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DEPARTMENT OF THE INTERIOR
GEOLOGICAL SURVEY

Basic Elements of Ground-Water Hydrology With Reference to Conditions in North Carolina

By Ralph C. Heath

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Preface

Ground water is one of North Carolina's most valuable natural resources. It is the primary source of water supplies in rural areas and is also widely used by industries and municipalities, especially in the Coastal Plain. However, its use is not increasing in proportion to the growth of the State's population and economy. Instead, the present emphasis in water-supply development is on large regional systems based on reservoirs on large streams.

The value of ground water as a resource not only depends on its widespread occurrence but also on its generally excellent chemical quality. Thus, in most cases, ground-water does not require treatment prior to use, except as a precaution against unsuspected pollution. However, in an effort to control stream pollution, greater and greater emphasis is being placed on the land disposal of liquid and solid wastes. One result of this may be increasingly widespread deterioration of ground-water quality.

The development of ground-water supplies and the protection of ground-water quality requires knowledge of the occurrence of ground water and knowledge of how ground-water systems function. The lack of such knowledge among those involved in the development, management, and regulation of water supplies has been an important element in the avoidance of ground water as a source of large supplies and in the increasing occurrence of ground-water pollution.

This report was prepared as an aid to developing a better understanding of the ground-water resources of the State. It consists of 46 essays grouped into five parts. The topics covered by these essays range from the most basic aspects of ground-water hydrology to the identification and correction of problems that affect the operation of supply wells. The essays were designed both for self study and for use in workshops on ground-water hydrology and on the development and operation of ground-water supplies.

Relative to the use of this report in workshops, selected essays have been used in workshops for the staff of the North Carolina Department of Natural Resources and Community Development and in short courses for operators of water systems that utilize wells as a source of supply. The essays used in each workshop and short course were selected on the basis of the background and needs of the group.

Finally, most of the essays contain sketches that illustrate the main points covered in the text. However, to facilitate the rearrangement of essays for use in different workshops, they are not numbered and titled as in most technical reports. For those who find this disconcerting, I suggest that the text of each essay simply be viewed as an expanded explanation of the sketches.

Ralph C. Heath

Definitions of Terms

Aquifer (p. 7)[1] A water-bearing bed that will yield water in a usable quantity.

Bedrock (p. 20) A collective term for the metamorphic and igneous rocks underlying the Piedmont and mountains.

Capillary fringe (p. 5) The zone above the water table in which water is held by surface tension. Water in the capillary fringe is under a pressure less than atmospheric.

Conductivity, hydraulic (p. 12) The capacity of a unit cube of rock to transmit water.

Cone of depression (p. 49) The depression of heads around a pumping well caused by the withdrawal of water.

Confining bed (p. 7) A layer of rock that hampers the movement of water.

Dispersion (p. 40) The extent to which a substance injected into an aquifer spreads as it moves through the aquifer.

Drawdown (p. 52) The reduction in head at a point caused by the withdrawal of water from an aquifer.

Equipotential line (p. 42) A line (on a map) or cross section along which total heads are the same.

Flow line (p. 42) The idealized path followed by particles of water.

Flow net (p. 42) The grid pattern formed by a network of flow lines and equipotential lines.

Formation (p. 22) A distinct rock layer named for a locality near which it occurs.

Gradient, hydraulic (p. 10) Change in head per unit of distance measured in the direction of the steepest change.

Head, total (p. 10) The height above a datum plane of the surface of a column of water. It is composed in a ground-water system of elevation head and pressure head.

Porosity (p. 8) The openings in a rock.

Saprolite (p. 20) The soil-like rock that occurs between land surface and bedrock in the Piedmont and mountains.

Specific retention (p. 9) The volume of water retained in a rock after gravity drainage.

Specific yield (p. 9) The volume of water that will drain under the influence of gravity from a saturated rock.

Storage coefficient (p. 38) The volume of water released from storage in a unit prism of an aquifer when the head is lowered a unit distance.

Stratification (p. 36) The layered structure of sedimentary rocks.

Transmissivity (p. 37) The capacity of an aquifer to transmit water. It equals the hydraulic conductivity times the aquifer thickness.

Water table (p. 5) The level in the zone of saturation at which the pressure is equal to the atmospheric pressure.

Zone, saturated (p. 5) Subsurface zone in which all interconnected openings are full of water.

Zone, unsaturated (p. 5) Subsurface zone, usually starting at the land surface, that contains both water and air.

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[1]Page numbers refer to page on which the term is discussed.

Contents

Acknowledgments

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Cover Photograph

Servicing a continuous water-level recorder on an observation well operated by the U.S. Geological Survey in the Cape Hatteras National Seashore.
(Photograph by Ralph C. Heath, U.S. Geological Survey)

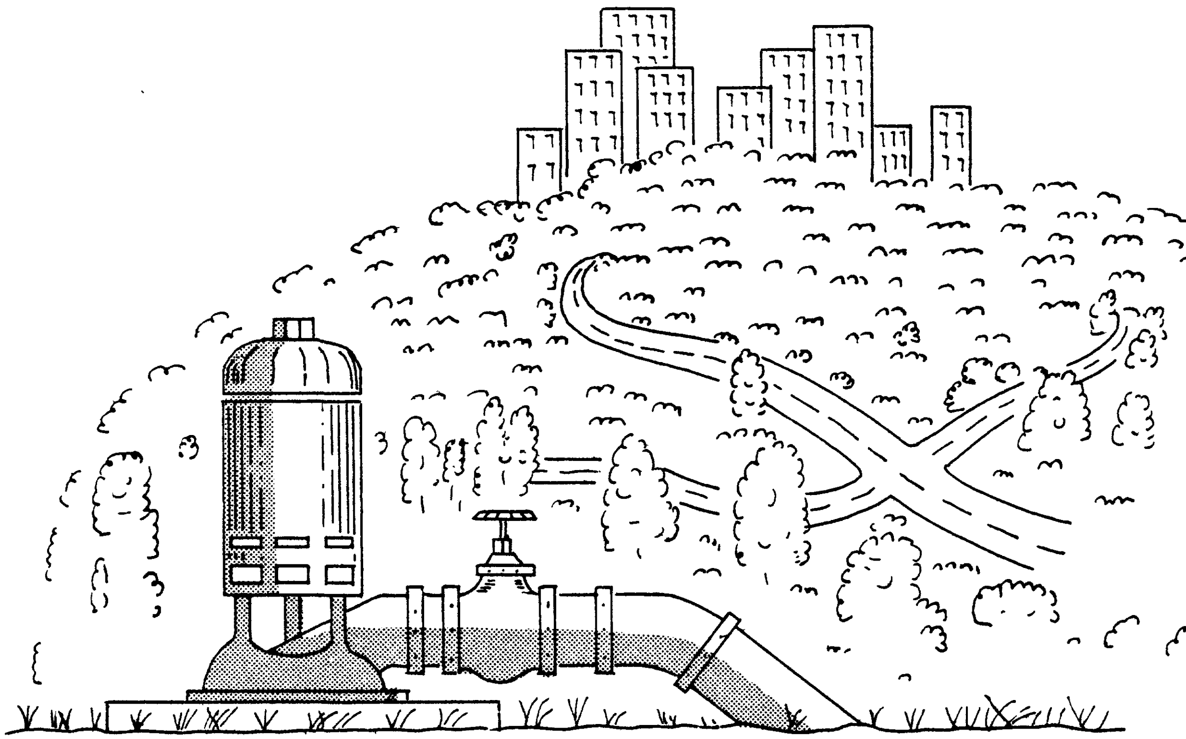
PART I. INTRODUCTION TO GROUND-WATER HYDROLOGY

"... all concerned recognize that the quality of water from the ... River is normally far superior to that of the wells ..."

"However, all concerned are aware of the fact that there are times when the river will be unsuitable as a source of potable water. To provide for such situations, it was agreed that the ... wells ... will be retained and used as the backup supply ... when the chloride concentration in the River becomes too high ..."

-Excerpts from a letter regarding a change in the source of a municipal supply from wells to a stream.

Ground Water As A Resource in North Carolina



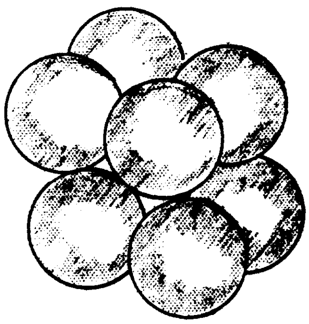
In 1975, North Carolina had 224 municipal water systems serving 500 or more customers. Of these systems, 102 obtained a part or all of their water from wells. These systems serve more than 500,000 people with 70 Mgal/day (million gallons per day) of water. These figures show the importance of ground water in North Carolina as a source of municipal supplies. *They do not, however, show the overall importance of ground water.*

In addition to the people served ground water through municipal systems, more than 2,400,000 people living in rural areas (of a total State population of more than 5,000,000) obtain their water supplies from wells and springs. Ground water is also an important source of water for industries and agriculture.

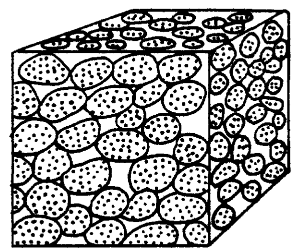
The U.S. Geological Survey estimates that 540 Mgal/day of ground water were being used in North Carolina in 1975. Most of this use is in the Coastal Plain region where ground water is obtained from extensive and productive sand and or limestone aquifers. Although the aquifers underlying the Piedmont and mountains are much less productive than those in the Coastal Plain, ground water is an important source of supply for small cities and industries and is the primary source for rural homes and farms.

Rocks and Water

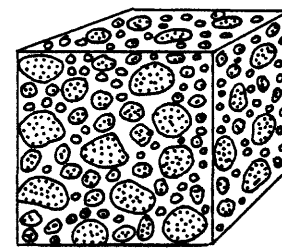
PRIMARY OPENINGS



POROUS MATERIAL

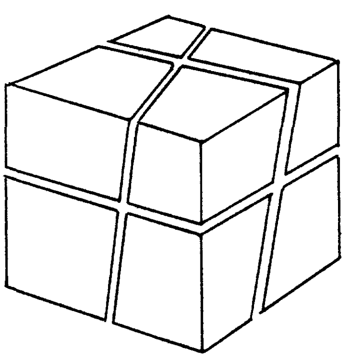


WELL-SORTED SAND

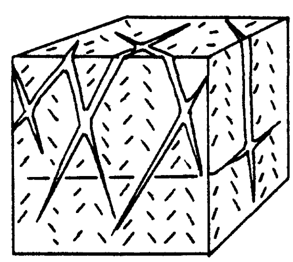


POORLY-SORTED SAND

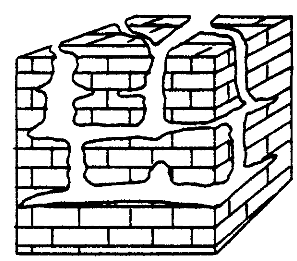
SECONDARY OPENINGS



FRACTURED ROCK



FRACTURES IN GRANITE



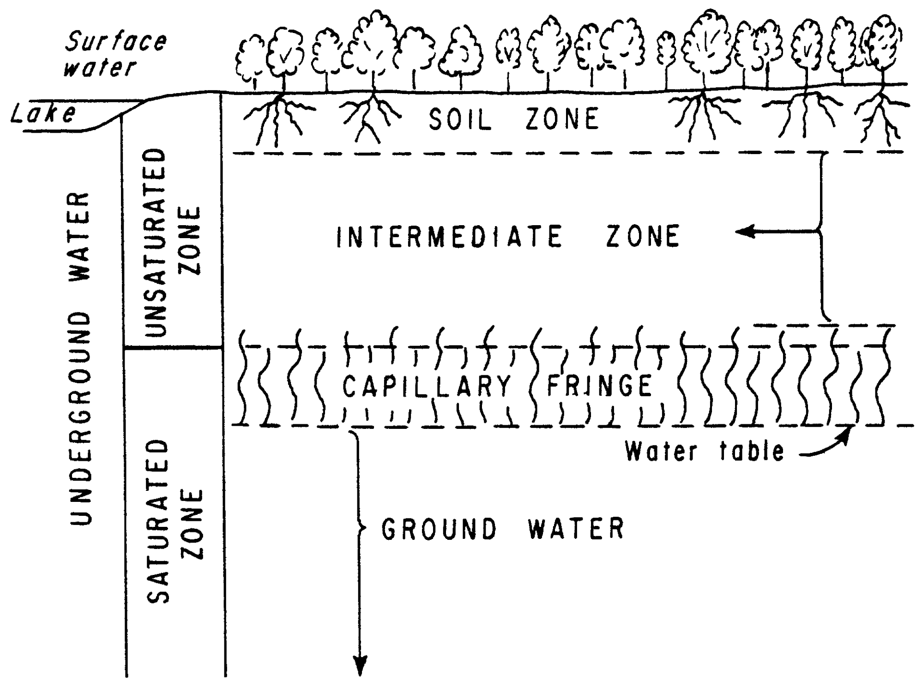
CAVERNS IN LIMESTONE

Most of the rocks near the Earth's surface are composed of both solids and holes. The solid part is, of course, much more obvious than the holes, but without the holes there would be no underground water to supply wells and springs.

There are different kinds of holes in rocks and it is sometimes useful to be aware of these. If the holes were formed at the same time as the rock, they are referred to as *primary openings*. The pores in sand and gravel are primary openings. The sand aquifers underlying the Coastal Plain contain water in primary openings.

If the holes were formed after the rock was formed, they are referred to as *secondary openings*. The fractures in granite and other igneous and metamorphic rocks are secondary openings. Caverns in limestone, which are formed as ground water slowly dissolves the rock, are an especially important type of secondary opening. Ground water in the bedrock underlying the Piedmont and mountains occurs in secondary openings, as does much of that in the limestones underlying the Coastal Plain.

Underground Water



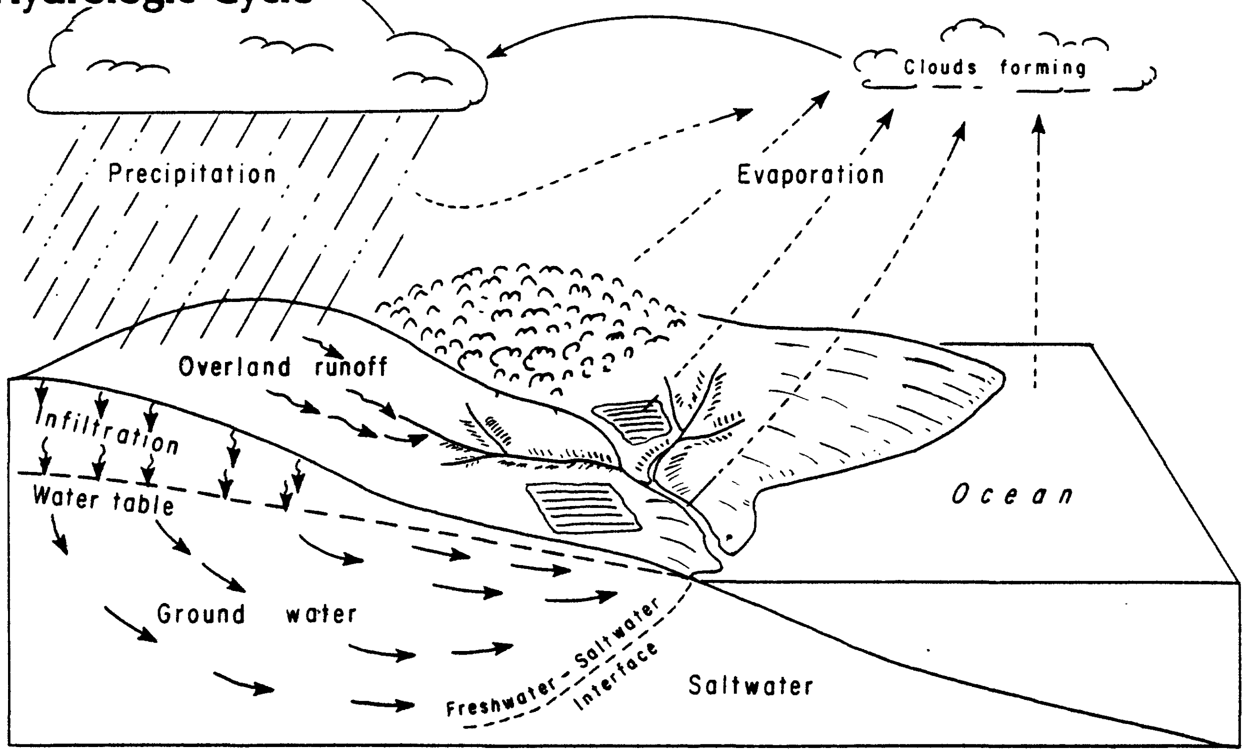
All water beneath the land surface is referred to as *underground water*. The equivalent term for water on the land surface is *surface water.* Underground water occurs in two different zones. One zone, which occurs immediately below the land surface in most areas, contains both water and air and is referred to as the *unsaturated zone*. The unsaturated zone is almost invariably underlain by a zone in which all interconnected openings are full of water. This zone is referred to as the *saturated zone*.

Water in the saturated zone is the only underground water that is available to supply wells and springs and is the only water to which the name *ground water* is correctly applied. Recharge of the saturated zone occurs by percolation of water from the land surface through the unsaturated zone. The unsaturated zone is, therefore, of great importance to ground-water hydrology. This zone may be divided usefully into three parts (or subzones); (1) the soil zone, (2) the intermediate zone, and (3) the capillary fringe.

The upper part - from the land surface to a depth of several feet - is referred to as the *soil zone.* The soil zone is the zone that supports plant growth. It is crisscrossed by living roots, by holes left by decayed roots of earlier vegetation, and by animal and worm burrows. This zone tends to have a higher porosity and to be more permeable than the underlying material. The soil zone is underlain by the *intermediate zone*, which differs in thickness from place to place depending on the thickness of the soil zone and the depth to the capillary fringe.

Moving to the lowest part of the unsaturated zone, the boundary between it and the saturated zone is occupied by the *capillary fringe.* The capillary fringe results from the attraction between water and rocks. As a result of this attraction, water clings as a film on the surface of rock particles and rises up small-diameter pores against the pull of gravity. Water in the capillary fringe and in the overlying part of the unsaturated zone is under a negative hydraulic pressure - that is, it is under a pressure less than atmospheric. The *water table* is the level in the saturated zone at which the hydraulic pressure is equal to atmospheric pressure.

Hydrologic Cycle



The term *hydrologic cycle* is used to refer to the constant movement of water above, on, and below the Earth's surface. The concept of the hydrologic cycle is central to an understanding of the occurrence of water and the development and management of water supplies.

Although the hydrologic cycle has neither a beginning nor an end, it is convenient to discuss its principal features by starting with evaporation from vegetation, from exposed surfaces including the land surface, and from the ocean. This moisture forms clouds which, under favorable conditions, return the water to the land surface or oceans in the form of precipitation.

Precipitation occurs in several forms, including rain, snow, and hail, but we will consider only rain in this discussion. The first rain wets vegetation and other surfaces and then begins to infiltrate into the ground. *Infiltration* rates vary widely, depending on land use, from possibly as much as an inch per hour in mature forests to a tenth of an inch per hour in silty soils under cultivation. When and if the rate of precipitation exceeds the rate of infiltration, *overland flow* occurs.

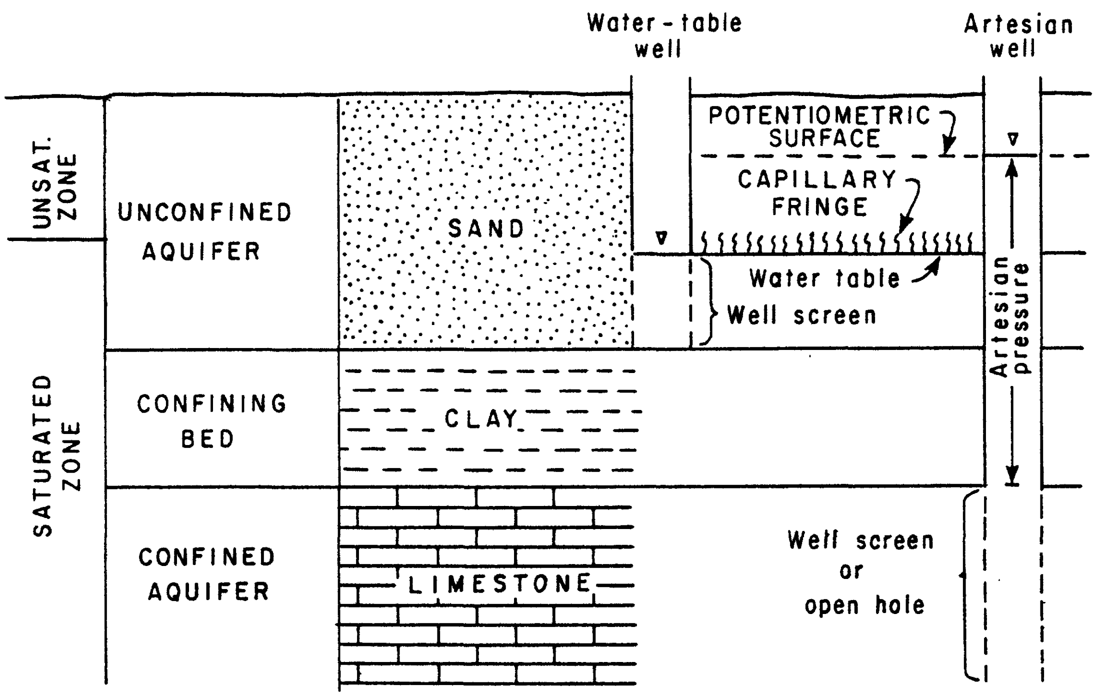
The first infiltration replaces soil moisture and thereafter the excess percolates slowly across the intermediate zone to the zone of saturation. The water in the zone of saturation moves downward and laterally to sites of ground-water discharge such as springs on hillsides or seeps in the bottoms of streams and lakes or beneath the ocean.

Water reaching streams, both by overland flow and from ground-water discharge, moves to the sea where it is again evaporated to perpetuate the cycle.

Movement is, of course, the key element in the concept of the hydrologic cycle. Some "typical" rates of movement are shown in the following table, along with the distribution of the Earth's water supply.

Location	Rate of movement	Distribution of Earth's water supply (percent)
Atmosphere	100s of miles per day	0.001
Water on land surface	10s of miles per day	.02
Water below the land surface	feet per day	.52
Ice caps and glaciers	feet per day	1.88
Oceans	--------	97.58

Aquifers and Confining Beds



From the standpoint of ground-water occurrence, all rocks underlying the Earth's surface are classified either as aquifers or confining beds. An *aquifer* is a rock unit that will yield water in a usable quantity to a well or spring. (In geological usage, "rock" includes unconsolidated sediments.)

A *confining bed* is a rock unit that restricts the movement of ground water either into or out of adjacent aquifers.

Ground water occurs in aquifers under two different conditions. Where water only partly fills an aquifer, the upper surface of the saturated zone is free to rise and decline. The water in such aquifers is said to be *unconfined* and the aquifers are referred to as *unconfined aquifers.*

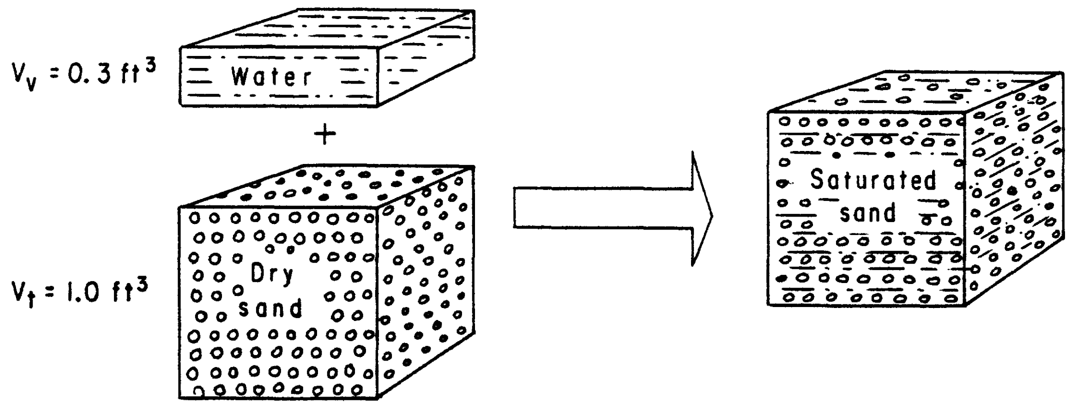
Where water completely fills an aquifer that is overlain by a confining bed, the water in the aquifer is said to be *confined.* Such aquifers are referred to as *confined aquifers.*

Wells open to unconfined aquifers are referred to as *water-table wells.* The water level in these wells indicates the position of the water table in the surrounding aquifer.

Wells drilled into confined aquifers are referred to as *artesian wells.* The water level in artesian wells stands at some height above the top of the aquifer but not necessarily above the land surface.

The static water level in tightly cased wells open to a confined aquifer stands at the level of the *potentiometric surface* of the aquifer.

Porosity of Soils and Rocks



$$\text{Porosity } (\eta) = \frac{\text{Volume of voids } (V_v)}{\text{Total volume } (V_t)} = \frac{0.3\ ft^3}{1.0\ ft^3} = 0.30$$

The openings (holes) in a soil or rock are referred to as its *porosity*. We express porosity either as a decimal fraction or as a percent. Thus

$$\eta = \frac{V_t - V_s}{V_t} = \frac{V_v}{V_t} \qquad (1)$$

where
n = porosity, as a decimal fraction,
V_t = total volume of a soil or rock sample
V_s = volume of solids in the sample, and
V_v = volume of holes (voids).

If we multiply the porosity determined with equation 1 by 100, the result is porosity expressed as a percent.

Soils are among the most porous of natural materials because soil particles tend to form loose clumps and because of the presence of root holes and animal burrows.

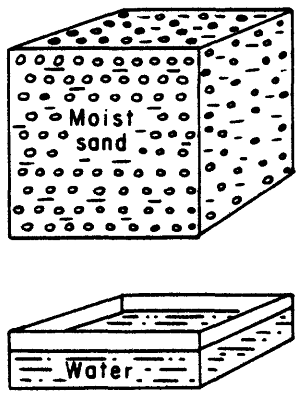
Selected values of porosity
(Percent by volume)

Material	Primary openings	Secondary openings
Equal-size spheres (marbles)		
Loosest packing	48	---
Tighest packing	26	---
Soil	55	---
Clay	50	---
Sand	25	---
Gravel	20	---
Limestone (Castle Hayne)	10	10
Marble (mountains)	--	2
Granite, gneiss, and schist (Piedmont and mountains).	--	0.1

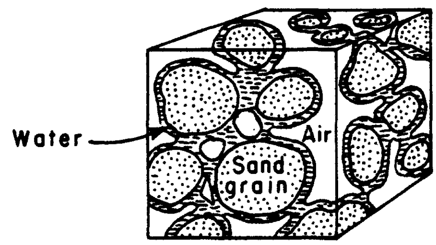
Specific Yield and Specific Retention

$S_r = 0.1 \text{ ft}^3$

$S_y = 0.2 \text{ ft}^3$

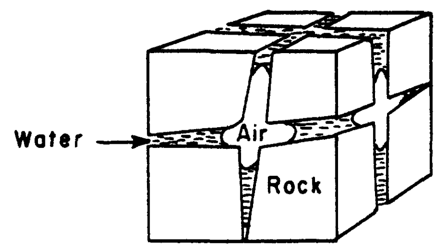


$$n = S_y + S_r = \frac{0.2 \text{ ft}^3}{1 \text{ ft}^3} + \frac{0.1 \text{ ft}^3}{1 \text{ ft}^3} = 0.30$$



GRANULAR MATERIAL

Water retained as a film on rock surfaces and in capillary-size openings after gravity drainage.



FRACTURED ROCK

Porosity is important in ground-water hydrology because it tells us the maximum amount of water a rock can contain when it is saturated. However, it is equally important to know that only a part of this water is available to supply a well or a spring.

Hydrologists divide porosity into the part that will drain under the influence of gravity (called *specific yield*) and the part that is retained as a film on rock surfaces and in very-small openings (called *specific retention*). It should be noted that the physical forces that control specific retention are the same forces involved in the thickness and moisture content of the capillary fringe.

Specific yield tells us how much water is available for man's use and specific retention tells us how much water remains in the rock after it is drained.

