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Conservation
Service**

**Part 631
National Engineering Handbook**

Chapter 30 Groundwater Hydrology and Geology

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Preface

The NRCS National Engineering Handbook (NEH), Part 631.30, Groundwater Hydrology and Geology, is derived from the following publication:

NEH Section 18, Ground Water, released by SCS, April 1968.

Note the following changes (the canceled documents are replaced by the new documents):

Canceled documents

- NEH, Section 18, Ground Water (June 1978)
- Technical Release No. 36, Ground Water Recharge (June 1967)
- NEH, Part 631, Chapter 33, Investigations for Ground Water Resources Development (November 1998)

New documents

- NEH, Part 631 chapters:
  - 631.30, Groundwater Hydrology and Geology
  - 631.31, Groundwater Investigations
  - 631.32, Well Design and Spring Development
  - 631.33, Groundwater Recharge

Chapter 30 Groundwater Hydrology and Geology

Contents

Tables

Figures

Chapter 30 Groundwater Hydrology and Geology

631.3000 Introduction

(a) Purpose and scope

The purpose of this chapter is to present information on groundwater as it relates to Natural Resources Conservation Service (NRCS) programs. This material was compiled to help NRCS personnel plan and conduct groundwater studies under established NRCS standards and policies.

Investigations are made to determine the availability and suitability of groundwater for beneficial use and to provide groundwater information needed to plan, design, and construct works of improvement. NRCS does not make groundwater surveys or studies for the sole purpose of collecting basic data.

In many areas, basic groundwater data have been collected and compiled by other Federal, State, and private agencies. This information may be available in published documents or in unpublished field reports. When these data are either insufficient in detail or outdated, the NRCS must conduct further investigations.

Because problems associated with groundwater are complex, a flexible pattern of investigational procedures is needed. Identifiable groundwater characteristics help to determine the best procedures to use.

At present, about 90 percent of the freshwater supply in the United States is from groundwater sources (National Groundwater Association).

(b) Groundwater rights

State laws regarding the use of groundwater must be followed. Three principal sets of rules or "doctrines" form the basis for these laws:

- *Absolute ownership*, or the common-law rule, which states that the owner of the land is the absolute owner of all underground waters under their property. The owner may develop and use their groundwater without regard to effects on groundwater supplies of adjacent landowners, subject to qualifications in some States. This doctrine does not recognize flowing groundwater or the effect that its misuse may have on other landowners using the same source.
- *Ownership with reasonable use* is similar to the common-law ownership, but limits the owner to a reasonable use related to the needs of other owners of lands overlying a common groundwater source.
- The *appropriation doctrine* follows the rule that, where groundwater limits or boundaries can be reasonably established, the subsurface waters are public waters and subject to appropriation. Priority rights are issued from a designated State agency after an examination of the intent of use. The appropriation system emphasizes beneficial use and conservation, security of investment, and responsibility for administrative guidance.

Most States regulate the development and administration of groundwater resources in the public interest.

Common State regulations include the following:

- All persons drilling wells for others must be licensed.
- Drilling permits, logs, and any work performed on wells must be reported to the State on prescribed forms.
- Wells furnishing domestic or municipal water must be properly constructed and finished to prevent contamination.
- Flowing wells must be suitably capped and regulated to avoid waste.
- Abandoned wells must be sealed.
- Air-conditioning and cooling waters must be returned to the ground through recharge wells.
- Disposal of any contaminants, such as brines or industrial wastes, which affect the quality of public water supplies, can be restricted.

631.3001 Groundwater hydrology

(a) The hydrologic cycle

Water exists in the atmosphere, on the Earth's surface (the hydrosphere), and under the Earth's surface (the lithosphere). A continual interchange produces a closed system that is known as the hydrologic cycle.

(b) The hydraulic properties of materials

Water in the lithosphere is in the form of free water, water vapor, or ice, or is chemically combined with earth materials. Uncombined water occupies the void spaces. The interstices may be occupied by air or other gases or by water or other liquids.

A rock or soil is said to be porous or to have porosity if it contains interstices or voids. Porosity can be quantitatively expressed as the ratio of the total volume of voids to the total volume of the rock or soil. It is usually given as a percentage. Primary porosity is a result of the processes that formed rock or soil. Secondary porosity is produced by fracture, solution, or recrystallization. Groundwater occurs in both primary and secondary voids.

Voids can be divided into three classes, based on their size and interaction with water.

- *Capillary*—Groundwater will rise in a capillary interstice by surface tension to a given height above the potentiometric surface. The amount of rise depends on the size, shape, and composition of the walls of the void. The upper limit of capillary width is about 3 millimeters.
- *Supercapillary*—Void spaces too wide to allow the rise of water by capillary action.
- *Subcapillary*—A subcapillary interstice is so small that water in it is held by adhesion to the sides of the interstice. The adhesive forces exceed the cohesive forces of the water. Movement is impossible except by external forces which greatly exceed those normally found in the zone of groundwater.

A rock or soil which has communicating voids of capillary or supercapillary size is said to be permeable. The permeability of a material is its capacity to transmit fluids under pressure.

While a material must have porosity in order to be permeable, there is no direct relationship between total porosity and permeability. This is illustrated in table 30–1.

An **aquifer** is a geologic unit that contains sufficient saturated and permeable material to yield significant quantities of water. Aquifer boundaries do not necessarily coincide with formation boundaries. An aquifer may include an entire formation, a group of formations, or part of a formation. It may consist of earth materials ranging from rock to unconsolidated sediments.

An **aquiclude** is porous and may contain groundwater but will not transmit it fast enough to be of consequence as a water supply. What is considered to be an aquifer in one area may be considered to be an aquiclude in another area because of differences in demand and availability of alternate supplies.

An aquiclude may also be relatively impermeable and form a confining layer, forming a **confined aquifer**.

The **potentiometric surface** is the elevation to which water will rise up in a tube or well that penetrates a confined aquifer.

Table 30–1 Comparison of typical porosities and coefficients of permeability for common materials (from various sources)

Material	Porosity	Coefficient of permeability, $ft^3/ft^2/d$ or ft/d
Poorly graded gravels	30–40	> 3000
Poorly graded sands	30–40	300–3000
Well-graded sands	20–35	15–300
Fine sands	30–35	15–150
Silty sands	30–40	0.3–6
Silts	40–50	0.03–1
Clays	45–60	< 0.003

The water table is more properly termed the potentiometric surface in an unconfined aquifer, also called a water table aquifer.

(c) The occurrence of groundwater

The upper part of the lithosphere consists of soils, regolith (broken, unconsolidated rock) and bedrock. Soil and rock are also collectively known as earth materials. With increased depth, the force applied by the weight of the overlying rock increases (lithostatic force). When lithostatic forces exceed the strength of rock, deformation closes void spaces, becoming impermeable. Weak rocks deform at moderate depths, and even the strongest rocks will deform at greater depths. Figure 30–1 shows divisions of subsurface water.

In the upper part of the lithosphere, where void spaces exist, groundwater may be present in two broad zones: the **zone of saturation** and **zone of aeration.**

(1) Zone of saturation

All interconnected voids are filled with water under hydrostatic pressure. In an unconfined aquifer, the upper surface of this zone is called the water table, which is actually a specific type of potentiometric surface.

(2) Zone of aeration

Voids are filled or partially filled with water, air, or other fluids or gases. Water may be moving downward by gravity, may be stationary and held in place by capillary action, or may be drawn upward by plants or evaporation.

The zone of aeration is divided into three subzones: capillary fringe, zone of soil water, and an intermediate zone.

Capillary fringe—immediately overlying the zone of saturation, and continuous with the water in the zone of saturation, is the capillary fringe. It is held above the zone of saturation by capillary action. All interstices of capillary size may be filled with water, but no hydrostatic head will be exhibited. Since the height to which water is held by capillary action is inversely proportional to the diameter of the interstitial space, the thickness of the capillary fringe varies with the texture of the rock or soil, all other things being equal. If the material has only supercapillary openings, the capillary fringe will be practically absent. If the openings are all subcapillary, the material is impermeable. Most materials have some capillary interstices, however. Some typical values for the height of capillary rise in common materials are presented in table 30–2.

At equilibrium, any water reaching the capillary fringe from above by gravity flow will cause the immediate discharge of an equivalent amount of water to the zone of saturation. In effect, this raises the elevation of the water table.

Zone of soil water—the soil-water zone extends from the surface to slightly below the depth of root penetration. Water in this zone is available for transpiration by plants or for direct evaporation. The soil-water zone is not saturated except temporarily when excess water is applied. Water is held by surface tension and is moved by capillary action. If excess water is available, it drains through the soil under the influence of gravity.

The thickness of the zone varies greatly with different types of soil and vegetation. Water in this zone is commonly called soil water.

Intermediate zone—the zone of soil water and the capillary fringe may or may not be separated by an intermediate zone. The intermediate zone is the residual part of the zone of aeration. It does not exist where the capillary fringe or the water table approaches the surface and may be several hundred feet thick in deep water table areas.

Water in the intermediate zone cannot be brought back to the soil-water zone by capillary action, but it has not yet reached the capillary fringe. It is held in place by surface tension or is moving downward under the force of gravity.

The amount of stationary or pellicular water in the intermediate zone depends on the nature of the soil or rock and is equivalent to the field capacity of the same materials in the soil-water zone and to the specific retention of an aquifer. Water in excess of this amount, if and when it becomes available from the soil zone, moves downward under the force of gravity to the capillary fringe, where it displaces fringe water into the zone of saturation.

