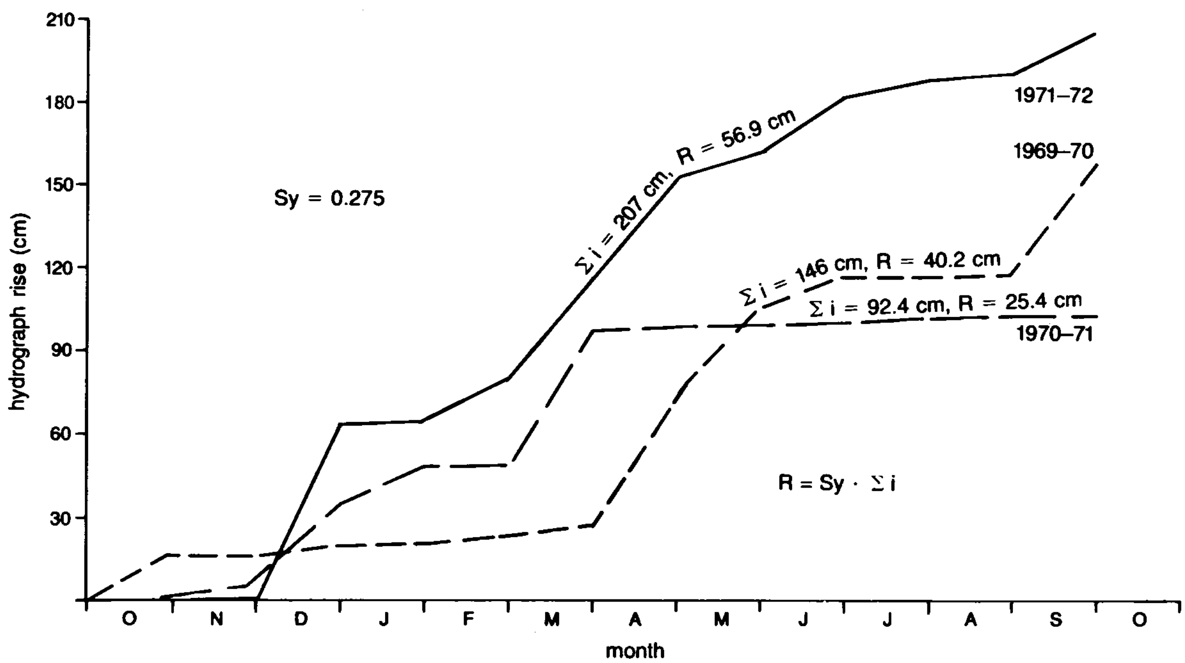


Figure 12 Example of recharge estimate based on cumulative water-table rise (from Hughes et al. 1976)

several landfills in Illinois (fig. 12). More recently, Nielsen and Widjaya (1989) used this method in a comparative study of methods to estimate recharge. Recharge rates estimated by this method were comparable with rates estimated from flow-net analysis and from a numerical model (methods discussed in a later section). The water-table fluctuation method assumes recharge occurs in events that cause the water table to rise and the unit is not steadily recharged in such a way that a rise in the water table does not occur. Results from this analysis will be most accurate if continuous water level measurements are used.

Soil-moisture budgets Soil-moisture budgets are most suitable for calculating recharge in humid regions (Lerner 1990). The Penman–Grindley model (Penman 1950, Grindley 1967, 1969, discussed in Lerner 1990) is a commonly used soil-moisture budgeting equation. Estimation of recharge using this method requires reliable estimates of soil moisture and soil-moisture deficits, vegetation constants, evapotranspiration, and precipitation (fig. 13). All soil-moisture models require calibration against in situ soil-moisture data or other recharge estimates (Lerner 1990). Kitching et al. (1977) compared soil-moisture budget estimates of recharge with lysimeter data and found that the mean annual recharge measured in the lysimeters was 175% of the values obtained with the soil-moisture budget. Rushton and Ward (1979) noted that the Penman–Grindley model generally underestimates the amounts of late summer and early autumn recharge when soil-moisture deficits occur regionally. They suggested improvements to this model by factoring in direct recharge or reducing the value of the vegetation constant.

Soil analysis Mills and Zwarich (1986) used soil profiles to differentiate recharge areas from discharge areas. They suggested that a soil profile with a well-developed B horizon and a deep solum that is free of carbonates develops when there is a long-term downward flux of water, such as in a recharge area. A poorly developed profile with an accumulation of carbonates or salts near the surface develops in areas with a net upward flux of water, as would occur in a discharge

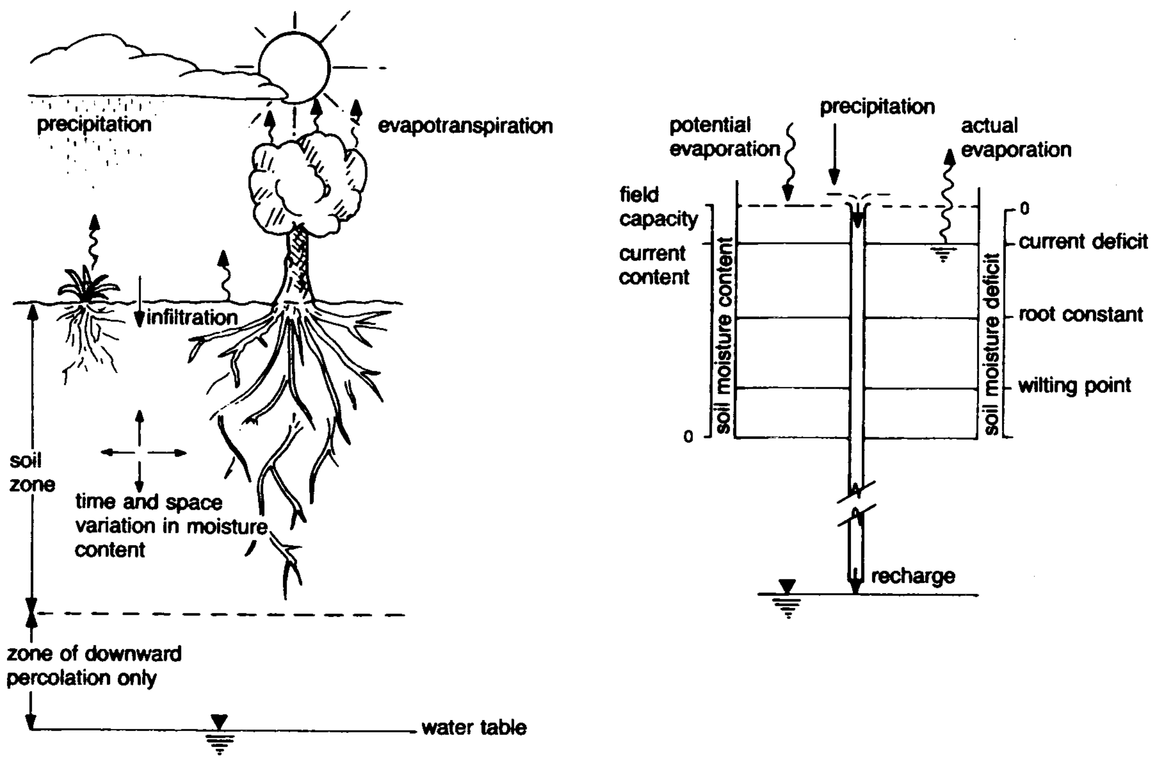


Figure 13 Schematic drawing of hydrogeologic parameters used in determining a soil-moisture budget (Lerner 1990). The thickness of the soil zone, rate of infiltration, rate of evapotranspiration, and initial soil-moisture conditions must be known. When the soil-moisture content is below field capacity, a soil-moisture deficit exists, and infiltrating water will replenish this deficit. Significant downward drainage and recharge occur only when moisture content exceeds field capacity. Therefore, the soil-moisture characteristics of the profile must also be known.

area. This method could not be used to quantify recharge rates, but it could indicate whether recharge or discharge is occurring.

Basinwide Estimates of Recharge

Regional water-table rise Lerner (1990) presented a method of calculating aquifer recharge rates based on the water-table rise of aquifers with well-defined recharge seasons. The method accounts for the loss of water to pumping wells and discharge from the aquifer. The first step is to map and contour the water-table rise for the entire aquifer. Recharge is then calculated from the following equations:

$$r = \frac{(\delta s + \sum Q_a \delta t + V_D)}{A \delta t} \tag{3}$$

$$\delta s = S_y \sum_i 0.5(c_i + c_{i+1})A_i \tag{4}$$

where r = total recharge rate per unit area (L/T),
s = volume of water stored between lowest and highest water-table positions (L^3),
Q_a = pumping rate from wells during water-table rise (L^3/T),
t = time interval between high and low water-table conditions (T),
A = area of aquifer (L^2),
V_D = volume of water discharged to the surface (L^3),
S_y = specific yield,
c_i = value of water-table rise contour i (L),
A_i = area between water-table rise contours (L^2).