Thus

$$\eta = S_y + S_r$$

and

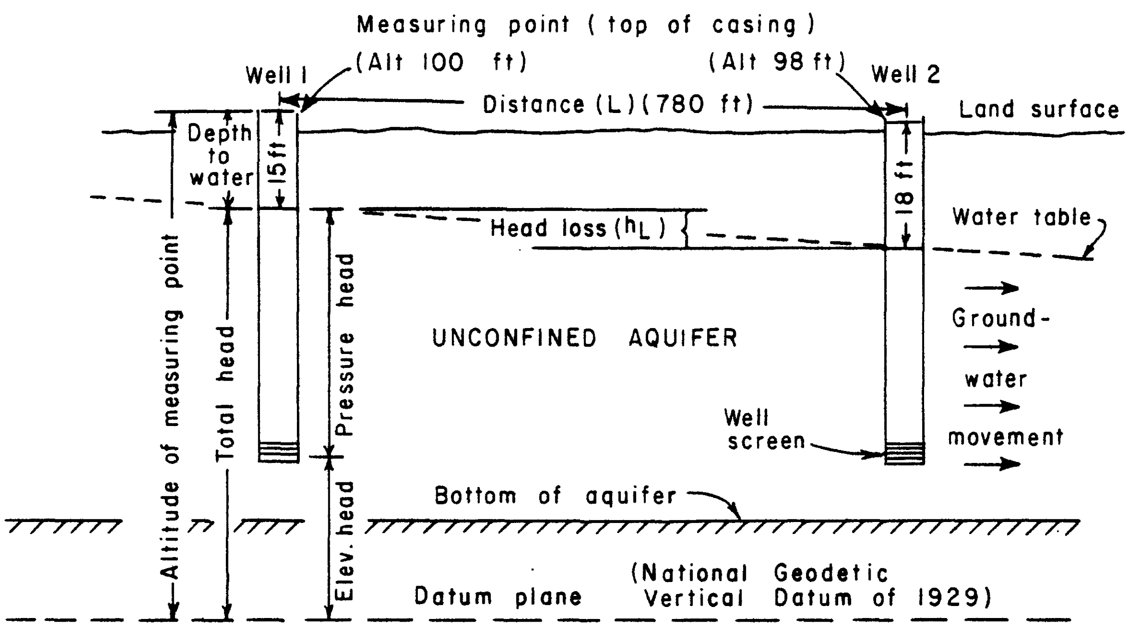
$$S_y = \frac{V_d}{V_t}, \qquad S_r = \frac{V_r}{V_t},$$

where S_y = specific yield,
S_r = specific retention,
V_d = volume of water that drains from a total volume of V_t,
V_r = volume of water retained in a total volume of V_t, and
V_t = total volume of a soil or rock sample.

Selected values of porosity, specific yield, and specific retention
(Percent by volume)

Material	Porosity	Specific yield	Specific retention
Soil	55	40	15
Clay	50	2	48
Sand	25	22	3
Gravel	20	19	1
Limestone (Castle Hayne)	20	18	2
Marble	2	1.8	0.2
Granite, gneiss, and schist	0.1	0.09	0.01

Heads and Gradients



In many ground-water investigations it is either necessary or desirable to know the depth to the water table and the direction and rate of ground-water movement. Both of these require the measurement of the position of the water level in wells.

The first step is to identify (and describe) a fixed point - *measuring point* (MP) - that all measurements will be referred to. The depth to the water level below the measuring point can be measured by any of several means. (See MEASUREMENTS OF WATER LEVELS AND PUMPING RATES.) In order to determine the direction and rate of ground-water movement, it is necessary to measure the *depth to water* (D to W) in at least three wells and to determine the altitude of their *measuring points* with respect to a common *datum plane* - usually mean sea level (now officially the National Geodetic Vertical Datum of 1929).

If the D to W in a nonflowing well is subtracted from the altitude of the MP, the result is the *total head* at the well. Total head, as we know from fluid mechanics, is composed of *elevation head, pressure head,* and *velocity head.* Because ground water moves relatively slowly, we can ignore velocity head. Therefore, the total head at an observation well involves only two components - elevation head and pressure head. Ground water moves in the direction of decreasing total head; this may or may not be in the direction of decreasing pressure head.

The equation for total head (h_s) is

$$h_s = z + h_p \qquad (1)$$

where z is elevation head and is the distance from the datum plane to the point where the pressure head is h_p.

All other factors being constant, the rate of ground-water movement depends on the *hydraulic gradient.* The hydraulic gradient is the change in head per unit of distance in a given direction. If not specified, the direction is understood to be in the direction in which the maximum rate of decrease in head occurs.

If we assume that the movement of ground water is in the plane of the preceding drawing - in other words, that it moves from well 1 to well 2 - we can calculate the hydraulic gradient from the information given on the drawing. The hydraulic gradient is h_L/L where h_L is the head loss between wells 1 and 2 and L is the distance between them, or

$$\frac{h_L}{L} = \frac{(100\text{ ft} - 15\text{ ft}) - (98\text{ft} - 18\text{ft})}{780\text{ ft}} =$$

$$\frac{85\text{ ft} - 80\text{ ft}}{780\text{ ft}} = \frac{5\text{ ft}}{780\text{ ft}}$$

Gradients are usually expressed in feet/mile or feet per 1000 feet. Thus

$$\frac{5}{780} = \frac{x}{5280} \text{ or } x = 33.8\text{ ft/mile.}$$

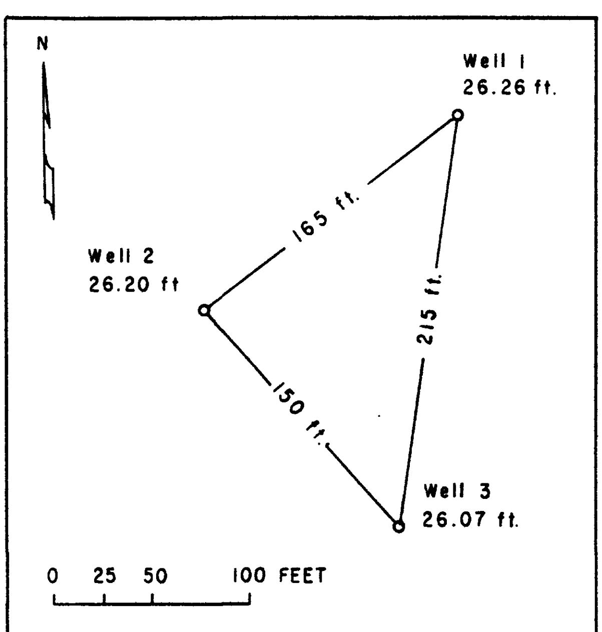
Both the direction of ground-water movement and the hydraulic gradient can be determined if the following data are available for three wells located in any triangular arrangement such as that shown on Sketch A.

1. The relative geographic position of the wells,
2. The distance between the wells, and
3. The relative position of the water level in each well.

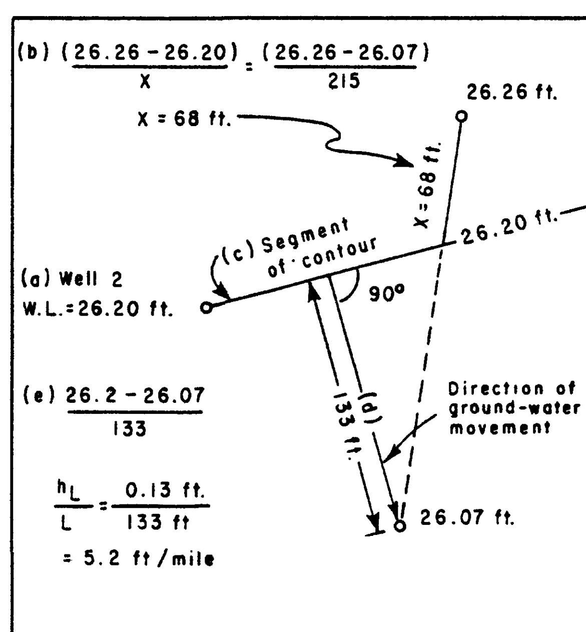
Steps in the solution are outlined below and illustrated on Sketch B.

(a) Identify the well that has the intermediate water level - that is, neither the highest nor the lowest water level.
(b) Calculate the position between the wells having the highest and lowest water levels at which the ground-water level is the same as in the intermediate well.
(c) Draw a straight line between the intermediate well and the point identified in step b between the wells having the highest and lowest water level. This line represents a segment of the water-level contour along which the total head is the same as in the intermediate well.
(d) Draw a line perpendicular to the water-level contour and through either the well with the highest or the lowest water level. This line parallels the direction of ground-water movement.
(e) Divide the difference in water level between the well and the contour by the distance between the well and the contour. The answer is the hydraulic gradient.

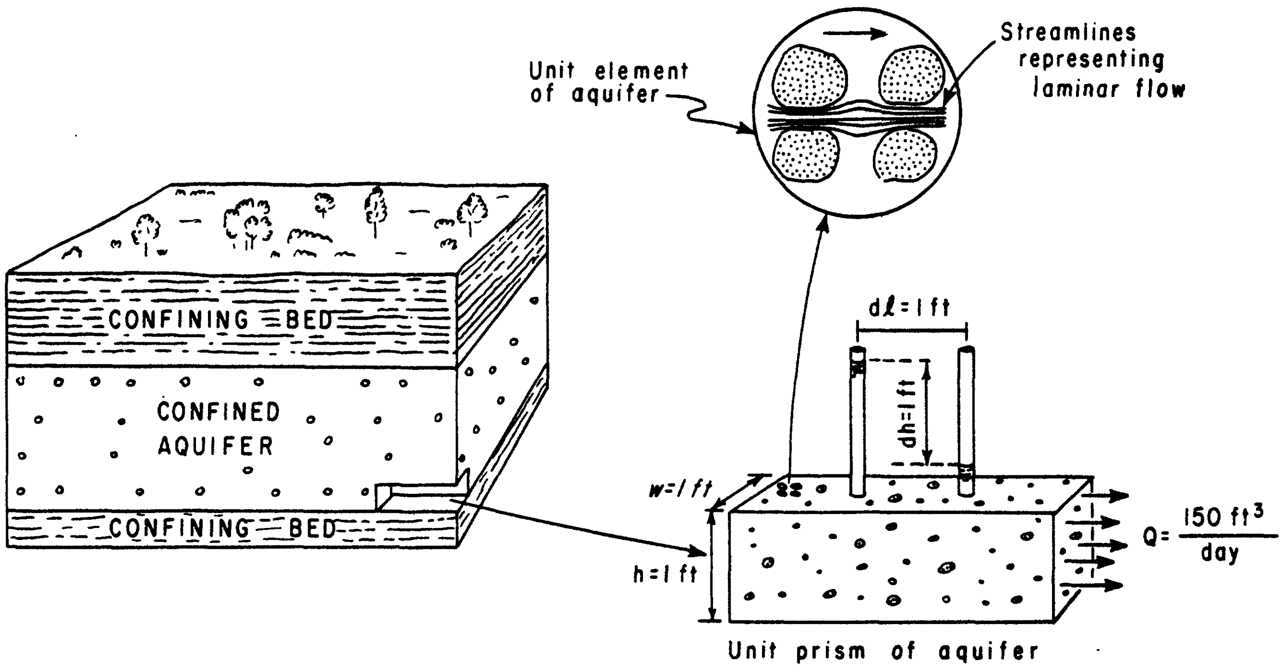
SKETCH A



SKETCH B



Hydraulic Conductivity



Aquifers transmit water from recharge areas to discharge areas and thus function as pipelines. The factors controlling ground-water movement were first expressed in the form of an equation by Henry Darcy in 1856. Darcy's law is

$$Q = KA\frac{dh}{dl} \qquad (1)$$

where

Q is the quantity of water,
K is hydraulic conductivity and depends on the size and arrangement of the water-transmitting openings (pores and fractures),
A is cross-sectional area through which the flow occurs, and
dh/dl is the hydraulic gradient.[1]

[1] Where hydraulic gradient is discussed as an independent entity, as in the preceding discussion of HEADS AND GRADIENTS, it is shown symbolically as h_L/L. Where hydraulic gradient appears as one of the factors in an equation, as in equation 1, it is shown symbolically as dh/dl to be consistent with other ground-water literature.

Because the quantity of water (Q) is directly proportional to the hydraulic gradient (dh/dl), we know that ground-water flow is *laminar* - that is, water particles tend to follow discrete streamlines and not mix with particles in adjacent streamlines.

Hydraulic conductivity is expressed in terms of a unit hydraulic gradient (such as foot per foot) in order to permit ready comparison of the water-transmitting capacity of different materials.

The units of hydraulic conductivity are those of velocity. Thus, if we rearrange equation 1 for K, we obtain

$$K = \frac{Qdl}{Adh} = \frac{(ft^3\ day^{-1})(ft)}{(ft^2)\ (ft)} = \frac{ft}{day} \qquad (2)$$

Hydraulic Conductivity of Selected Rocks

Material	Hydraulic conductivity (rounded values) (ft/day)	[(gal/day)/ft²]	(meters/day)
Coarse sand	200	1500	60
Medium sand	130	1000	40
Silt	1	5	0.2
Clay	0.001	0.01	0.0004
Limestone (Castle Hayne)	300	2000	80
Saprolite	5	50	2
Granite and gneiss	5	50	2
Slate	3	25	1

Hydraulic conductivity replaces the term "field coefficient of permeability" and should be used when referring to the water-transmitting characteristic of material in quantitative terms. It is still permissible to refer in qualitative terms to "permeable" and "impermeable" material.

PROBLEM - Determine the hydraulic conductivity of the confined aquifer shown in the preceding drawing in both feet per day and gallons per day per square foot.

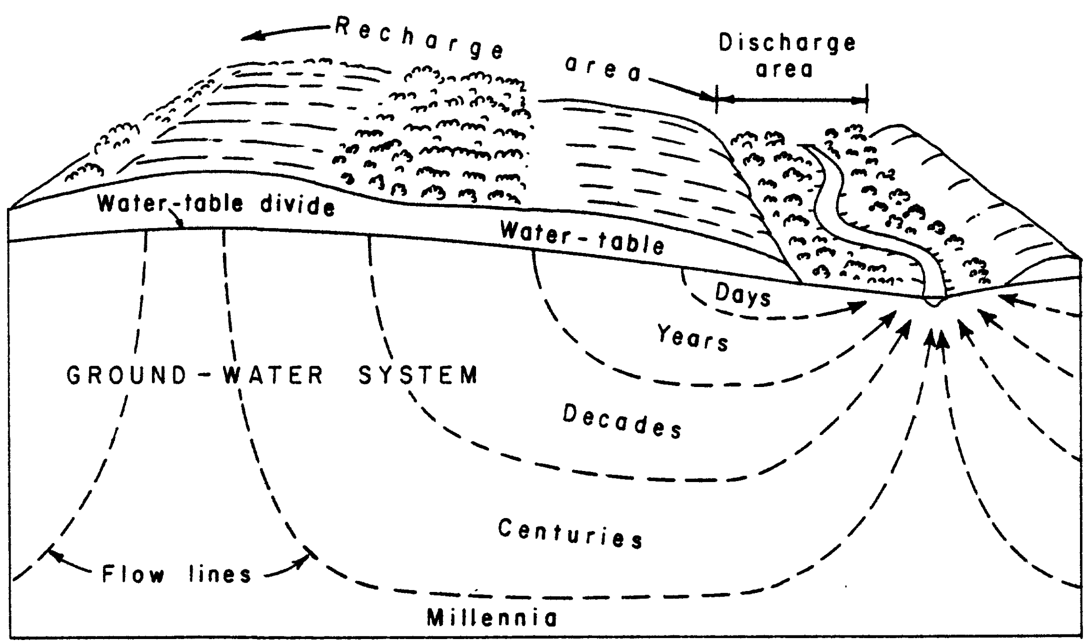
(1) Solution in feet per day

$$K = \overbrace{\frac{Qdl}{Adh}}^{\text{(Equation)}} = \overbrace{\frac{150\ ft^3}{day}}^{(Q)} \times \overbrace{\frac{1}{ft^2}}^{(A)} \times \overbrace{\frac{1\ ft}{1\ ft}}^{(dl/dh)} = \frac{150\ ft^3}{day\ ft^2} = 150\ ft/day$$

(2) Conversion of feet per day to gallons per day per square foot

$$\frac{150\ ft^3}{day\ ft^2} \times \frac{7.5\ gal}{ft^3} = 1125\ (gal/d)/ft^2$$

Functions of Ground-Water Systems



The aquifers and confining beds underlying any area comprise the *ground-water system* of the area. Hydraulically, this system serves two functions: (1) it stores water to the extent of its porosity, and (2) it transmits water from recharge areas to discharge areas. Thus, a ground-water system serves both as a reservoir and as a pipeline. With the exception of cavernous limestones and lava flows, ground-water systems are more effective as reservoirs than as pipelines.

Water enters ground-water systems in *recharge areas* and moves through them, as dictated by hydraulic gradients and hydraulic conductivities, to *discharge areas.*

The identification of recharge areas is becoming increasingly important because of the expanding use of the land surface for waste disposal. In a humid area, such as North Carolina, recharge occurs in all interstream areas - that is, in all areas except along streams and their adjoining flood plains. The streams and flood plains are, under most conditions, discharge areas.

Recharge rates are generally expressed in terms of volume (such as gallons or ft^3), per unit of time (such as a day or a year), and per unit of area (such as a square mile or acre). When the units are reduced to their simpliest form, the result is recharge expressed as a depth of water on the land surface per unit of time. Recharge rates vary from year to year, depending on the amount of precipitation, its seasonal distribution, air temperature, and other factors. Among the other factors are land use. For example, recharge rates are much higher in forest than in cities.

Relatively few estimates of recharge rates have been made in North Carolina. The information presently available suggests that rates in the Piedmont and mountains range from about 100,000 gallons per day per square mile in the areas underlain by Triassic rocks to about 250,000 $(gal/d)/mi^2$ in areas underlain by granite and gneiss. Rates in the Coastal Plain are believed to range from about 250,000 $(gal/d)/mi^2$ in areas underlain by clayey soils to 1,000,000 $(gal/d)/mi^2$ in areas underlain by thick sandy soils.

The rate of movement of ground water from recharge areas to discharge areas depends on the hydraulic conductivities of the aquifers and confining beds through which the water moves and on the hydraulic gradients. (See GROUND-WATER VELOCITY.) A convenient way of showing the rate is in terms of the time required for ground water to move from different parts of a recharge area to the nearest discharge area. The time ranges from a few days in the zone adjacent to the discharge area to thousands of years (millennia) for water that moves from the central part of the recharge area through the deeper parts of the system.

Natural discharge from ground-water systems not only includes the flow of springs and the seepage of water into stream channels, but also evaporation from the upper part of the capillary fringe where it occurs within a few feet of the land surface. Large amounts of water are also withdrawn from the capillary fringe and zone of saturation by plants during the growing season. Thus, discharge areas not only include the channels of perennial streams but also the adjoining flood plains and other low-lying areas.

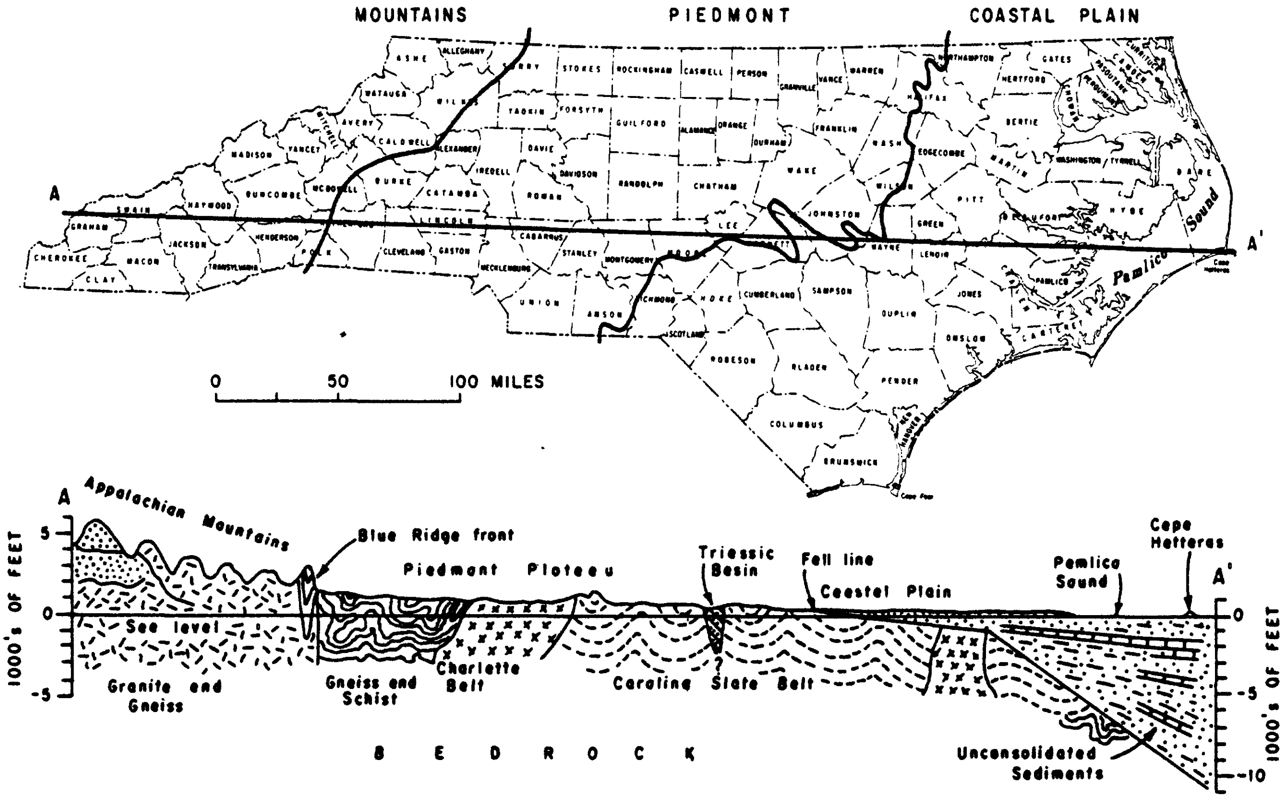
One of the most significant differences between recharge areas and discharge areas is that discharge areas are invariably much smaller in areal extent that recharge areas. This shows, as we would expect, that discharge areas are more "efficient" than recharge areas. Recharge involves unsaturated movement of water in the vertical direction; in other words, in the direction in which the hydraulic conductivity is generally the lowest, whereas discharge involves saturated movement, much of it in the horizontal direction - that is, in the direction of the largest hydraulic conductivity.

PART II. GROUND-WATER GEOLOGY OF NORTH CAROLINA

"It is time that we return to examining hydrologic systems and attempting to describe the systems in a more realistic, quantitative manner. When one comes to solving problems of chemical contamination, it is necessary to quantify the distribution of porosity, dispersivity, and other factors. Dispersivity measured at only a few field locations tends to be large; from three to more than five orders of magnitude larger than those measured in the laboratory. This suggests that the geologic complexity of real aquifers greatly complicates the flow field, causing increased hydrodynamic dispersion.

-From remarks by Dr. John Bredehoeft at hearings on "Ground-water quality research and development" before the Subcommittee on Environment and the Atmosphere, 95th Congress, 2nd Session, April 1978, p. 236.

Physical Setting of the Ground-Water System



From the standpoint of ground-water hydrology, North Carolina may be divided into two zones, one zone consisting of the Coastal Plain and the other consisting of the Piedmont Plateau and the Appalachian Mountains. Because differences in the ground-water system coincide with the different topographic divisions of the State, it will be useful to briefly review these divisions.

As Jasper L. Stuckey, former North Carolina State Geologist, has said, "The State of North Carolina extends from the crest of the Great Smoky and Unaka mountains on the west, to the Atlantic Ocean on the east and lies across three major topographic provinces of the United States. As a result, it is divided into three natural divisions—the Coastal Plain on the east, the Piedmont Plateau in the center, and the Appalachian Mountains on the west. Beginning at sea level at the eastern edge of the State the surface of North Carolina rises gradually in elevation and increases in irregularity until it reaches its maximum height and ruggedness in the Appalachian Mountains on the west."

The *Coastal Plain* includes almost one-half of the area of the State and extends west from the Atlantic Ocean to the *Fall Line.* The Fall Line is not a line but a zone 30 to 40 miles wide that is marked by discontinuous rapids where major streams leave the bedrock areas of the Piedmont and flow onto the unconsolidated sediments of the Coastal Plain. Altitudes in the Coastal Plain range from sea level at the coast to about 300 to 500 ft. along the Fall Line. The Coastal Plain can conveniently be divided into the Tidewater Region, in which the effect of tides and other oceanic influences are apparent, and the Inner Coastal Plain which, though underlain by unconsolidated (Coastal Plain) sediments, is not subject to direct oceanic effect.

The *Piedmont Plateau* contains about 20,000 mi^2, or two-fifths of the land area of the State. It lies between the Coastal Plain on the east and the Appalachian Mountains on the west. Altitudes in the Piedmont range from about 500 ft above sea level along the Fall Line to about 1500 to 2000 ft. along its western border. The Piedmont consists of well-rounded hills and long-rolling ridges with a northeast-southwest trend. Parts of the Piedmont contain prominent hills referred to as mountains, including the Uwharrie Mountains in Montgomery and Randolph Counties, the South Mountains in Burke and Rutherford Counties, and the Brushy Mountains in Wilkes County.

The Appalachian Mountains are bounded on the east by the Blue Ridge Mountains and on the west by the Great Smoky and Unaka Mountains. The mountain slopes are gentle, presenting smooth rounded outlines. The mountain region of North Carolina contains the highest peak east of the Mississippi, Mt. Mitchell at 6,684 ft., 43 peaks above 6,000 ft, and 82 peaks between 5,000 and 6,000 ft. in altitude. The eastern Continental Divide follows the Blue Ridge Mountains so that most of the mountain area drains west to the Gulf of Mexico. The streams are well graded and cascades and waterfalls are only locally abundant.

Reference: Stuckey, Jasper L., 1965, North Carolina: its *geology and mineral resources*: North Carolina Department of Conservation and Development, 550 p.

Water-Bearing Rocks

The rocks underlying the surface of North Carolina form the environment in which ground water occurs and moves.

Geologists divide all rocks exposed at the Earth's surface into one of two great classes: (1) igneous, or (2) sedimentary. *Igneous rocks* are those that have formed from a molten or partially-molten state. Some types of igneous rocks, including granite, solidify at great depth below the land surface and are referred to as intrusive igneous rocks. Other igneous rocks form from lava or volcanic ash ejected onto the surface and are referred to as extrusive igneous rocks.