Figure 30–1 Divisions of subsurface water

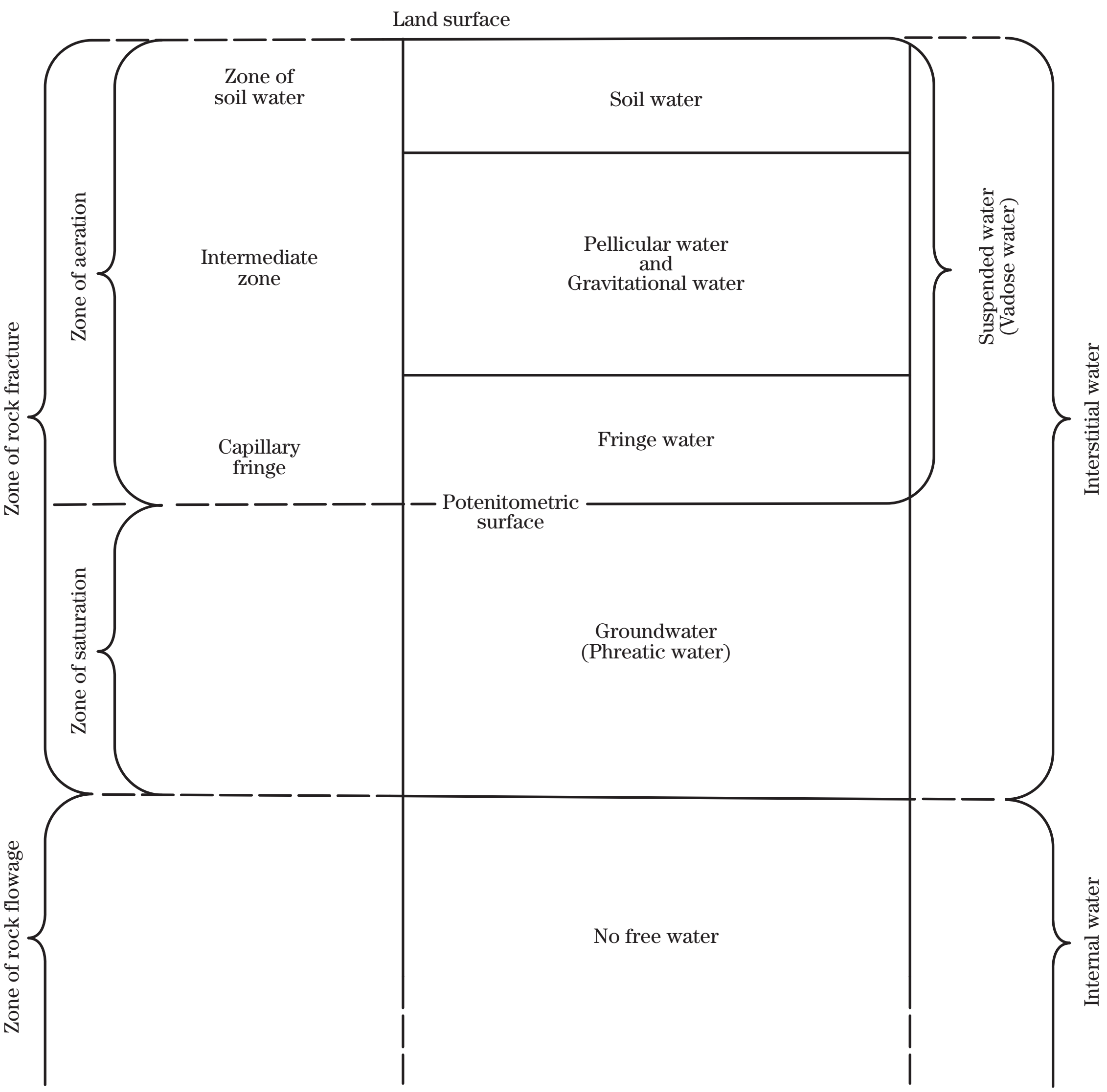


Table 30–2 Estimated values of capillary rise

Soil type	Height of capillary rise, inches
Coarse sand	½–2
Sand	5–14
Fine sand	14–28
Silt	28–60
Clay	80–160 or more

(3) The piezometric surface

The piezometric surface is the elevation or depth of water in a well. The piezometric surface coincides with the water table in an unconfined aquifer. Recharge to an artesian aquifer occurs if the water table is higher than the piezometric surface or if the permeable formation extends above the piezometric surface, in which case, the aquifer is a phreatic aquifer.

Water table—The piezometric surface in an unconfined aquifer is generically called the water table. The shape and slope of the water table depends on areas of recharge, discharge, and changes in hydraulic conductivity. The water table generally mirrors surface topography but with less relief.

Artesian wells—Artesian conditions or artesian pressure exists when the upper surface of the saturated zone is overlain by an aquiclude, and the piezometric surface in a well penetrating the saturated zone is above the surface of the saturated zone. An aquifer such as this is called a confined aquifer or an artesian aquifer, even though water does not necessarily flow at the surface.

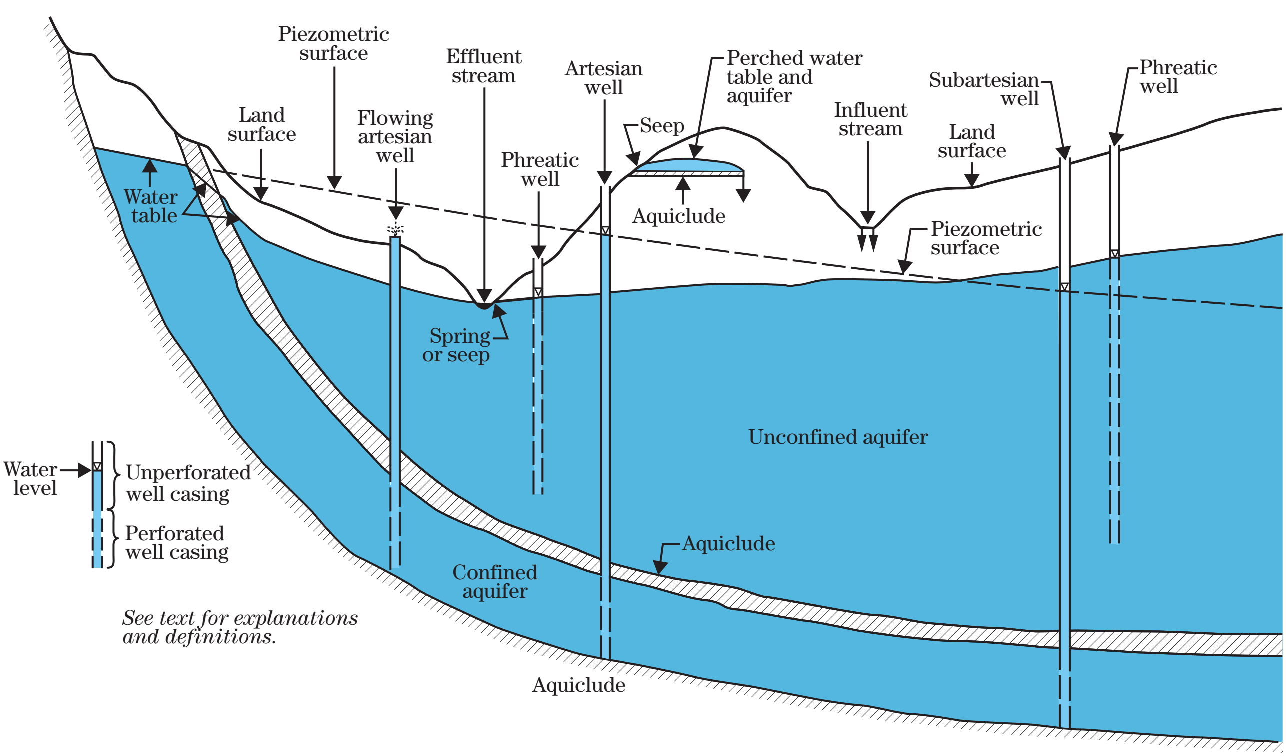
Recharge to an artesian aquifer occurs where the permeable formation is exposed at the surface at an elevation above the piezometric surface. Recharge can also occur where the confining layer is more permeable, and the water table in an overlying aquifer is higher that the piezometric surface in the confined aquifer. The same aquifer may be unconfined in some ares, such as a recharge area, and confined in others. Changes in the elevation of the piezometric surface indicate changes in pressure rather than changes in the volume of water in storage.

The groundwater relationships are illustrated in figure 30–2.

(4) Perched water table

Groundwater is said to be perched if it is separated from the main water table by unsaturated materials. The upper surface of this perched zone of saturation

Figure 30–2 Groundwater relationships



is a perched water table. Water may be perched either temporarily or permanently. It is underlain by a negative confining bed which stops or retards the downward movement of water under the force of gravity. Many islands have perched water tables, which are important sources of fresh water. Fresh groundwater floats as a lens-shaped body on seawater, extending below sea level approximately 40 times the height of the freshwater table above seal level (Ghyben-Herzberg Principle). This 1 to 40 relation occurs because freshwater is slightly less dense than seawater (1,000 grams per cubic centimeter (g/cm^3) versus 1.025 g/cm^3).

(5) Aquifer characteristics

Many major aquifers are composed of unconsolidated materials, chiefly sands and gravels. Less important aquifers are found in heterogeneous alluvial deposits, loess, and glacial till.

Buried preglacial valleys and glacial outwash deposits may also be important groundwater sources.

Sedimentary rocks yield most of their water through joints and fractures and through passages opened by weathering or solution. Some less thoroughly cemented sandstones yield water from primary porosity.

Extrusive igneous rocks yield water from shrinkage cracks, joints, flow breccias, and lava tubes. Clinker zones in basalt flows may be the source of very large amounts of groundwater.

Intrusive igneous rocks and metamorphic rocks are generally unproductive, but may yield small supplies for domestic use from fractures and weathered zones.

(d) Groundwater movement

(1) Darcy's law

In 1856, Henri Darcy observed that the volume of flow through a porous medium when flow is laminar is directly proportional to the head loss and inversely proportional to the length of the flow path. This is now known as Darcy's law. It can be expressed mathematically as:

$$v = \frac{k\Delta h}{\ell} = kI \qquad \text{(eq. 30–1)}$$

where:
v = volume of water per unit cross-sectional area of a column of permeable material, expressed as a velocity (e.g., ft/d)
Δh = difference in pressure head at the ends of the column (ft)
ℓ = length of flow path (ft)
k = coefficient of permeability—a constant based on the character of the material, ($ft^3/ft^2/d$)
I = $\Delta h/\ell$ = hydraulic gradient (dimensionless)

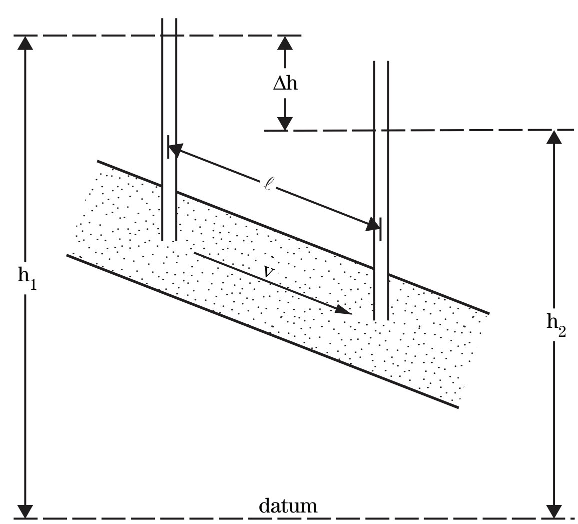
These relationships are illustrated in figure 30–3. In the equation, V represents the rate of motion of a solid column of water the size of the test section.