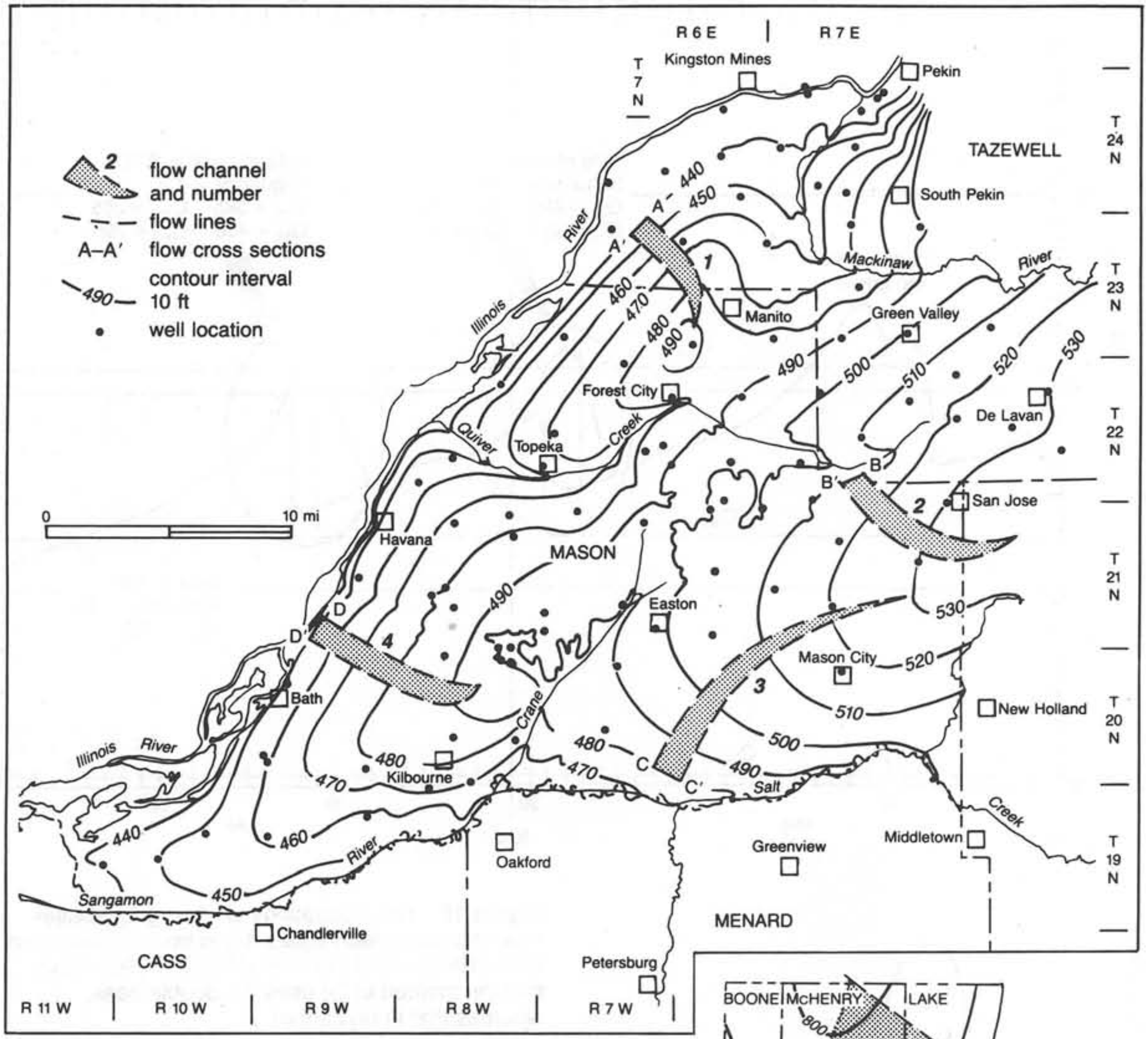
Water budget Recharge is estimated as the residual term in a water budget. A simple water-budget equation is

$$R = P - RO - ET \tag{5}$$

where R = recharge,
P = precipitation,
RO = runoff to surface water,
ET = evapotranspiration.

This method assumes that there is no interbasin flow of groundwater, and that the change in aquifer storage over time is negligible. Schicht and Walton (1961) and Steenhuis et al. (1985) present examples of water budgets. Steenhuis et al. (1985) calculated recharge for a small sandy area on Long Island, New York, using a Darcy flux estimate and a water budget. They found that the two methods gave comparable results during spring, summer, and fall, but the methods gave different results during the winter months. Because the groundwater component of a water budget is a residual and there is a degree of error in the calculation of the other parameters, particularly evapotranspiration, water budgets are less reliable than other methods of estimating recharge. Winter (1981) cautioned that water-budget estimates of groundwater fluxes may be off by more than 100 percent.

Flow-net analysis A flow net is a two-dimensional graphical representation of groundwater flow. Thorough discussions on the art and science of creating flow nets are given in Freeze and Cherry (1979) and Cedergren (1989) as well as in most other groundwater textbooks. Freeze and Witherspoon (1968) used a cross-sectional flow net to determine vertical hydraulic gradients, which they then used in a Darcy calculation to obtain estimates of recharge. Walton (1965) used areal flow nets to calculate recharge to aquifers. In one of Walton's calculations, the hydraulic conductivity of the unconfined aquifer and the horizontal gradient were used to calculate the flux



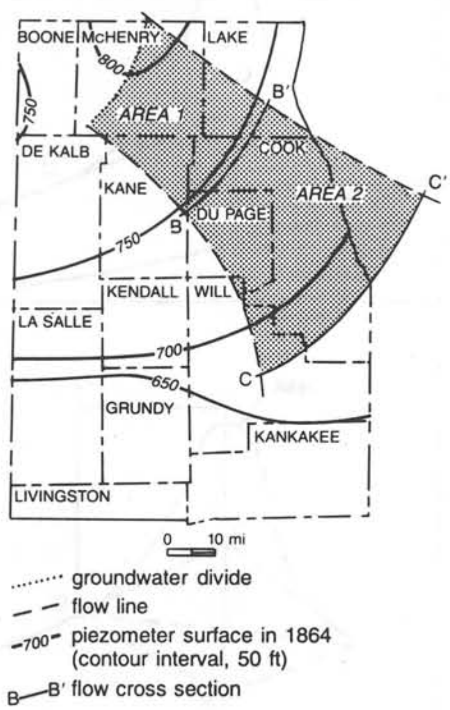
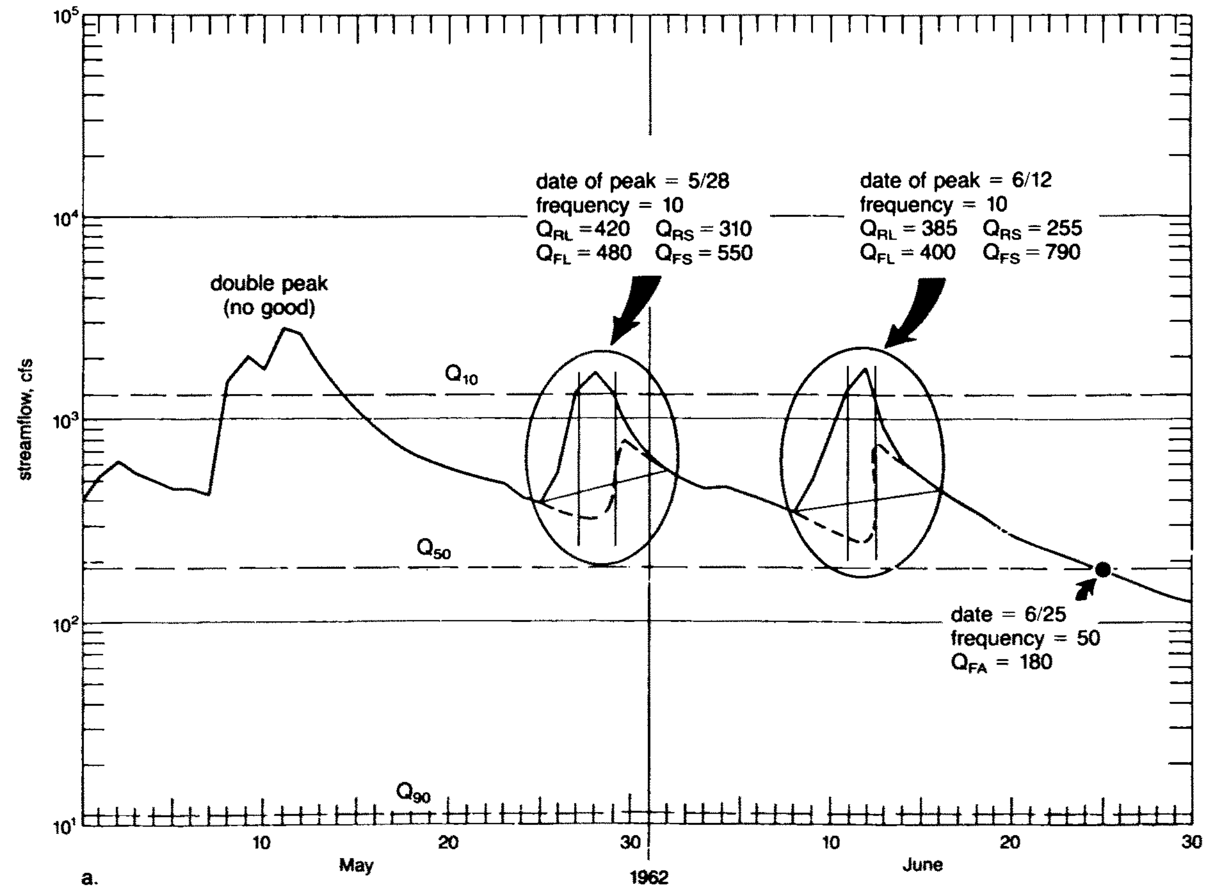


Figure 14 Examples of recharge estimates based on flow nets (from Walton 1965): (a) flow channels (flow lines converge at an upgradient point): recharge is equal to groundwater flow volume at cross section (e.g., A–A′) divided by area between flow lines; (b) recharge is equal to change in groundwater flow volume from cross sections B–B′ to C–C′.



a.