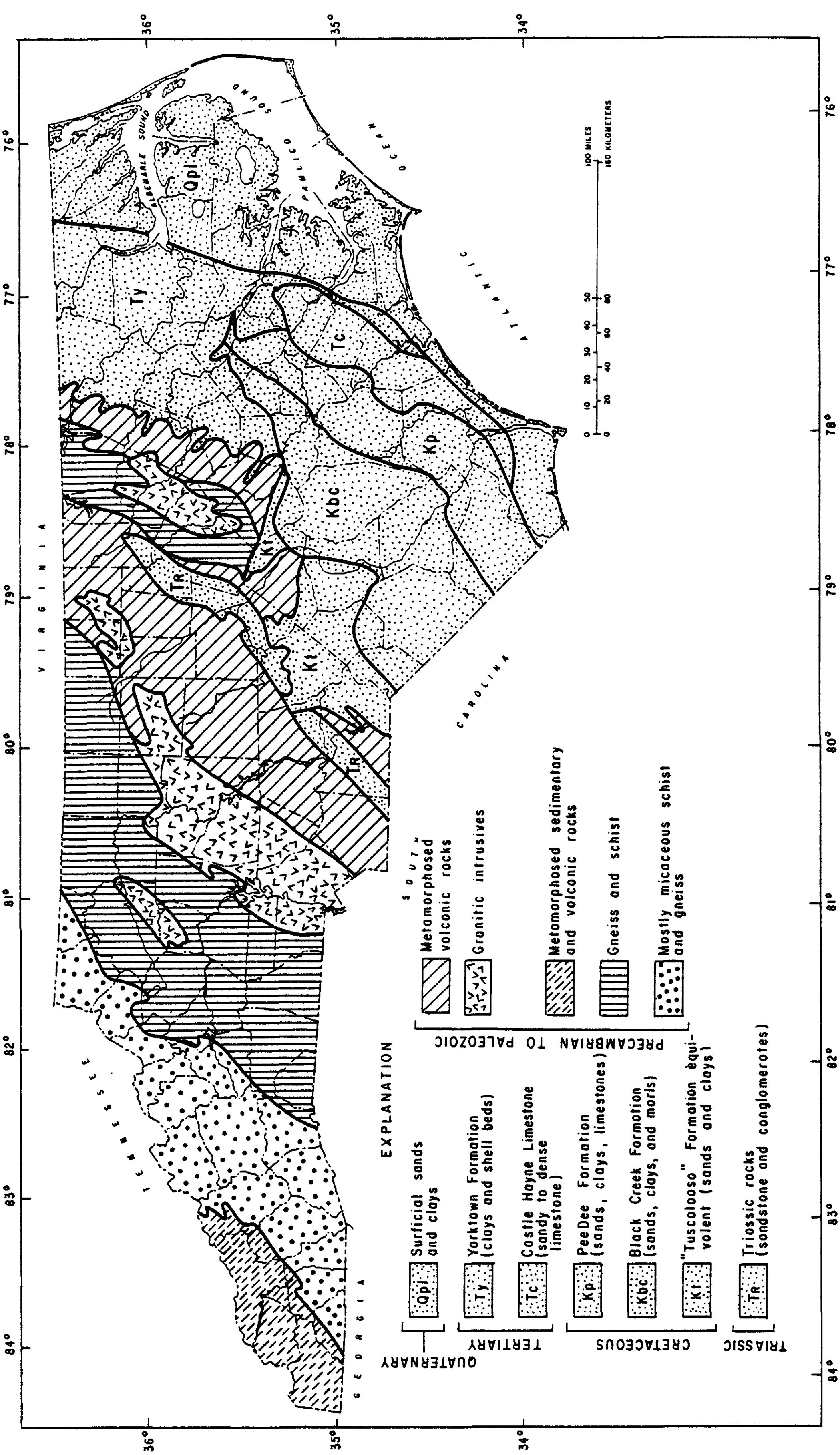
Sedimentary rocks are rocks formed by the accumulation of sediment in water or from the air. Most sedimentary rocks are unconsolidated (soil-like) at the time of formation. If they are, in time, buried deeply enough, or if they undergo certain chemical changes, they may become consolidated.

Both igneous and sedimentary rocks may, over the course of geologic time, reach depths beneath the Earth's crust at which they are subjected to great heat and pressure. This may alter both their structural characteristics and their mineral composition to such an extent that they are changed into *metamorphic rocks.* Depending on their original mode of origin, they may be referred to, for example, as metavolcanic or metasedimentary rocks.

North Carolina is underlain by an unusually large number of different types of rocks, including representatives of both the igneous and sedimentary classes and types of both classes that have been subjected to metamorphism. The major types of rocks are shown on the accompanying generalized geological map.

The Piedmont and mountain regions are underlain by igneous and metamorphosed igneous and sedimentary rocks that are referred to collectively as *bedrock.* They form broad northeast - southwest trending zones in which the rocks are of similar composition and origin. Most of these rocks were formed in the Precambrian and Paleozoic Erasof the Earth's history and thus are at least several hundred million years old. The bedrock in the Piedmont and mountains is exposed at the surface along steep hillsides and stream channels and in roadcuts. In most other areas they are covered by unconsolidated material formed from the breakdown of the bedrock in the process referred to by geologists as weathering. This layer of weathered material is referred to as *saprolite* or *residuum.*

The Coastal Plain region is also underlain by the same types of igneous and methamorphic rocks as those present in the Piedmont. However, in the Coastal Plain they are covered by unconcolidated sedimentary deposits which range in thickness from a few feet along the Fall Line to about 10,000 ft. at Cape Hatteras. (See the geologic section in PHYSICAL SETTING OF THE GROUND-WATER SYSTEM.) The sediments underlying the Coastal Plain include sand, clay, beds composed of seashells, and limestone.


EXPLANATION
QUATERNARY
Qpl Surficial sands and clays
TERTIARY
Ty Yorktown Formation (clays and shell beds)
Tc Castle Hayne Limestone (sandy to dense limestone)
CRETACEOUS
Kp PeeDee Formation (sands, clays, limestones)
Kbc Black Creek Formation (sands, clays, and marls)
Kt "Tuscaloosa" Formation equivalent (sands and clays)
TRIASSIC
TR Triassic rocks (sandstone and conglomerates)
PRECAMBRIAN TO PALEOZOIC
Metamorphosed volcanic rocks
Granitic intrusives
Metamorphosed sedimentary and volcanic rocks
Gneiss and schist
Mostly micaceous schist and gneiss
VIRGINIA
TENNESSEE
GEORGIA
SOUTH CAROLINA
ATLANTIC OCEAN
ALBEMARLE SOUND
PAMLICO SOUND
100 MILES
160 KILOMETERS

Rock Units and Aquifers in the Coastal Plain

The Coastal Plain of North Carolina is underlain by sedimentary rocks that were deposited in water in several different layers which geologists refer to as formations. Formations are commonly given names for places near which they are exposed at the land surface, for ease in referring to them in geologic literature.

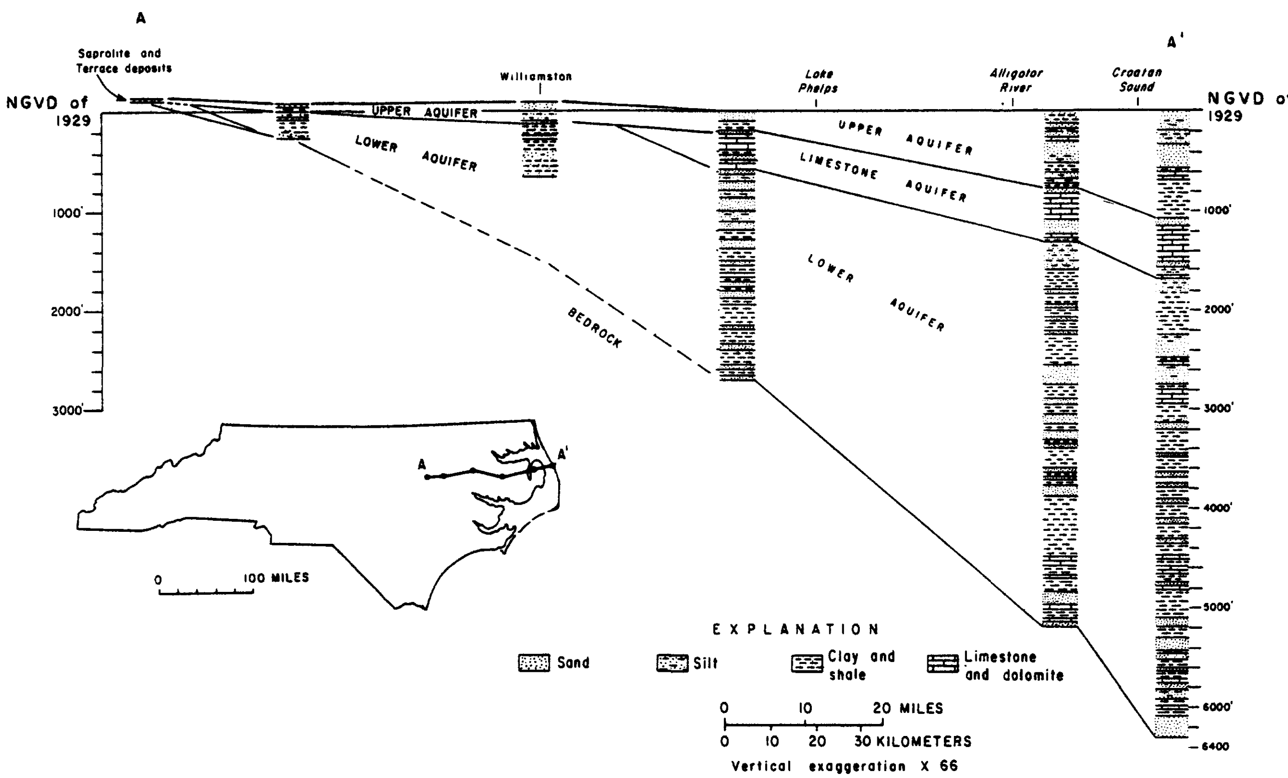
Rock layers are normally given names by geologists if they have a distinct composition or, if of variable composition, include materials deposited during a particular segment of geologic time. Named rock units may or may not coincide with hydrologic units so that in ground-water reports some aquifers may be referred to by the formal geologic names used by geologists, such as the Castle Hayne Limestone and Yorktown Formation, and others may be given more informal names, such as the "Upper aquifer", "Surficial aquifer" or "post-Miocene deposits." The name "Surficial aquifer" indicates the aquifer in any area that is closest to land surface and thus is clearly identifiable, regardless of any other names that may have been assigned to that rock unit.

Non-geologists concerned both with ground-water problems and with ground-water studies in the Coastal Plain are probably confused by the different named applied to the different hydrologic units. The following chart was prepared in an effort to eliminate some of this confusion. We should note, however, that in preparing the chart we have neither tried to include all formation names nor been overly concerned with the relative geologic age of the formations. The names used in the last two columns can be confusing to the extent that all units are referred to as "aquifers." Confining beds composed of clay occur in all of the formations and in the formations of Cretaceous age clay comprises about half of the total thickness.

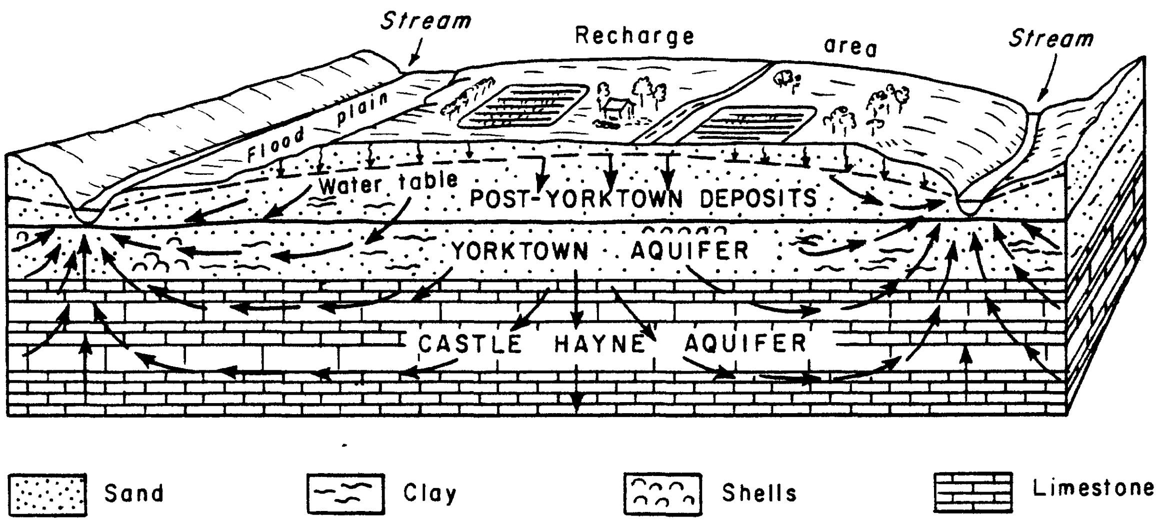
The two most important aquifers in the Coastal Plain are the upper aquifer and the limestone aquifer. Recharge of the ground-water system is from precipitation on the land surface. Therefore, the surficial aquifer has the largest yield in terms of rate per unit area (for example, gallons per minute per square mile). The upper aquifer is also most subject to pollution from land-surface waste disposal. The limestone aquifer is the most productive aquifer in North Carolina in terms of yields of individual wells. Wells capable of yielding more than 1000 gal/min can readily be developed in this aquifer.

Geologic age	Formation or geologic name	Names used in some ground-water reports	Simpliest useful hydrologic names
Pleistocene	Pleistocene deposits	Post-Miocene deposits	Upper aquifer
	Croatan Formation		
Pliocene	Yorktown Formation	Yorktown aquifer	
Miocene	Pungo River Formation		
	Belgrade Formation		
Oligocene	River Bend Formation	Castle Hayne aquifer	Limestone aquifer
Eocene	Castle Hayne Limestone		
Paleocene	Beaufort Formation	Beaufort aquifer	Lower aquifer
Cretaceous	Peedee Formation	Cretaceous aquifer	
	Black Creek Formation		
	"Tuscaloosa" Formation		

The complex interlaying of the sediments underlying the Coastal Plain is shown in the following cross section. It will also be observed from the cross section that the rock layers (and formations) underlying the Coastal Plain dip toward the coast at a rate of about 15 ft./mi. As a result, each formation occurs at a greater depth below land surface toward the coast.



Ground-Water Situation in the Coastal Plain



Recharge of the ground-water system in the Coastal Plain occurs in the "upland" areas above the flood plains of perennial streams. Water reaching the saturated zone moves downward and laterally through the system to discharge areas.

Ground-water discharge occurs by seepage through the bottoms and sides of streams and drainage ditches and also through evaporation from the top of the capillary fringe in flood plains and other areas in which the water table is within several feet of the land surface During the growing season, ground water is also used by plants whose roots reach the capillary fringe or saturated zone. In the area adjacent to the coast, ground water also discharges by seepage into the sides and bottoms of estuaries and the ocean.

The presence of clay layers in the Coastal Plain formations hampers recharge to the deeper aquifers, so that most of the recharge tends to move laterally to discharge areas through the shallowest aquifers. Recharge to the deepest aquifers occurs only in the central part of the interstream recharge areas. This is an important point relative to waste disposal, in that if pollution of the deeper aquifers is to be avoided, waste disposal sites should be located as close as possible to perennial streams.

Prior to the construction of drainage ditches into the central part of the interstream areas, the water table reached the land surface in these areas during the fall, winter, and early spring recharge season. As a result, water was ponded on the surface for periods of several months each year in high-level swamps referred to as *pocosins*.

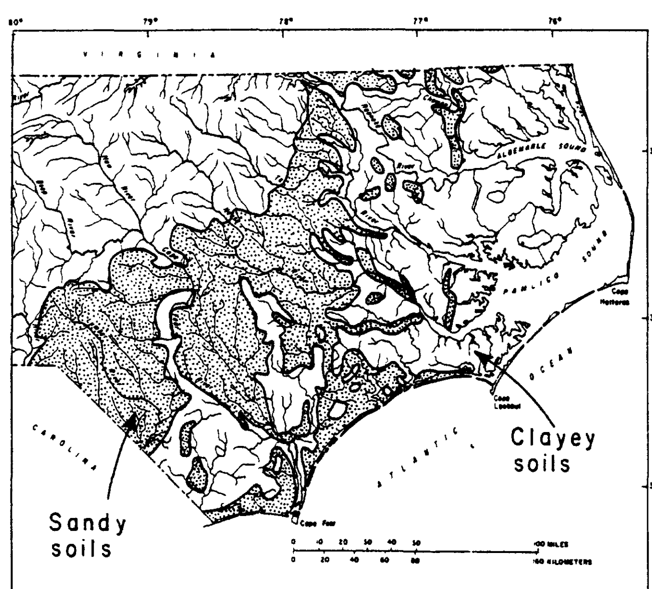
Two regional aspects of the Coastal Plain are of primary importance from the standpoint of ground-water occurrence and availability. The first is the nature of the surficial materials, which controls the recharge to the ground-water system. The second is the geologic conditions that control the occurrence of aquifers and confining beds.

The materials forming the surface of the Coastal Plain can usefully be divided into sandy and clayey soils on the basis of their effect on ground-water recharge. Recharge in the areas underlain by sandy soils is much more effective than in the areas underlain by clayey soils. One of the consequences of this is a much larger sustained base flow of streams in these areas.

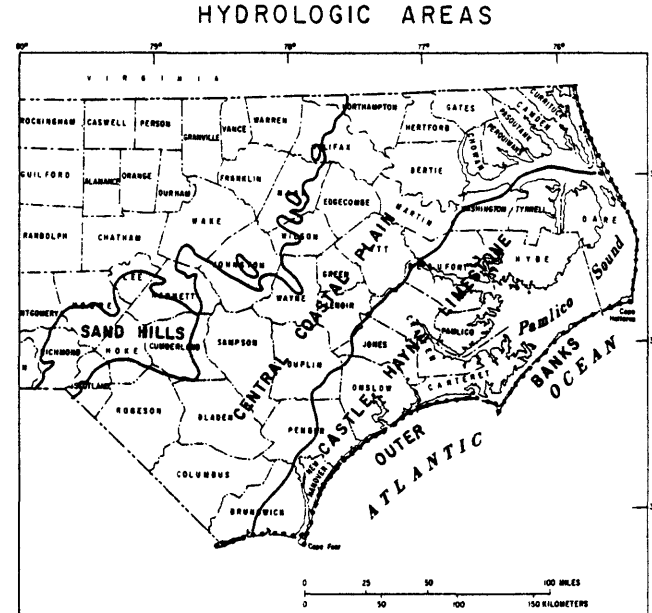
The geologic conditions in the area of about 25,000 mi² occupied by the Coastal Plain differ significantly from one part of the region to another. These differences affect both the occurrence and the availability of ground water. As an aid to understanding the Coastal Plain ground-water system, it is useful to divide the region into four hydrologic areas. Information on these is summarized in the following table in which the areas are listed in order from the simpliest to the most complex.

Hydrologic area	Geohydrologic characteristics	Yield of the most productive wells (gal/min)	Remarks
Sand Hills	Productive water-bearing sand at the land surface and overlying, for the most part, much less-productive material.	50-250	Water obtained from both bored wells and from screened drilled and driven wells. Water only slightly mineralized.
Outer Banks	Productive water-bearing sand at the land surface containing fresh water in contact with sea water.	25-100	Water obtained from both shallow, vertical, screened wells, and from horizontal collectors. Fresh-water zone subject to salt-water encroachment both from above and below.
Castle Hayne Limestone	Productive limestone overlain and underlain by less-productive sand interbedded with clay.	more than 1000	Drilled open-hole wells. Water moderately hard. Aquifer is confined and large withdrawals affect a large area.
Central Coastal Plain	Numerous thin layers of water-bearing sand complexly interbedded with clay.	250-1000	Sediments of Cretaceous age comprise most productive zones and is tapped by multiple-screened drilled wells. Water of excellent quality and, in places, naturally softened. Surficial sand aquifer also widely used for domestic supplies.

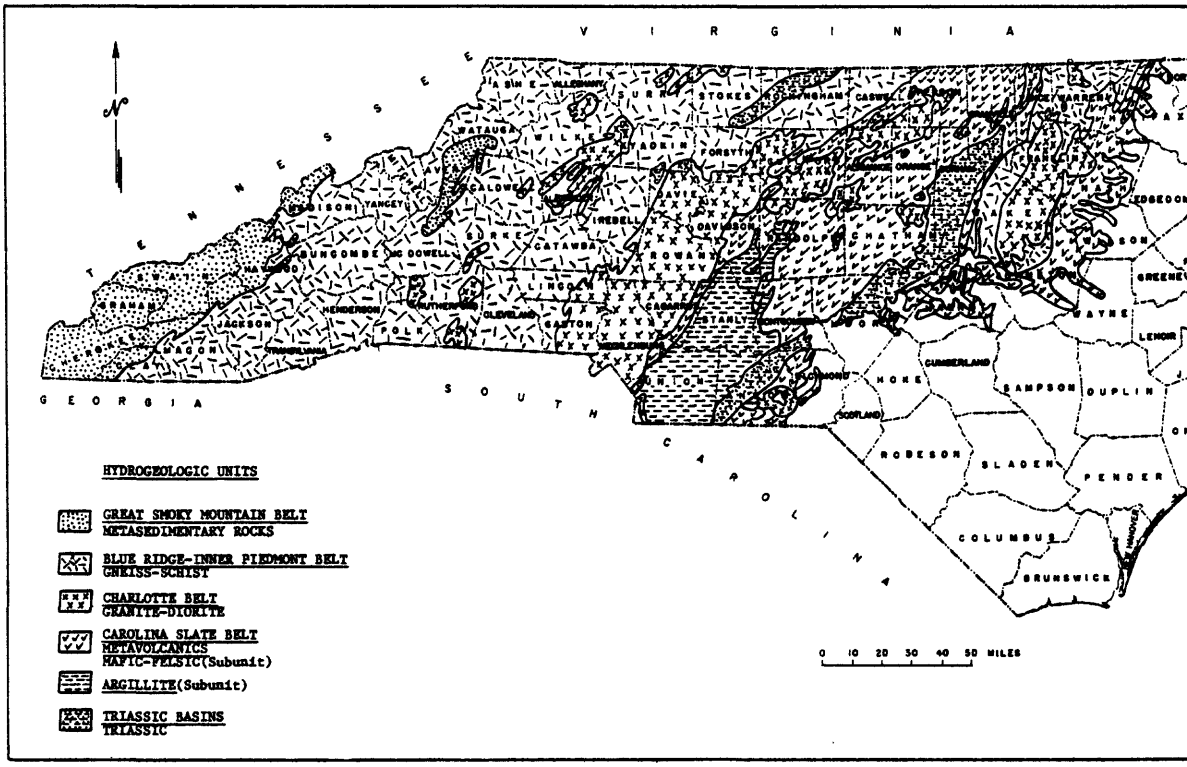
SOILS



HYDROLOGIC AREAS



Rock Units and Aquifers in the Piedmont and Mountains



The rocks underlying the Piedmont and mountains can be divided into two groups: (1) bedrock, and (2) saprolite (or residuum). The saprolite underlies the land surface and ranges in thickness from a foot or two near bedrock outcrops to more than 100 ft. Bedrock underlies the saprolite and is the parent rock from which the saprolite was derived in the process referred to as weathering.

Many stream valleys, especially those of larger streams, are underlain by a layer of material similar in composition to saprolite. This material, which has been deposited by the streams during floods, is correctly referred to as *alluvium*. However, to avoid unnecessary complications, we will lump the alluvium in with the saprolite for the purpose of this discussion.

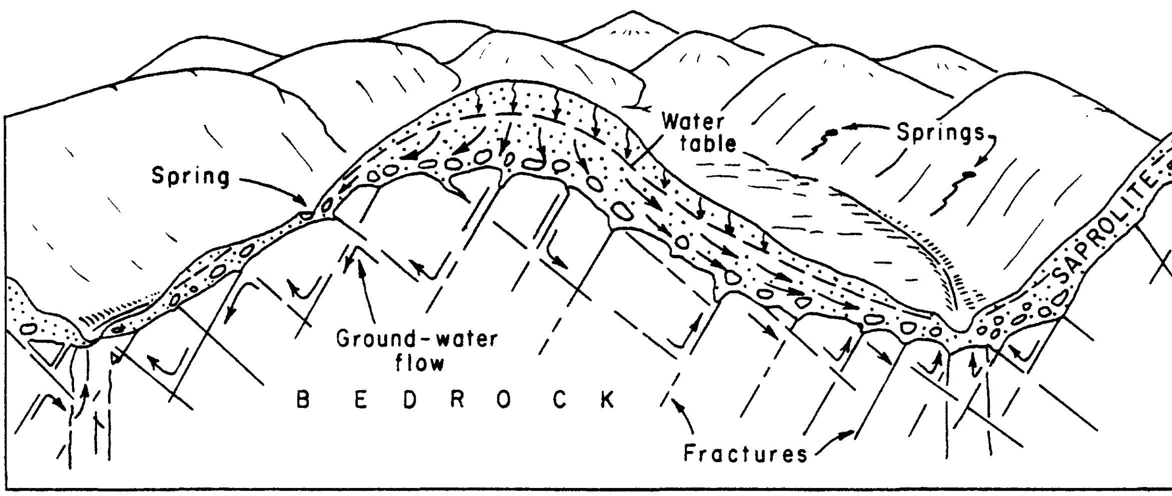
The bedrock underlying the Piedmont and mountains consists of many different types of igneous and metamorphosed igneous and sedimentary rocks. The Generalized Geological Map of North Carolina accompanying the discussion of WATER-BEARING ROCKS divides the bedrock in the Piedmont and mountains into six units. The 1:500,000 scale Geologic Map of North Carolina, published in 1958, divides the bedrock in the same area into 48 different units. But, a much larger number of units have been identified and are shown on large scale geologic maps.

The bedrock units differ from each other in mineral composition and other geologic characteristics. Fortunately, these differences do not result in large differences in hydraulic characteristics so that it is possible to combine the bedrock units into a relatively small number of hydrogeologic units.

The accompanying map shows the hydrogeologic units into which the bedrock in the Piedmont and mountains has been divided by the U.S. Geological Survey and the North Carolina Groundwater Section.

The most productive hydrogeologic units are the Great Smoky Mountain belt and the Blue Ridge-Inner Piedmont belt. The least productive units are the Carolina Slate Belt and the Triassic Basins. The Charlotte Belt is intermediate in productivity.

Ground-Water Situation in the Piedmont and Mountains



The *saprolite* (weathered rock) that forms the land surface in the Piedmont and mountains consists of unconsolidated granular material. It thus contains water in the pore spaces between rock particles.

The *bedrock*, on the other hand, does not have any significant intergranular (primary) porosity. It contains water, instead, in sheet-like openings formed along fractures (that is, breaks in the otherwise "solid" rock). Fractures in bedrock are of two types: (1) *joints*, which are breaks along which there has been no differential movement; and (2) *faults*, which are breaks along which the adjacent rocks have undergone differential movement.

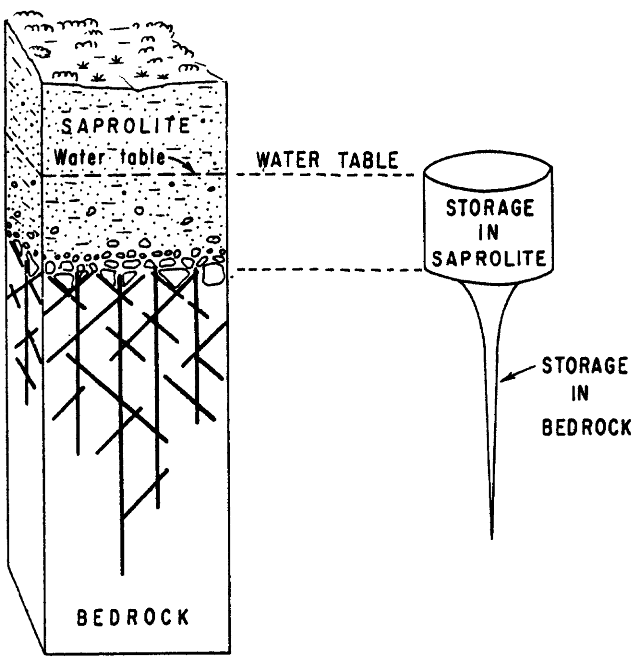
Faults are formed during earthquakes and generally contain larger and more extensive openings than those developed along joints. Joints, however, are far more numerous than faults.

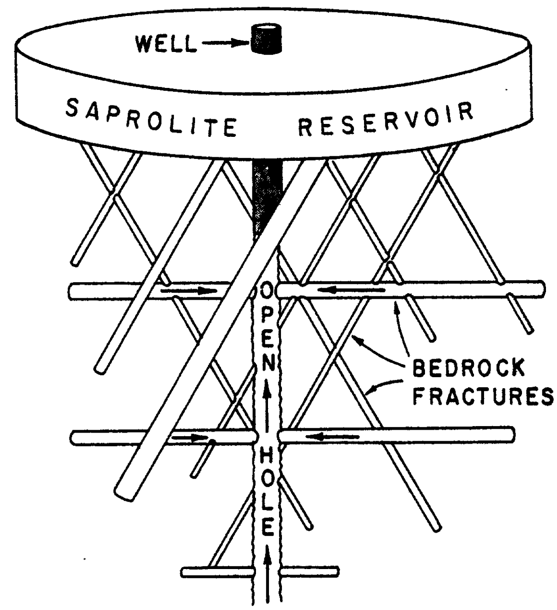
Fractures (joints and faults) are more abundant under valleys, draws, and other surface depressions than under hills. In fact, geologists assume that it is the presence of fractures that determined the position of valleys in the first place. Fractures tend to be more closely-spaced and the openings developed along them tend to be larger near the surface of the bedrock. Most fractures appear to be non water-bearing below a depth of 300 to 400 ft. Large water-bearing openings, penetrated below this depth are probably associated with faults.