The actual rate of movement of the water, as measured with dye tracers for instance, is the velocity v in equation 30–1 divided by porosity:

$$V = \frac{v}{p} = \frac{kI}{p} \qquad \text{(eq. 30–2)}$$

where:
V = actual rate of movement (ft/d)
p = effective porosity (dimensionless decimal)
V, k, and I are defined in equation 30–1.

Figure 30–3 Factors involved in Darcy's law



It naturally follows that:

$$k = \frac{pV}{I} \qquad \text{(eq. 30–3)}$$

This is a useful concept in estimating field permeabilities with the use of dye tracers.

(2) Coefficients of hydraulic conductivity and transmissibility

The coefficient of hydraulic conductivity, k, of a given material is the volume of water that will flow through a unit cross-sectional area in unit time, under unit hydraulic gradient and at a standard temperature. NRCS commonly expresses the coefficient of hydraulic conductivity in cubic feet of water per day through a 1-foot-square cross-sectional area of the aquifer, under a hydraulic gradient of 1 foot per foot at a temperature of 20 degrees Celsius. This is:

$$k = \frac{\text{ft}^3 \text{ at } 20° \text{ C}}{(1 \text{ d})(1 \text{ ft}^2)(1 \text{ ft/ft})}$$

The U.S. Geological Survey (USGS) commonly expresses the coefficient of hydraulic conductivity in Meinzer's units. This is in gallons per day at 60 degrees Fahrenheit, with the other terms remaining the same or:

$$Pm = \frac{\text{gal at } 60° \text{ F}}{(1 \text{ d})(1 \text{ ft}^2)(1 \text{ ft/ft})}$$

The viscosity of water is a function of temperature. For this reason, hydraulic conductivity is adjusted to a standard temperature. The temperature range of most aquifers, however, is small, and it is common practice to ignore the temperature correction and compute field coefficients of hydraulic conductivity at the prevailing temperature.

The coefficient of transmissibility, T, is the rate of flow of water through a vertical strip of unit width, which extends the entire saturated thickness of the aquifer under unit hydraulic gradient and at the prevailing water temperature. This is expressed as:

$$T = \frac{\text{ft}^3}{(1 \text{ d})(1 \text{ ft})(\text{thickness-m ft})(1 \text{ ft/ft})}$$

or

$$T = \frac{\text{gal}}{(1 \text{ d})(1 \text{ ft})(\text{thickness-m ft})(1 \text{ ft/ft})}$$

In a uniform aquifer, the coefficient of transmissibility is equal to the field coefficient of hydraulic conductivity times the saturated thickness. Figure 30–4 illustrates the relationships of hydraulic conductivity and transmissibility.

A useful form of Darcy's law is given by the formula:

$$Q = kIA = TIL \qquad \text{(eq. 30–4)}$$

where:
Q = discharge (ft^3/d)
k = coefficient of hydraulic conductivity ($ft^3/ft^2/d$)
I = hydraulic gradient (ft/ft)
A = cross-sectional area through which discharge occurs (ft^2)
T = coefficient of transmissibility ($ft^3/ft/d$)
L = width of section through which discharge occurs (ft)

Coefficients of transmissibility determined in the field can be converted to coefficients of hydraulic conductivity by dividing by the thickness of the aquifer. This represents the overall average hydraulic conductivity of the aquifer tested.

Methods of determining k and T by pump tests are discussed in the section 631.3002(e), Hydraulics of wells.

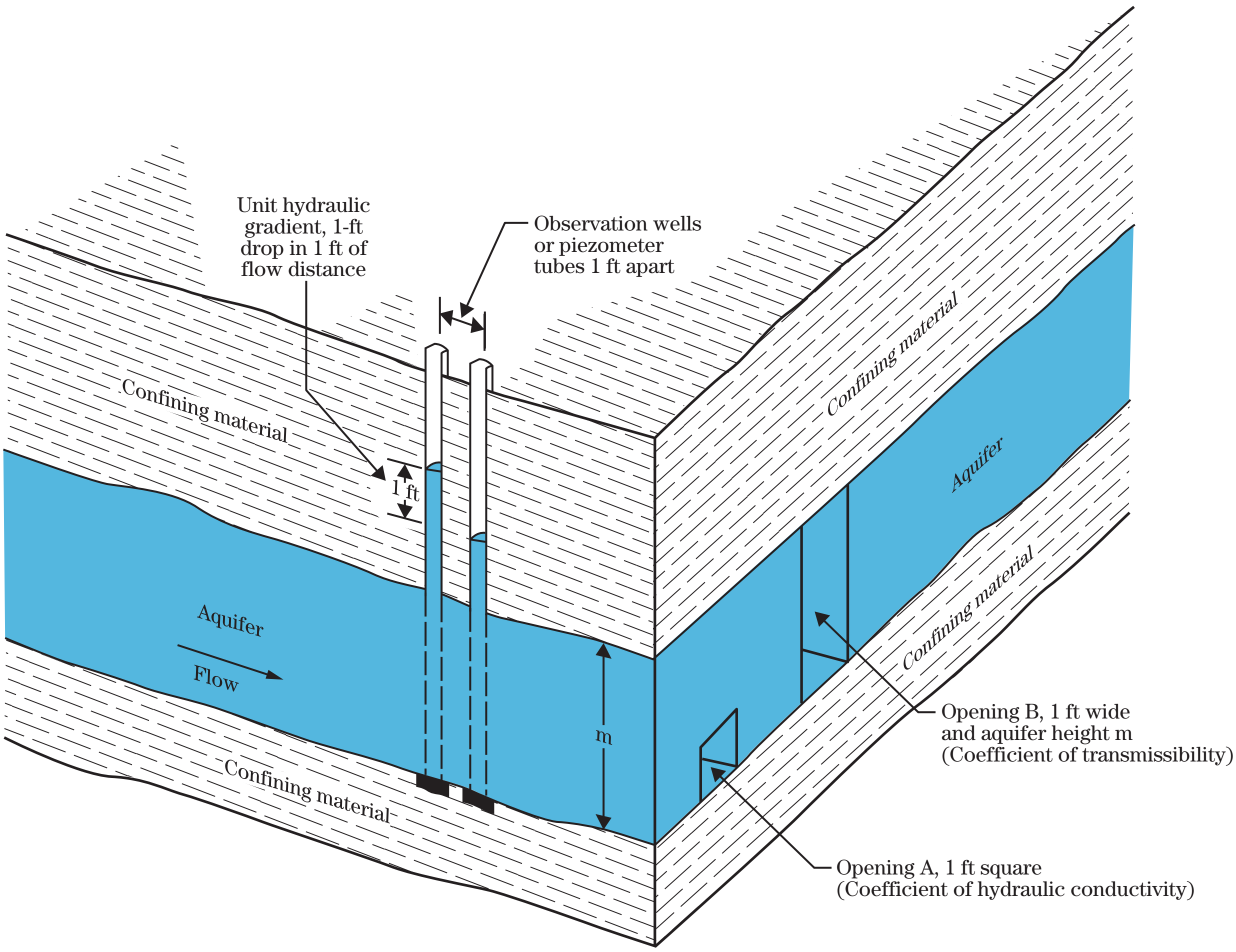
(3) Flow lines and equipotential lines

A streamline or flow line is the mean path of flow of water as it moves through an aquifer in the direction of decreasing head.

Equipotential lines in an aquifer represent contours of equal head. They intersect flow lines at right angles. Groundwater contours in an unconfined aquifer and equal pressure lines in a confined aquifer are equipotential lines.

A **flow net** is a graphical representation of a flow pattern and is composed of families of flow lines and equipotential lines. The actual flow pattern contains an infinite number of flow lines and equipotential lines. The graphical representation is constructed using only a few of these lines, selected so that the quantity of flow is equal between adjacent pairs of flow lines, and the drop in head is equal between adjacent pairs of equipotential lines.

Figure 30–4 Coefficients of hydraulic conductivity and transmissibility (from Theory of Aquifer Tests by J.G. Ferris et al., 1962, U.S. Geological Survey Water Supply Paper 1536–E)



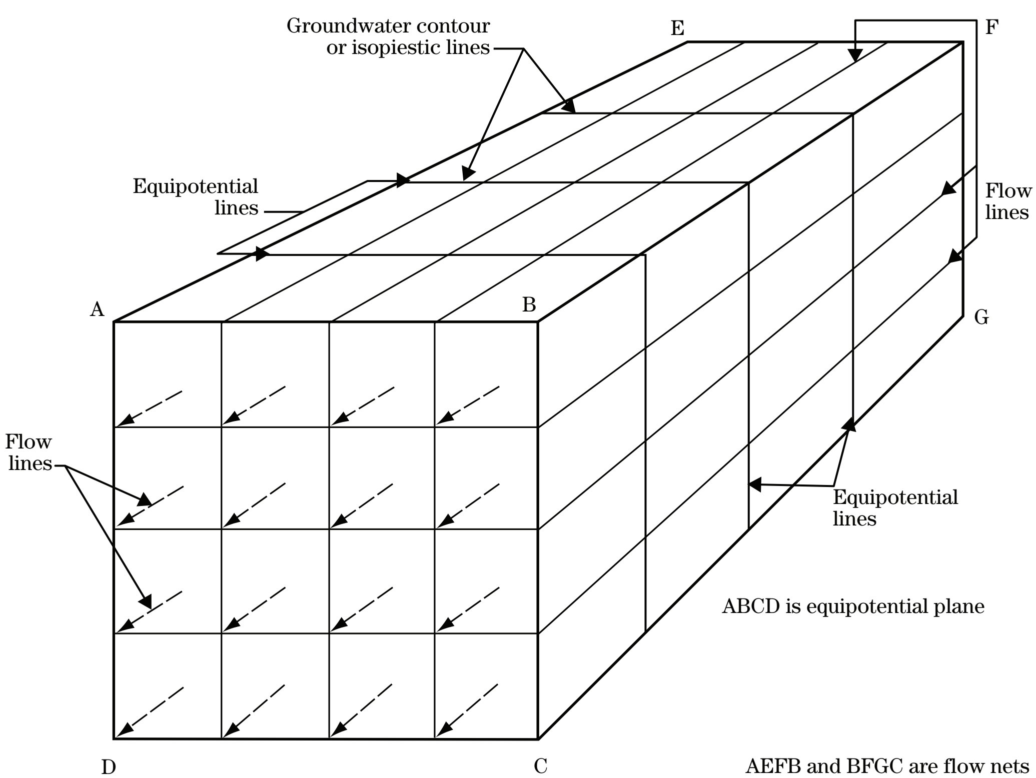
When constructing a flow net, the flow lines are drawn orthagonal to the equipotential lines, resulting in a pattern of rectangles (or squares), with the ratio of the mean dimensions of each rectangle equal.