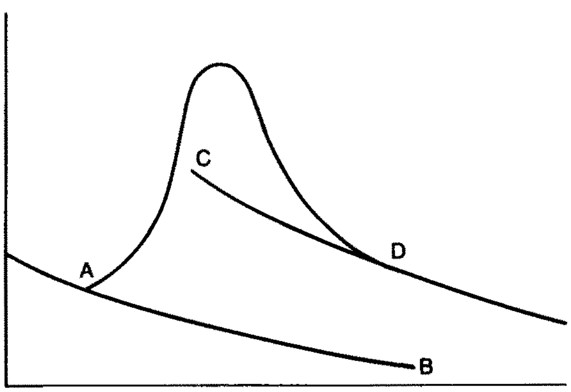
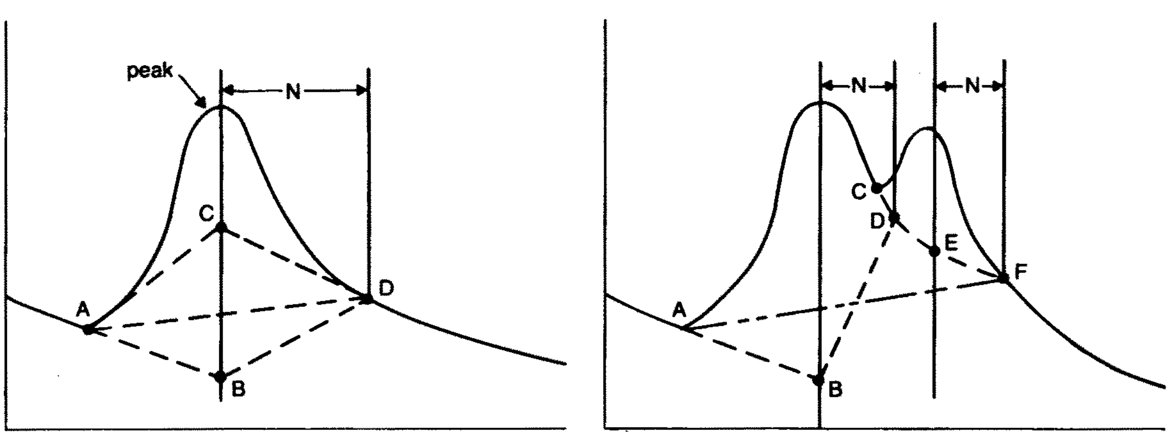


Figure 15 Two illustrations of hydrograph base-flow separation techniques: (a) is from O'Hearn and Gibb (1980) and (b) is from USEPA (1990). Note that the method in (b) uses the double peak, whereas that in (a) cannot.

b.



between pairs of flow lines that converged at an upgradient point (fig. 14a). Walton also used the flow net to calculate recharge to a confined aquifer based on the change in groundwater flux between two flow lines (fig. 14b). This calculation required an aquifer storage term as well as transmissivity and gradient data. The flow-net methods require a large number of groundwater observation points, and knowledge of hydraulic conductivity and anisotropy, so that the net may be properly constructed.

Base-flow separation Base flow is the portion of stream flow that comes from groundwater, as opposed to the portion from runoff. There are several methods for separating *base flow* from stream hydrographs, based on the assumption that the volume of base flow to a stream equals the volume of groundwater recharged in the stream's watershed. For many small watersheds, this is a valid assumption. If there are numerous groundwater withdrawal wells within the watershed or there is a downward gradient to a regional aquifer that does not discharge into the stream, however, this assumption will be erroneous.

The base-flow separation method requires the following assumptions:

- there is no long-term change in storage,
- all water recharged in the watershed discharges to the stream,
- all groundwater discharged into the stream was recharged in the watershed.

This method should only be used in watersheds for which records of stream discharge have been maintained for years so that long-term trends can be analyzed. It also has the disadvantage that there are no standard criteria for separating base flow from stream hydrographs (fig. 15), so comparisons cannot always be made between work of different authors (Nielsen and Widjaya 1989). Examples of this technique are given in Nielsen and Widjaya (1989), O'Hearn and Gibb (1980), and Houston (1990).

Depression-cone analysis Walton (1965) used the areal extent of the cone of depression caused by water wells to calculate recharge to aquifers. To make this calculation, he determined the volume of water pumped and divided it by the capture area of the pumping cone surrounding the well(s) (fig. 16). This method assumes that the volume of recharge in the capture area equals the volume of water being pumped.

Other Methods

Numerical methods Numerical methods of calculating recharge are generally derived from one or more of the preceding hydrologic methods, usually the water budget or flow net. Numerical models can be used to provide an integration of limited data over large areas. Because the validity of these methods depends upon the accuracy of the input data and the applicability of the assigned boundary conditions, numerical models should be used as a supplement to, rather than a substitute for, field data and field-based estimates of recharge.

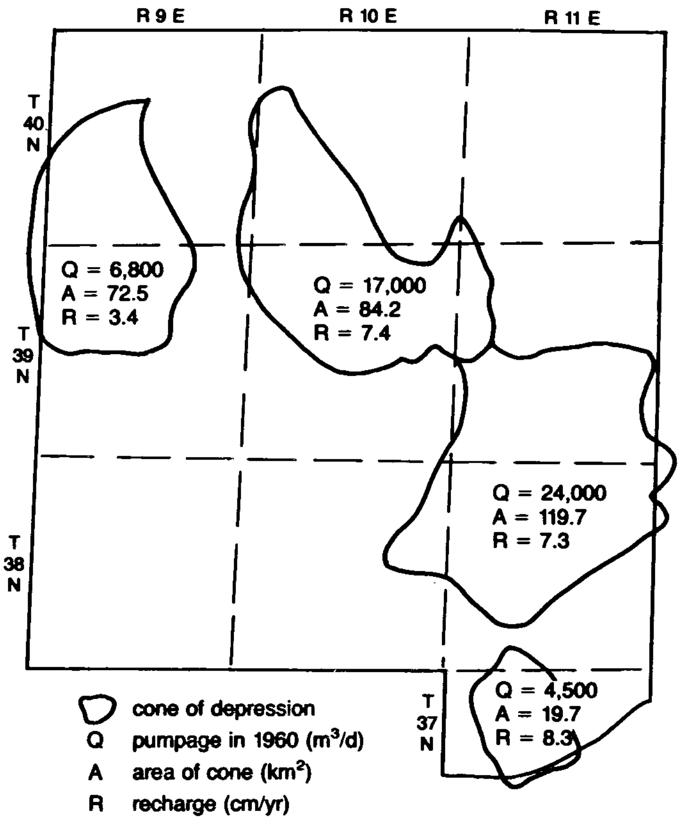


Figure 16 Recharge estimate based on the cone of depression around a groundwater pumping center. Recharge is the volume of water pumped divided by the area of the cone.

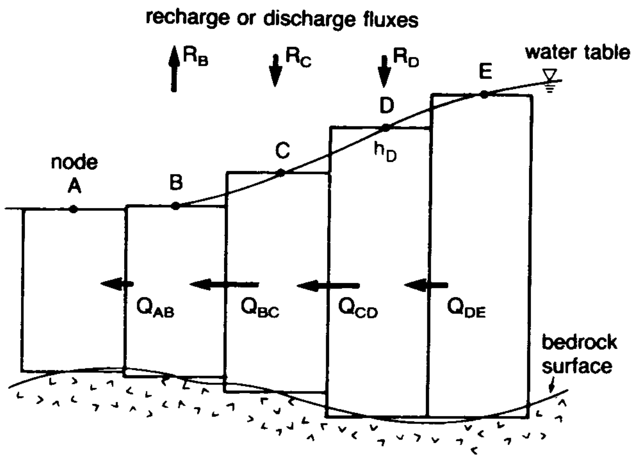


Figure 17 Schematic example of the mass-balance relationship used in a numerical estimate of recharge (Stoertz and Bradbury 1989). Recharge in cell A is Q_{CD}-Q_{BC}.

Stoertz and Bradbury (1989) used a water (or mass) balance routine in a two-dimensional model to estimate recharge. After entering the hydrogeological parameters of their study area, they specified the elevation of the water table as a constant head boundary. Because the head value of every node in the model domain was a constant, the model simply calculated the mass-balance of each node based on nodal values of hydraulic conductivity and constant head. Cells with a negative mass-balance are recharge areas because water must be added to maintain the constant head, and cells with a positive mass-balance are discharge areas (fig. 17).

A digital overlay technique for evaluation of recharge to a confined aquifer was demonstrated by Anderson and Giovannelli (1980). They assumed that recharge to an aquifer is proportional to the head difference between the aquifer and its confining layer, and inversely proportional to the vertical hydraulic conductivity and thickness of the confining layer. Thus, if the elevation of the water table in the confining layer is higher than the potentiometric surface of the aquifer, and the confining layer is thin or has fairly high hydraulic conductivity, recharge to the underlaying aquifer will be high. If the elevation of the potentiometric surface is higher than the water table or the confining layer is very thick and has very low hydraulic conductivity, little or no recharge will occur. Application of this method requires detailed mapping of the elevation of the water table, thickness and hydraulic conductivity of the confining layer, and potentiometric surface of the confined aquifer.

One-dimensional models of vertical fluid movement through the unsaturated zone have been used to estimate point recharge. The numerical codes for these models often combine calculations of the Darcy flux in the unsaturated zone with soil-moisture budgeting over time to estimate recharge rates. Johansson (1987) used such a model in a comparative study with a soil-moisture budget to calculate recharge. Annual recharge values calculated from the two methods were within 30% of each other. Johansson concluded that soil-water models could be valuable tools for studying temporal variations in recharge rates.

Estimates based on water chemistry The basic concept of mapping recharge on the basis of groundwater chemistry is that the chemistry and ionic composition of groundwater changes as it travels through the earth. Recharge areas will have water that has been in the subsurface for a relatively short time so that it is similar to atmospheric water. Over time, rock–water interactions change the chemistry of groundwater so that it is no longer similar to atmospheric water. Examples of chemicals that distinguish older water from water more recently introduced into the ground are environmental isotopes such as tritium and agricultural chemicals (Lloyd et al. 1981), both of which have largely been introduced to the environment in measurable quantities during the last 50 years.