The ground-water system in the Piedmont and mountains is recharged by precipitation on the interstream areas. A part of the precipitation infiltrates through the unsaturated zone to the water table, which normally occurs in the saprolite.

Ground water moves laterally and downward through the saprolite to points of ground-water seepage (springs) on the hillsides and to the streams in the adjacent valleys. Some of the water in the saprolite also moves downward into the bedrock and, thereafter, through the fractures to the adjacent valleys.

Hydraulic Characteristics of the Piedmont and Mountain Ground-Water System





One of the most basic concepts of ground-water hydrology is that aquifers function both as reservoirs, in which water is in storage, and as pipelines, which transmit water from one point to another. This is referred to as the *reservoir-pipeline concept.* This concept forms a useful basis on which to discuss the hydraulic characteristics of the Piedmont and mountain ground-water system.

The reservoir (storage) function of aquifers depends on the porosity. The pipeline function depends on the hydraulic conductivity and the thickness of the aquifer. The approximate range in porosity and hydraulic conductivity for the saprolite and bedrock is shown in the following table.

Rock type	**Porosity in percent**	**Hydraulic conductivity in feet per day**
Saprolite	20-30	1-20
Bedrock	0.1-1	1-20

The above values suggest that the principal difference between saprolite and bedrock is in water-storage capacity. In other words, the saprolite has the capacity to store a much larger quantity of water than does the bedrock. This is not the entire story, however.

As we noted above, the capacity of an aquifer to transmit water depends both on hydraulic conductivity and on aquifer thickness. The part of the bedrock containing water-bearing fractures is several times thicker than the saprolite.

We can then, without great error, view the ground-water system in the Piedmont and mountains as consisting of a saprolite reservoir overlying a bedrock pipeline consisting of numerous small, interconnected pipes. In the vicinity of a pumping well the bedrock fractures ("pipes") convey water from the saprolite reservoir to the well.

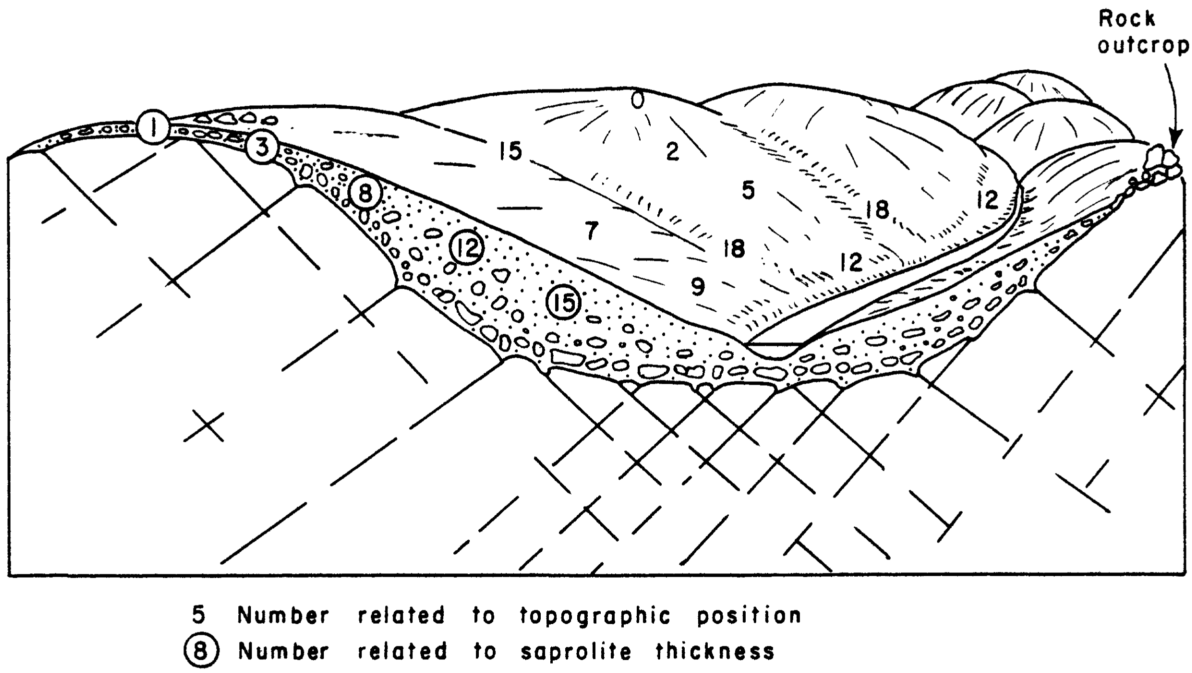
The yield of a well drawing from fractured bedrock depends on several factors. The most important of these are believed to be:

1. The number, size, areal extent, and degree of interconnection of the fractures penetrated by the well,
2. The thickness of saturated saprolite in the vicinity of the well and the specific yield of the saprolite, and
3. The hydraulic conductivity of the saprolite and the nature of the hydraulic connection between the saprolite and the bedrock.

The number and the size of the fractures control the rate at which water can enter the well. The areal extent and degree of interconnection of the fractures control the size of the area that supplies water to the well.

The thickness and the specific yield of the saprolite determines the volume of water available from storage in the saprolite. The hydraulic conductivity of the saprolite and the nature of the hydraulic connection between the saprolite and the bedrock determines the rate at which water can drain from the saprolite into the bedrock fractures.

Selecting Well Sites in the Piedmont and Mountains



Most ground-water supplies in the Piedmont and mountains are obtained from wells that are cased through the saprolite and finished with open holes in the bedrock. The yield of these wells depends on the number and size of the fractures they penetrate. Therefore, where moderate to large supplies of water are needed, well sites should be selected at the places where fractures appear to be most abundant.

H. E. LeGrand, of the U.S. Geological Survey, attempted in 1967 to indicate the relative favorability of well sites in the Piedmont and mountains by assigning point values to areas on the basis of saprolite thickness and topographic position. The point values assigned by LeGrand to features of the land surface that suggest thickness of saprolite and to different topographic positions are shown below. Selected values of each are also indicated on the above sketch.

Features of land surface related to saprolite thickness	Point value
Bare rock-almost no soil	0-2
Some rock outcrops-very thin soil	2-6
A few rock outcrops-thin soil	6-9
No fresh outcrops-moderately thick soil	9-12
No rock outcrops-thick soil	12-15

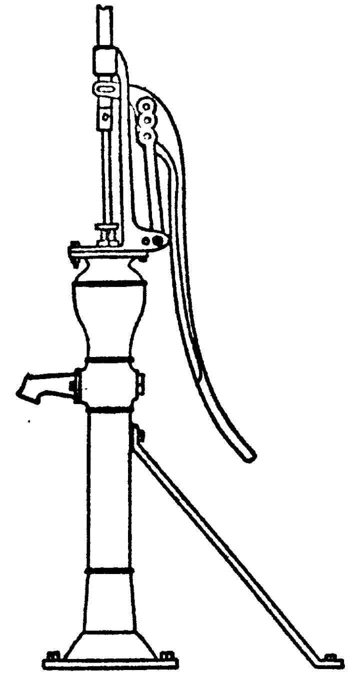
Topographic position	Point value
Steep ridge top	0
Upland steep slope	2
Rounded upland	4
Midpoint of ridge slope	5
Gentle upland slope	7
Broad flat upland	8
Lower part of upland slope	9
Valley bottom of flood plain	12
Draw with small catchment	15
Draw with large catchment	18

The point values for saprolite thickness and topographic position are added and the total is used in conjunction with the following table to estimate the chance of obtaining different yields. (The following table is an abbreviated and slightly modified version of the table prepared by LeGrand.)

Total points assigned to a site	Average yield (gal/min)	Chance of success, in percent, for a well to yield at least—				
		3 gal/min	10 gal/min	25 gal/min	50 gal/min	75 gal/min
5	3	50	20	5	3	---
10	6	65	40	15	15	---
15	15	80	55	30	15	---
20	25	90	70	50	25	20
25	40	93	80	65	45	35
30	60	96	90	75	55	45

Most of the wells used in LeGrand's analysis were drilled to obtain water for domestic needs at the sites most convenient to the well owners. Thus, no special attempt was made to select the most favorable sites. We know that the chances of success can be greatly improved if wells are not only located in valleys but also at places where the topography suggests the presence of intersecting fractures. In fact, recent studies suggest that where best technology is applied in the selection of well sites, an average yield of 150 gal/min can be expected.

Reference: LeGrand, H. E., 1967, Ground water of the Piedmont and Blue Ridge Provinces in the Southeastern States: U.S. Geological Survey Circular 538.

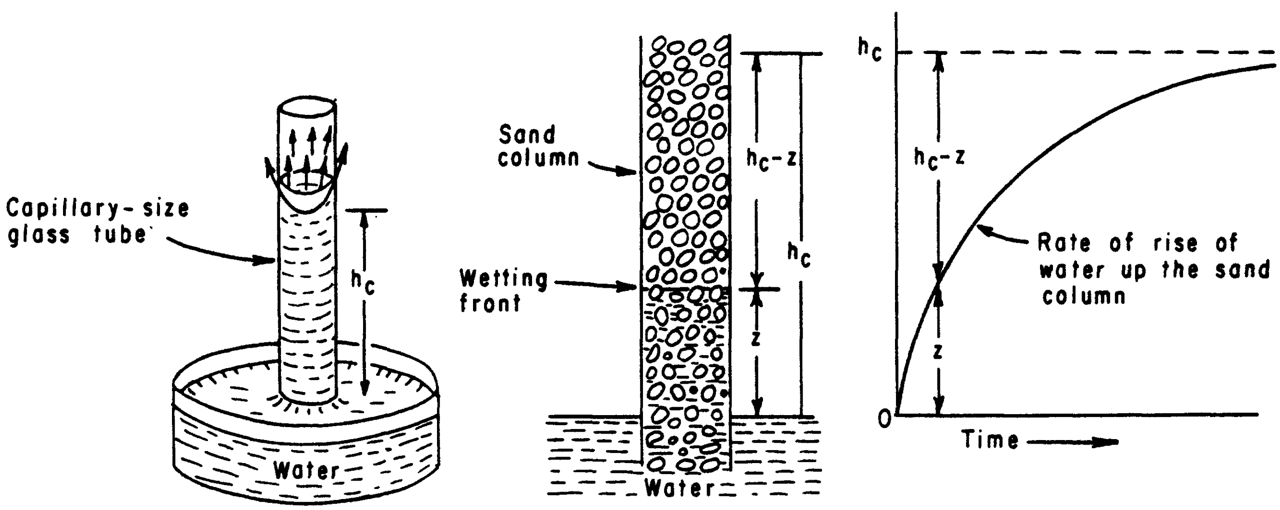


PART III. BASIC HYDRAULIC CONCEPTS AND METHODS

"The principal reason for failure of most (not all) Federal, State, and local regulatory planning, or management agencies to develop the necessary grasp of ground-water occurrence is that staffs are commonly trained and experienced mainly in surface-water concepts, including legal and engineering elements. They lack familiarity with the slow migration of water and pollutants under ground with the attendant "out-of-sight, out-of-mind" syndrone."

-From remarks by Dr. John Bredehoeft at hearings on "Ground-water quality research and development" before the Subcommittee on Environment and the Atmosphere, 95th Congress, 2nd Session, April 1978, p. XXXIV.

Capillarity and Unsaturated Flow



Most recharge of ground-water systems occurs during the percolation of water across the unsaturated zone. The movement of water through the unsaturated zone is controlled by both gravitational and capillary forces.

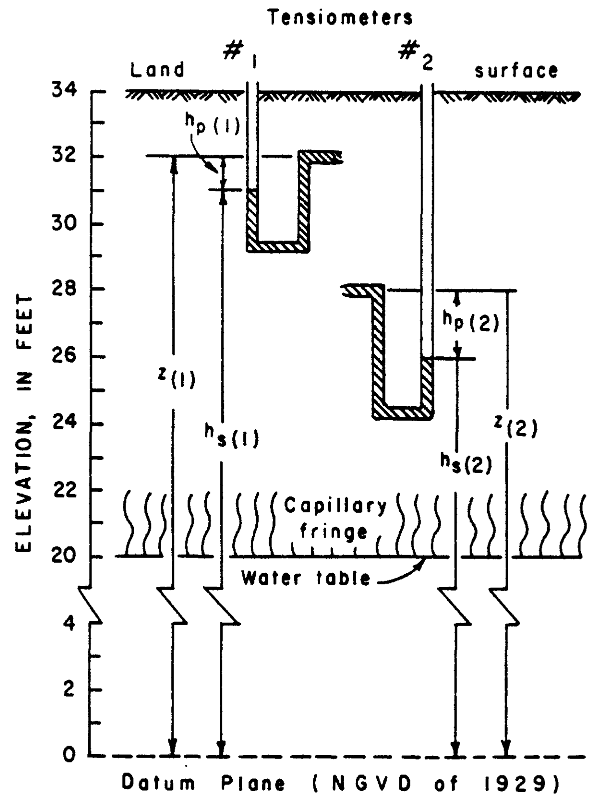
Capillarity results from two forces: (1) the mutual attraction (cohesion) between water molecules, and (2) the molecular attraction (adhesion) between water and different solid materials. As a consequence of these forces, water will rise up small-diameter glass tubes to a height h_c above the water level in a large container.

Most pores in granular materials are of capillary size and, as a result, water is pulled upward into a capillary fringe above the water table.

Height of capillary rise (h_c) in granular materials

Material	Rise in inches
Coarse sand	5
Medium sand	10
Fine sand	15
Silt	40

Steady-state flow of water in the unsaturated zone can be determined from a modified form of Darcy's law. Steady state in this context refers to a condition in which the moisture content remains constant as, for example, beneath a waste-disposal pond whose bottom is separated from the water table by an unsaturated zone.



Steady-state, unsaturated flow is proportional to the effective hydraulic conductivity (K_e), the cross-sectional area through which the flow occurs (A), and gradients due both to capillary and gravitational forces. Thus

$$Q = K_e A\left(\frac{h_c - z}{Z} \pm \frac{dh}{dl}\right) \qquad (1)$$

where Q is quantity of water,

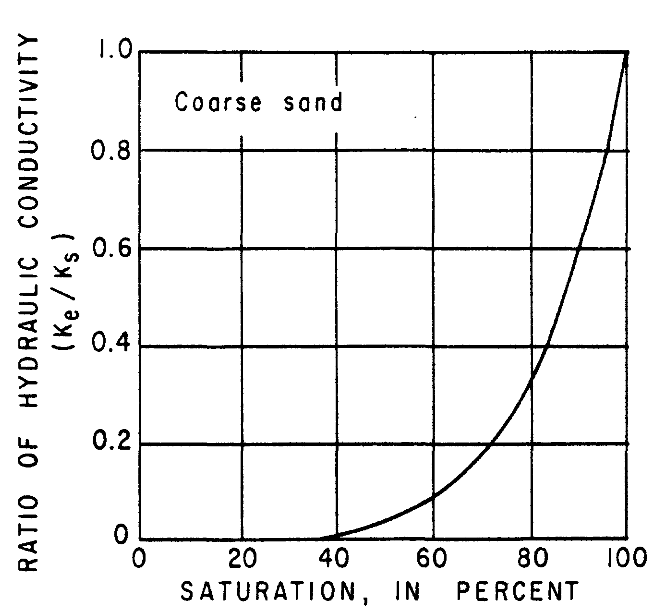
K_e is the hydraulic conductivity under the degree of saturation existing in the unsaturated zone,

$(h_c-z)/z$ is the capillary gradient, and

dh/dl is the gravitational gradient.

The $\pm$ sign is related to the direction of movement; + if downward and - if upward. For movement in a vertical direction, either up or down, the gravitational gradient is 1/1 or 1. For lateral (horizontal) movement in the unsaturated zone the term for the gravitational gradient drops out.

The capillary gradient at any time depends on the length of the water column (z) supported by capillarity in relation to the maximum possible height of capillary rise (h_c). If the lower end of a sand column is suddenly submerged in water, the capillary gradient is maximum and the rate of rise of water is fastest. As the wetting front advances up the column, the capillary gradient declines and the rate of rise decreases.



The hydraulic gradient (consisting of the combined effect of the capillary gradient and the gravitational gradient) can be determined from tensiometer measurements of pressure heads. In order to determine the gradient, it is necessary to measure the negative pressures at two levels in the unsaturated zone, as shown on the accompanying drawing, and use these pressures to calculate total heads. The equation for total head is

$$h_s = z + h_p. \qquad (2)$$

Substituting values in this equation for tensiometer No. 1, we obtain

$$h_s = 32 + (-1) = 32-1 = 31 \text{ ft.}$$

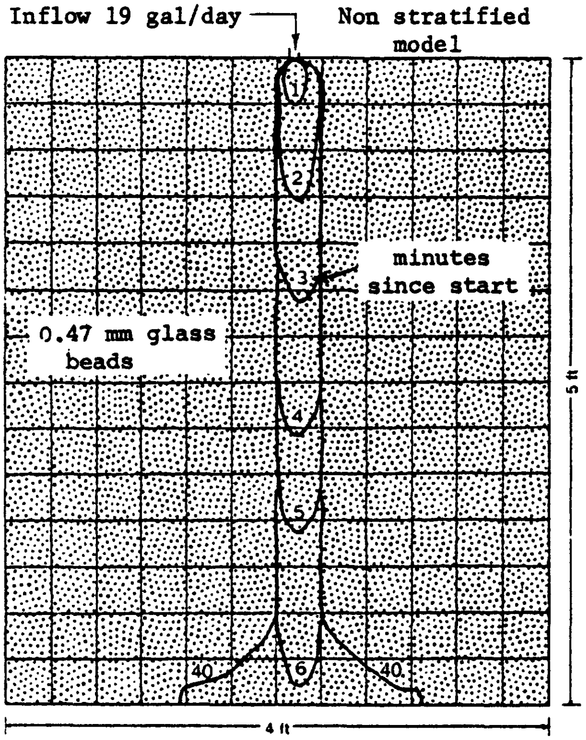
The total head at tensiometer No. 2 is 26 ft. The vertical distance between the tensiometers is 32 ft minus 28 ft or 4 ft. Because the hydraulic gradient equals the head loss divided by the distance between tensiometers, the gradient is

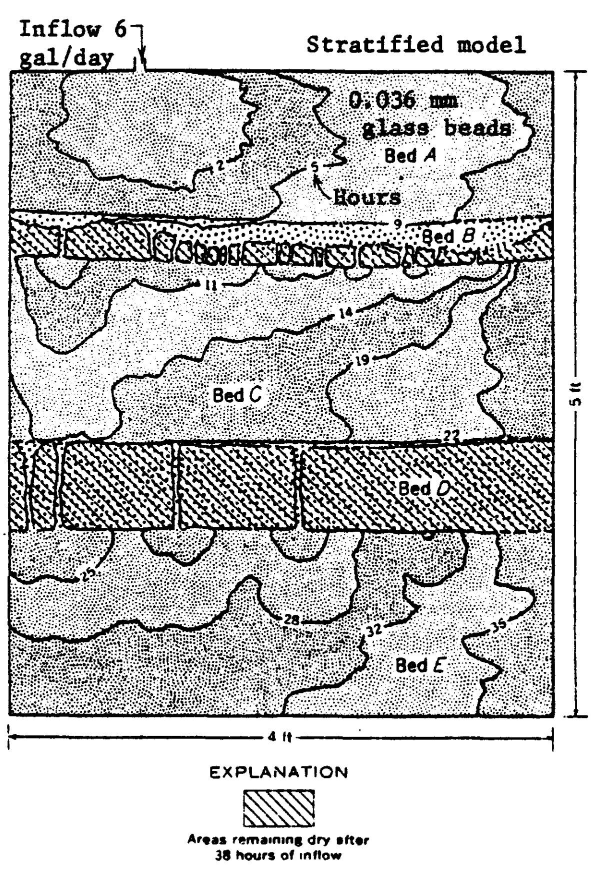
$$\frac{h_L}{L} = \frac{h_{s(1)} - h_{s(2)}}{z_{(1)} - z_{(2)}}$$

$$= \frac{31-26}{32-28} = \frac{5 \text{ ft}}{4 \text{ ft}} = 1.25 \text{ ft}$$

The effective hydraulic conductivity (K_e) is the hydraulic conductivity of material that is not completely saturated. It is thus less than the (saturated) hydraulic conductivity (K_s) for the material. The accompanying graph shows the relation between degree of saturation and the ratio of saturated and unsaturated hydraulic conductivity for coarse sand. The hydraulic conductivity (K_s) of coarse sand is about 200 ft./day.

Stratification and Unsaturated Flow





Most sediments are deposited in layers (beds) that have a distinct grain-size, sorting, or mineral composition. Where adjacent layers differ in one or more of these characteristics, the deposit is said to be *stratified* and its layered structure is referred to as *stratification.*

The layers comprising a stratified deposit commonly differ from each other in both grain size and sorting and, consequently, differ from each other in hydraulic conductivity. These differences in hydraulic conductivity significantly affect both the percolation of water across the unsaturated zone and the movement of ground water.

In most areas the unsaturated zone is composed of horizontal or nearly-horizontal layers. The movement of water, on the other hand, is predominantly in a vertical direction.

The manner in which water moves across the unsaturated zone has been studied with models containing glass beads. One model contained beads of a single size representing a nonstratified deposit and another consisted of five layers, three of which were finer grained and more impermeable than the other two. The dimensions of the models were 5 ft. X 4 ft. X 3 in.

In the nonstratified model, water introduced at the top moved vertically downward through a zone of constant width to the bottom of the model.

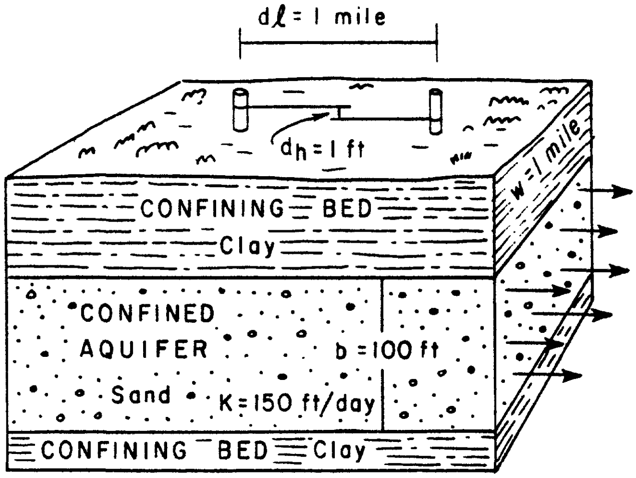
In the stratified model, beds A, C, and E consisted of silt-size beads having a capillary height (h_C) of about 40 inches and a hydraulic conductivity (K) of 2.7 ft./day. Beds B and D consisted of medium-sand-size beads having a capillary height of about 10 inches and a hydraulic conductivity of 270 ft./day.

Because of the strong capillary force and low hydraulic conductivity in bed A, the water spread laterally at almost the same rate as it did vertically and it did not begin to enter B until 9 hours after the start of the experiment. At that time the capillary saturation in bed A had reached a level where the unsatisfied (remaining) capillary pull in bed A was the same as in bed B. In other words, z in bed A at that time equalled 40-10 or 30 inches.

Because the hydraulic conductivity of bed B was 100 times that of Bed A, water moved across bed B through narrow vertical zones. We can guess that the glass beads in these zones were packed somewhat more tightly than in other parts of the beds.

Reference: Palmquist, W. N., and Johnson, A. I., 1962, Vadose flow in layered and nonlayered materials: U.S. Geological Survey Prof. Paper 450-C.

Transmissivity



The capacity of an aquifer to transmit water is referred to as its transmissivity. The *transmissivity* (T) of an aquifer is equal to the hydraulic conductivity of the material comprising the aquifer multiplied by the thickness of the aquifer. Thus

$$T = Kb \qquad (1)$$

where T is transmissivity,
K is hydraulic conductivity, and
b is aquifer thickness.

As was the case with hydraulic conductivity, transmissivity is also defined in terms of a unit hydraulic gradient.

If we combine equation 1 with Darcy's law (see HYDRAULIC CONDUCTIVITY), the result is an equation that can be used to calculate the quantity of water moving through a width, w, of an aquifer. Thus,

$$Q = KA\frac{dh}{dl} = K(bw)\frac{dh}{dl} = (Kb)w\frac{dh}{dl}$$

$$Q = Tw\frac{dh}{dl} \qquad (2)$$

Equation 2 is also used to calculate transmissivity, where the quantity of water (Q) discharging from a known width of aquifer can be determined as, for example, with streamflow measurements. Rearranging terms, we obtain

$$T = \frac{Q\,dl}{w\,dh} \qquad (3)$$

The units of transmissivity, as can be demonstrated with the preceding equation, are

$$T = \frac{(ft^3\ day^{-1})\ (ft)}{(ft)\ \ (ft)} = \frac{ft^2}{day}$$

Because transmissivity depends both on K and b, its value is different in different aquifers and from place to place in the same aquifer.

Average Values of Hydraulic Conductivity, Thickness, and Transmissivity for Selected Aquifers in North Carolina

Aquifer	Hydraulic Conductivity (ft./day)	Thickness (ft.)	Transmissivity (ft^2/day)
Post-Yorktown deposits	50	20	1000
Yorktown Formation	50	40	2000
Castle Hayne Limestone	300	100	30000
Cretaceous deposits	20	200	4000
Saprolite	5	50	250
Granite and gneiss	5	200	1000

Transmissivity replaces the term "coefficient of transmissibility" because, by convention, an aquifer is transmissive and the water in it is transmissible.

PROBLEM - Determine the quantity of water (Q) moving through the segment of the confined aquifer shown in the preceding drawing in both ft^3/day and gal./day.

(1) **Calculation of transmissivity**

$$T = Kb = \frac{150\ ft}{day} \times \frac{100\ ft}{1} = \frac{15{,}000\ ft^2}{day}$$

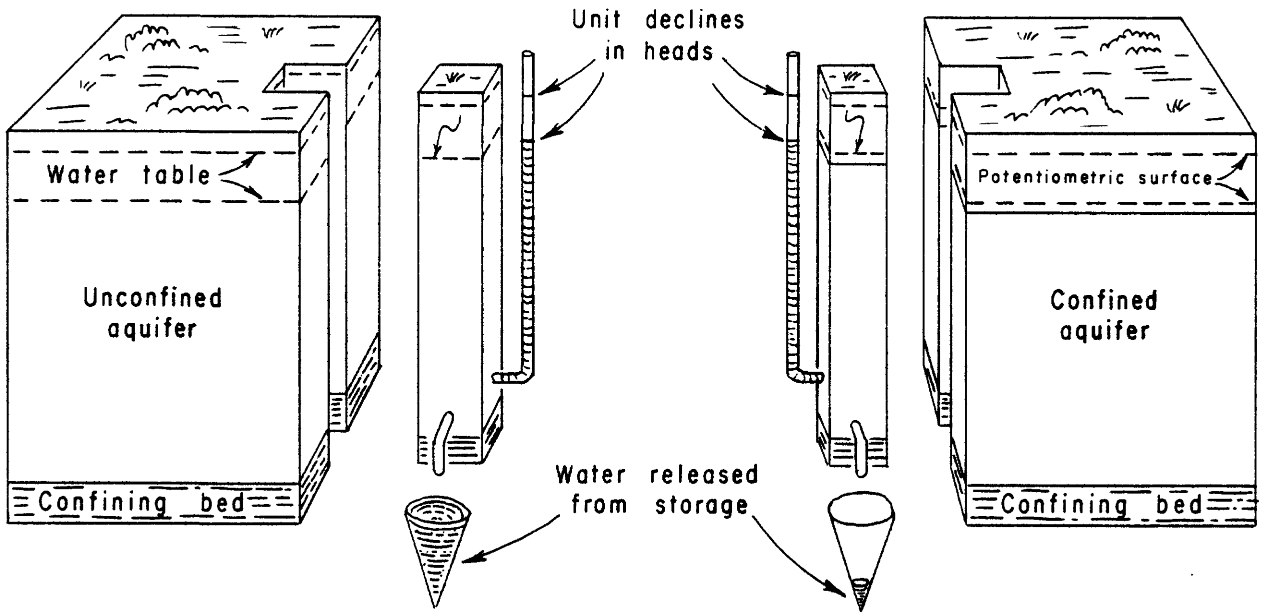
(2) **Solution in ft^3/day**

$$Q = Tw\frac{dh}{dl} = \frac{15{,}000\ ft^2}{day} \times \frac{5280\ ft}{1} \times \frac{1\ ft}{5280\ ft} = \frac{15{,}000\ ft^3}{day}$$

(3) **Conversion of ft^3/day to gal./day**

$$\frac{15{,}000\ ft^3}{day} \times \frac{7.5\ gal}{ft^3} = 112{,}500\ gal/day$$

Storage Coefficient



The abilities (capacities) of water-bearing materials to store and to transmit water are their most important hydraulic properties. Depending on the intended use of the information, these properties are given either in terms of a unit cube of the material or in terms of a unit prism of an aquifer.