A flow net can represent a vertical plane through an aquifer (BFGC in fig. 30–5) or any plane parallel to the flow (AEFB in fig. 30–5).

In three dimensions, each equipotential line becomes a plane. Under ideal conditions (an isotropic aquifer), this plane is vertical. The upper edge of the plane is a line on the upper surface of the aquifer, and the lower edge is a line on the bottom of the aquifer. Piezometers set at any point on this plane will have the same water level. An equipotential plane is represented by plane ABCD in figure 30–5. Figure 30–5 represents a prism taken from an ideal aquifer where the equipotential lines are straight and equally spaced, and the flow lines are equally spaced. The discharge from the small squares on the face of the prism are all equal, as are their areas. Under field conditions, the discharges would be equal, but the areas would probably vary.

Since flow cannot cross an impermeable boundary, flow lines adjacent to a boundary must parallel the boundary. In an unconfined aquifer, the water table is also a bounding surface. Flow lines adjacent to the water table must parallel the surface of the water table. Since groundwater contours are actually equipotential lines, flow lines must be perpendicular to the contours. The relative spacing of the flow lines depends on the water table gradient and the hydraulic conductivity and thickness of the aquifer.

Figure 30–5 Flow nets



(4) Coefficient of storage

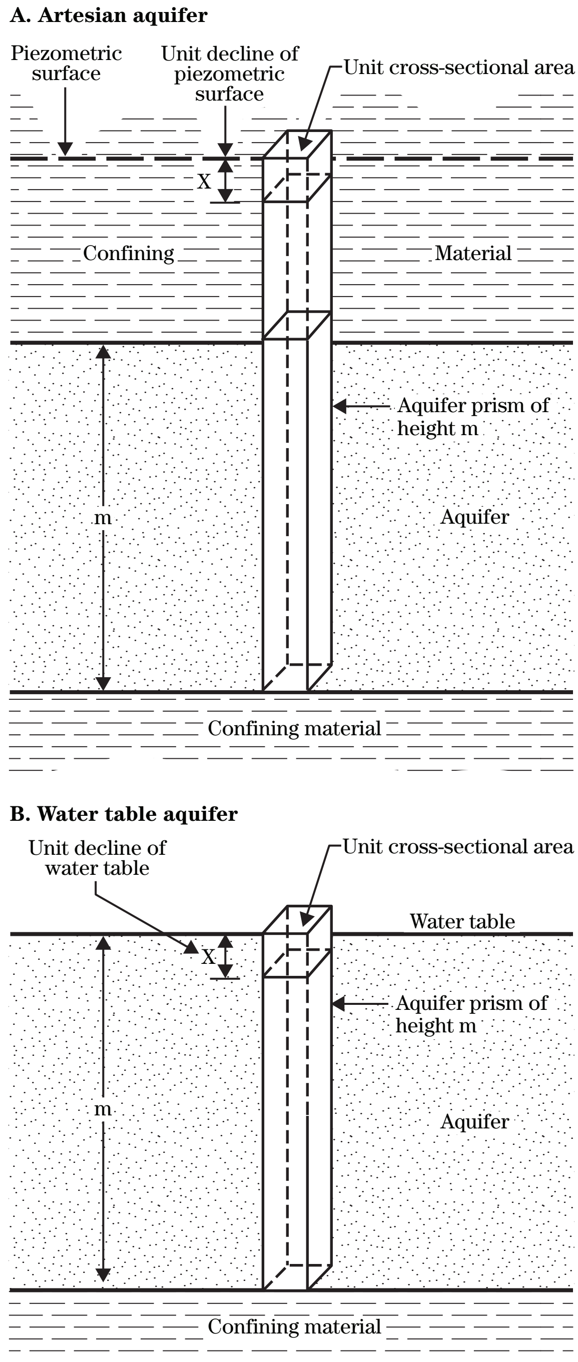
The coefficient of storage of an aquifer is the volume of water it releases or takes into storage, per unit surface area of the aquifer, per unit change in the component of head normal to that surface. This is illustrated in figure 30–6.

In an unconfined aquifer, the coefficient of storage is equivalent to the specific yield. It is the volume of water, in cubic feet, that a cubic foot of the aquifer will yield to gravity drainage. It will range from about 0.05 to 0.30.

The storage coefficient of a confined aquifer equals the volume of water in cubic feet released from a vertical column 1 foot square extending through the aquifer, when the piezometric surface declines 1 foot. This volume is attributable to the compressibility of the aquifer material and of the water. The storage coefficients of artesian aquifers usually range from 0.00001 to 0.001.

An unconfined aquifer yields 50 to 30,000 times more water per foot decline in head than will a confined aquifer, other things being equal. So, a decline in the piezometric surface over large areas is required to produce significant amounts of water from a confined aquifer.

Figure 30–6 Coefficient of storage (from Theory of Aquifer Tests by J.G. Ferris, et al., 1962, U.S. Geological Survey Water Supply Paper 1536–E)



(e) Hydraulics of wells

A groundwater reservoir tends toward a state of equilibrium, with natural discharge balancing natural recharge. The installation of a discharging well disturbs this balance, and a new equilibrium state is reached by the propagation of a cone of depression. As the cone of depression extends outward and deepens, natural recharge is increased, or natural discharge and well discharge are decreased, until equilibrium is again established. The extent of the cone of depression is called the area of influence.

(1) Equilibrium conditions

In an ideal, homogeneous, isotropic aquifer of finite thickness and infinite extent, a completely penetrating discharging well can be considered as being surrounded by a series of concentric cylinders. At equilibrium, the discharge through the wall of each of these cylinders is equal to the discharge from the well. At successively greater distances from the well, the side areas of the cylinders increase at an increasing rate, since the area of a cylinder equals $2\pi rh$.

Since, according to Darcy's law:

$$Q = kIA$$

and therefore,

$$k = \frac{Q}{IA}$$

the permeability of the aquifer can be determined if the hydraulic gradient of the groundwater is known at a distance, r, from a well discharging a known volume of water. At this distance, the area of the cylinder through which flow occurs is $2\pi rm$ for a confined aquifer, where m is the thickness of the aquifer. For an unconfined aquifer, the area of the cylinder is $2\pi r$ (m–s), where m is saturated thickness before pumping, and s is drawdown.

Since the hydraulic gradient cannot be measured at a point, a mathematical treatment of Darcy's equation was developed that substitutes the drawdown in two observation wells for the hydraulic gradient. This is known as the Thiem formula (Theim 1906):

$$k = \frac{Q \log_e \frac{r_2}{r_1}}{2\pi m_{(s_1 - s_2)}} \qquad \text{(eq. 30–5)}$$

where:
k = coefficient of permeability ($ft^3/ft^2/d$)
Q = discharge from pumped well (ft^3/d)
r_1, r_2 = distances from pumped well to two observation wells (ft)
s_1, s_2 = drawdown in the observation wells (ft)
m = saturated aquifer thickness before pumping (ft)

The formula can be used to compute permeabilities under equilibrium conditions, assuming the following:

- The aquifer is isotropic, of infinite areal extent and rests on a horizontal impermeable bed.
- The discharging well penetrates the entire saturated thickness of the aquifer.
- The observation wells are placed in line with the discharging well.
- Pumping has continued at a constant rate for sufficient time for the drawdown cone to have reached a steady state.
- Flow to the well is radial.
- The nonpumped water table was horizontal.

The Thiem formula applies to confined aquifers and unconfined aquifers, if the difference in drawdown between the two observation wells is small compared to the saturated thickness. Figure 30–7 illustrates the various relationships in the Thiem formula.