Lawrence and Upchurch (1982) used factor analysis to differentiate waters on the basis of the geochemistry associated with source areas. For example, one type of water measured had high concentrations of calcium and alkalinity, probably resulting from the dissolution of limestone and indicating that the water had been in the aquifer for a relatively long time. Another type of water had constituents derived from leaching through surficial phosphatic sands; thus, the water had recently been introduced to the aquifer as recharge through the sand.

Tritium (^3H) was introduced to the environment in large quantities during atmospheric tests of thermonuclear devices from 1953 to 1969. Water that entered the groundwater flow system prior to the thermonuclear tests should have tritium levels less than 2 tritium units; post-test concentrations may be thousands of tritium units (Hendry 1983). By observing tritium concentrations with depth and accounting for tritium's half-life of 12.3 years, it is sometimes possible to estimate the rate of groundwater recharge on the basis of the depth of the tritium peaks. This method is more effective in silty-clayey soils or in the unsaturated zone where flow is downward rather than in sandy soils or where horizontal flow is predominant.

The concentration of carbon-14 in dissolved organic carbon can also be used to estimate the age of groundwater. Large volumes (20 to 200 liters) of water are needed to obtain sufficient carbon for the analysis, and an estimate of inorganic carbon that entered groundwater by dissolution of calcite and dolomite must be obtained for this method to be effective (Freeze and Cherry 1979). In many cases, it is also necessary to estimate the input of inorganic carbon from the bacterial decay of buried organic matter (Hackley et al. 1991).

RECHARGE IN TYPICAL ILLINOIS TERRANES

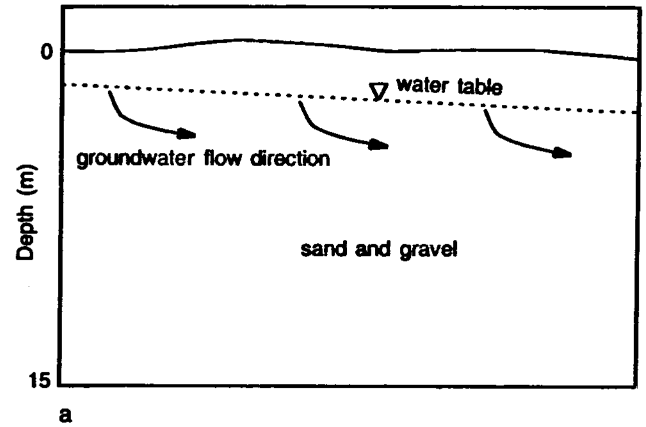
A geologic terrane is a physiographic feature with a predictable sequence of geologic materials (Kempton and Cartwright 1984). Accordingly, any given terrane should have relatively uniform and predictable general recharge characteristics. Several geologic terranes common to Illinois can be distinguished on the basis of their potential for recharge. Other factors such as precipitation patterns and land use are ignored in this generalized analysis, but must be taken into account when estimating the recharge potential of a specific area. Kempton and Cartwright (1984) used a similar approach when mapping recharge in northeastern Illinois.

Areas with High Potential for Recharge

High rates The potential for recharge will be greatest in terranes where water from precipitation or irrigation will rapidly infiltrate to the saturated zone. Two such terranes are common in Illinois.

Generally level or gently sloping uplands with coarse-grained sand and gravel or highly permeable, possibly fractured, bedrock at or near the surface, and overlain by coarse-grained soils Infiltration capacity in these areas will be high, so little runoff will occur (fig. 18a). If the water table is not excessively deep, most precipitation not used by plants will eventually reach the saturated zone. In North Dakota, Rehm et al. (1982) observed that the depth to the water table in such areas was shallow because the volume of recharge was high. This relationship should also hold throughout Illinois, which has a more humid climate than North Dakota.

Areas of karst topography Karst terranes commonly occur in upland areas and are typified by numerous sinkholes (fig. 18b).



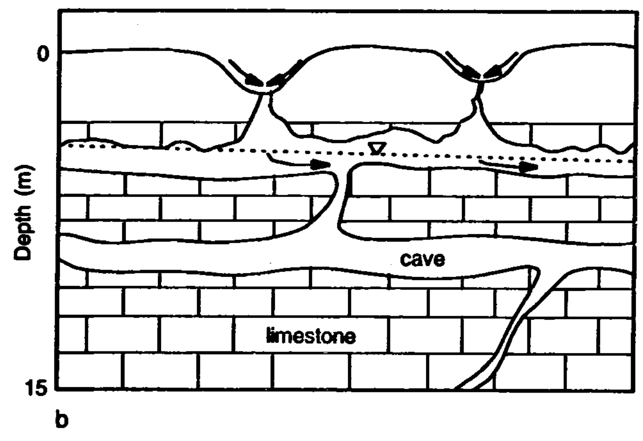


Figure 18 Hydrogeologic terranes with high potential for recharge at high rates: (a) uplands with thick surficial sand deposits, and (b) karst areas.

Large volumes of water can flow through the caverns in the carbonate bedrock, so there are usually few surface water drainage features other than the rivers where the water from the caverns discharges. Often, karst aquifers are overlain by a thin mantle of soil that may have high permeability. Runoff in karst areas typically flows into sinkholes and directly recharges the carbonate aquifer.

Low to moderate rates In Illinois, recharge will occur in most upland areas, regardless of the infiltration capacity of the soil, although the rate of recharge will be affected by factors such as slope and soil permeability. Three examples of these terranes are listed below. Recharge will probably occur in these terranes; however, the rate of recharge will be less than in the terranes described in the previous section.

Generally level or gently sloping uplands with fine-grained clay or silty materials at the surface Infiltration capacities in these soils are low, so some water will run off into depressions where evaporation as well as additional infiltration will occur (fig. 19a). Water tables in areas with thick sequences of fine-grained materials are generally shallow (Rehm et al. 1982) and may be inverted. Because infiltration rates are low, infiltrating water is more likely to be returned to the atmosphere through evaporation and transpiration from fine-grained soils than it is from sandy soils; thus, less water is available for recharge.

Generally level or gently sloping uplands where highly permeable materials are overlain by a thin layer of low -permeability, fine-grained ,or poorly drained soils Rehm et al. (1982) observed that a layer of fine-grained materials overlying coarse-grained materials greatly reduced recharge, as compared with a continuous profile of coarse-grained materials (fig. 19b). The low hydraulic conductivity of the fine-grained surficial material will reduce the infiltration rate, increasing the likelihood of water being returned to the atmosphere through evapotranspiration. Also, the water table in such a sequence is likely to be deep, further reducing the potential for recharge because water percolating downward through a deep, unsaturated soil profile will replenish soil-moisture deficits.

Dissected uplands with generally steep slopes underlain by thick sequences of coarse-grained materials or a thin mantle of soil over permeable bedrock Infiltrating water may replenish moisture deficits in the thick unsaturated zone that often exists under the upper slopes; thus, the amount of water potentially available for recharge is reduced (fig. 19c). Recharge rates may be high at the base of the slope because the water table will generally be closer to land surface, and because runoff may pond at the base of the slope.

Areas with Moderate Potential for Recharge

Some terranes have transitory recharge characteristics. They may alternate between recharge and discharge areas depending on seasonal or diurnal patterns of precipitation and evapotranspiration or variation in topography. Patterns of recharge in the four terranes described below are generally variable, as opposed to the fairly constant recharge of the previously described terranes.

Slopes with thick, coarse-grained deposits adjacent to lowlands Recharge characteristics at the top of the slope will be similar to that of the terrane in figure 19c, but the base of the slope will probably be a groundwater flowthrough or discharge area (fig. 20a). Recharge at the base will be unlikely except during periods of intense precipitation.

Slopes with a thin layer of coarse-grained, highly permeable materials overlying less permeable materials Subsurface storm flow often occurs on slopes of this type. Infiltrating water flows along the contact of the low-permeability layer, discharging to a stream (fig. 20b). Significant recharge in such areas is likely to occur only during prolonged periods of upper soil saturation.

Dissected uplands and steeply sloping areas with thick, fine-grained deposits at the surface The fine-grained materials typically have low hydraulic conductivity and infiltration capacity, causing very low infiltration rates and a tendency for ponding and runoff (fig. 20c). Rills, gullies, and the other surface drainage features that are common on such slopes will carry surface runoff to streams and lowlands; thus, less water is available for recharge. If gullies are deeply incised, water that recharges elsewhere on a slope may discharge into gullies.