Property	Unit cube of material	Unit prism of aquifer
Transmissive capacity	hydraulic conductivity (K)	transmissivity (T)
Available storage	specific yield (S_y)	storage coefficient (S)

The storage coefficient (S) is defined as the volume of water an aquifer releases from or takes into storage per unit surface area of the aquifer per unit change in head. The storage coefficient is a dimensionless unit as shown by the following equation in which the units in the numerator and denominator cancel.

$$S = \frac{\text{(Vol. of Water)}}{\text{(unit area)} \quad \text{(unit head change)}}$$

$$= \frac{(ft^3)}{(ft^2)\ (ft)} = \frac{ft^3}{ft^3}$$

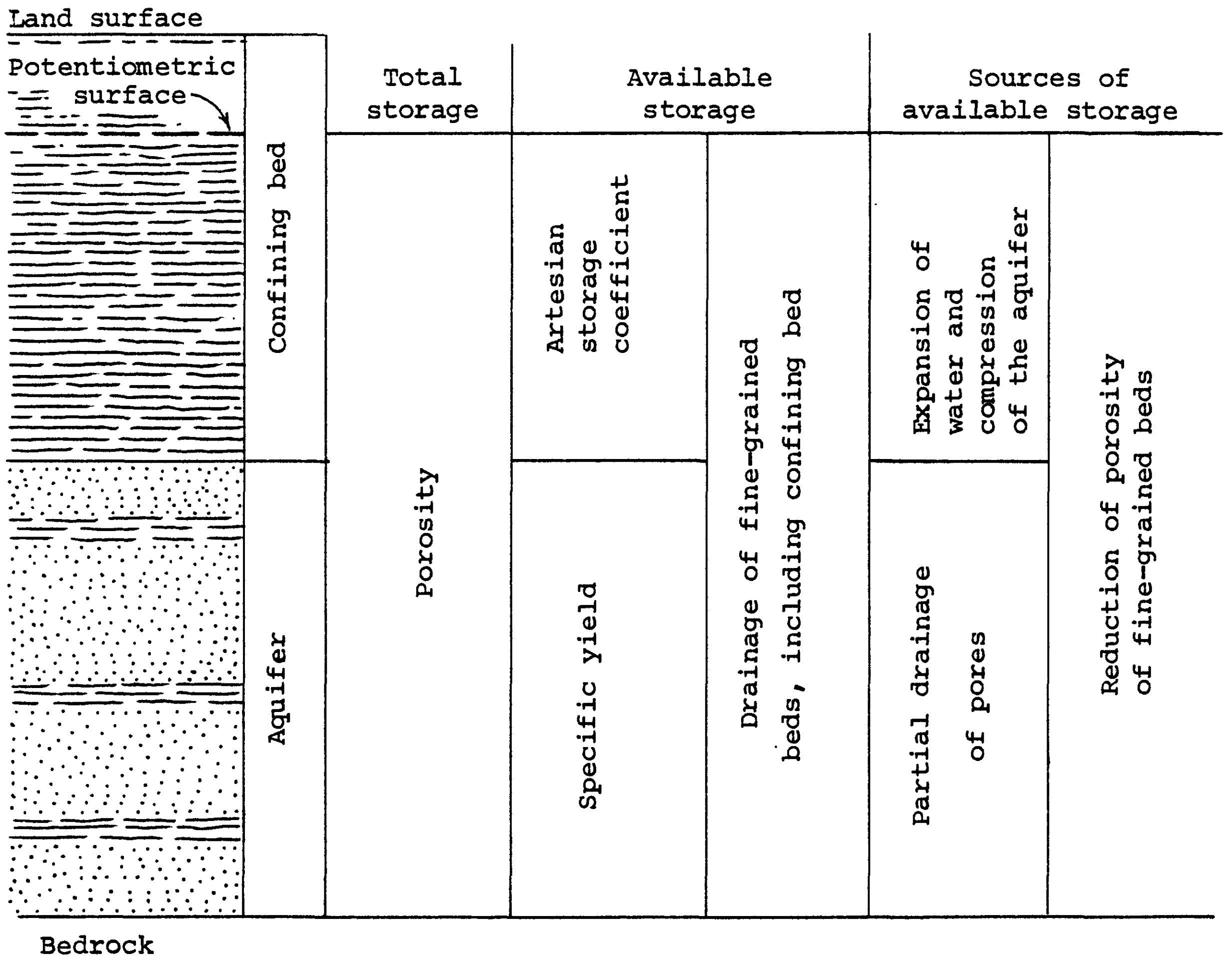
The value of the storage coefficient depends on whether the aquifer is confined or unconfined. If the aquifer is confined, the water released from storage when the head declines comes from expansion of the water and from compression of the aquifer.

If the aquifer is unconfined, the predominant source of water is from gravity drainage of the sediments through which the decline in the water table occurs. In an unconfined aquifer the volume of water derived from expansion of the water and compression of the aquifer is negligible. Thus, in such an aquifer the storage coefficient is virtually equal to the specific yield.

Because of the differences in the sources of storage, the storage coefficient of unconfined aquifers is 1,000 to 100,000 times the storage coefficient of confined aquifers. However, if water levels in an area are reduced to the point where an aquifer changes from a confined to an unconfined condition, the storage coefficient of the aquifer immediately increases from that of a confined aquifer to that of an unconfined aquifer.

Long-term withdrawals of water from most confined aquifers result in drainage of water from both clay layers within the aquifer and from adjacent confining beds. Subsidence of the land surface is one of the manifestations of such drainage.

The potential sources of water in a two-unit ground-water system consisting of a confining bed and a confined aquifer are shown in the following sketch. The sketch is based on the assumption that water is removed in two separate stages; first, while lowering the potentiometric surface to the top of the aquifer and, second, by dewatering the aquifer.



The differences in storage coefficient of confined and unconfined aquifers are of great importance in determining the response of the aquifers to stresses such as withdrawals through wells.

PROBLEM — Calculate the volume of ground water released from storage per square mile of area when the head declines 1 ft. in a confined aquifer with an S of 0.0002 and in an unconfined aquifer with an S of 0.2.

(1) Storage released from the confined aquifer

$V(\text{ft}^3) = \text{Area}(\text{ft}^2) \times \text{W.L. decline (ft)} \times S$

$= (5280 \text{ ft})^2 \times 1 \text{ ft} \times 0.0002 = 5576 \text{ ft}^3$

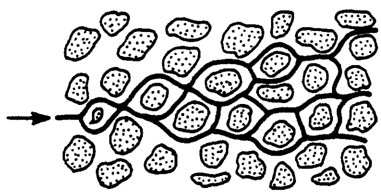
$V(\text{gal}) = \frac{5576 \text{ ft}^3}{1} \times \frac{7.5 \text{ gal}}{\text{ft}^3} = 41{,}820 \text{ gal}$

(2) Storage released from the unconfined aquifer

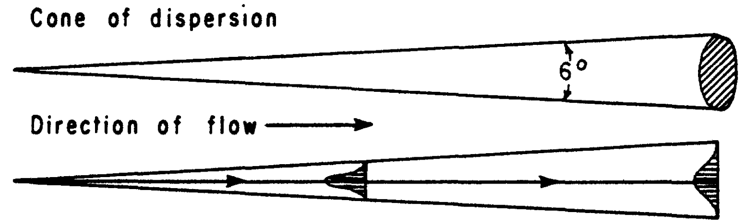
$V(\text{ft}^3) = (5280 \text{ ft})^2 \times 1 \text{ ft} \times 0.2 = 5{,}576{,}000 \text{ ft}^3$

$V(\text{gal}) = 41{,}820{,}000 \text{ gal}$

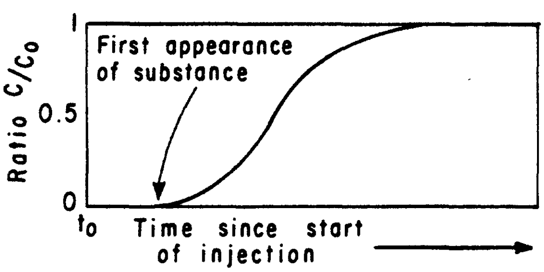
Ground-Water Flow And Dispersion



Dispersion in a granular deposit



Changes in concentration in the dispersion cone



Ground-water flow under natural gradients is laminar, except in large openings such as those in gravel, lava flows, and limestone caverns. However, even under conditions of laminar flow in most granular deposits there is at least some intermingling of stream lines. Thus, streamlines tend to converge in the narrow necks between particles and diverge where the pores widen. This results in lateral dispersion — that is, dispersion at right angles to the direction of ground-water flow.

There are also differences in velocity resulting from friction between the water and rock particles. The slowest rate of movement occurs adjacent to the particles and the fastest rate in the center of pores. This results in longitudinal dispersion — that is, dispersion in the direction of flow.

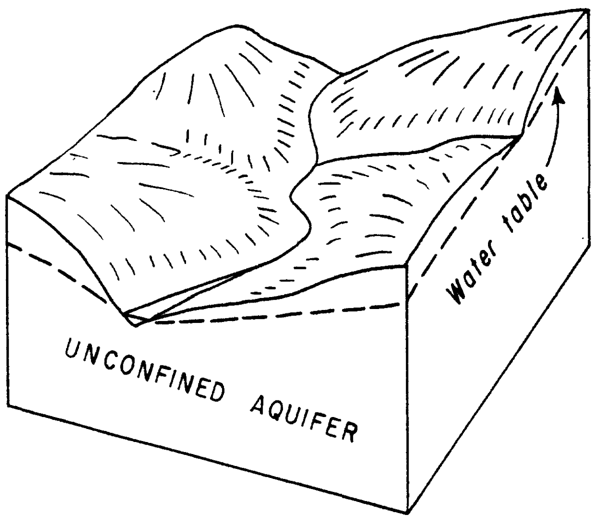
Danel (1953) found that dye injected at a point in a homogenous and isotrophic granular medium dispersed laterally in the shape of a cone about 6° wide. He also found that the concentration over a plane at any given distance from the inlet point is a bell-shaped curve similar to the normal probability curve. Because of lateral and longitudinal dispersion the peak concentration decreases in the direction of flow.

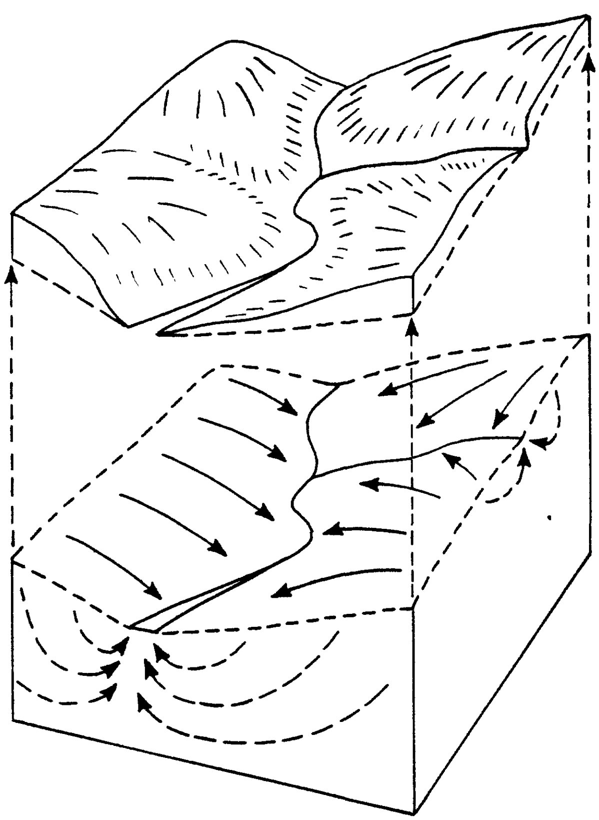
The effect of longitudinal dispersion can be observed from the change in concentration of a substance downstream from a point at which the substance is being injected constantly at a concentration of c_O. The concentration rises slowly at first as the "fastest" streamlines arrive, then rises rapidly until the concentration reaches about 0.7 c_O at which point the rate of increase in concentration begins to decrease.

Disperson is important in the study of ground-water pollution. However, it is difficult to measure in the field because the rate and direction of movement of wastes are also affected by stratification, ion exchange, filtration, and other conditions and processes.

Reference: Danel, Pierre, 1953, The measurement of ground-water flow: Proc. Ankara Symposium on Arid Zone Hydrology, UNESCO, Paris, p. 99-107.

Ground-Water Movement And Topography





Arrows show direction of ground-water movement.

It is desirable, wherever possible, to determine the position of the water table and the direction of ground-water movement by measuring the depth of water level in shallow wells. However, in humid areas like North Carolina, general, but very valuable, conclusions about the direction of ground-water movement can be derived from observations of land-surface topography.

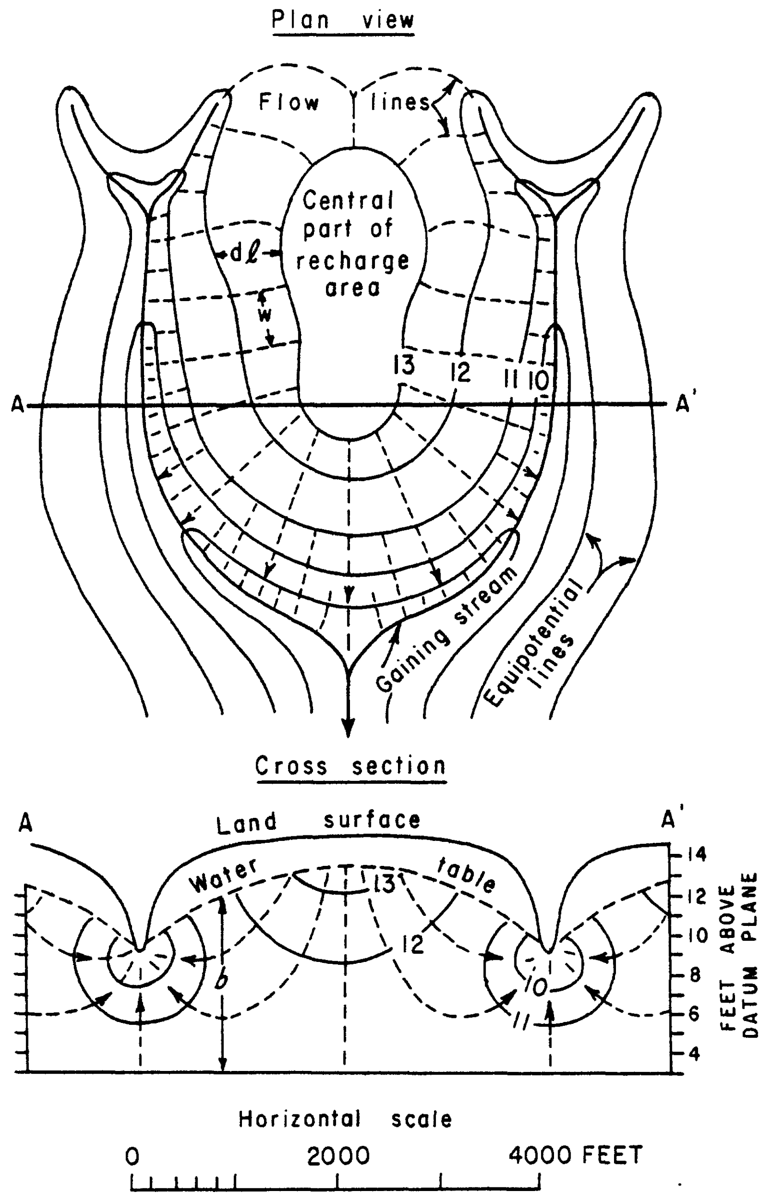
We know that gravity is the dominant driving force in ground-water movement and that under natural conditions ground water moves "downhill" until, in the course of its movement, it reaches the land surface at a spring or through a seep along the side or bottom of a stream channel or an estuary.

Thus, ground water in the shallowest part of the saturated zone moves from interstream areas toward streams or the coast. If we ignore minor surface irregularities, we find that the land surface also slopes in these directions. The depth to the water table is greatest along the divide between streams and least beneath the floodplain. In effect, *the water table is a subdued replica of the land surface.*

In areas where ground water is used for domestic and other needs requiring good-quality water, septic tanks, sanitary landfills, waste ponds, and other waste disposal sites should not be located uphill from supply wells.

The potentiometric surface of confined aquifers, like the water table, also slopes from recharge areas to discharge areas. Shallow confined aquifers, which are relatively common along the Atlantic Coastal Plain, share both recharge and discharge areas with the surficial unconfined aquifer. This may not be the case with the deeper confined aquifers. The principal recharge areas for these are probably in their outcrop areas near the western border of the Coastal Plain. Their discharge areas are probably near the heads of the estuaries along the major streams. Thus, movement of water through these aquifers is in a general west to east direction, where it has not been modified by withdrawals.

Ground-Water Flow Nets



Ground water is normally hidden from view and, as a consequence, many people have difficulty visualizing its occurrence and movement. This adversely affects their ability to understand and to deal effectively with ground-water related problems. This difficulty is partly solved through the use of flow nets, which are one of the most effective means yet devised for illustrating conditions in ground-water systems.

Flow nets consist of two sets of lines. One set, referred to as *equipotential* lines, connect points of equal head and thus represent the height of the water table, or the potentiometric surface of a confined aquifer, above a datum plane. The second set, referred to as *flow lines*, depict the idealized paths followed by particles of water as they move through the aquifer.

Because ground water moves in the direction of the steepest hydraulic gradient, flow lines in isotropic aquifers are perpendicular to equipotential lines — that is, flow lines cross equipotential lines at right angles.

In any aquifer there are an infinite number of equipotential lines and flow lines. However, for purposes of flow-net analysis, it is necessary to draw only a few of each set. Equipotential lines are drawn so that the drop in head is the same between adjacent pairs of lines. Flow lines are drawn so that the flow is equally divided between adjacent pairs of lines and together with the equipotential lines form a series of "squares."

Flow nets not only show the direction of ground-water movement but, if drawn with care, can be used to estimate the quantity of water in transit through an aquifer. According to Darcy's law, the flow through any "square" is

$$q = (Kb)(w)\frac{dh}{dl} = Tw\frac{dh}{dl} \qquad (1)$$

and the total flow through any set or group of "squares" is

$$Q = nq \qquad (2)$$

where K is hydraulic conductivity,

b is aquifer thickness at the midpoint between equipotential lines,

T is transmissivity or Kb,

dh is the difference in head between equipotential lines,

w is the distance between flow lines,

dl is the distance between equipotential lines, and

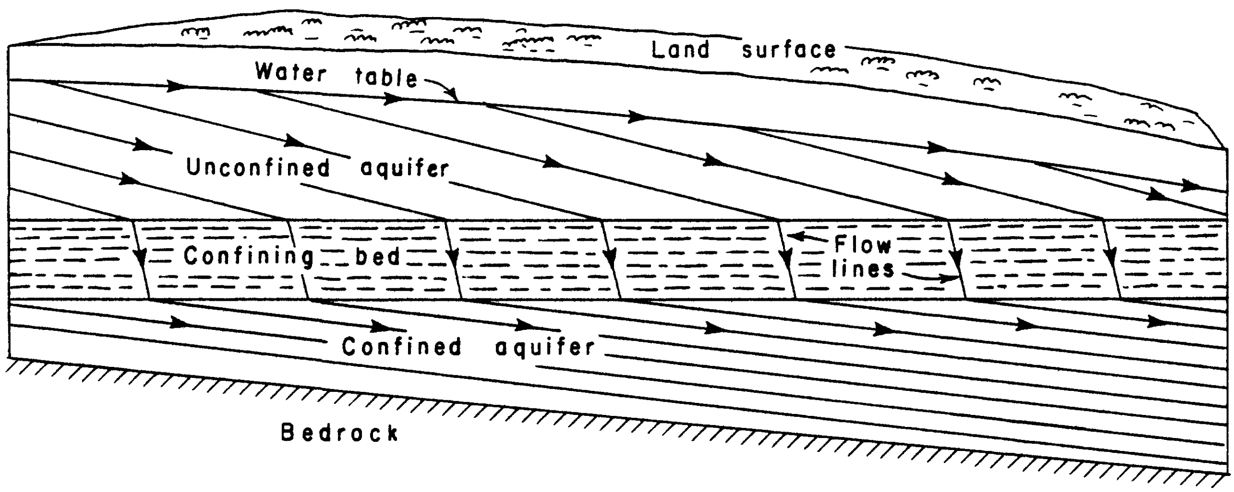
n is the number of squares through which the flow occurs.

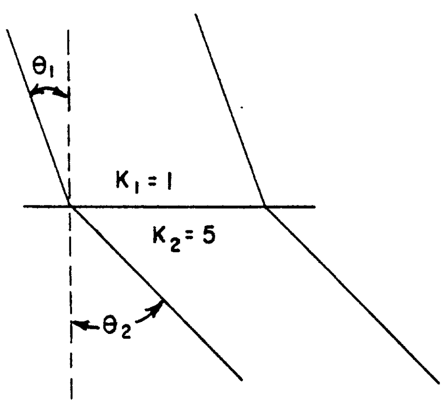
The drawings show a flow net in both plan view and profile for an area underlain by an unconfined aquifer composed of sand. The sand overlies a horizontal confining bed, the top of which occurs at an elevation 3 ft. above the datum plane. Note that some flow lines originate in the area in which heads exceed 13 ft. This indicates the presence of recharge to the aquifer in this area. The relative positions of land surface and the water table suggest that recharge occurs throughout the area, except along the stream valleys. This is confirmed by the fact that flow lines also originate in areas where heads are less than 13 ft.

Flow lines tend to diverge (get further apart) from the central part of recharge areas and converge from different directions in discharge areas. Closed contours (equipotential lines) indicate the central part of recharge areas but do not normally indicate the limits of the areas.

In the cross section view it should be noted that heads decrease downward in the recharge area and decrease upward in the discharge area. Consequently, the deeper a well is drilled in a recharge area, the lower the water level in the well stands below land surface. The reverse is true in discharge areas. Thus, in a discharge area if a well is drilled deeply enough in an unconfined aquifer, the well may flow above land surface.

Ground-Water Movement And Stratification





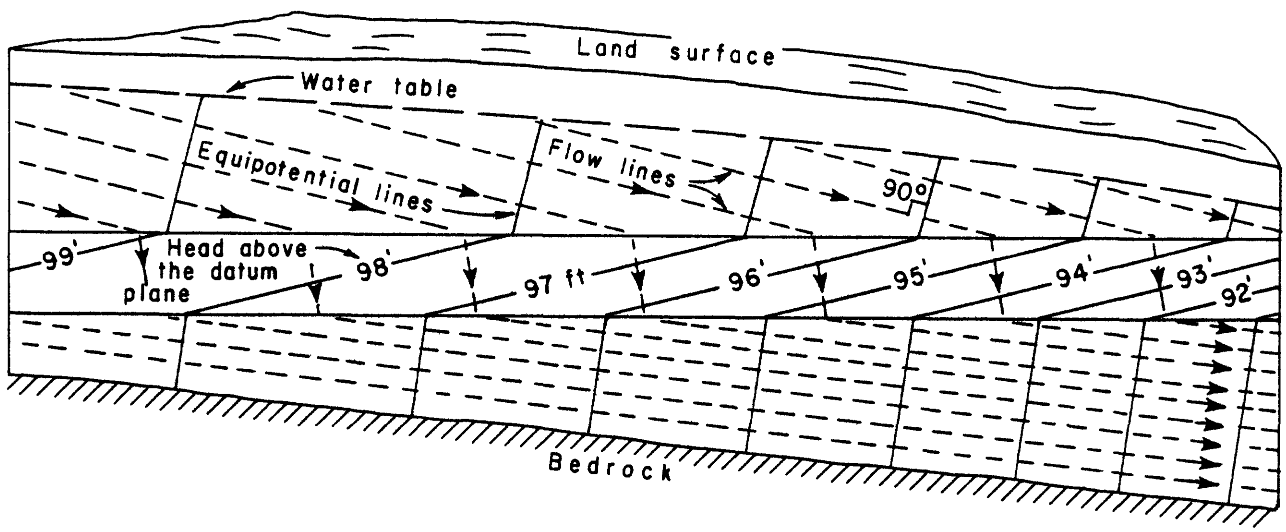
Nearly all ground-water systems include both aquifers and confining beds. Thus, ground-water flow through these systems involves movement not only *through* the aquifers but also *across* the confining beds.

The hydraulic conductivities of aquifers are tens to thousands of times those of confining beds. Thus, aquifers offer the least resistance to flow with the result that for a given rate of flow the head loss per unit of distance along a flow line is tens to thousands of times less in aquifers than in confining beds. Consequently, flow lines tend to "concentrate" in aquifers and be parallel to aquifer boundaries.

Differences in hydraulic conductivities of aquifers and confining beds cause a refraction of flow lines at their boundaries. As flow lines move from aquifers into confining beds, they are refracted towards a direction perpendicular to the boundary. In other words, they are refracted in the direction that produces the shortest flow path in the confining bed. As the flow lines emerge from the confining bed they are refracted back towards a direction parallel to the boundary.

The angles of refraction (and the spacing of flow lines in adjacent aquifers and confining beds) are proportional to the differences in hydraulic conductivities such that

$$\frac{\tan \theta_1}{\tan \theta_2} = \frac{K_1}{K_2}$$

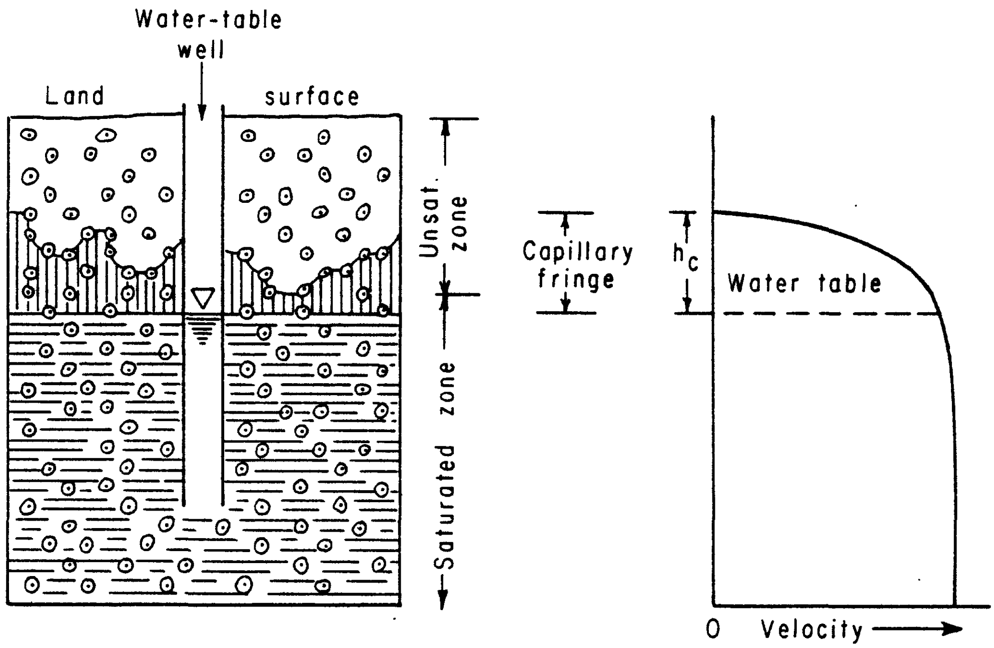


The water table is a somewhat unique flow line. It represents a bounding surface for the ground-water system and, thus, in the development of many ground-water flow equation it is assumed to be coincident with a flow line. However, during periods when recharge is arriving at the top of the capillary fringe, the water table is also the point of origin of flow lines.