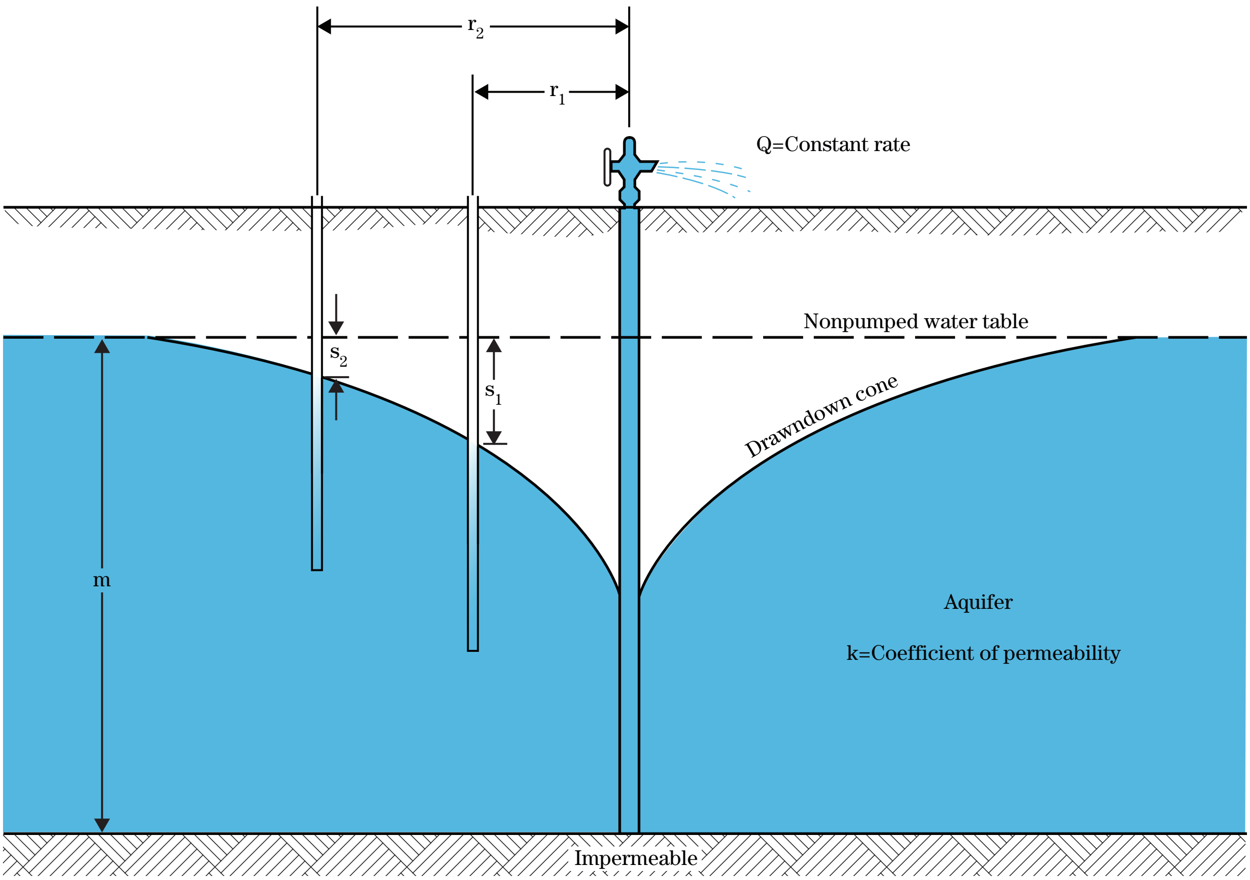
In practice, one procedure for using the Thiem formula is to measure the drawdown in the observation wells at a time when equilibrium of the cone of depression has been established by pumping at a constant rate. These measurements are plotted on semilog scale with the drawdown, s, on the arithmetic scale and distance, r, on the logarithmic scale. When two or more observation wells are used, all values of r and s at a given time, t, should fall on a straight line on the semilog plotting, if equilibrium has been attained. Convenient values of r and r^2, and s and s^2 can then be picked from the curve, and the Thiem formula solved for k. An example of these procedures is given in NEH631.31.

(2) Nonequilibrium conditions

If equilibrium of the drawdown cone is not established, a nonequilibrium equation can be applied to determine the aquifer constants. This equation, developed by C.V. Theis in 1935, is based on the assumption that hydraulic conditions in an aquifer and thermal conditions in a thermal system are analogous in mathematical theory. The following aquifer conditions are assumed in the application of the equation:

- aquifer is homogeneous and isotropic
- aquifer has infinite areal extent
- discharging well penetrates the entire thickness of the aquifer
- coefficient of transmissibility is constant at all times and all places
- well has an infinitesimally (reasonably) small diameter

Figure 30–7 Aquifer pump test



- water removed from storage is discharged instantaneously with decline in head

The final Theis equation for the drawdown of the water level in the vicinity of a discharging well is:

$$s = \frac{114.6Q}{T} \int_u^\infty \frac{e^{-u}}{u} du \qquad \text{(eq. 30–6)}$$

where:

s = drawndown at a point r distance from a pumping well (ft)
Q = discharge from the well (gal/min)
T = coefficient of transmissibility (gal/d/ft)
u = $\frac{1.87\, r^2 S}{Tt}$
r = distance from pump well to point where drawdown s is determined (ft)
s = coefficient of storage (dimensionless)
t = time that pump well has been discharging (d)

The expression:

$$\int_u^\infty \frac{e^{-u}}{u} du$$

is an exponential integral, and cannot be integrated directly. The notation for this integral is W(u), which is read "well function of u," and its value is given by the series:

$$W(u) = -0.577216 - \log_3 u + u - \frac{u^2}{2\times 2!} + \frac{u^3}{3\times 3!} - \frac{u^4}{4\times 4!} \cdots \qquad \text{(eq. 30–7)}$$

Gallon is the unit used here instead of cubic feet because most literature concerning pumps and pump tests is in terms of gallons. Appropriate conversion factors can be used after computations have been completed.

Equation 30–6 can then be rewritten as:

$$s = \frac{114.6Q}{T} W(u) \qquad \text{(eq. 30–8)}$$

Selected values of W(u) for given values of u between 10^{-15} and 9.5 are given in table 30–3.

If the coefficient of transmissibility and coefficient of storage are known, the drawdown on the cone of depression at any time and any distance from the well can be determined after the well starts discharging. This is done by substituting the known values of S and T and the desired values of t and r in the equation:

$$u = \frac{1.87 r^2 S}{Tt} \qquad \text{(eq. 30–9)}$$

The equation is then solved to determine u. The value of W(u) for u is obtained from table 30–3 and equation 30–8 is then solved for the drawdown, s.

The nonequilibrium equation can also be solved for transmissibility, T, and storage coefficient, S, if several values of drawdown, s, at times, t, are known for one distance, r, from the pumped well or several values of s and r are known for one value of t. The solution requires the graphical determination of W(u), u, s, and either r^2/t or the reciprocal of time (1/t).

A type curve of W(u) versus u for the values shown in table 30–3 is plotted on a log-log scale. Figure 30–8 is an example. Data from the observation wells are then plotted on log-log scale at the same scale as the type curve. Values of s are plotted against r^2/t (or 1/t if only one observation well is used and r therefore is constant).

Table 30–3 Selected values of u and W(u)

u	W(u)	u	W(u)
9.5	0.0000072	0.01	4.04
6.0	0.00036	0.005	4.73
4.0	0.0038	10^{-3}	6.33
3.0	0.013	10^{-4}	8.63
2.0	0.049	10^{-5}	10.94
1.5	0.10	10^{-6}	13.24
1.0	0.22	10^{-7}	15.54
0.75	0.34	10^{-8}	17.84
0.5	0.56	10^{-9}	20.15
0.4	0.70	10^{-10}	22.45
0.3	0.91	10^{-11}	24.75
0.2	1.22	10^{-12}	27.05
0.1	1.82	10^{-13}	29.36
0.075	2.09	10^{-14}	31.66
0.05	2.47	10^{-15}	33.96
0.025	3.14		

The curve of the observed data is superimposed on the type curve and with the axes of the curves kept parallel the superimposed sheet is shifted until a section of the two curves matches. The values of W(u), u, s, and r^2/t of an arbitrary common point on the matched section of the curves are recorded and the equation:

$$T = \frac{114.6Q}{s} W(u) \qquad \text{(eq. 30–10)}$$

is solved for T. The equation:

$$u = \frac{1.87r^2S}{Tt} \qquad \text{(eq. 30–9)}$$

can then be solved for S.

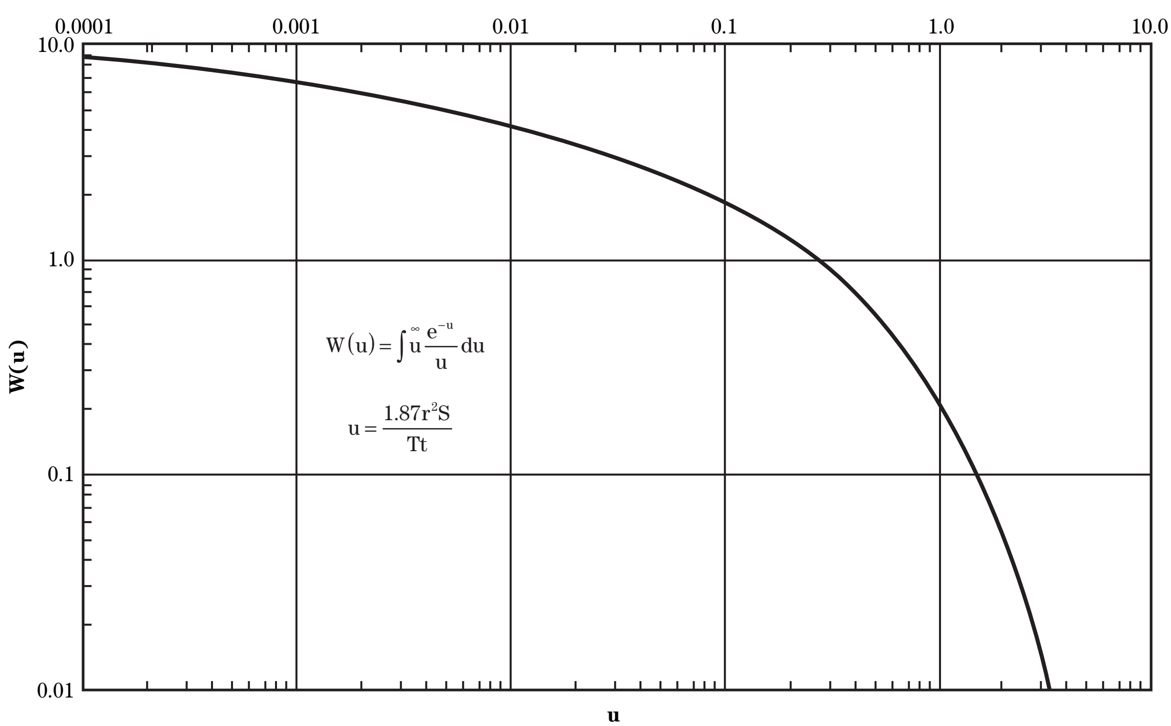
Many modifications, simplifications, and alterations can be applied to the nonequilibrium equations. Conditions such as impermeable boundaries, recharge boundaries, leaky aquifers, partial penetration of the aquifer by discharging wells, and other conditions encountered in the field require adjustments and modifications of the basic equations. An example of a pump-out test is given in NEH631.31.

(f) Quality of groundwater

Groundwater quality depends on the total amounts and kinds of dissolved minerals and the intended use of the water. The quality of groundwater is influenced by local geology; land use; climate; and disposal of domestic, municipal, and industrial wastes.

The U.S. Environmental Protection Agency has set standards for drinking water that have been adopted by most States. A complete detailed listing of primary and secondary drinking water standards are available at http://www.epa.gov/safewater/contaminants/index.html#mcls).