Lowlands, characterized by streams, rivers, and lakes, in areas with coarse-grained materials or highly permeable bedrock at or near the surface These will usually be areas of groundwater flowthrough or discharge (fig. 20d). High influxes of precipitation, however, may cause a

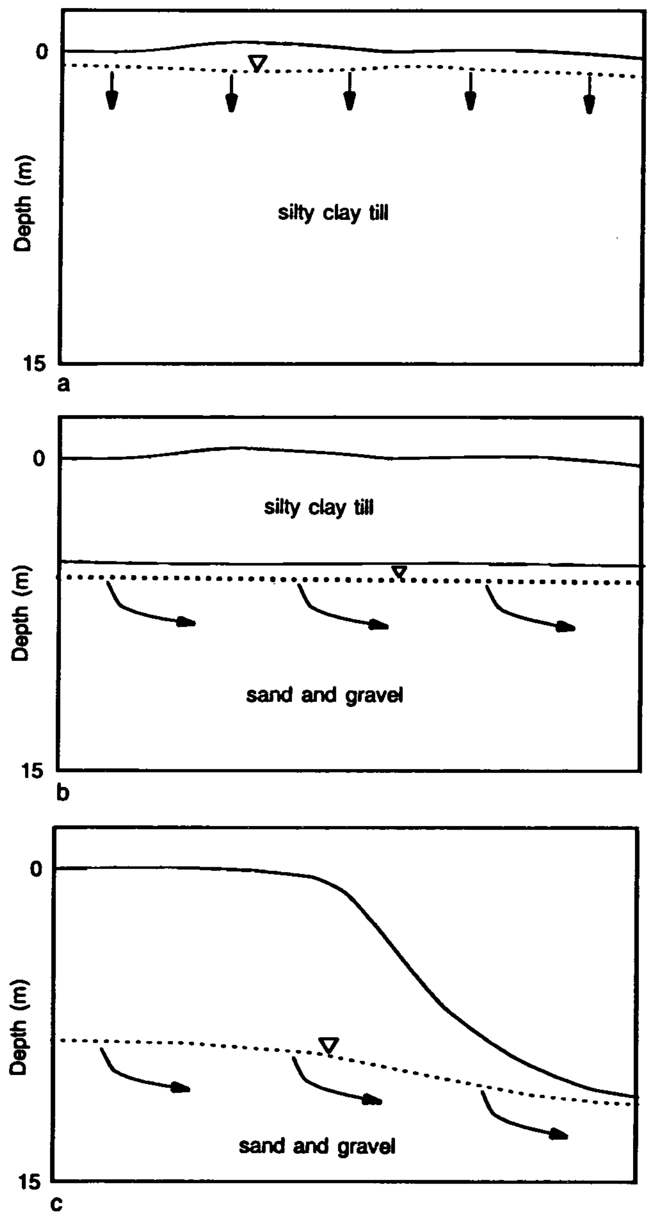


Figure 19 Hydrogeologic terranes with high potential for recharge at low to moderate rates: (a) uplands with surficial deposits of silt or clay and a shallow water table; (b) upland sand and gravel deposits overlain by thin clay or silt; and (c) upland slopes with coarse sand and gravel deposits.

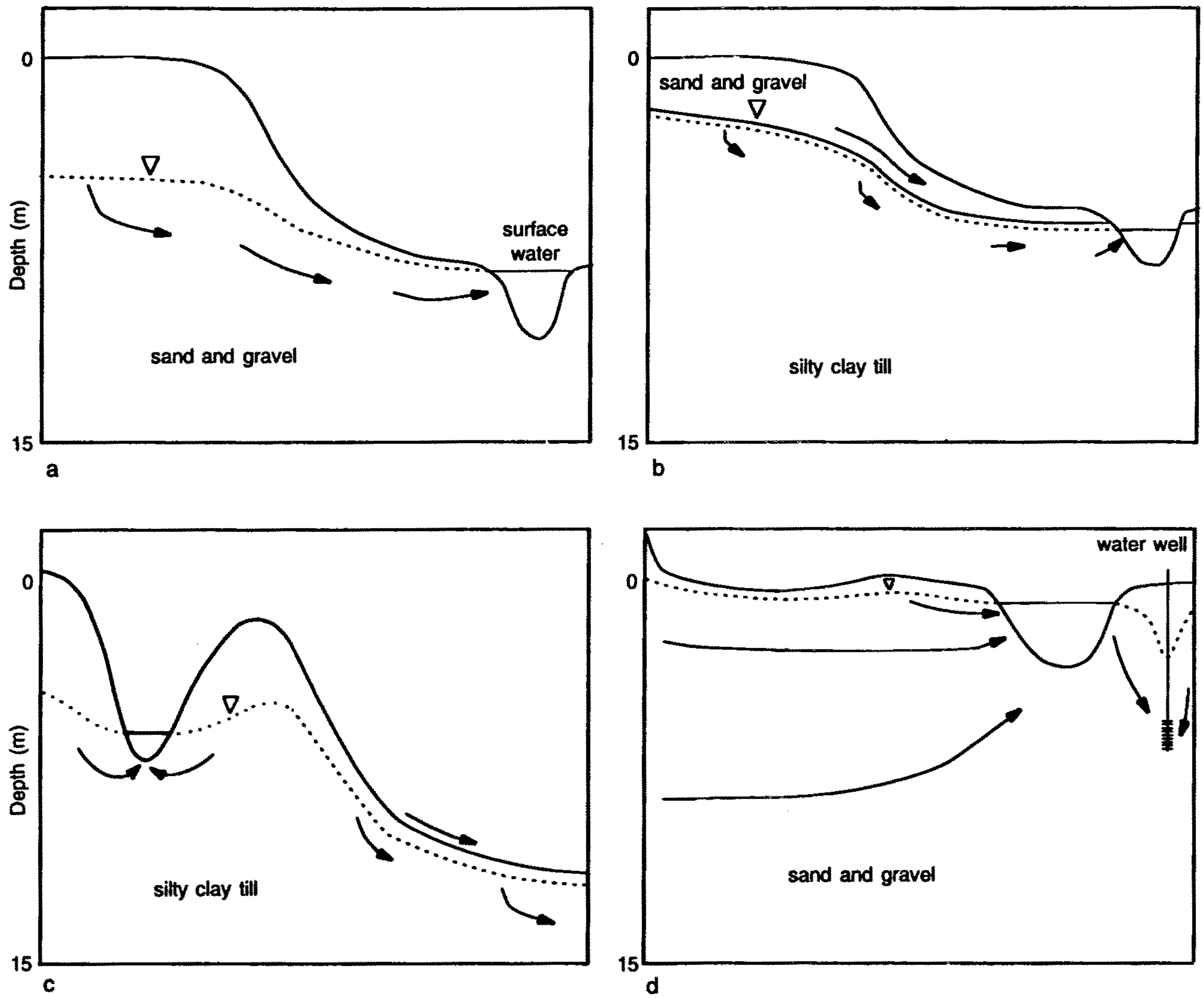


Figure 20 Hydrogeologic terranes with moderate potential for recharge: (a) slopes with sand and gravel; slopes with thin sand and gravel overlying finer grained deposits; (c) slopes consisting of clay or silt deposits; and (d) lowland areas with thick deposits of coarse-grained sand and gravel.

temporary gradient reversal so that groundwater recharge occurs. Gradient reversals may even occur in rivers where rapid rises in the water level often cause recharge of river water to the adjoining banks. The cone of depression caused by pumping wells can also cause such lowlands to be areas of groundwater recharge. In many cases, pumping cones have been known to cause induced recharge from lakes and rivers.

Areas with Low Potential for Recharge

Lowlands, characterized by rivers, lakes, and marshes, with thick deposits of fine-grained materials (fig. 21) are almost always areas of groundwater flowthrough or discharge. The hydraulic gradient is typically upward, and the water table is commonly within 1 m (3 ft) of the ground surface. High precipitation events generally cause ponding, but the ponded water typically runs off to surface water bodies rather than infiltrates surface materials. In areas with poorly developed surface water drainage, or in closed depressions, marshes may develop.

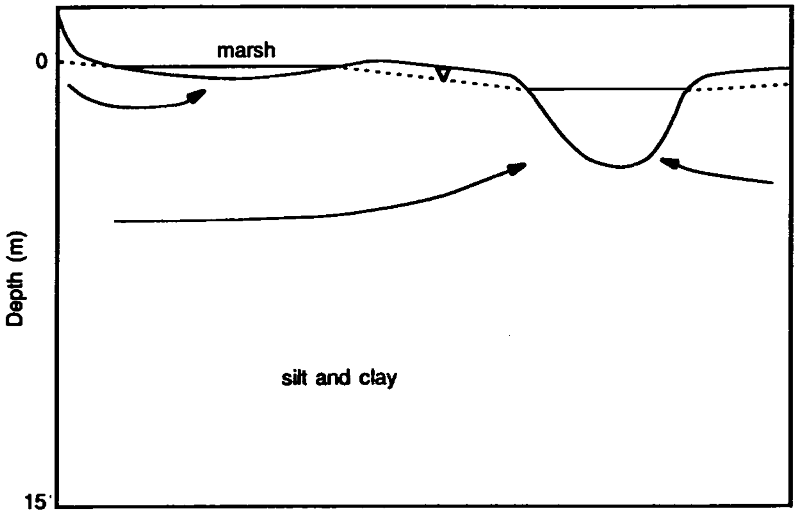


Figure 21 Hydrogeologic terranes with low potential for recharge. Lowland areas with surficial deposits of silt and clay.

Example of Recharge Terrane

Figure 22 shows the distribution of these terranes in an east–west cross section across McHenry County, Illinois. This cross section from Specht and Westerman (1976) illustrates how the terrane concept may be used to map recharge potential in Illinois on a regional scale.