The movement of water through ground-water systems is controlled by the transmissivities of the aquifers, the hydraulic conductivities and thickness of the confining beds, and the hydraulic gradients. The maximum difference in head exists between the central part of recharge areas and discharge areas. Because of the relatively large head loss that occurs as water moves across confining beds, the most vigorous circulation of ground water normally occurs through the shallowest aquifers. Movement becomes more and more lethargic with increasing depth.

The most important exceptions to the general situation described in the preceding paragraph are those systems in which one or more of the deeper aquifers have transmissivities significantly larger than the surficial and other shallower aquifers. Thus, in eastern North Carolina, the Castle Hayne Limestone is the dominant aquifer because of its very large transmissivity, although it is overlain in most of the area by an unconfined surficial aquifer and by the Yorktown aquifer.

Ground-Water Velocity



The rate of movement of ground water is important in many problems, particularily those related to pollution. For example, if a harmful substance is accidentally introduced into an aquifer up gradient from a supply well, it becomes a matter of great urgency to estimate when the substance will reach the well.

The rate of movement of ground water is greatly overestimated by many people, including those who think in terms of ground water moving through "veins" and underground rivers at the rates commonly observed in surface streams. It would be more appropriate to compare the rate of movement of ground water to the movement of water through a very large lake being drained by a very small stream.

The *ground-water velocity* equation can be derived from a combination of Darcy's law and the basic velocity equation of hydraulics.

Darcy's law $Q = KA\frac{dh}{dl}$

velocity equation $Q = Av$

where Q is rate of flow or volume per unit of time,

K is hydraulic conductivity,

A is cross-sectional area through which the flow Q occurs,

dh/dl is the hydraulic gradient, and

v is the velocity.

Combining these equations, we obtain

$$Av = KA\frac{dh}{dl}$$

Canceling the area terms, we find that

$$v = K\frac{dh}{dl}$$

Because this equation contains terms only for hydraulic conductivity and gradient, it is not yet a complete expression of ground-water velocity. The missing term is porosity, η, because, as we know, water moves only through the openings in a rock. Adding the porosity term, we obtain.

$$v = \frac{Kdh}{\eta dl} \qquad (1)$$

It is important to note that movement in unconfined aquifers is not limited to the zone below the water table or to the saturated zone. Water in the capillary fringe is subjected to the same hydraulic gradient as exists at the water table and water in the capillary fringe moves, therefore, in the same direction as the ground water.

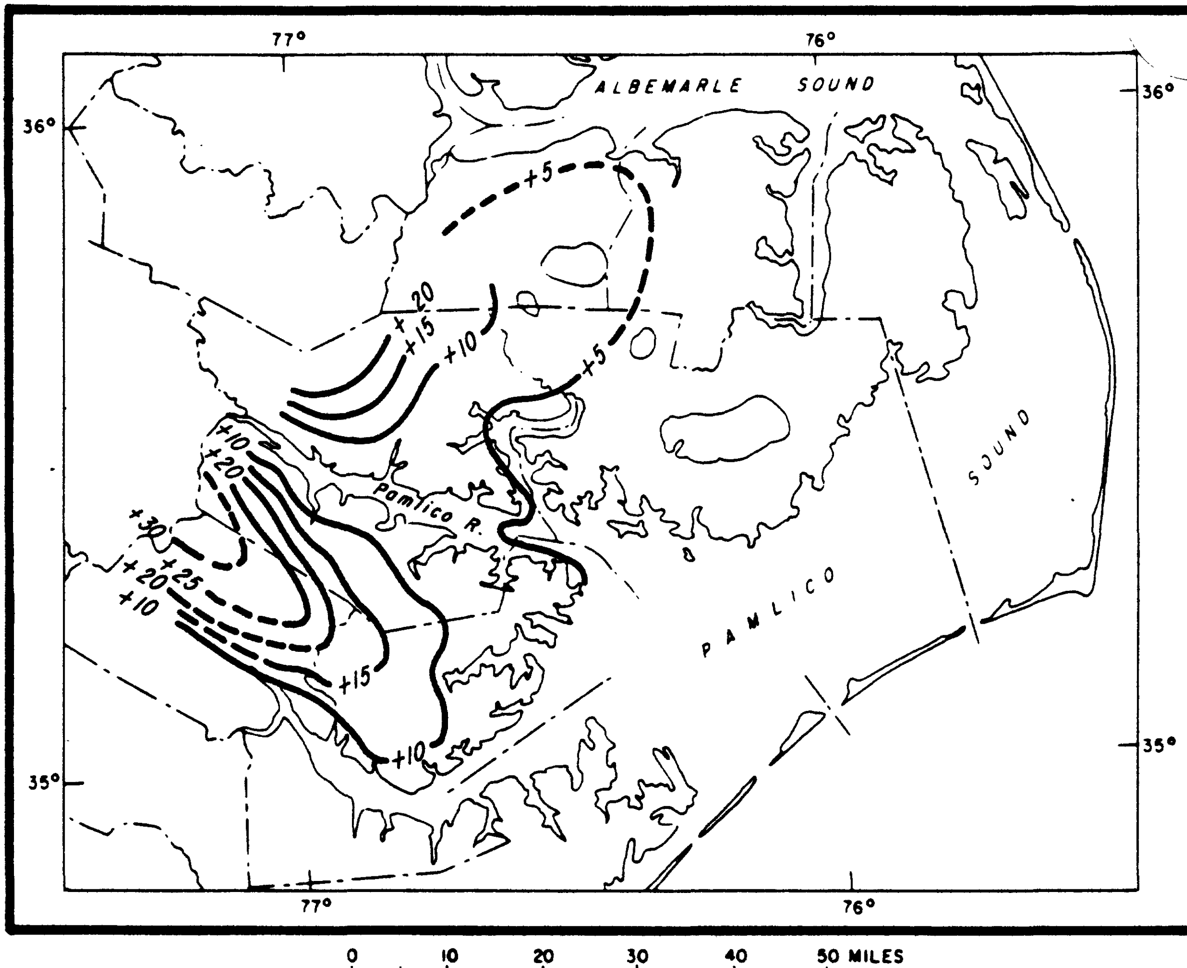
As shown in the preceding sketch, the rate of lateral movement in the capillary fringe decreases in an upward direction and becomes zero at the top of the fringe. This is an important consideration where unconfined aquifers are polluted with gasoline and other substances less dense than water.

PROBLEM — The map on page 48 which was prepared by the North Carolina Groundwater Section, shows the potentiometric surface of the Castle Hayne Limestone in June 1965 prior to the start of phosphate mining in Beaufort County, N.C. At that time, water in the Castle Hayne in a part of the area south of the Pamlico River was moving toward the river in response to a hydraulic gradient of about 5 ft/mi. The hydraulic conductivity and porosity of the Castle Hayne are estimated to be 300 ft/day and 0.20, respectively. At what rate was the ground water moving?

Solution

$$v = \frac{Kdh}{\eta dl} = \frac{300 \text{ ft}}{\text{day}} \times \frac{1}{0.20} \times \frac{5 \text{ ft}}{5280 \text{ ft}} = \frac{1500 \text{ ft}^2}{1056 \text{ ft day}} = 1.4 \text{ ft/day}$$

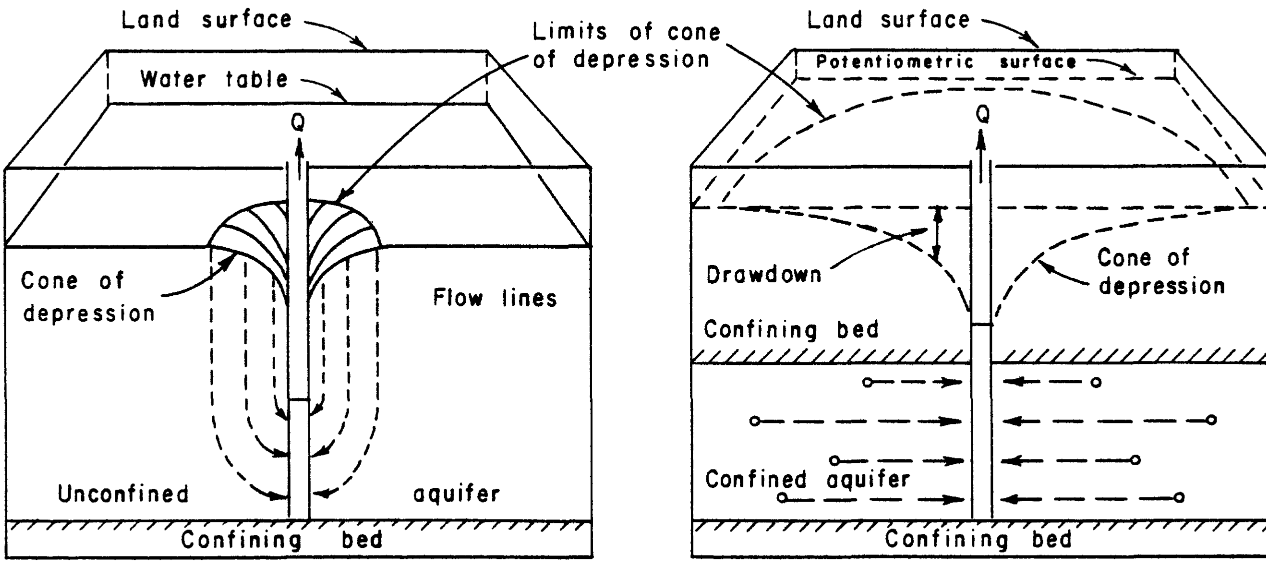
POTENTIOMETRIC SURFACE OF CASTLE HAYNE LIMESTONE, JUNE 1965



EXPLANATION

+10 POTENTIOMETRIC CONTOUR--shows altitude of potentiometric surface. Dashed where approximately located. Contour interval 5 feet. Datum is mean sea level.

The Cone of Depression



Both wells and springs serve as sources of ground-water supply. However, springs having yields large enough to meet municipal, industrial, and large commercial and agricultural needs occur only in areas underlain by cavernous limestones and lava flows. Therefore, most ground-water needs are met by withdrawals from wells.

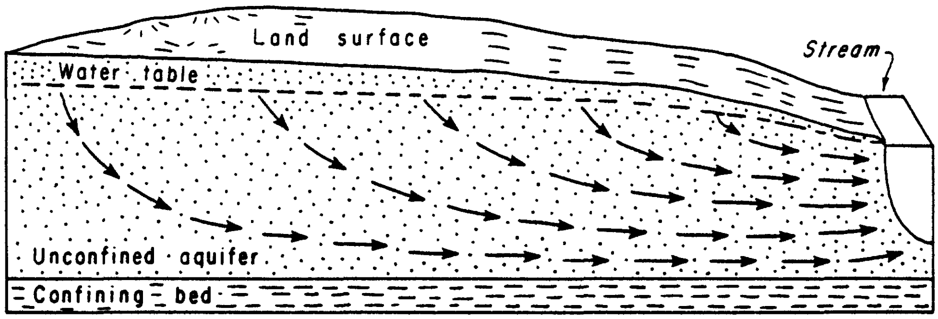
The response of aquifers to withdrawals from wells is an important topic. When withdrawals start, the water level in the well begins to decline as water is removed from storage in the well. This lowers the head in the well below the level in the surrounding aquifer. As a result, water begins to move from the aquifer into the well. The water level in the well continues to decline and the rate of flow into the well from the aquifer continues to increase until the rate of inflow equals the rate of withdrawal.

The movement of water from an aquifer into a well results in the formation of a *cone of depression*. Because water must converge on the well from all directions and because the area through which the flow occurs decreases toward the well, the hydraulic gradient must increase toward the well.

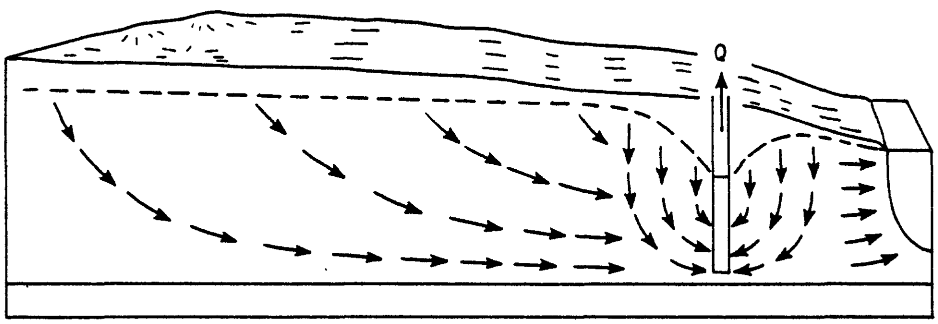
There are several important differences between the cones of depression in confined and unconfined aquifers. Withdrawals from an unconfined aquifer result in a drainage of water from the rocks through which the water table declines as the cone of depression forms. Because the storage coefficient of an unconfined aquifer equals the specific yield of the aquifer material, the cone of depression expands very slowly. On the other hand, dewatering of the aquifer results in a decrease in transmissivity which causes, in turn, an increase in drawdown both in the well and in the aquifer.

Withdrawals from a confined aquifer cause a drawdown in artesian pressure but do not (normally) cause a physical dewatering of the aquifer. The water withdrawn from a confined aquifer is derived from expansion of the water and compression of the solid skeleton of the aquifer. The very small storage coefficient of confined aquifers results in a very rapid expansion of the cone of depression. Consequently, the mutual interference of expanding cones around adjacent wells is a more serious problem in confined aquifers than in unconfined aquifers.

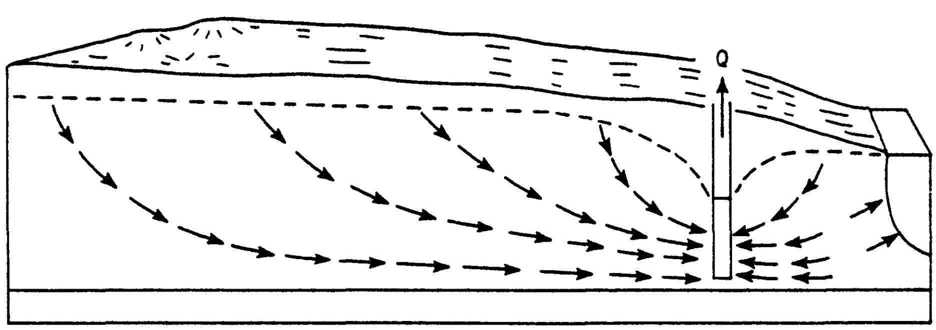
Source of Water Derived from Wells



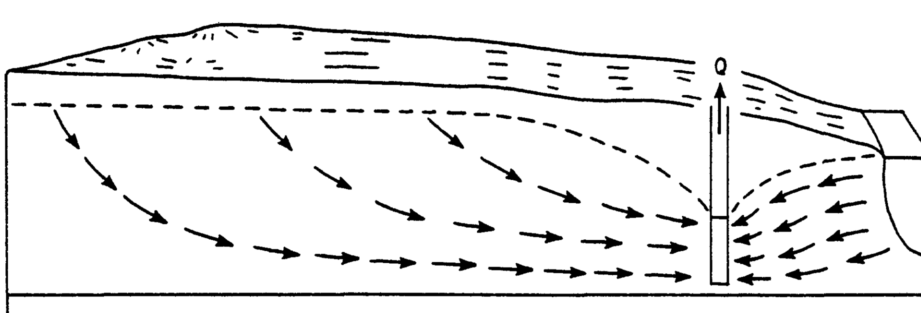
Discharge (D) = Recharge (R)



Withdrawal (Q) = Reduction in storage (ΔS)



Withdrawal (Q) = Reduction in storage (ΔS) + Reduction in discharge (ΔD)



Withdrawal (Q) = Reduction in discharge (ΔD) + Increase in recharge (ΔR)

Both the economical development and the effective management of any ground-water system requires an understanding of the response of the system to withdrawals from wells. The first concise description of the hydrologic principles involved in this response was presented by C. V. Theis in a paper published in 1940.

Theis pointed out that the response of an aquifer to withdrawals from wells depends on:

1. The rate of expansion of the cone of depression caused by the withdrawals, which depends on the transmissivity and the storage coefficient of the aquifer;
2. The distance to areas in which water discharges naturally from the aquifer; and
3. The distance to recharge areas in which the rate of recharge can be increased.

Over a sufficiently long period of time under natural conditions — that is, prior to the start of withdrawals — the discharge from every ground-water system equals the recharge to it. In other words,

Natural discharge (D) = Natural recharge (R)

In the eastern part of the United States, including North Carolina, the amount and distribution of precipitation is such that the period of time over which discharge and recharge balance is usually a year or less. Over shorter periods of time differences between discharge and recharge involve changes in ground-water storage. In other words,

1. When discharge exceeds recharge, ground-water storage (S) is reduced by an amount ΔS equal to the difference between discharge and recharge. Thus,

   $D = R + \Delta S$.
2. Conversely, when recharge exceeds discharge, ground-water storage is increased. Thus,

   $D = R - \Delta S$.

When withdrawal through a well begins, water is removed from storage in its vicinity as the cone of depression develops. Thus, the withdrawal (Q) is balanced by a reduction in ground-water storage. In other words,

$Q = \Delta S$.

As the cone of depression expands outward from the pumping well, it may reach an area where water is naturally discharging from the aquifer. This will reduce the hydraulic gradient toward the discharge area and cause a decrease in the rate of natural discharge. To the extent that the decrease in natural discharge compensates for the pumpage, the rate of expansion of the cone of depression will decline. If and when the reduction in natural discharge equals the rate of withdrawal, a new balance will be established in the aquifer. This balance in symbolic form is

$$(D - \Delta D) + Q = R$$

where ΔD is the reduction in the natural discharge.

Conversely, if instead of expanding into a natural discharge area, the cone of depression expands into a recharge area, the hydraulic gradient between the recharge area and the pumping well will be increased. If under natural conditions more water was available in the recharge area than the aquifer could accept (the condition Theis referred to as one of *rejected recharge*), the increase in the gradient away from the recharge area will permit more recharge to occur and the rate of growth of the cone of depression will decrease. If and when the increase in recharge equals the rate of withdrawal, a new balance will be established in the aquifer and expansion of the cone of depression will cease. The new balance in symbolic form is

$$D + Q = R + \Delta R$$

where ΔR is the increase in recharge.

In the eastern part of the United States, streams to which ground-water discharges are relatively closely spaced and areas in which rejected recharge occurs are relatively unimportant. In this region, the rate of growth of cones of depression are commonly affected first by a reduction in natural discharge. If the pumping wells are near a stream or if the withdrawals are continued long enough, natural discharge may be stopped entirely in the vicinity of the wells and water may be induced to move from the stream into the aquifer. In other words, in this region the tendency is for withdrawals to change discharge areas into recharge areas. This is an important consideration where the streams contain brackish or polluted water.

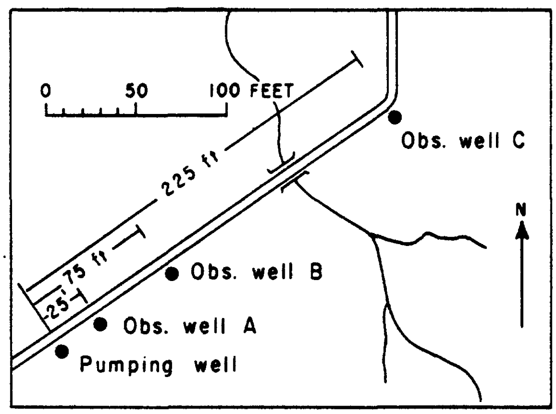
To summarize, the withdrawal of ground water through a well reduces the water in storage in the source aquifer during the growth of the cone of depression. When and if the cone of depression ceases to expand, the rate of withdrawal is being balanced by a reduction in the rate of natural discharge and/or by an increase in the rate of recharge. Under this condition

$$Q = \Delta D + \Delta R$$

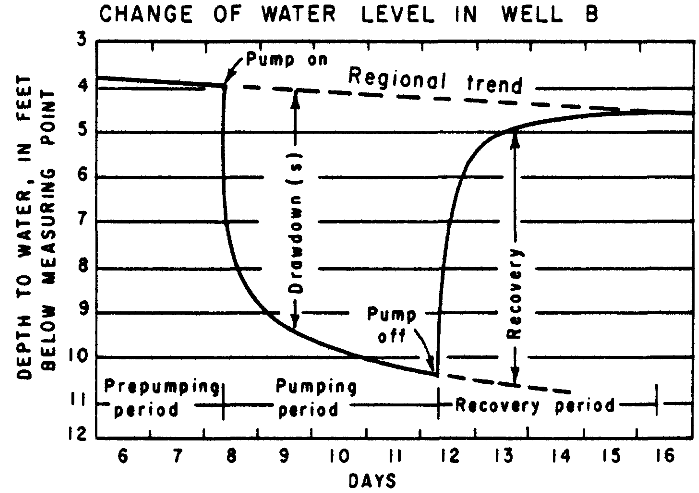
Reference: Theis, C. V., 1940, The source of water derived from wells, essential factors controlling the response of an aquifer to development: *Civil Engineering*, p. 277-280 (May).

Aquifer Tests

MAP OF AQUIFER TEST SITE



CHANGE OF WATER LEVEL IN WELL B



Determination of the yield of ground-water systems and evaluation of the movement and fate of ground-water pollutants requires knowledge of

1. The position and thickness of aquifers and confining beds,
2. The transmissivity and storage coefficient of the aquifers,
3. The hydraulic characteristics of the confining beds, and
4. The position and nature of the aquifer boundaries.

Acquisition of knowledge on these factors requires both geologic and hydrologic studies. One of the most important hydrologic studies involves analyzing the change in water levels (or pressure) in an aquifer in response to withdrawals through wells. This type of study is referred to as an *aquifer test* and, in most cases, includes pumping a well at a constant rate for a period ranging from several hours to several days and measuring the change in water level in observation wells located at different distances from the pumping well.

Successful aquifer tests require, among other things,

1. Determination of the prepumping water-level trend (that is, the regional trend),
2. A carefully controlled constant pumping rate, and
3. Accurate water-level measurements made at precisely-known times during both the drawdown and recovery periods.

Drawdown is the difference between the water level at any time during the test and the position at which the water level would have been if withdrawals had not started. Drawdown is very rapid at first. As pumping continues and the cone of depression expands, the rate of drawdown decreases.

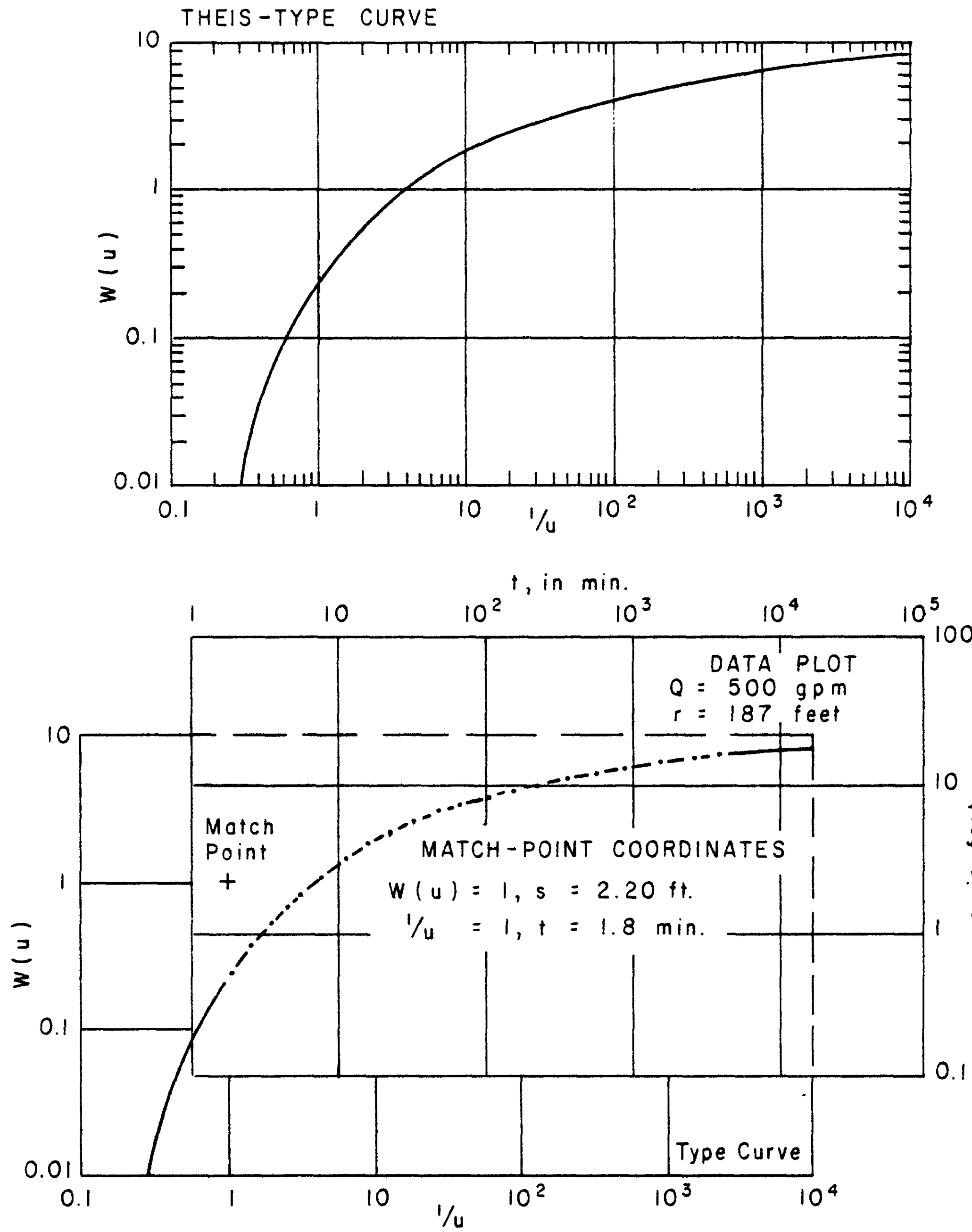
The *recovery* of the water level under ideal conditions is a mirror image of the drawdown. The change in water level during the recovery period is the same as if withdrawals had continued at the same rate from the pumping well but, at the moment of pump cutoff, a recharge well had begun recharging water at the same point at the same rate. Therefore, the recovery of the water level is the difference between the actual measured level and the projected pumping level.

Selected values of W(u) for values of 1/u

$^1/_u$	10	7.69	5.88	5.00	4.00	3.33	2.86	2.5	2.22	2.00	1.67	1.43	1.25	1.11
10^{-1}	0.219	0.135	0.075	0.049	0.025	0.013	0.007	0.004	0.002	0.001	0.000	0.000	0.000	0.000
1	1.82	1.59	1.36	1.22	1.04	0.91	0.79	0.70	0.63	0.56	0.45	0.37	0.31	0.26
10	4.04	3.78	3.51	3.35	3.14	2.96	2.81	2.68	2.57	2.47	2.30	2.15	2.03	1.92
10^2	6.33	6.07	5.80	5.64	5.42	5.23	5.08	4.95	4.83	4.73	4.54	4.39	4.26	4.14
10^3	8.63	8.37	8.10	7.94	7.72	7.53	7.38	7.25	7.13	7.02	6.84	6.69	6.55	6.44
10^4	10.94	10.67	10.41	10.24	10.02	9.84	9.68	9.55	9.43	9.33	9.14	8.99	8.86	8.74
10^5	13.24	12.98	12.71	12.55	12.32	12.14	11.99	11.85	11.73	11.63	11.45	11.29	11.16	11.04
10^6	15.54	15.28	15.01	14.85	14.62	14.44	14.29	14.15	14.04	13.93	13.75	13.60	13.46	13.34
10^7	17.84	17.58	17.31	17.15	16.93	16.74	16.59	16.46	16.34	16.23	16.05	15.90	15.76	15.65
10^8	20.15	19.88	19.62	19.45	19.23	19.05	18.89	18.76	18.64	18.54	18.35	18.20	18.07	17.95
10^9	22.45	22.19	21.92	21.76	21.53	21.35	21.20	21.06	20.94	20.84	20.66	20.50	20.37	20.25
10^{10}	24.75	24.49	24.22	24.06	23.83	23.65	23.50	23.36	23.25	23.14	22.96	22.81	22.67	22.55
10^{11}	27.05	26.79	26.52	26.36	26.14	25.96	25.80	25.67	25.55	25.44	25.26	25.11	24.97	24.86
10^{12}	29.36	20.09	28.83	28.66	28.44	28.26	28.10	27.97	27.85	27.75	27.56	27.41	27.28	27.16
10^{13}	31.66	31.40	31.13	30.97	30.74	30.56	30.41	30.27	30.15	30.05	29.87	29.71	29.58	29.46
10^{14}	33.96	33.70	33.43	33.27	33.05	32.86	32.71	32.58	32.46	32.35	32.17	32.02	31.88	31.76

Examples: When $1/u = 10 \times 10^{-1}$, $W(u) = 0.219$; when $1/u = 3.33 \times 10^2$, $W(u) = 5.23$

Analysis of Aquifer-Test Data



In 1935, C. V. Theis of the New Mexico District of the U.S. Geological Survey, developed the first equation that could be used to analyze the effect of withdrawals from a well that included time of pumping as a factor. Thus, the *Theis equation* permitted, for the first time, the determination of the hydraulic characteristics of an aquifer prior to the development of steady-state conditions.