Figure 30–8 Relation between u and W(u)



631.3002 Groundwater geology

(a) Unconsolidated materials

Unconsolidated materials consist of soil and recent deposits of rockslide debris, talus, fans, mud flows, alluvium, lake sediments, loess, dune sand, and glacial drift. They are important to the planning and development of water resources wherever they are involved in site conditions.

Rock-slide debris is generally very porous and highly permeable. It is difficult to develop as a source of groundwater, but is usually excellent for groundwater recharge developments.

Mudflow deposits may have low permeability and are of little value for water developments.

Talus deposits and alluvial fans in stream valleys may be good sources of groundwater, depending on their size and topographic position. The following description of alluvium also applies to these deposits.

(1) Alluvium

Alluvium is stream-deposited sediment. Alluvium varies in thickness, texture, degree of sorting, stratification, porosity, and permeability. Understanding principles of valley alluvial deposition, and as related to glacial influences, is important in the interpretation and evaluation of site conditions for groundwater development.

During the Pleistocene, degradation in the stream systems occurred when sea level was lowered by increasing accumulation of ice in the continental glaciers. Conversely, aggradation occurred in the stream systems with the retreat of the continental glaciers.

This degradation and aggradation that accompanied each glacial advance and retreat represents one cycle of sedimentation. Seven such cycles have been recognized in nonglaciated as well as glaciated areas.

Deep channel and terrace deposits in the Mississippi River (fig. 30–9) indicate that sea level may have been as much as 400 feet lower during the peak of glacial ice accumulation than it is now.

In glaciated areas, some preglacial and interglacial channels were blocked and filled by succeeding glaciers. In many instances, the present drainage system differs from the antecedent. Many of these antecedent or ancient drainage systems have been discovered by drilling groundwater test wells. They constitute an important reservoir for the accumulation and storage of groundwater. One example is the Teays River Valley, which was a major river that flowed from east to west generally across Ohio, Indiana, and Illinois (fig. 30–10).

Permeability and specific yield of alluvial deposits vary considerably with the aquifer's texture. Textural information can be obtained from the logs of test wells and the mechanical analyses of the water-bearing materials. This information is important for the proper design and development of wells.

Natural replenishment to alluvial water-bearing materials is influenced by intensity and duration of precipitation, intake rates of soil and subsoil, runoff from the drainage basin, and in some areas, by groundwater flow from nearby drainage basins. Groundwater replenishment is also influenced by land slope, plant cover, land use, and available storage capacity. Frozen ground is an effective barrier against infiltration.

(2) Lake sediments

Lacustrine deposits consist of clays, silts, sands, gravels, marl, tufa, peat, and evaporites. Sands and gravels occur as deltaic and shoreline or beach deposits, whereas silts and clays may be lakewide in occurrence. Limonite (hydrated iron oxide) and manganese dioxide (MnO_2) may occur as nodular and concretionary deposits of irregular distribution. MnO_2 deposits may also take the form of pyrolusite dendrites. Peat is common along the margins, and sapropel (black muds rich in organic matter) is deposited in the central parts of lakes. Marl and tufa are precipitated in the greatest temperature change zone. Evaporites are common in lake deposits of arid regions. Some lake deposits have a predominance of carbonates.

Deltas and beaches are groundwater catchment areas. Water quality is usually good because of rapid intake from precipitation.

Water developments from the central portions of lake plains are less favorable because of low permeabilities of silts and clays and accumulations of dissolved solids.

Figure 30–9 Geomorphology and quaternary geologic history of the Lower Mississippi Valley (Plate 16), prepared by Roger T. Saucier, U.S. Army Engineer Waterways Experiment Station Vicksburg, Mississippi, December 1994

(a)

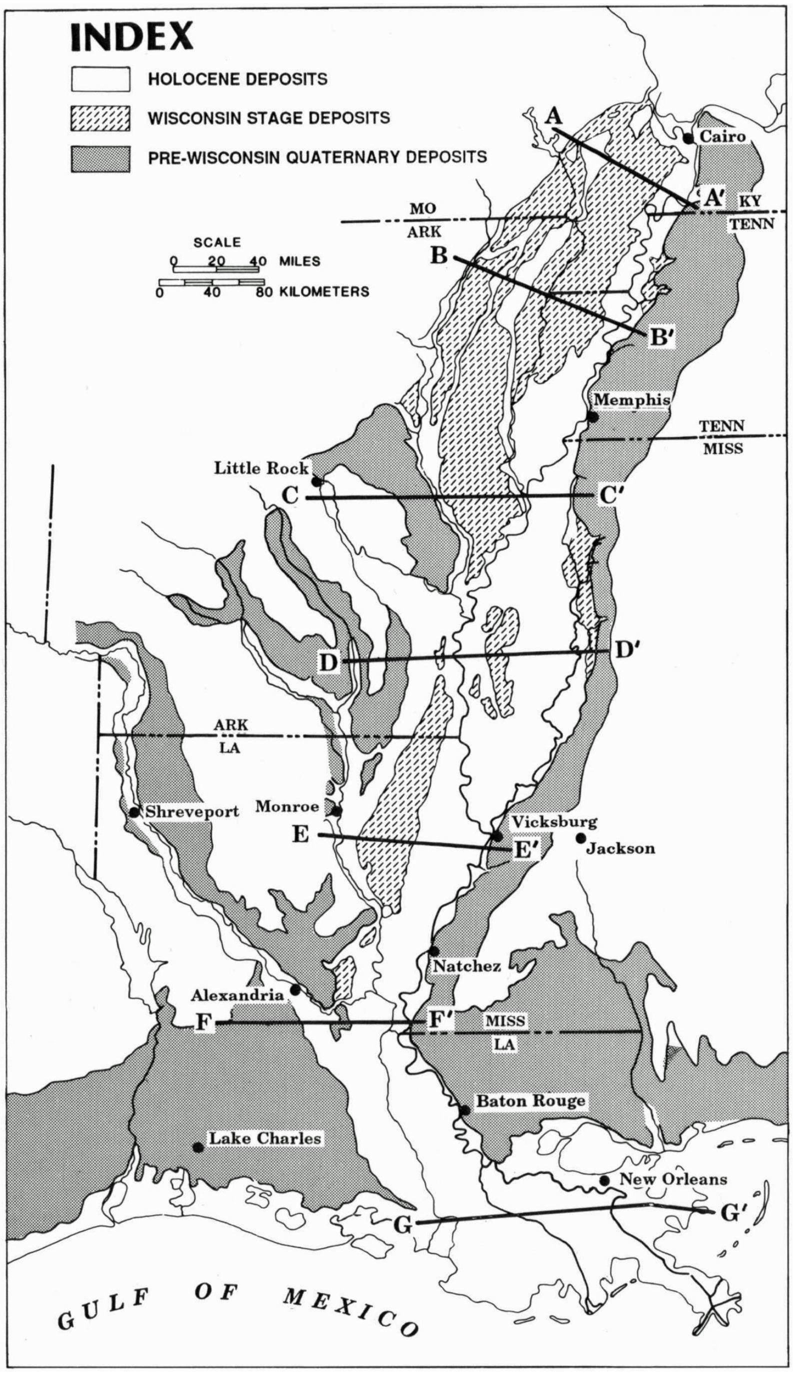
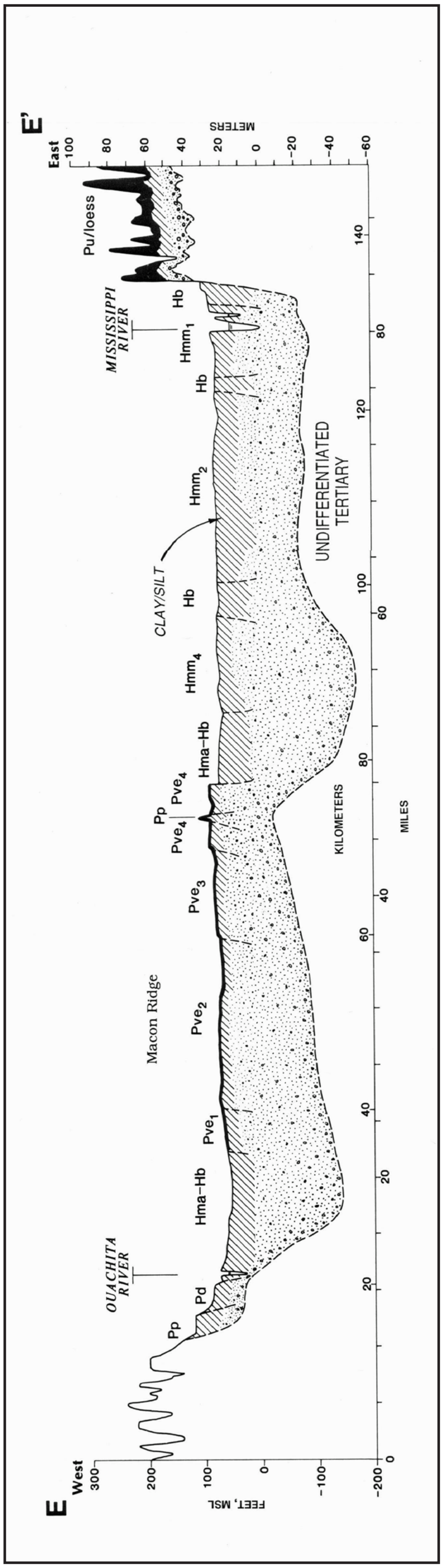


Figure 30–9 Geomorphology and quaternary geologic history of the Lower Mississippi Valley (Plate 16), prepared by Roger T. Saucier, U.S. Army Engineer Waterways Experiment Station Vicksburg, Mississippi, December 1994—Continued

(b)



(c)