SUMMARY

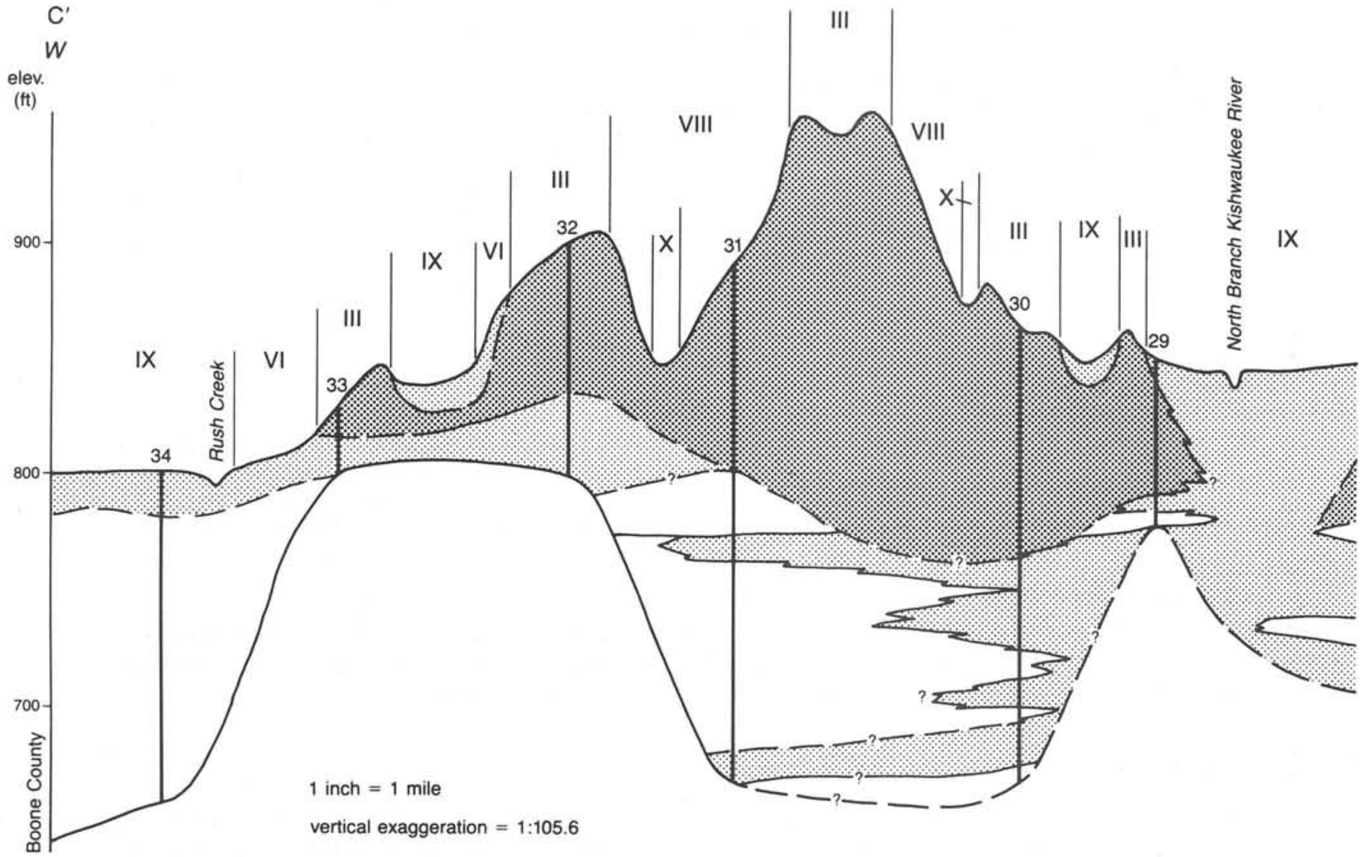
Recharge and infiltration are related hydrological processes. Infiltration occurs when water is supplied to the ground surface by rainfall, snow melt, or irrigation. If the rate of water supply is lower than the infiltration capacity of the soil, water will infiltrate without ponding, otherwise some water will infiltrate and some will pond or runoff. The moisture content of the soil will increase and the infiltration capacity will decrease as infiltration occurs. Thus, ponding and runoff will be more likely as the duration of precipitation and infiltration increases. If the soil becomes saturated, the infiltration capacity will be a function of the saturated hydraulic conductivity of the soil. Infiltration into coarse-grained sand and gravel is generally greater than into fine-grained silt and clay. In fine-grained materials, infiltration through macropores may occur during heavy rain or snow melt when the macropores are in contact with free-standing water such as a temporary pond or saturated soil.

Once water has infiltrated, it may return to the atmosphere as evapotranspiration, replenish moisture deficits in unsaturated soil/rock, flow toward a nearby surface water body as subsurface storm flow, or continue downward to recharge groundwater.

Groundwater recharge does not occur during every infiltration event. It will most likely occur when the soil is wet because less water is used to replenish moisture deficits. Other factors affect the rate and probability of groundwater recharge in a given area:

- hydrogeologic characteristics of the geologic materials in the unsaturated zone,
- depth to the water table,
- hydraulic gradient in the unsaturated zone,
- precipitation and evapotranspiration patterns,
- land use.

In Illinois, recharge most commonly occurs in late winter and early spring when precipitation is plentiful, evapotranspiration is low, and soil moisture is high. Recharge occurs infrequently during the summer because evapotranspiration commonly uses more soil water than precipitation supplies.



Areas with high potential for recharge at high rates

I — Generally level or gently sloping uplands with coarse-grained sand and gravel or highly permeable, possibly fractured bedrock at or near the surface, and overlain by coarse-grained soils

II — Karst topography (none present on figure)

Areas with high potential for recharge at low to moderate rates

III — Generally level or gently sloping uplands with fine-grained clay or silty materials at the surface

IV — Generally level or gently sloping uplands where highly permeable materials are overlain by a thin layer of low permeability, fine-grained or poorly drained soils (none present on figure)

V — Dissected uplands with generally steep slopes underlain by thick sequences of coarse-grained materials or a thin mantle of soil over permeable bedrock

Areas with moderate potential for recharge

VI — Slopes with thick, coarse-grained deposits adjacent to lowlands

VII — Slopes with a thin layer of coarse-grained, highly permeable materials overlying less permeable material (none present on figure)

VIII — Dissected uplands and steeply sloping areas with thick, fine-grained deposits at the surface

IX — Lowlands, characterized by streams, rivers, and lakes with coarse-grained materials and/or highly permeable bedrock at or near the surface

Areas with low potential for recharge

X — Lowlands, characterized by rivers, lakes, and marshes, with thick deposits of fine-grained materials

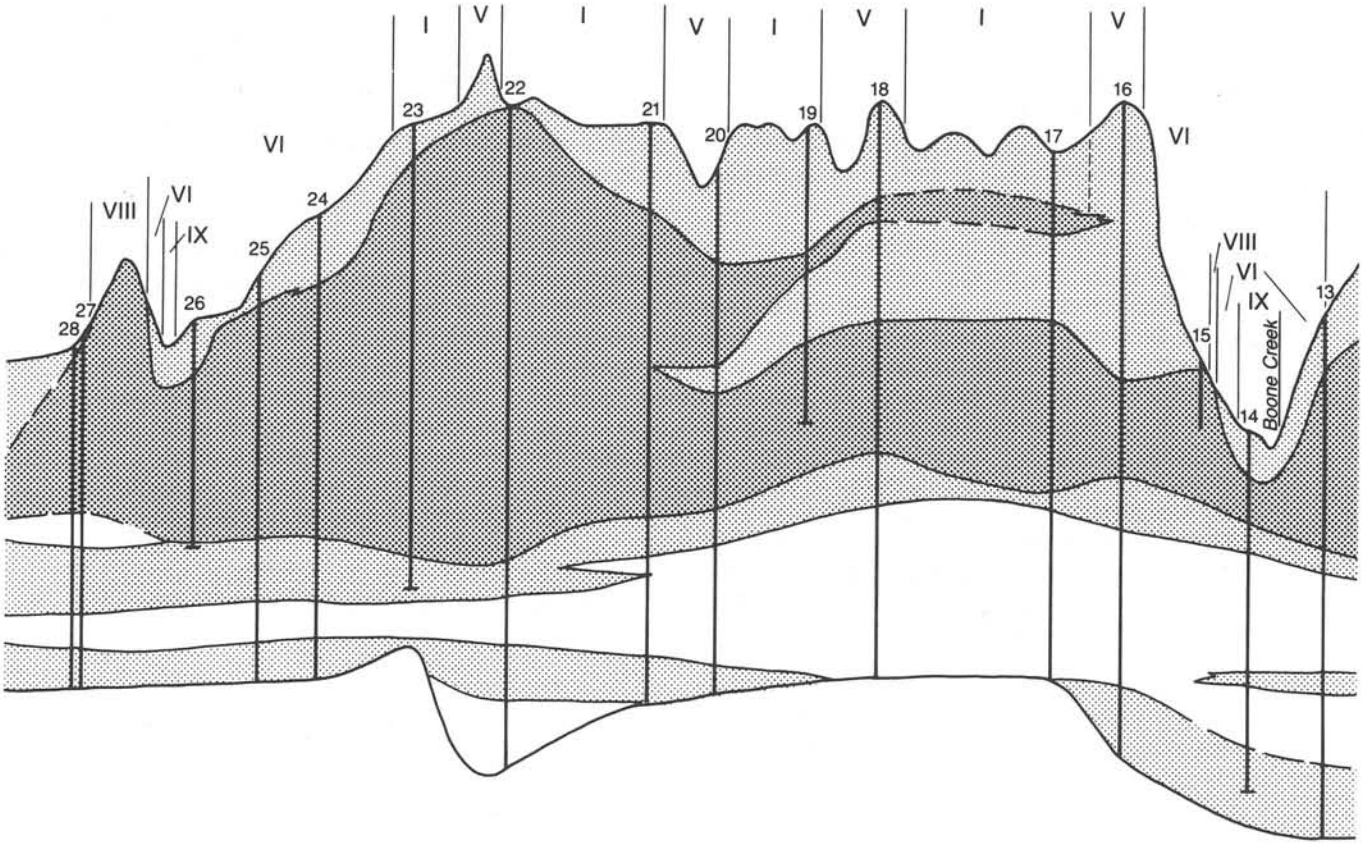
Sandy/gravelly materials that typically have high hydraulic conductivity and permeability

Clayey/silty materials that typically have low hydraulic conductivity and permeability

Organic materials (e.g., peat and marl) that typically have low hydraulic conductivity and permeability

Undetermined materials

Figure 22 Geologic cross section through McHenry County, Illinois, shows distribution and relationship of recharge terranes (cross section from Specht and Westerman 1976).