Theis assumed in the development of the equation that:

1. The transmissivity of the aquifer tapped by the pumping well is constant during the test to the limits of the cone of depression,
2. The water withdrawn from the aquifer is derived entirely from storage and is discharged instantaneously with the decline in head, and
3. The discharging well penetrates the entire thickness of the aquifer and its diameter is small compared to the pumping rate so that storage in the well is negligible.

These assumptions are most nearly met by confined aquifers at sites remote from their boundaries. However, if certain precautions are observed, the equation can also be used to analyze tests of unconfined aquifers.

The forms of the Theis equation used to determine the transmissivity and storage coefficient are:

$$T = \frac{Q\,W(u)}{4\pi s} \qquad (1)$$

and

$$S = \frac{4Ttu}{r^2}, \qquad (2)$$

where T is transmissivity,
Q is pumping rate,
s is drawdown,
t is time,
r is distance from the pumping well to the observation well, and

$$W(u) = 0.577216 - \log_e u + u - \frac{u^2}{2X2!} + \frac{u^3}{3X3!} - \frac{u^4}{4X4!} \ldots \infty$$

The form of the Theis equation is such that it cannot be solved directly. To overcome this problem Theis devised a convenient graphical method of solution. This method involves matching a data plot of drawdown versus time (or drawdown versus t/r^2) to a type curve of W(u) versus 1/u. At some convenient point on the overlapping part of the data plot and type curve, values of s, t (or t/r^2), W(u), and 1/u are noted. These values are then substituted in Equations (1) and (2) which are solved for T and S, respectively.

A Theis type curve of W(u) versus 1/u can be prepared from the values given in the accompanying table. The data points are plotted on logarithmic graph paper - that is, on graph paper having logarithmic divisions in both the x and y directions.

The fundamental units of transmissivity are L^2/t where t is a day. Traditionally in the United States, L^2 has been expressed in terms of gal./ft. The common practice now is to report transmissivity in units of ft.²/day. If Q is measured in gallons per minute, as is normally the case, and drawdown is in feet, as is also normally the case, we can modify Equation (1) to obtain T in ft.²/day as follows:

$$T = \frac{Q\,W(u)}{4\pi s} = \frac{gal}{min} X \frac{1440\ min}{day} X \frac{ft^3}{7.48\ gal} X \frac{1}{ft} X \frac{W(u)}{4\pi}$$

or

$$T(\text{in } ft^2/day) = \frac{15.3Q\,W(u)}{s}$$

(Q in gal/min and s in ft).

Storage coefficient is dimensionless. Therefore, if T is in ft.²/day, t is in minutes, and r is in feet then

$$S = \frac{4Ttu}{r^2} = \frac{4}{1} X \frac{ft^2}{day} X \frac{min}{ft^2} X \frac{day}{1440\ min},$$

or

$$S = \frac{Ttu}{360\ r^2}\ (T \text{ in } ft^2/day,\ t \text{ in min, and } r \text{ in ft}).$$

Analysis of aquifer-test data using the Theis equation involves plotting both the type curve and the test data on logarithmic graph paper. If the aquifer and the conditions of the test satisfy Theis' assumptions, the type curve has the same shape as (1) the cone of depression along any line radiating away from the pumping well, and (2) the drawdown graph at any point in the cone of depression.

Use of the Theis equation for unconfined aquifers involves two considerations. First, if the aquifer is relatively fine grained, water is not released instantaneously with the decline in head but slowly over a period of hours or days. Therefore, the S determined from a short-period test may be too small.

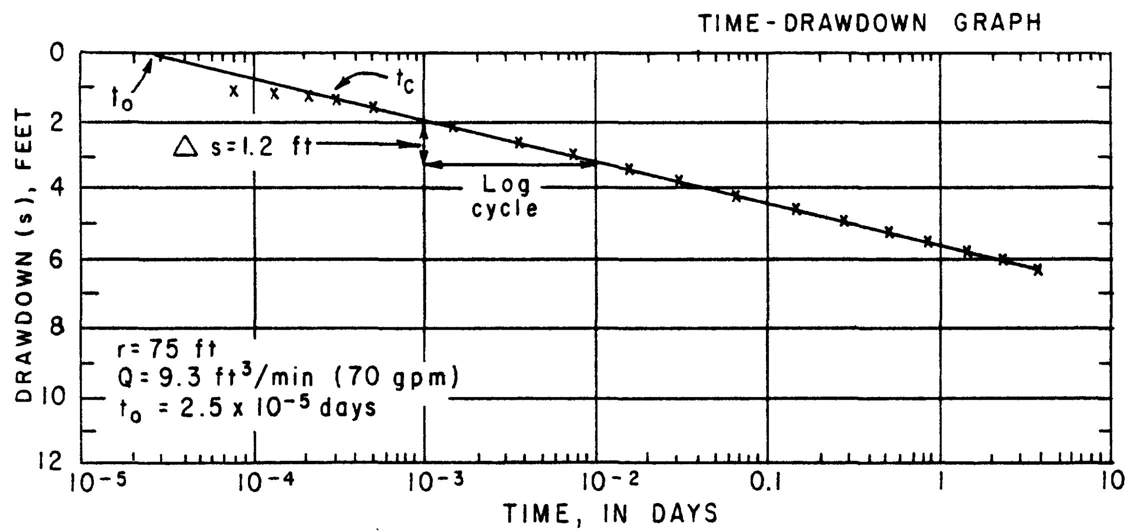
Second, if the pumping rate is large and the observation well is near the pumping well, dewatering of the aquifer may be significant and the assumption that the transmissivity of the aquifer is constant is not satisfied. The effect of dewatering of the aquifer can be eliminated with the following equation:

$$s' = s - \frac{s^2}{2b}, \qquad (3)$$

where s is the observed drawdown in the unconfined aquifer,
b is the aquifer thickness, and
s' is the drawdown that would have occurred if the aquifer had been confined.

To determine the transmissivity and storage coefficient of an unconfined aquifer, a data plot consisting of s' versus t (or t/r^2) is matched with the Theis type curve of W(u) versus 1/u.

Time-Drawdown Analysis



The Theis equation is only one of several methods that have been developed for the analysis of aquifer-test data. (See ANALYSIS OF AQUIFER-TEST DATA.) Another method, and one that is somewhat more convenient to use, was developed by C. E. Jacob (1950) from the Theis equation. The greater convenience of the Jacob method derives partly from the use of semi-logarithmic-graph paper instead of the logarithmic paper used in the Theis method and from the fact that under ideal conditions the data plot along straight lines rather than as a curve.

However, it is essential to note that whereas the Theis equation applies at all times and places (if the assumptions are met), *Jacob's method applies only under certain additional conditions.* These conditions must be satisfied also in order to obtain reliable answers.

To understand the limitations of Jacob's method we must consider the changes that occur in the cone of depression during an aquifer test. The changes that are of concern involve both the *shape of the cone* and the *rate of drawdown*. As the cone of depression migrates outward from a pumping well its shape and, therefore, the hydraulic gradient at different points in the cone, changes. We can refer to this condition as *unsteady shape*. At the start of withdrawals the entire cone of depression has an unsteady shape. After a test has been underway for some time, the cone of depression begins to assume a *steady shape*, first at the pumping well and then gradually to greater and greater distances. If withdrawals continue long enough for increases in recharge or reductions in discharge to balance the rate of withdrawal, drawdowns cease and the cone of depression is said to be in a *steady state*.

The Jacob method is applicable only to the zone in which steady-shape conditions prevail or to the *entire* cone only after steady-state conditions have developed. The time at which steady-shape conditions develop at the outermost observation well can be determined with the following equation:

$$t_c = \frac{7200\ r^2 S}{T} \qquad (1)$$

where t_0 is the time, in minutes, at which steady-shape conditions develop,

r is the distance from the pumping well, in feet,

S is the estimated storage coefficient (dimensionless), and

T is estimated transmissivity, in ft^2/day.

After steady-shape conditions have developed, the drawdowns at an observation well begin to fall along a straight line on semi-logarithmic graph paper, as shown on the accompanying drawing. Prior to that time the drawdowns plot below the extension of the straight line. In preparing a time-drawdown graph, drawdowns are plotted on the vertical (arithmetic) axis versus time on the horizontal (logarithmic) axis.

The slope of the straight line is proportional to the pumping rate and to the transmissivity. Jacob derived the following equations for determination of the transmissivity and storage coefficient from the time-drawdown graphs:

$$T = \frac{2.3Q}{4\pi\Delta s} \qquad (2)$$

$$S = \frac{2.25Tt_0}{r^2} \qquad (3)$$

where Q is the pumping rate,

Δs is the drawdown across one log cycle,

t_0 is the time at the point where the straight line intersects the zero-drawdown line, and

r is the distance from the pumping well to the observation well.

Equations 2 and 3 are in consistent units. Thus if Q is in $ft.^3$/day and s is in ft., T is in $ft.^2$/day. S is dimensionless so that in equation 3 if T is in $ft.^2$/day, then r must be in ft. and t_0 must be in days.

It is still common practice in the United States to express Q in gal./min., s in ft., t in min., r in ft., and T in $ft.^2$/day. We can modify equations 2 and 3 for direct substitution of these units as follows:

$$T = \frac{2.3\,Q}{4\pi\Delta s} = \frac{2.3}{4\pi} \times \frac{gal}{min} \times \frac{1440\ min}{day} \times \frac{ft^3}{7.48\ gal} \times \frac{1}{ft}$$

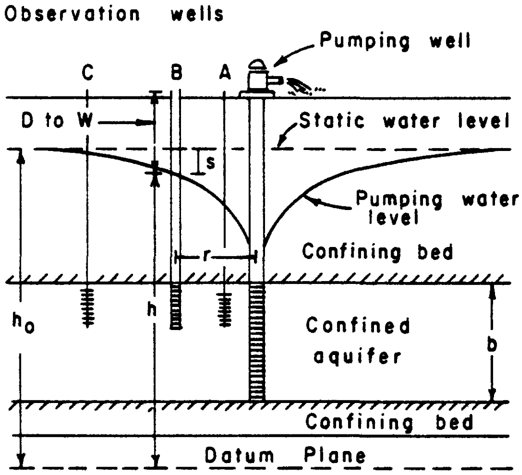
$$T = \frac{35\,Q}{\Delta s} \quad \frac{(\text{T is in ft}^2\text{/day,}}{\text{Q is in gal/min, s is in ft)}} \qquad (4)$$

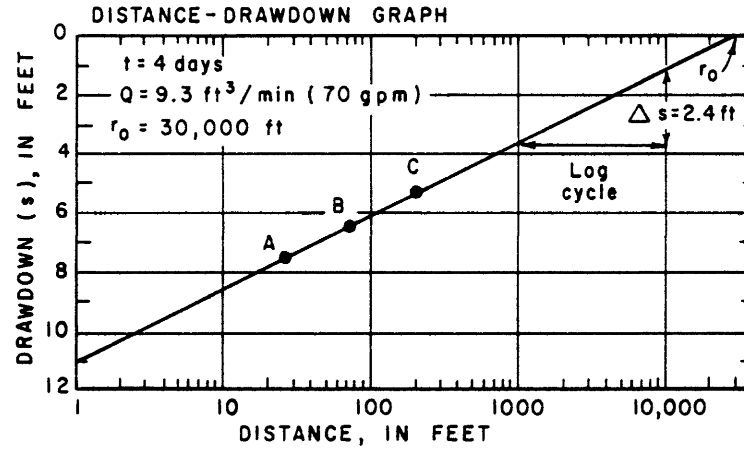
$$S = \frac{2.25\,Tt_0}{r^2} = \frac{2.25}{1} \times \frac{ft^2}{day} \times \frac{min}{ft^2} \times \frac{day}{1440\ min} = \frac{Tt_0}{640r^2}$$

$$S = \frac{Tt_0}{640r^2} \quad \frac{(\text{T is in ft}^2\text{/day,}}{t_0\text{ is in min, r is in ft)}} \qquad (5)$$

Reference: Jacob, C. E., 1950, Flow of ground water, Chap. 5 *in* Rouse, Hunter, Engineering Hydraulics: New York, John Wiley & Sons.

Distance - Drawdown Analysis





It is desirable in aquifer tests to have at least three observation wells located at different distances from the pumping well. If conditions permit, it is also desirable that the wells be located in the same direction from the pumping well. Drawdowns measured at the same time in these wells can be analyzed with the Theis equation and type curve to determine the aquifer transmissivity and storage coefficient.

After the test has been underway long enough, drawdowns in the wells can also be analyzed by the Jacob method, either through the use of a time-drawdown graph using data from individual wells or through the use of a distance-drawdown graph using "simultaneous" measurements in all of the wells. To determine when sufficient time has elapsed, see TIME-DRAWDOWN ANALYSIS.

In the Jacob method, drawdowns are plotted on the vertical (arithmetic) axis versus distance on the horizontal (logarithmic) axis. If the aquifer and test conditions satisfy the Theis assumptions and the limitation of the Jacob method, the drawdowns measured at the same time in different wells should plot along a straight line.

The slope of the straight line is proportional to the pumping rate and to the transmissivity. Jacob derived the following equations for determination of the transmissivity and storage coefficient from distance-drawdown graphs:

$$T = \frac{2.3Q}{2\pi\Delta s} \qquad (1)$$

$$S = \frac{2.25Tt}{r_o^2} \qquad (2)$$

where Q is the pumping rate,

Δs is the drawdown across one log cycle,

t is the time at which the drawdowns were measured, and

r_o is the distance from the pumping well to the point where the straight line intersects the zero-drawdown line.

Equations 1 and 2 are in consistent units. For the inconsistent units still in relatively common use in the United States, equations 1 and 2 should be used in the following forms:

$$T = \frac{70\,Q}{\Delta s} \quad \left(\frac{\text{T in ft}^2/\text{day,}}{\text{Q in gal/min, }\Delta\text{s in ft}}\right) \qquad (3)$$

$$S = \frac{Tt}{640r_o^2} \quad \left(\frac{\text{T in ft}^2/\text{day,}}{\text{t in min, } r_o \text{ in ft}}\right) \qquad (4)$$

(See TIME-DRAWDOWN ANALYSIS.)

Note that the distance r_o does not indicate the outer limit of the cone of depression. Because nonsteady-shape conditions exist in the outer part of the cone, the Jacob method does not apply to that part. If the Theis equation were used to calculate drawdowns in the outer part of the cone, it would be found that they would plot below the straight line. In other words, the measurable limit of the cone of depression is beyond the distance r_o.

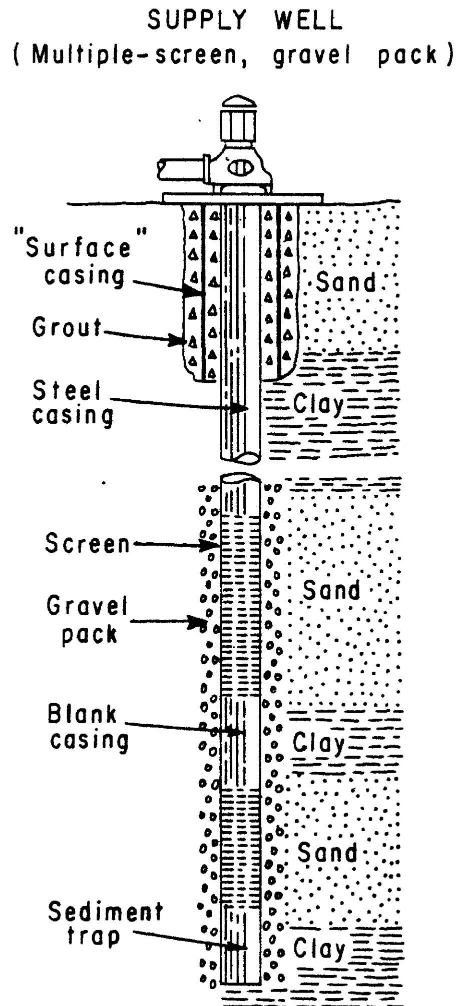
If the straight line of the distance-drawdown graph is extended inward to the radius of the pumping well, the drawdown indicated at that point is the drawdown in the aquifer outside of the well. If the drawdown inside the well is found to be greater than this, the difference is attributable to *well loss*. (See WELL-ACCEPTANCE TESTS.)

PART IV. BASIC ASPECTS OF GROUND-WATER DEVELOPMENT

"As noted earlier, ground-water supplies are considerably more economical to develop and to operate than surface-water supplies are, assuming that the quantity of water needed can be developed from either source. The principal reasons for this are that aquifers function not only as reservoirs but also as pipelines and treatment plants. Where ground-water is not adequate to meet expected needs, the only alternative is to develop a supply from surface sources; however, in the search for new or additional water supplies there are advantages to considering ground water first."

-North Carolina Engineer, Jan. 1973, p. 25.

Well-Construction Methods



Suitability of different well-construction methods to geologic conditions in North Carolina

Characteristics	Dug	Bored	Driven	Jetted	Drilled: Percussion (Cable-tool)	Drilled: Rotary, Hydraulic	Drilled: Rotary, Air
Maximum practical depth, ft.	50	100	50	100	1000	1000	750
Range in diameter, inches	3 ft-20 ft	2-30	1¼-2½	2-12	4-18	4-24	4-10
Type of material:							
Coastal Plain							
Silt	X	X		X	X	X	
Sand		X	X	X	X	X	
Shell beds		X	?	?	X	X	
Limestone		?[1]		?	X	X	
Piedmont and mountain							
Saprolite	X	X	?		X	X	
Bedrock	?				X	X	X

[1]Question marks (?) in the table indicate that the method may be suitable if cemented layers or boulders are not encountered.

The seven different methods of well construction in relatively common use are listed in the above table. The first four methods are limited to relatively shallow depths and are most commonly employed in the construction of domestic wells. One of the last three methods is usually employed in the construction of municipal and industrial wells and domestic wells drawing from bedrock.

The objectives of well construction is to excavate a hole, usually of small diameter, to an aquifer and to provide a means for water to enter the hole while excluding rock material. The means of excavating the hole is different for the different methods.

Dug wells constructed with a pickaxe and shovel were relatively common in the rural areas of North Carolina prior to World War II. Such wells are relatively effective in fine-grained materials, such as saprolite and thinly bedded sand and clay. The large irrigation ponds that are now being dug by bulldozer or dragline in the Coastal Plain are the modern version of the dug well.

Bored wells are constructed with earth augers turned either by hand or power equipment and are the modern equivalent of the small-diameter "hand-dug" well. Bored wells are relatively effective in material of low hydraulic conductivity, such as saprolite, and in areas underlain by thin surficial layers of silty and clayey sand.

Driven wells are constructed by driving a casing equipped with a screened drive point. Because of their relatively small diameter, they are suitable only for relatively permeable surficial aquifers. They are widely used as sources of domestic and farm water supplies in those parts of the Coastal Plain underlain by permeable sand. In rural areas, driven wells are commonly referred to as "pumps."

Jetted wells are constructed by excavating a hole with a high pressure jet of water. In dense clays, shell beds, and partially-cemented layers, it may be necessary to attach a chisel bit to the jet pipe and alternately raise and drop the pipe to cut a hole.

The *percussion drilling method* (commonly referred to as the cable-tool method) consists of alternately raising and dropping a heavy weight equipped with a chisel bit. This shatters the rock at the bottom of the hole which, together with water, forms a slurry that is removed with a bailer. In unconsolidated material the casing is driven a few feet at a time ahead of the drilling. After drilling to the maximum depth to be reached by the well, a screen is "telescoped" inside the casing and held in place while the casing is pulled back to expose the screen. The top of the screen is sealed against the casing by expanding a lead packer. In bedrock wells the normal practice is to firmly "seat" the casing in the top of the rock and drill an open hole to the depth required to obtain the needed yield. (See the sketches of wells in WATER-WELL DESIGN.)

The hydraulic-rotary method excavates a hole by rotating a drill pipe to which one of several types of drag or roller bits are attached. Water containing clay is circulated down the drill pipe, in the "normal-rotary" method, and up the annular space, both to cool the bit and to remove the rock cuttings. In the "reverse-rotary" method the drilling fluid is circulated down the annular space and up the drill pipe. Clay in the drilling fluid adheres to the side of the hole and prevents caving of the formation material. Thus, in the hydraulic-rotary method it is not necessary to install permanent-well casing during the drilling process. When the hole reaches the desired depth, a line of casing containing sections of screens at the desired intervals is lowered in the well. Hydraulic-rotary is the method most-commonly employed in drilling large-yield wells in the Coastal Plain. Because the deeper aquifers in most of the area consist of alternating thin beds of sand and clay, the common practice is to install a gravel envelope around the screens. Such wells are referred to as *gravel-packed*. (See the preceding sketch of a multiple-screened, gravel-packed supply well.)

The *air-rotary method* is similar to the hydraulic-rotary method, except that the drilling fluid is air rather than mud. The air-rotary method is suitable only for drilling in consolidated rocks and is, therefore, used only in the Piedmont and mountains. Most air-rotary rigs are also equipped with mud pumps which permit them to be used in the hydraulic-rotary mode when drilling through a thick, permeable section of saprolite. This method is widely used in the construction of bedrock wells for domestic needs.

Upon completion of the construction phase, it is necessary to begin the phase referred to as *well development*. The objective of this phase is to remove clay, silt, and fine-grained sand from the area adjacent to the screen or open hole so that the well will produce sediment-free water. The simpliest method of development is to pump water from the well at a gradually increasing rate with the final rate being larger than the planned production rate. This method is not normally successful in screened and gravel-packed wells drilled with the hydraulic-rotary method. For these it is necessary to use a surge block or some other means to alternately force water into the formation and pull it back into the well. One of the most effective methods is to pump water under high pressure through orifices directed at the inside of the screen. The coarser-grained particles pulled into the well during development tend to settle to the bottom of the well and must be removed with a bailer or pump.

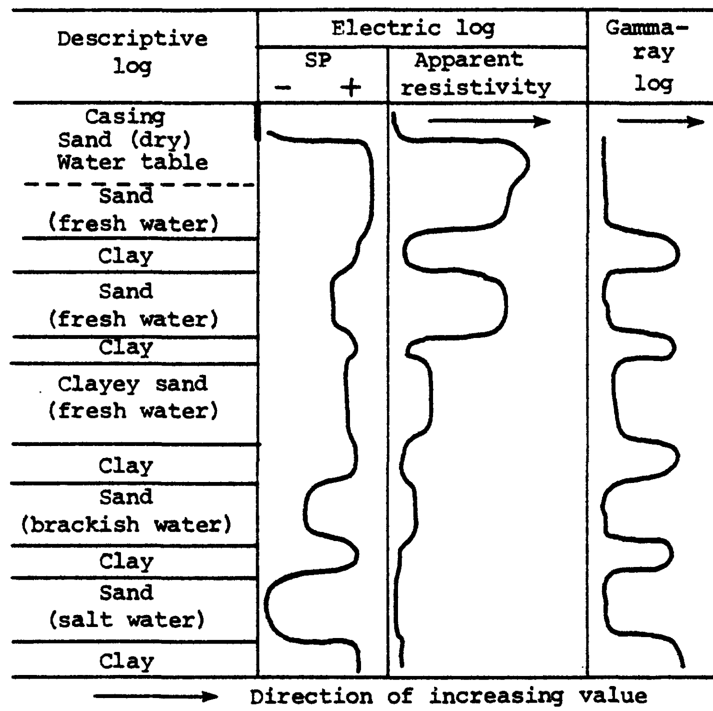
References: Campbell, M. C., and Lehr, J. H., 1973, Water well technology: New York, McGraw-Hill Book Co.
U.S. Environmental Protection Agency, 1974, Manual of individual water supply systems: EPA-430/74-007.

Well Logs

Drillers log

From (ft)	To (ft)	Rock material
0	4	Black soil
4	50	Sandy clay, yellow
50	80	White fine sand
80	185	Red and green clay
185	195	Red sandy clay
195	225	Blue slate
225	704	Blue slate

Idealized geophysical logs



An important part of well construction is to determine the character and the thickness of the different layers of material penetrated by the well. This information is essential for the installation of casing and for the proper placement of screens. Information on materials penetrated is recorded in the form of "logs." The logs most commonly prepared for supply wells are (1) drillers logs, and (2) geophysical (electric) logs. Copies of logs should be carefully preserved as a part of the file on each well.

Drillers logs consist of written descriptions of the material penetrated by wells. These descriptions are based both on samples of rock cuttings brought to the surface during drilling operations and by changes in the rate of penetration of the drill and changes in vibration of the rig. The well driller may also collect samples of the rock cuttings for study by geologists on his staff or those on the staff of State Geological Surveys or Federal and State water-resources agencies. Descriptions of these samples utilizing a microscope and other aids are commonly referred to as a *geologic log* to differentiate them from the driller's log. If the well is to be finished with a screen, the well driller will retain samples of material from the principal water-bearing zones for use in selecting the slot size of screens.

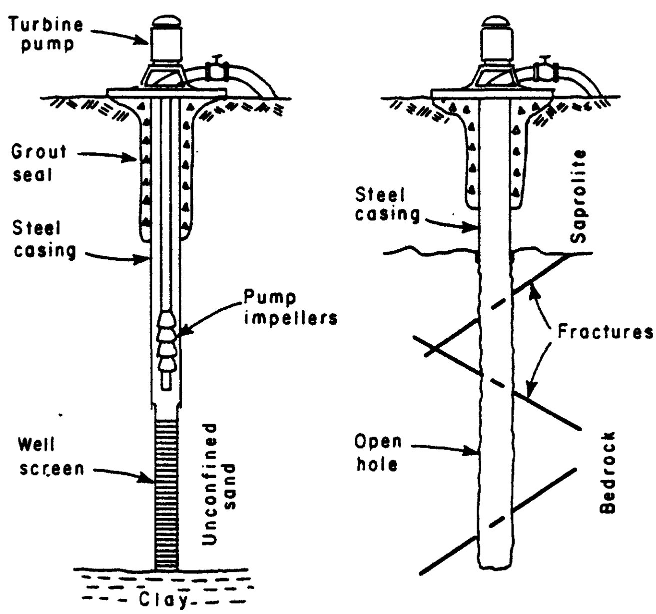
Geophysical logs provide indirect information on the character of rock layers. The most common type of geophysical log, the type normally referred to as an *electric log*, consists of a record of the apparent electrical resistivity of the rock units and of the spontaneous electrical potentials generated in the borehole. Several types of electric loggers are available but nearly all provide continuous graphs of spontanous potential and resistivity as a sensing device is lowered into and removed from the borehole. Electric logs can be made only in the uncased portion of drill holes. The part of the hole to be logged must also contain drilling mud or water.

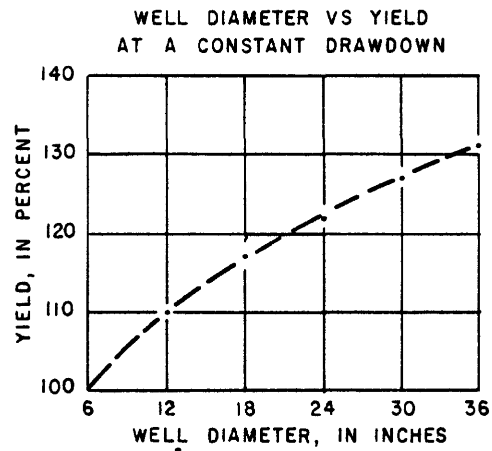
The spontaneous potential log (which is usually referred to as the SP log) is a record of the differences in voltage between an electrode at the land surface and an electrode in the borehole. Variations in voltage occur as a result of electrochemical and other spontaneous electrical effects. The SP graph is relatively featureless in shallow water wells that penetrate only the fresh-water zone. The right-hand boundary of an SP log generally indicates impermeable beds such as clay, shale, and bedrock. The left-hand boundary generally indicates sand, cavernous limestone, and other permeable layers.