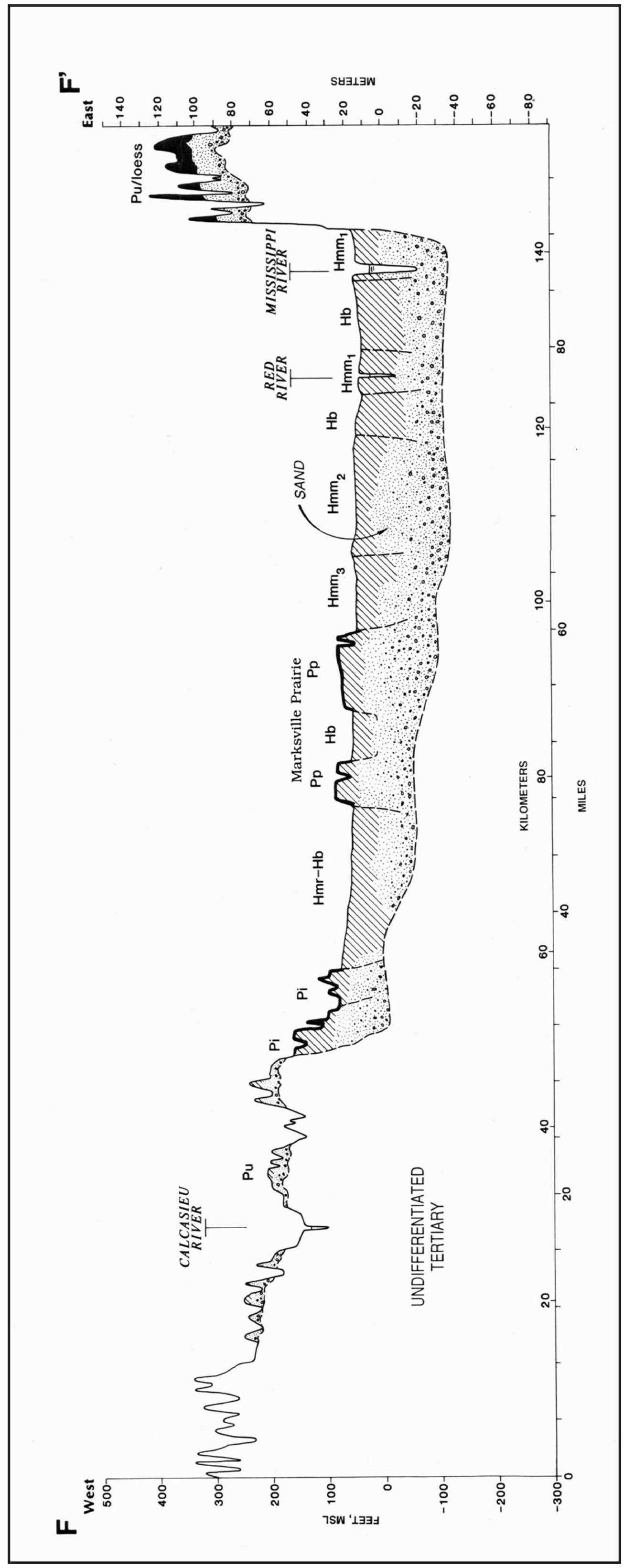


Figure 30–9 Geomorphology and quaternary geologic history of the Lower Mississippi Valley (Plate 16), Prepared by Roger T. Saucier, U.S. Army Engineer Waterways Experiment Station Vicksburg, Mississippi, December 1994—Continued

(d)

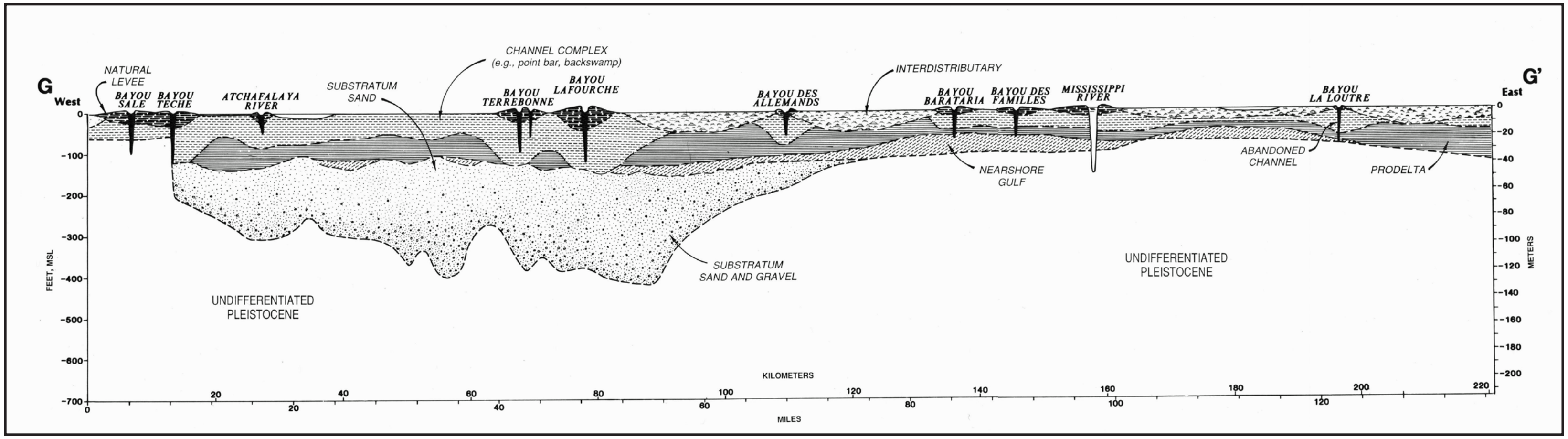
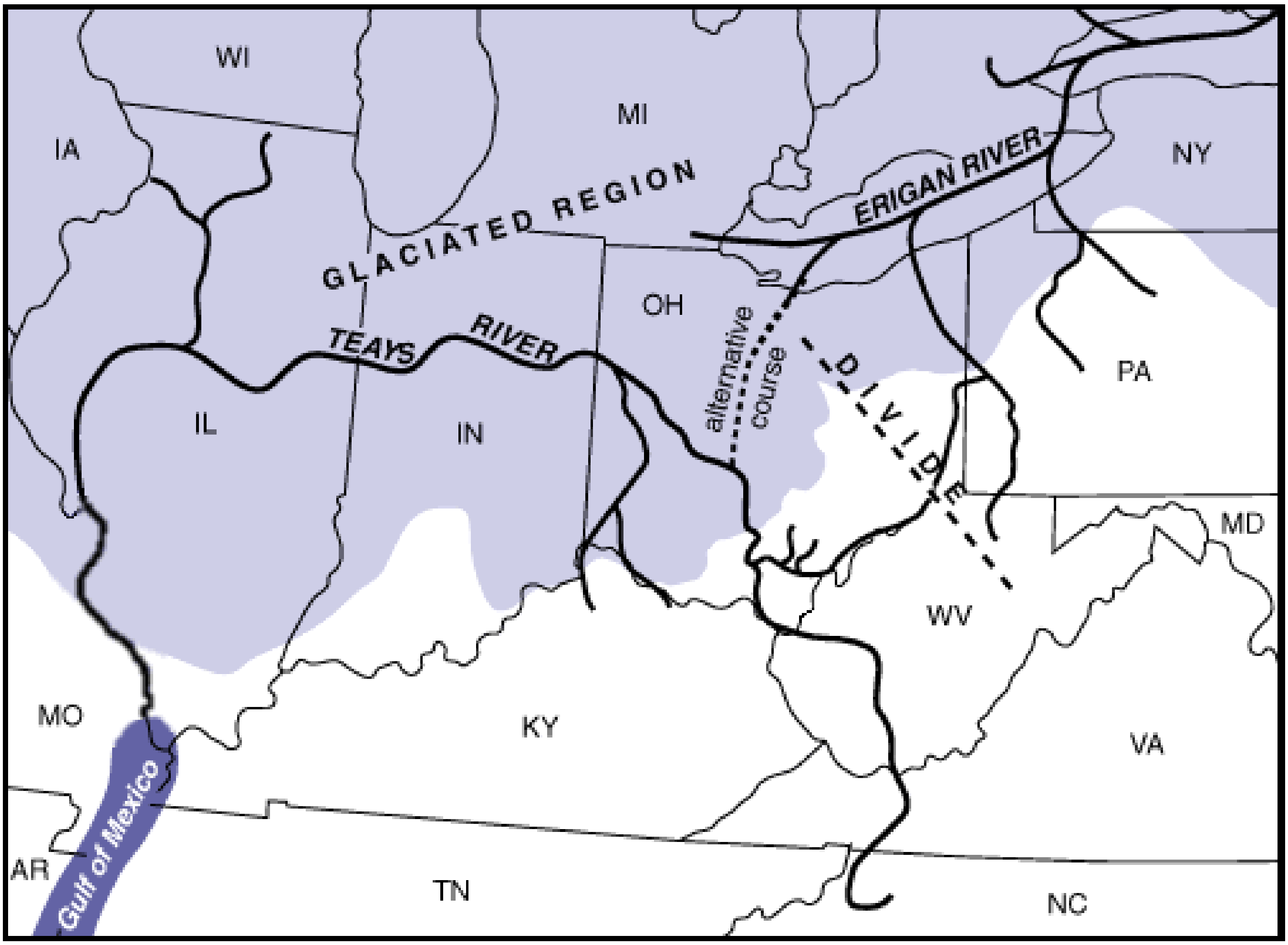


Figure 30–10 Preglacial Teays River Valley. Significant groundwater resources exist in the buried channel deposits (Ohio Department of Natural Resources, Division of Geological Survey, 2003)



(3) Loess

Loess is an eolian or windblown deposit that consists mainly of silts that are generally loose and noncohesive and may contain clay and sand. The clay content of some loessial soils is affected by soil development and associated chemical changes.

Loess occurs in intermontane valleys, hilltops, throughout great lengths of tablelands and uplands, terraces of stream systems, and flood plains and streambeds of some drainage systems. It may contain buried soils and tends to mantle antecedent topography.

Porosity, permeability, and specific yield of loess varies texture or size of particles and degree of sorting. The coefficient of permeability of loess is normally too low to be considered an aquifer.

(4) Dune sand

Eolian sands may include gypsum, oolites, and shell fragments. The St. Peter, Tensleep, and Navajo sandstones are examples of wind-worked sands. Eolian sands act as good intake areas for groundwater recharge to underlying formations but may themselves be above the main zone of saturation.

(5) Glacial deposits

Glacial deposits consist of glacial drift as ground, terminal, lateral, and recessional moraines and outwash deposits from the melting of glacial ice. Unstratified drift consists of a heterogeneous mixture of clay, sand, gravel, cobbles, and boulders. The glacial drift of mountainous areas and the New England States is predominantly an unstratified mass of boulders, whereas that of the northern Great Plains contains considerable clay derived from glacial gouging of Cretaceous shales. Some glacial drift in the northern Great Plains contains inclusions of stratified silts, sands, and gravels of various cross-sectional dimensions, with the bedding planes tilted from the horizontal. Some of these sand and gravel inclusions are large enough to be reservoirs for groundwater.