I
V
I
V
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VI
VIII
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IX
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VIII
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Boone Creek
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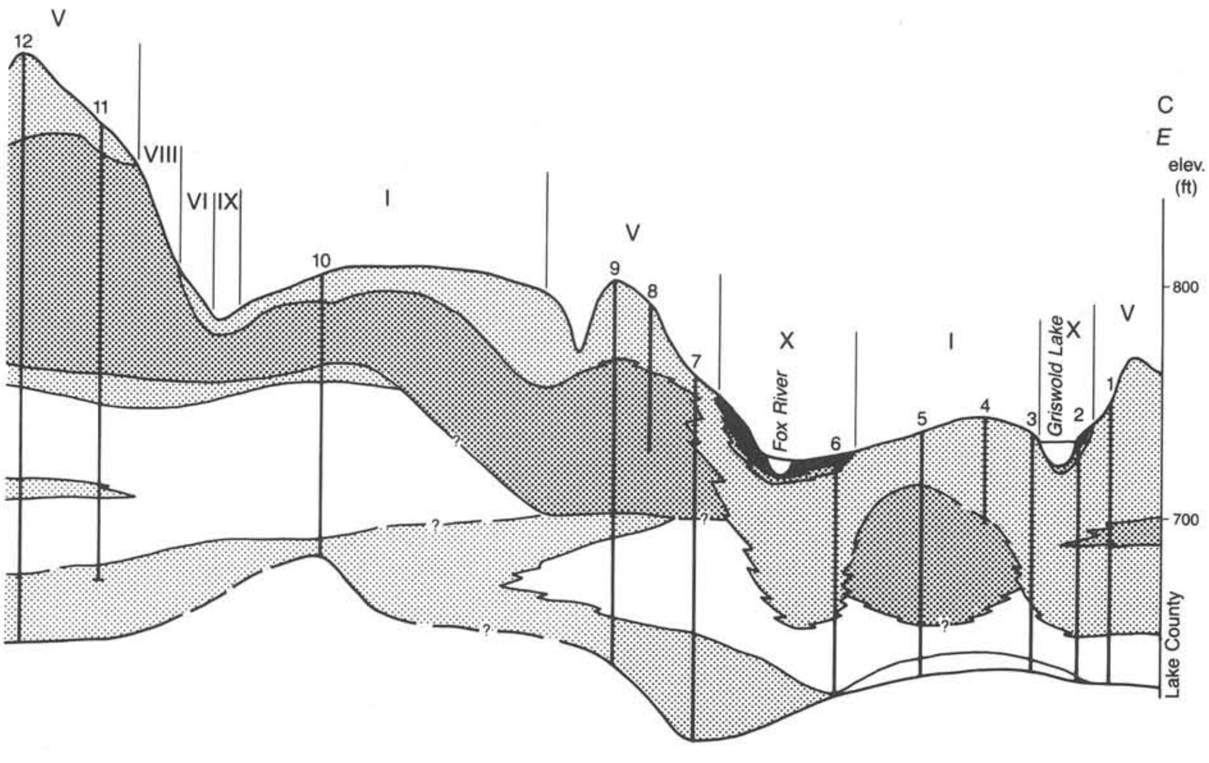

V
12
11
VIII
VI
IX
I
10
V
9
8
7
X
Fox River
6
I
5
4
3
Griswold Lake
X
2
V
1
C
E
elev. (ft)
800
700
Lake County
?

Groundwater recharge will typically occur on uplands where hydraulic head is high and hydraulic gradients are downward. The rate of recharge will be greatest in areas with thick deposits of coarse sand and gravel or in areas of karst topography. Lowlands, characterized by marshes, lakes, and rivers, where hydraulic head is low and hydraulic gradients are upwards, are typically areas of groundwater discharge. Localized recharge may occur in some lowlands, usually where formations of sand and gravel are thick, especially if a water supply well is finished in the formation.

No single, easy, and accurate method can be used to directly measure the rate or amount of recharge entering the groundwater flow system. Variations in patterns of precipitation and geologic and soil conditions reduce even the most exacting methods to mere estimates of recharge. Direct measurements of point recharge may be made with lysimeters. Indirect estimates may be made by calculating the Darcy flux in the saturated or unsaturated zone or by observing fluctuations of the water table in an observation well. Indirect measurements of areal recharge over a basin may be made by measuring the rise of the water table over a long time, calculating a water budget, or by stream hydrograph base-flow separation. Numerical models and analysis of groundwater chemistry can be used to estimate recharge at point and basinwide scales. Point measurements are more exacting than basin measurements and thus are generally more accurate; however, many point measurements in all types of terranes are needed to estimate basin-wide recharge. Basin measurements are more prone to error and do not reflect differences in recharge characteristics that may occur from location to location within the basin, but basin measurements are still the most practical way to estimate regional recharge rates and volumes.

REFERENCES

Anderson, M. W., and R. F. Giovannelli, 1980, A digital overlay technique for evaluation of potential infiltration and recharge: Water Resources Bulletin, v. 16, p. 93–99.

Bowman, J. A., and M. A. Collins, 1987, Impacts of Irrigation and Drought on Ground-Water Resources: Illinois State Water Survey, Report of Investigations 109, 31 p.

Cartwright, K., T. H. Larson, B. L. Herzog, T. M. Johnson, K. A. Albrecht, D. L. Moffett, D. A. Keefer, and C. J. Stohr, 1988, Trench Covers to Minimize Infiltration at Waste Disposal Sites: Illinois State Geological Survey, Circular 541, 88 p.

Cedergren, H. R., 1989, Seepage, Drainage, and Flow Nets, third edition: John Wiley, New York, 465 p.

Cherkauer D. A., and J. M. McBride, 1988, A remotely operated seepage meter for use in large lakes and rivers: Ground Water, v. 26, p. 165–171.

Dunne, T., 1978, Field studies of hill slope flow processes, *in* M. J. Kirkby, editor, Hillslope Hydrology: John Wiley, New York, p. 227–293.

Ehlers, W., 1975, Observations on earthworm channels and infiltration on tilled and untilled loess soil: Soil Science, v. 119, p. 242–250.

Freeze, R. A., 1969, Mechanism of natural ground-water recharge and discharge, 1. one-dimensional, vertical, unsteady, unsaturated flow above a recharging or discharging ground-water flow system: Water Resources Research, v. 5, no. 1, p. 153–171.

Freeze, R. A., and J. A. Cherry, 1979, Groundwater: Prentice-Hall, Englewood Cliffs, NJ, 604 p.

Freeze, R. A., and P. A. Witherspoon, 1968, Theoretical analysis of regional ground water flow, 3. quantitative interpretations: Water Resources Research, v. 4, no. 3, p. 581–590.

Frind, E. O., W. Gilham, and J. P. Pickens, 1977, Application of unsaturated flow properties in the design of geologic environments for radioactive waste storage facilities, *in* W.G. Gray, G.F. Pinder, and C.A. Breddia, editors, Finite Elements in Water Resources: Pentech Press, Plymouth, England, p. 3.133–3.163.

Gerhart, J. M., 1986, Ground-water recharge and its effects on nitrate concentration beneath a manured field site in Pennsylvania: Ground Water, v. 26, p. 483–489.

Germann, P. F., and K. Beven, 1985, Kinematic wave approximation to infiltration into soils with sorbing macropores: Water Resources Research, v. 21, no. 7, p. 990–996.

Glass, R. J., T. S. Steenhuis, and J.-Y. Parlange, 1988, Wetting front instability as a rapid and far-reaching hydrologic process in the vadose zone: Journal of Contaminant Hydrology, v. 3, p. 207–226.

Grindley, J., 1967, The estimation of soil-moisture deficits: Meterological Magazine, v. 96, p. 97–180.

Grindley, J., 1969, The calculation of actual evaporation and soil moisture deficits over specified catchment areas: Meterol. Off. Bracknell, Hydrology Mem., no. 38, 3 p.

Hackley, K. C., C. L. Liu, and D. D. Coleman, 1992, Radiocarbon dating of groundwater containing microbial methane: Radiocarbon, in press.

Hendry, M. J., 1983, Groundwater recharge through a heavy textured soil: Journal of Hydrology, v. 63, p. 201–209.

Hillel, D., 1980, Applications of Soil Physics: Academic Press, New York, 385 p.

Hillel, D., 1982, Introduction to Soil Physics: Academic Press, New York, 364 p.

Horton, R. E., 1940, An approach toward a physical interpretation of infiltration capacity: Soil Science Society of America Proceedings, v. 5, p. 399–417.

Houston, J., 1990, Rainfall - runoff - recharge relationships in the basement rocks of Zimbabwe, *in* D. L. Lerner, A. S. Issar, and I. Simmers, editors, Groundwater Recharge: a Guide to Understanding and Estimating Natural Recharge: International Association of Hydrogeologists, International Contributions to Hydrogeology, v. 8, p. 271–283.

Hughes, G. M., J. A. Schleicher, and K. Cartwright, 1976, Supplement to the Final Report on the Hydrogeology of Solid Waste Disposal Sites in Northeastern Illinois: Illinois State Geological Survey, Environmental Geology Notes, 80, 24 p.

Johansson, P.-O., 1987, Estimation of groundwater recharge in sandy till with two different methods using groundwater level fluctuations: Journal of Hydrology, v. 90, p. 183–198.

Kempton, J. P., and K. Cartwright, 1984, Three-dimensional geologic mapping; a basis for hydrogeologic and land-use evaluations: Bulletin of the Association of Engineering Geologists, v. 21, p. 317–335.

Kitching, R., and L. Bridge, 1974, Lysimeter installations in the sandstone at Styrrup, Nottinghamshire: Journal of Hydrology, v. 23, 219–232.

Kitching, R., T. R. Shearer, and S. L. Shedlock, 1977, Recharge to the Bunter Sandstone determined from lysimeters: Journal of hydrology, v. 33, p. 217–232.

Lawrence F. W., and S. B. Upchurch, 1982, Identification of recharge areas using geochemical factor analysis: Ground Water, v. 20, p. 680–687.

Lehr, J. H., 1988, The misunderstood world of unsaturated flow: Groundwater Monitoring Review, v. 8, no. 2, p. 4–6.

Lerner, D. L., 1990, Part III: Techniques, *in* D. L. Lerner, A. S. Issar, and I. Simmers, editors, Groundwater Recharge: a Guide to Understanding and Estimating Natural Recharge: International Association of Hydrogeologists, International Contributions to Hydrogeology, v. 8, p. 101–228.

Lloyd, J. W., D. Harker, and R. A. Baxendale, 1981, Recharge mechanisms and groundwater flow in the chalk and drift deposits of east Anglia: Quarterly Journal of Engineering Geology, v. 14, no. 2, p. 87–96

Meyboom, P., 1966, Unsteady groundwater flow near a willow ring in hummocky moraine: Journal of Hydrology, v. 4, p. 38–62.

Mills, J. G., and M. A. Zwarich, 1986, Transient groundwater flow surrounding a recharge slough in a till plain: Canadian Journal of Soil Science, v. 66, no. 1, p. 121–134.

Musgrave, G. W., 1955, How much rain enters the soil?, *in* A. Stefferud, editor, Water—the Yearbook of Agriculture: U.S. Department of Agriculture, p. 151–159.

Nielsen, G. L., and J.N. Widjaya, 1989, Modeling of ground-water recharge in southern Bali, Indonesia: Ground Water, v. 27, p. 473–480.

O'Hearn, M., and J. P. Gibb, 1980, Groundwater Discharge to Illinois Streams: Illinois State Water Survey, Contract Report 246, 32 p.

Penman, H. L., 1950, The water balance of the Stour catchment area: Journal of the Institute of Water Engineering, v. 4, p. 457–569.

Philip, J. R., 1991, Hillslope Infiltration: planar slopes: Water Resources Research, v. 27, p. 109–117.

Rehm, B. W., G. H. Groenewold, and W. M. Peterson, 1982, Mechanisms, distribution, and frequency of groundwater recharge in an upland area of western North Dakota: North Dakota Geological Survey, Report of Investigations 75, 72 p.

Rushton, K. R., and C. Ward, 1979, The estimation of groundwater recharge: Journal of Hydrology, v. 41, p. 345–361.

Schicht, R. J., and W. C. Walton, 1961, Hydrologic Budgets for Three Small Watersheds in Illinois: Illinois State Water Survey, Report of Investigations 40, 40 p.

Smettem K.R.J., and N. Collis-George, 1985, The influence of cylindrical macropores on steady-state infiltration in a soil under pasture: Journal of Hydrology, v. 79, p. 107–114.

Specht, S. A., and A. A. Westerman, 1976, Geology for planning in Northeastern Illinois, Part III, Geology for planning in McHenry County: Illinois State Geological Survey, Open File Report 1976-3, 36 p.

Startz, R. S., 1969, Soil-water movement as affected by deep freezing: Proceedings of Soil Science Society of America, v. 33, p. 333–337

Steenhuis, T. S., C. D. Jackson, S.K.J. Kung, and W. Brutsaert, 1985, Measurement of recharge on eastern Long Island, New York, U.S.A.: Journal of Hydrology, v. 79, p. 145–169.

Stoertz, M.W., and K.R. Bradbury, 1989, Mapping recharge areas using a ground-water flow model - a case study: Ground Water, v. 27, pp. 220–228.

Storey, H. C., 1955, Frozen soil and spring and winter floods, *in* A. Stefferud, editor, Water—The Yearbook of Agriculture. U.S. Department of Agriculture, p. 179–184.

USEPA, 1990, Handbook—ground water, volume 1: ground water and contamination: United States Environmental Protection Agency, Center for Environmental Research Information, Cincinnati, Ohio, EPA/625/6-90/016a, 144 p.

Walton, W. C., 1965, Ground-water Recharge and Runoff in Illinois: Illinois State Water Survey, Report of Investigations 48, 55 p.

Williams, R. E., 1968, Flow of groundwater adjacent to small, closed basins in glacial till: Water Resources Research, v. 4, p. 777–783.

Williams, R. E., and R. N. Farvolden, 1967, The influence of joints on the movement of ground water through glacial till: Journal of Hydrology, v. 5, p. 163–170.

Winter, T. C., 1981, Uncertainties in estimating the water balance of lakes: Water Resources Bulletin, v. 17, p. 82–115.

Zebarth, B. J., E. De Jong, and J. L. Henry, 1989, Water flow in a hummocky landscape in central Saskatchewan, Canada, II. saturated flow and groundwater recharge: Journal of Hydrology, v. 110, p. 181–198.

APPENDIX: GLOSSARY

Adsorption As used in this text, adsorption is the attraction of water molecules to soil and water particles caused by hydrogen bonding.

Aquifer A unit of soil or rock capable of supplying adequate water to meet the demands of a given user.

Aquitard A unit that retards but does not prevent the flow of water to or from an aquifer; also known as a confining zone. In Illinois, aquitards are usually composed of continuous layers of fine-grained silty to clayey till and lacustrine deposits, shale, and massive to slightly fractured limestone and dolomite.

Base flow The portion of stream flow that comes from groundwater, as opposed to the portion from runoff.

Capillarity The surface tension of a liquid that causes it rise or fall in contact with a solid.

Cone of depression The drawdown in the water table or potentiometric surface caused by withdrawal of water at a well.

Confining zone See aquitard.

Darcy flux The volumetric flow of groundwater across a given area.

Elevation head One of the components of hydraulic head. It is a measure of the potential for water to move downward in response to gravity, expressed with reference to a standard datum such as sea level; it is sometimes referred to as gravity head.

Evapotranspiration The combined loss of water from a land area through evaporation from the soil surface and transpiration of plants.

Fingers Saturated water columns through which water moves downward in otherwise unsaturated soil.

Flow net A two-dimensional graphical representation of groundwater flow.

Flux The volumetric discharge of water over a unit area.

Gravity head See elevation head.

Groundwater discharge The movement of water from saturated geologic materials to the ground surface.

Groundwater recharge The advance of water into the saturated zone associated with groundwater flow away from the water table and within the saturated zone (Freeze and Cherry 1979).This definition identifies groundwater recharge as occurring when infiltrating water passes the water table, regardless of geologic and aquifer characteristics.

Hydraulic conductivity The ability of a soil/rock unit to transmit water. The hydraulic conductivity of a saturated material is fairly constant over time; however, the hydraulic conductivity of an unsaturated soil is a function of its moisture content. Hydraulic conductivity has distinct directional components and may vary from place to place within a formation.

Hydraulic gradient The change in hydraulic head over a given distance.

Hydraulic head A measure of the potential energy of groundwater at a given point beneath the surface. Groundwater in both the saturated and unsaturated zones moves from points of high head to points of low head. Hydraulic head has three components: pressure head, elevation head, and velocity head.

Hydrologic cycle The constant circulation of water as a gas, liquid, and solid from the atmosphere to land and back again through evaporation, transpiration, precipitation, runoff, infiltration, recharge, and discharge.

Infiltration The entry and associated downward flow of water from the ground surface into the unsaturated zone.

Infiltration capacity The maximum possible infiltration rate in soil/rock for a given soil-moisture condition; infiltration capacity is reached when the rate of water available at the ground surface exceeds the ability of the material to adsorb water. It is not constant and decreases as the moisture content of the soil increases (wetting) and increases as the soil drys (moisture content decreases).

Infiltration rate The volume flux of water that flows into a soil or rock per unit of ground surface area.

Leakage The movement of water through a confining zone into an aquifer. It may also be considered to be recharge to an aquifer.

Macropores Secondary pore openings, including fractures, animal burrows, and root channels.

Potentiometric surface The areal head distribution of a confined aquifer. Analogous to the water table of an unconfined aquifer.

Precipitation The discharge of water from the atmosphere in the form of rain, snow, hail, or sleet.

Pressure head Hydrostatic pressure expressed as the height of a column of water that the pressure can support. In unsaturated materials, suction forces create a negative pressure head that is usually expressed as tension or suction. In saturated materials, pressure head is positive because the water at any point is subjected to pressure from the weight of the water above.

Rain packing Rain drops falling on unprotected soil or intense tillage can break up soil aggregates, causing a dense soil layer to form. This dense layer will have low hydraulic conductivity and, subsequently, a low infiltration rate.

Recharge See groundwater recharge and recharge to an aquifer.

Recharge to an aquifer This occurs when water enters a hydrostratigraphic unit defined as an aquifer. Groundwater recharge and recharge to an aquifer are the same only when the water table occurs within the aquifer. Recharge to a confined aquifer is also termed leakage.

Runoff The overland flow of water during a precipitation event.

Saturated zone The area of soil and rock where pore spaces are completely filled with water.

Soil horizon A layer of soil, approximately parallel to the soil surface, with distinct characteristics produced by soil-forming processes. The upper two layers are labeled A and B, with B being beneath A.

Soil profile The vertical section of the soil through all its horizons and extending into the parent material.

Soil-moisture budget A procedure to account for all water that enters and exits the soil profile over a given period of time.

Solum The upper and most weathered parts of the soil profile.

Tension A result of suction forces that have two primary components: capillarity and adsorption.

Transpiration The process by which water absorbed by plants is evaporated into the atmosphere from the plant surface.

Unsaturated zone The layer of soil and rock where pore spaces are partially filled with water and partially filled with air.

Velocity head One of the three components of hydraulic head. The velocity head results from the inertial motion of water; because the movement of water is relatively slow, velocity head is negligible compared with the other heads and is usually ignored.

Vertical hydraulic conductivity Hydraulic conductivity in the vertical direction. It may be lower that horizontal hydraulic conductivity in some sedimentary formations.

Water table The surface that divides the saturated from the unsaturated zone. In many lowland situations, the water table is shallow or at the ground surface. Beneath uplands, the water table may be very deep.

Water budget A procedure to account for all water that enters and exits a given area over a given time. For a small basin, the water budget may be: precipitation - runoff - evapotranspiration = groundwater recharge.