The Nebraskan glacial till has been reported as "almost impervious to water," (Shimek 1910). Kansan and younger glacial deposits furnish water to many small wells in the north-central United States, although the permeability of the tills is very low.

Kames, eskers, and outwash sands and gravels are usually very porous and highly permeable. They can be excellent sources of groundwater for domestic and livestock use. High iron content may be a problem, however.

(b) Rocks

The consolidated materials of the Earth's crust are classified into three major groups: igneous rocks, formed by cooling from the molten state; sedimentary rocks, formed by lithification of unconsolidated material, chemical and organic action, or evaporation; and metamorphic rocks, formed from stress and increased temperature on sedimentary or igneous rocks.

(1) Igneous rocks

Molten rock may be extruded onto the surface as lava flows, blown out by volcanic eruptions, or intruded into the Earth's crust at various depths below the surface.

Pyroclastic rocks and some lava flows contain openings due to fragmentation, gas bubbles, or contraction during cooling. Locally, these openings may be connected so that groundwater can move through them. Most igneous rocks, however, are dense and lack voids. They are not considered potential sources for groundwater unless openings have developed by fracturing, jointing, or faulting. Where joint systems are well developed, as in some dense lava flows, large amounts of groundwater may be available. Jointed and fractured granitic masses may also yield some groundwater.

The Columbia River Basin in Washington, Oregon, and Idaho, drains about 50,000 square miles of lava. The lava occurs in individual flows that vary in thickness from about 5 to 150 feet. The total maximum thickness is about 5,000 feet, from about 1,100 individual flows. Tunnels may be formed by molten lava flowing out from beneath a solidified crust. The base of some flows contains local rubble that is permeable, and the tops of some are columnar and blocky from fracturing caused by cooling. Columnar structure contributes to permeability. Faulting transverse to the dip of bedrock has produced barriers to horizontal flow of water in some localities. The rock varies from dense massive columnar-jointed rock to vesicular and fractured masses on weathered surfaces of individual flows. Precipitation and runoff from the mountainous areas filter into surface fractures and migrate to lower strata along faults and major fractures. Most highly productive wells are located in structural troughs, and some are artesian.

(2) Sedimentary rocks

Sedimentary rocks are the products of chemical, organic, and mechanical weathering processes. Shale, sandstones, and conglomerates are formed by mechanical weathering processes.

Massive forms of indurated muds, silts, or clays are termed mudstone, siltstone, and claystone, respectively. Shales are laminated or fissile. Shales have very low porosity and are generally considered nonwater-bearing and as aquicludes.

Weathering and fracturing of shales produce secondary voids. Shales may also heave and rebound from release of load to produce fractures. Freezing, thawing, root action, and chemical changes tend to increase the size and extent of fractures. Wells sited in weathered shale may produce limited yields, primarily due to secondary permeability in fracture zones.

Poorly sorted or well-cemented sandstones have low porosity and low permeability and are considered only fair to poor as aquifers. Highly fractured, poorly sorted sandstones may be good aquifers. Well-sorted, poorly cemented, clean quartz sandstones are probably the most dependable reservoir rock for groundwater.

Sandstones famous for the dependability of water supplies are found in the Bresbach, Jordan, St. Peter, Dakota, Mesaverde, Brandywine, Ripley, Wilcox, and Ogallala formations.

The quality of water in sands and sandstones is influenced by structural geology, geomorphology, stratigraphy, sedimentation, porosity, permeability, infiltration, and circulation of groundwater. Where circulation is restricted or connate water is present, quality is poor to unsatisfactory.

Along a coastline, groundwater quality is influenced by its relation to sea level. A freshwater-saltwater interface is kept in balance by groundwater recharge and groundwater movement seaward. Withdrawal of groundwater in excess of recharge lowers the water table. When the water table is lowered, saltwater intrusion of the freshwater aquifer occurs.

Chemically precipitated sedimentary rocks that are important water-bearing formations in some areas are limestones, dolomites, chalk, and gypsum. Paleozoic limestones and dolomites of the northeastern United States yield water to springs and wells. Cretaceous limestones in the southern United States furnish water for many municipalities.

The porosity of limestones is influenced by granularity, matrix, cement, skeletal remains, reefs, fractures, dolomitization, and solution. Solution cavities in the rock follow fractures, joints, and bedding planes and may be well developed.

Development of water supplies from wells in limestones depend on fractures and solution cavities that are adequate to store and yield large volumes of water within economic pumping lifts.

Coal and lignite are genetically associated with sedimentary rocks. They are associated with valley flat deposits and some coal strata are in contact with sand. Many coal beds have partially burned and formed clinker beds that are extremely porous. Locally, they are aquifers, but the water is generally of poor quality.

(3) Metamorphic rocks

The texture and mineral composition of rocks change when they are subjected to high temperatures and pressures. Under metamorphic processes, the rocks are compressed and may exhibit foliation or banding as in slate, phyllite, schist, and gneiss. They are usually so dense that they lack interstices and are incapable of absorbing water. They may, however, become fractured and faulted during and after metamorphism. The fractures may create a system of connecting ruptures that will absorb and transmit water. Zones of fracture may be suitable for groundwater development.

Minerals in metamorphic and igneous rocks weather differentially. Weathered zones may be favorable for the development of groundwater supplies.

(c) Stratigraphy

Determining the capacity of strata to hold, yield, and transmit water is the object of subsurface investigations for water developments. A knowledge of stratigraphy is valuable in evaluating potential aquifers.

(1) Stratigraphic traps
Varieties of stratigraphic traps may confine groundwater. Examples are pinch-outs, overlaps, and changes in facies. Pinch-outs may result from certain conditions of sedimentation, erosion, truncation, and overlap. In alluvial deposits, they may be in the form of splays, deltas, tributary fans, and cut and filled channels and meander loops in valleys. Those of marine sediments may be in the form of offshore bars, deltas, and other shoreline features.

Change in facies is a lateral change in the composition or grain size of the deposited material. It may be due to a change in the source of the deposited material, distance from the source of the material, tectonic activity, or combinations of these. Perhaps the most important stratigraphic traps are those associated with channel deposits.

(d) Structure

Structure refers to the arrangement and relative position of rock masses. A description of rock features such as folds, faults, bedding, joints, flow banding, parting, cleavage, brecciation, pillow bodies, and blocky or ropey features are used to describe structure. Some structural features such as stratification, jointing, and blocky development are exhibited by unconsolidated materials and are a part of the structure of those bodies of material.

Structure may influence the occurrence and availability of groundwater. When a permeable stratum is confined between impermeable strata, water available to the permeable materials depends on intake along structurally high areas. When a permeable stratum is faulted so that the displacement terminates the permeable materials against impermeable, and the fault plane or fault zone is impermeable, the supply of water may be cut off to the side away from the recharge area. Wells on one side of the fault might yield artesian water in large quantities, whereas wells on the other side of the fault may be unsatisfactory.

(e) Major features

(1) Folds
Folds are warps, flexures, or bends in strata. They may be very gentle and simple patterns or very complex. They may be symmetrical, unsymmetrical, squeezed, overturned, or recumbent. Synclines are the most productive folds for storage and replenishment of groundwater.

(2) Faults
These are breaks in the rocks of the Earth's crust, along which movement and displacement can be identified. They may be seen on the surface and can sometimes be identified on aerial photos. Subsurface faults may also form important groundwater sources.

(3) Unconformities
Unconformities are surfaces that indicate periods of nondeposition or erosion. Evidence of unconformities includes discordant strata, basal conglomerate, basal black shale, fossil desert varnish, lag gravel, edgewise conglomerate, residual chert, concentration of glauconite, phosphatic pellets and nodules, manganese and iron concretions, clastic zones in nonclastic rocks, abrupt changes in heavy-mineral assemblages, sharp differences in lithology or fossil assemblages, and buried soils. Unconformities may influence the migration of water in aquifers.

(f) Minor features

(1) Cleavage
Cleavage is the tendency of rocks to split or break apart under stress along definite smooth, parallel, closely spaced planes. The openings transmit water, but are not considered important groundwater sources. Cleavage is a secondary phenomenon that influences weathering and rock disintegration.

(2) Fractures
Fractures are cracks or breaks in rock. They are openings along which the walls of the voids are distinctly separated. Joints, faults, and fissures are varieties of fractures. Fractures are good sources of groundwater in some geologic formations. Locally, they may yield enough for irrigation well developments. Even fractured quartzites and argillites may be important sources of groundwater. Open fractures in limestone lead to the development of solution channels.

(3) Joints
Fractures in rock along which there has been little or no transverse movement are called joints. They may be caused by expansion and contraction, resulting

from changes in moisture content or temperature or as a result of stress. A joint set consists of a group of breaks that are parallel or nearly so. A joint system is a group of two or more intersecting sets or any group of joints with a characteristic pattern such as radiating or concentric.

Joints are passageways for the movement of water. They are usually inadequate as reservoirs for other than minor quantities of groundwater unless well developed. They may, however, contribute to groundwater recharge, and where there are well-developed joint systems, as in some shales and basalts, adequate supplies of water may be available.

(4) Solution openings

Solution openings may be formed from the solution of carbonates by groundwater circulation along fractures, joints, and bedding planes. Some cavities become enlarged to form extensive caverns and caves. In limestone areas, groundwater gradients are usually low. Limestone voids are usually not saturated at elevations above the elevation of permanently flowing streams.

631.3003 References

Saucier, R.T. 1994. Geomorphology and quaternary geologic history of the Lower Mississippi Valley (Plate 16), U.S. Army Corps of Engineer Waterways Experiment Station, Vicksburg, MS.

Ohio Department of Natural Resources, Division of Geological Survey, 2003.

Theis, C.V. 1935. The relation between the lowering of the piezometric surface and the rate and duration of discharge of a well using groundwater storage. Trans. Amer. Geophys. Union, 2, pp. 519–524.

Thiem, G. 1906. Hydrologische Methoden. Gebhardt, Leipzig.

U.S. Environmental Protection Agency (USEPA), Office of Water, EPA 816–F–03–016, June 2003: http://www.epa.gov/safewater/consumer/pdf/mcl.pdf