The *resistivity log* is a record of the resistance to the flow of an alternating electric current offered by the rock layers and their contained fluids and the fluid in the borehole. There are several different electrode arrangements that measure the resistivity of different volumes of material but the arrangement most commonly used by the water-well industry is that referred to as the single-point electrode. The resistivity of water-bearing material primarily depends on the salt content of the water and the porosity of the material. Clay layers normally have a low resistivity because the water they contain tends to be relatively highly mineralized. In contrast, sand layers saturated with fresh water tend to have a high resistivity. Sand layers containing salty water, on the other hand, tend to have a low resistivity resembling that of clay layers. Such layers tend to have a strongly negative spontaneous potential which, viewed together with the resistivity, aids in identification of the layers.

Several other types of geophysical logs are available, including gamma-ray logs that record the rate of emission of gamma rays by different rock layers. In fact, geophysical logging is a complex topic that has been developed, largely by the oil industry, into an advanced technical field. It is being utilized to an increasing extent by the water-well industry, especially in conjunction with the construction of wells by the hydraulic-rotary method.

Water-Well Design





Water-well design is the first step in the construction of large-yield wells, such as those required by municipalities and industries. Before the design is started, it is necessary to know the yield desired from the well and the depth to, composition, and hydraulic characteristics of the aquifers underlying the area and the quality of water in the aquifers. The completed design should specify the (1) diameter, (2) total depth of the well and the position of the screen or open-hole sections, (3) method of construction, and (4) the materials to be used in the construction.

The *well diameter* is determined primarily by two factors; (1) the desired yield, and (2) the depth to the source aquifer. The diameter has a relatively insignificant effect on yield as shown in the above graph of the relative yield, at a constant drawdown, of wells of different diameters. For example, doubling the diameter from 6 inches to 12 inches only results in a 10 percent increase in yield.

The primary effect of well diameter on yield is related to the size of the pump that can be installed which, in turn, determines the pumping rate. Data on pumping rate, pump size, and well diameter are given in Table 1. In some designs, the upper part of the well will be made larger than the remainder of the well in order to accommodate the pump.

The well diameter, together with screen length, slot size, and pumping rate, also determines the velocity at which water passes through the screen (that is the so-called "entrance velocity"). The entrance velocity should not normally exceed about 6 ft/min. If the anticipated yield in ft^3/min shown in Table 1 is divided by 6 ft/min, the result is the open area of screen needed in ft^2.

The amount of open area in well screens depends on the diameter, the slot size, and the type of screen. Table 2 shows, for example, the open area of screens manufactured by the Edward E. Johnson, Co.[1] If the open area needed in ft^2 is divided by the open area per lineal foot, the result is the length of screen, in feet, required to provide the yield without exceeding the recommended entrance velocity.

The depth to the source aquifer also affects the well diameter to the extent that wells expected to reach aquifers more than a few hundred feet below land surface must be large enough to accept the larger-diameter cable-tool or drill rods required to reach these depths.

The *total depth* of the well depends on the depth below land surface to the bottom of the lowest section of screen or open hole.

[1]The use of a company name is for identification purposes only and does not empIy endorsement by the U.S. Geological Survey.

Table 1.—Data on yield, pump size, and well diameter.

Anticipated well yield		Nominal size of pump bowls, in inches	Optimum well diameter, in inches
in gal/min	in ft³/min		
Less than 100	Less than 13	4	6 ID[1]
75 to 175	10 to 23	5	8 ID
150 to 400	20 to 53	6	10 ID
350 to 650	47 to 87	8	12 ID
600 to 900	80 to 120	10	14 OD[2]
850 to 1300	113 to 173	12	16 OD
1200 to 1800	160 to 240	14	20 OD
1600 to 3000	213 to 400	16	24 OD

[1]ID = inside diameter. [2]OD = outside diameter.

Table 2.—Open areas of Johnson well screens.

Nominal Screen dia., in inches	Open areas per lineal foot of screen, in square feet [1]						
	Slot No. 10	Slot No. 20	Slot No. 40	Slot No. 60	Slot No. 80	Slot No. 100	Slot No. 150
4	0.1	0.18	0.31	0.40	0.40	0.51	0.61
5	.12	.23	.38	.50	.51	.65	.78
6	.14	.27	.45	.59	.60	.77	.92
8	.19	.35	.60	.78	.81	.91	1.11
10	.25	.45	.76	.99	1.02	1.15	1.41
12	.29	.53	.90	1.18	1.21	1.25	1.55
14	.26	.49	.85	1.13	1.23	1.38	1.74
16	.24	.48	.85	1.14	1.19	1.38	1.74

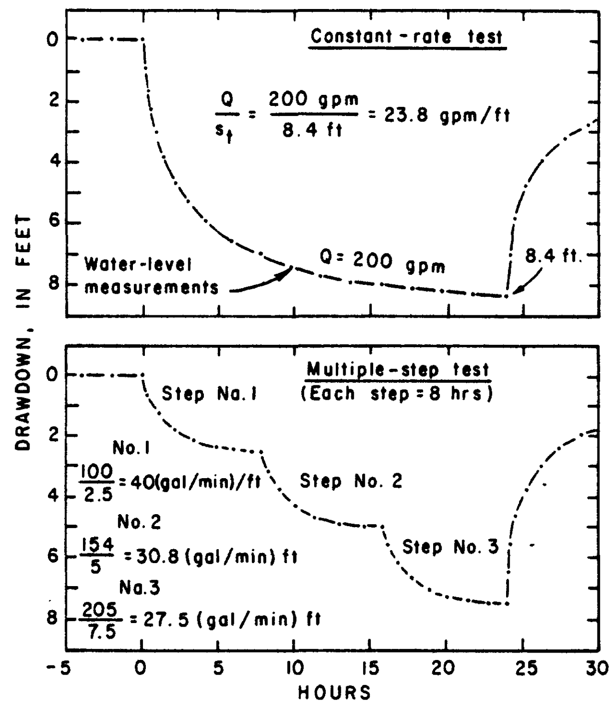
[1]Slot No. denotes width of screen opening in thousandths (1/1000) of an inch. Slot No. 10 indicates an opening 10/1000 or (0.01) inch wide.

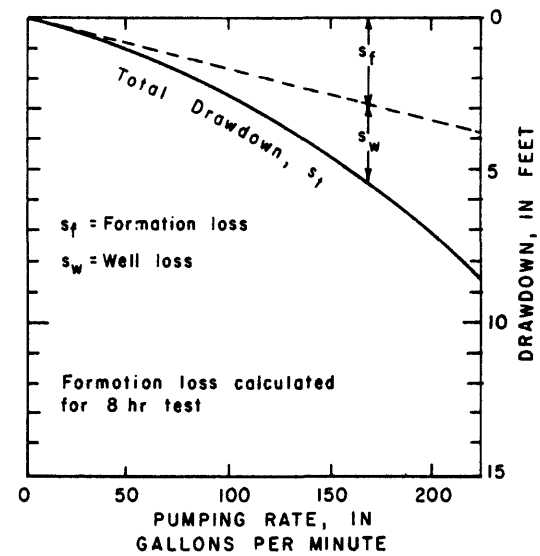
The *position of the screen* depends on the thickness and composition of the source aquifer and whether the well is being designed to obtain the maximum possible yield. Detailed treatment of this topic is beyond the scope of this discussion. However, screens are necessary in the Coastal Plain area. Wells drawing from the unconfined surficial aquifer in this area are screened through about the lower third of the aquifer. Wells drawing from the sediments of Cretaceous age, which consist of alternating thin beds of sand and clay, may be finished with several sections of screen set in the coarsest-grained layers. In the optimum situation, the total length of screens should be at least half the thickness of the aquifer. Most large-yield wells in the Cretaceous aquifer are also "gravel packed." (See WELL-CONSTRUCTION METHODS.)

Most wells in the Piedmont and mountains are cased through the saprolite and finished with open holes in the bedrock. The length of the open hole depends on the desired yield and the hydraulic conductivity of the fractures penetrated by the well. Thus, well design in this area is both simplier and the results are less predictable than in the Coastal Plain. The most important problem related to the development of ground-water supplies is selection of well sites. (See SELECTING WELL SITES IN THE PIEDMONT AND MOUNTAINS.)

Reference: Edward E. Johnson, Inc., 1966, Ground Water and wells: St. Paul, Minn.

Well-Acceptance Tests





Most supply-well contracts require a "guaranteed" yield and some stipulate that the well reach a certain level of "efficiency." Most contracts also specify the length of the "drawdown test" that must be conducted to demonstrate that the yield requirement is met. For example, the North Carolina Department of Human Resources requires that tests of public-supply wells be at least of 24-hrs. duration. Tests of most industrial and irrigation wells probably do not exceed about 8 hrs.

Well-acceptance tests, if properly conducted, can not only confirm the yield of a well but also provide information of great value in well operation and maintenance. Such tests should, therefore, be conducted with the same care as aquifer tests made to determine the hydraulic characteristics of aquifers. A properly conducted test will include:

1. Determination of well interference from nearby pumping wells, based on accurate water-level measurements made prior to the drawdown test.
2. A pumping rate that is either held constant during the entire test or is increased in steps of equal length. The pumping rate during each step should be held constant and the length of each step should be at least 2 hrs.
3. Accurate measurements of the position of the water level in the well prior to starting the pump and at regular intervals throughout the test. Measurements should be made at intervals of 1 to 2 min. at the beginning of the test and after each change in rate.
4. Accurately recorded watch time at which each water-level measurement and each change in pumping rate are made.

Of the requirements listed above, the constant, carefully regulated pumping rate or rates, and accurate water-level measurements are the most important. Most well acceptance tests are made with temporary pump installations, usually powered with a gasoline or diesel engine. Instead of maintaining a constant rate for the duration of the test, the engine is frequently stopped to add fuel, check the oil level, and numerous other reasons. The rate is also increased and decreased on an irregular, unplanned schedule so that at the end of the test the "yield" of the well may be more speculative than factual.

Upon completion of a well-acceptance test, the pumping rate and drawdown data are analyzed in terms of specific capacity. *Specific capacity* is the ratio of the pumping rate and the drawdown. Thus

$$\text{Specific capacity} = \frac{Q}{s_t},$$

where Q is the pumping rate (either a single rate for the entire test or the rate during each step), and

s_t is the drawdown in the well at the end of a constant-rate test or at the end of each step.

Specific capacity is most useful when expressed in terms of yield per unit of drawdown. This value is calculated by dividing the pumping rate by the drawdown.

The total drawdown in most, if not all, pumping wells consists of two components. One is the drawdown, s_f, in the aquifer and the other is the drawdown, s_w, that occurs as water moves from the aquifer into the well.

The drawdown in the aquifer is referred to as *formation loss* and it can be calculated if the hydraulic characteristics of the aquifer are known. (See DISTANCE-DRAWDOWN ANALYSIS.)

The drawdown across the well face is referred to as *well loss* and is caused by the development of turbulent flow both across the well face and up the well bore. In screened wells, well loss is probably due primarily to partial blockage of the screen openings with aquifer material. Well loss has also been observed, somewhat unexpectedly, in open-hole, bedrock wells. The reason for this is not known but it probably results from the development of turbulent flow in the fractures adjacent to the wells. The well loss can be determined by subtracting the calculated formation loss from the total drawdown.

The total drawdown in a pumping well can be expressed in the form of the following equations:

$$s_t = s_f + sw$$

$$s_t = BQ + CQ^2 \qquad (1)$$

where Q is the pumping rate,

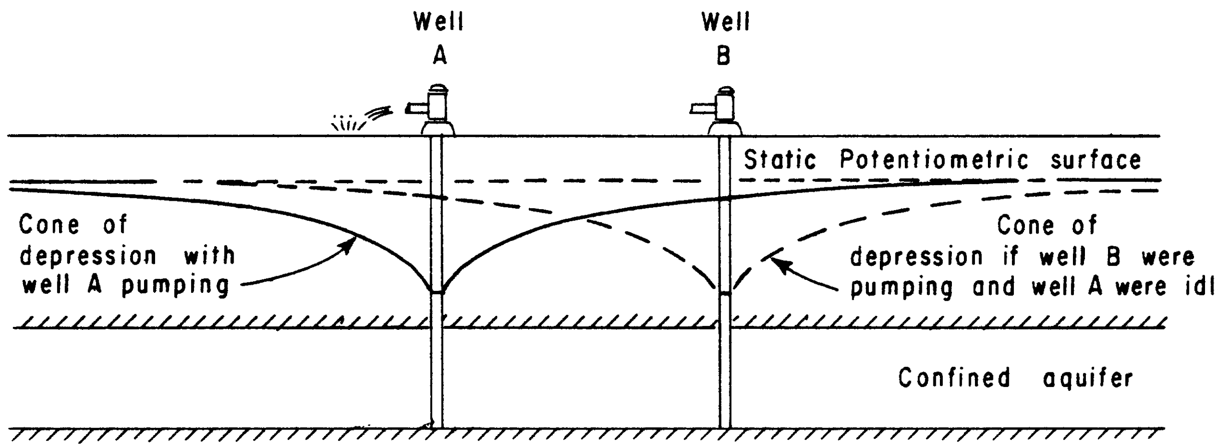
B is a constant related to the characteristics of the aquifer, and

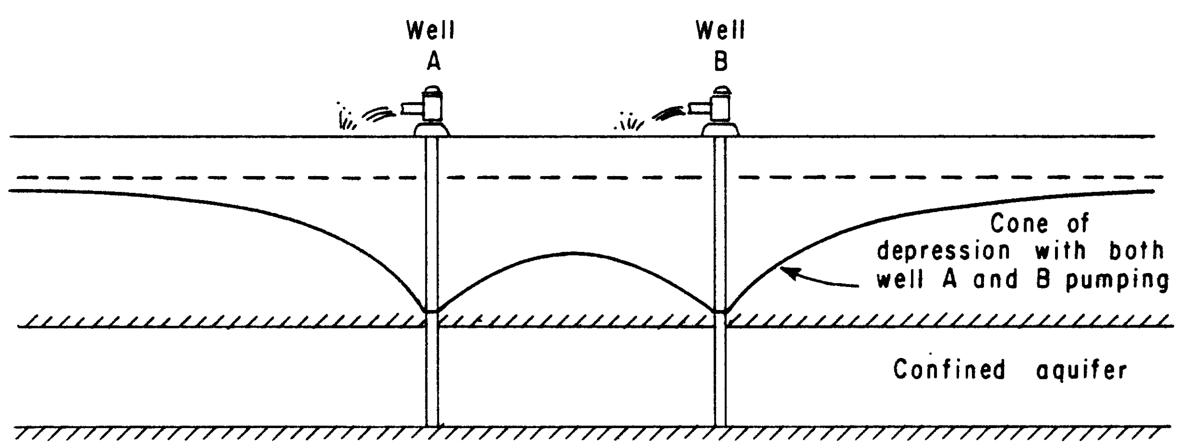
C is a constant related to the characteristics of the well.

Well efficiency is the ratio of the actual specific capacity determined from a well-acceptance test and the specific capacity that would exist if drawdown were caused only by formation loss. Therefore, referring to equation 1, if the total drawdown were due only to the factor BQ the well would be 100 percent efficient. The larger the well loss factor CQ^2, the more inefficient the well.

It is especially important to note that well loss is related to entrance velocity, which is proportional to the pumping rate. Thus, in equation 1 the drawdown due to well loss is proportional to the square of the pumping rate. Consequently, every effort is made in water-well design and in well development to keep well loss to a minimum. However, an efficiency of about 80 percent is the maximum that is normally achievable in most screened wells.

Well Interference





A pumping well causes a drawdown in the ground-water level in its vicinity. The drawdown in water level forms a conical-shaped depression in the water table or potentiometric surface which is referred to as a *cone of depression*. (See CONE OF DEPRESSION.)

The drawdown, s, caused by pumping at any point in an aquifer is directly proportional to the pumping rate, Q, and the length of time, t, pumping has been in progress; and inversely proportional to the transmissivity, T, storage coefficient, S, and the distance squared, r^2, between the pumping well and the point. In other words,

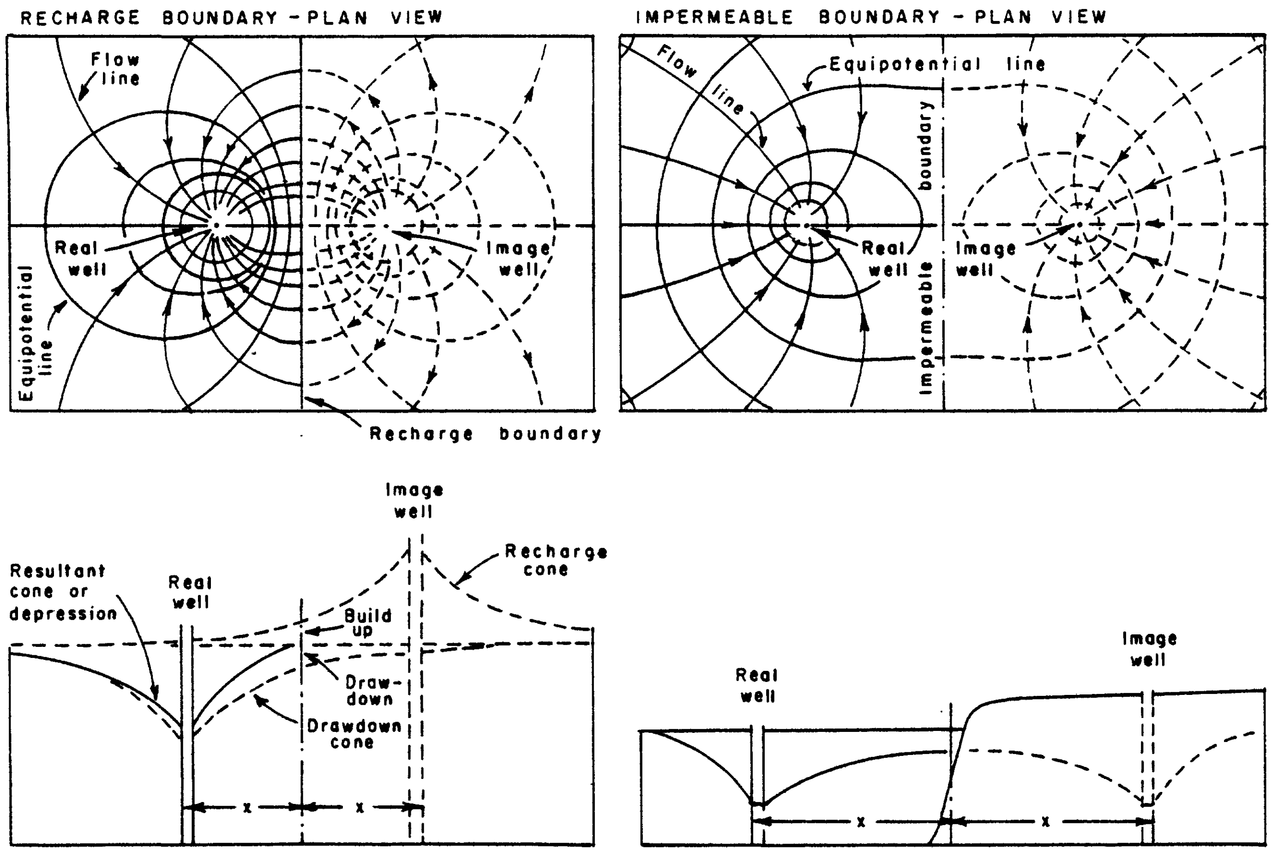
$$s = f\left(\frac{Q, t}{T, S, r^2}\right)$$

Where pumping wells are spaced relatively close together, pumping of one may cause a drawdown in the other. The drawdowns in pumping wells caused by withdrawals from other pumping wells are referred to as *well interference*.

We see from the above functional equation that drawdown is directly proportional to the pumping rate. Conversely, the maximum pumping rate is directly proportional to the *available drawdown.* For confined aquifers, available drawdown is normally taken as the distance between the static water level and the top of the aquifer. For unconfined aquifers, available drawdown is normally taken as about 60 percent of the aquifer thickness.

Where the pumping rate of a well is such that only a part of the available drawdown is utilized, well interference lowers the pumping level and results in an increase in pumping costs. In the design of a well field, the increase in pumping cost must be evaluated along with the cost of the additional water lines and power lines that must be installed if the spacing of wells is increased to reduce well interference. (See WELL-FIELD DESIGN.)

Aquifer Boundaries



One of the assumptions inherent in the Theis equation, and most other ground-water flow equations, is that the aquifer to which they are being applied is infinite in extent. Obviously, no such aquifer exists on Earth so that, from a practical standpoint, it is simply assumed that the data being analyzed are not affected by any of the aquifer boundaries.

All aquifers are bounded in both the vertical direction and in the horizontal direction. For example, vertical boundaries include the water table, the plane of contact between each aquifer and each confining bed and, at the base of the ground-water system, the plane marking the lower limit of the zone of interconnected openings.

Hydraulically, aquifer boundaries are of two types; recharge boundaries and impermeable boundaries. A *recharge boundary* is a boundary along which flow lines originate. In other words, such a boundary will, under certain hydraulic conditions, serve as a source of recharge to the aquifer. Examples of recharge boundaries include the zones of contact between an aquifer and a perennial stream or the ocean.

An *impermeable boundary* is a boundary which flow lines cannot cross. Such boundaries exist where aquifers terminate against "impermeable" material. Examples include the contact between an aquifer composed of sand and a laterally adjacent bed composed of clay.

The position and nature of aquifer boundaries is of critical importance in many ground-water problems, including the movement and fate of pollutants and the response of aquifers to withdrawals. Depending on the direction of the hydraulic gradient, a recharge boundary may either be the source or the destination of a pollutant.

Lateral boundaries have a profound effect on the response of an aquifer to withdrawals. In order to analyze, or to predict, the effect of a lateral boundary, it is necessary "to make" the aquifer appear to be of infinite extent. This feat is accomplished through the use of imaginary wells and the *theory of images.* The preceeding sketches show, in both plan view and profile, how image wells are used to compensate, hydraulically, for the effects of both recharging and impermeable boundaries.

The key feature of a recharge boundary is that withdrawals from the aquifer do not produce drawdowns across the boundary. We can, therefore, duplicate the hydraulic effect of the boundary by assuming that a recharging image well is present on the opposite side of the boundary from the real discharging well. Water is injected into the image well at the same rate and on the same schedule that water is withdrawn from the real well. Note that flow lines originate at the boundary and equipotential lines parallel the boundary.

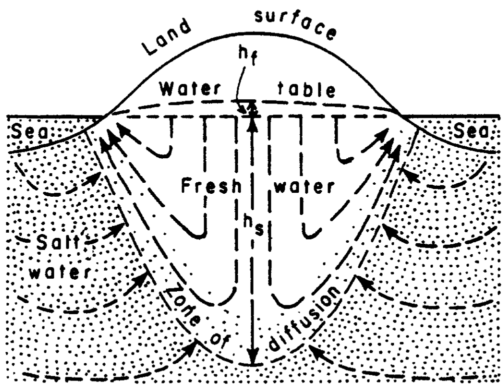
The key feature of an impermeable boundary is that no water can cross it. Such a boundary, therefore, resembles a divide in the water table. We can duplicate the effect of an impermeable boundary by assuming that a discharging image well is present on the opposite side of the boundary from the real discharging well. The image well withdraws water at the same rate and on the same schedule as the real well. Note that flow lines parallel an impermeable boundary and equipotential lines intersect it at a right angle.

The image-well theory is an essential tool in the design of well fields near aquifer boundaries. We can readily see from this theory that

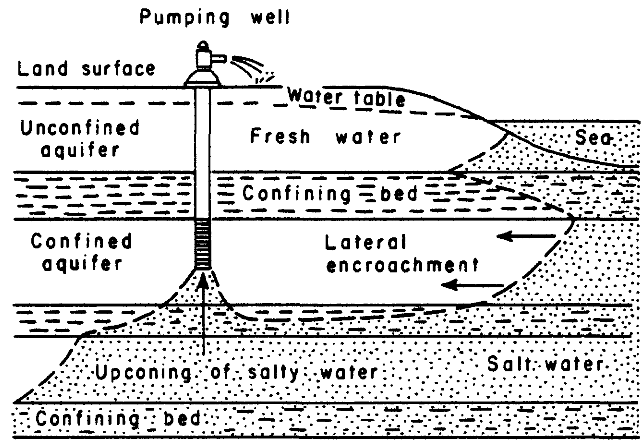
1. Pumping wells should be located parallel to and as close as possible to recharging boundaries, and
2. Pumping wells should be located perpendicular to and as far as possible from impermeable boundaries.

Salt-Water Encroachment

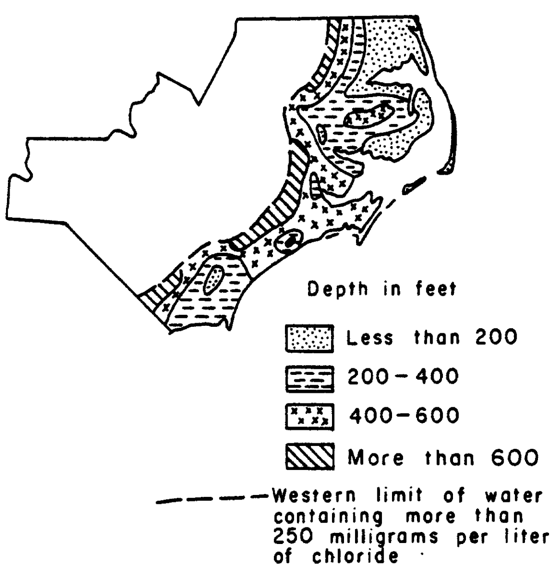
Fresh-water lens floating on salt water



Two aspects of salt-water encroachment



Depth to salty water



In coastal areas, fresh ground water derived from precipitation on the land comes in contact with and discharges into the sea, or into estuaries containing brackish water. The relation between the fresh water and the sea water, or brackish water, is controlled primarily by their differences in density.

The *density* of a substance is its mass per unit volume. The density of water is affected by the amount of minerals, such as common salt (NaCl), that the water contains in solution. Therefore, sea water has a larger density than fresh water. In metric units, the density of fresh water is about 1 gram per cubic centimeter and the density of sea water is about 1.025 gm/cm^3. Thus, fresh water, being less dense than sea water, tends to override or float on sea water.

On islands, such as the Outer Banks of North Carolina, precipitation forms a fresh-water lens that "floats on" the underlying salt water. The higher the water table stands above sea level, the thicker the fresh-water lens. This relation between the height of the water table and the thickness of the fresh-water lens was discovered, independently, by a Dutchman, Badon Ghyben, and a German, B. Herzberg, and is referred to as the *Ghyben-Herzberg relationship*. This relation, expressed as an equation is

$$h_s = \frac{\rho f}{\rho s - \rho f}(h_f) \qquad (1)$$

where h_s is the depth of fresh water below sea level,

ρ_f is the density of fresh water,

ρ_s is the density of sea water,

h_f is the height of the water table above sea level.

Based on equation 1 and the differences in density between fresh water and sea water, the fresh-water zone should extend to a depth below sea level, h_s, equal to 40 times the height, h_f, of the water table above sea level. The Ghyben-Herzberg relation strictly applies, however, only to a homogenous and isotropic aquifer in which the fresh water is static and is in contact with a tideless sea or brackish-water body.

Tides cause salt water to alternately invade and to retreat from the fresh-water zone, thereby forming a zone of diffusion across which the salinity changes from that of fresh water to that of sea water. A part of the sea water that invades the fresh-water zone is entrained in the fresh water and is flushed back to the sea by the fresh water as it moves to the sea to discharge.

Because both the sea water and the fresh water are in motion (not static), the thickness of the fresh-water zone is greater in a homogenous and isotropic aquifer than predicted by the Ghyben-Herzberg equation. On the other hand, in a stratified aquifer (and nearly all aquifers are stratified) the thickness of the fresh-water lens is less than predicted because of the head loss incurred as the fresh water moves across the least permeable beds.

When fresh-water heads are lowered by withdrawals through wells, the fresh water-salt water contact migrates toward the point of withdrawals until a new balance is established. The movement of salt water into zones previously occupied by fresh water is referred to as *salt-water encroachment.*

Salt-water encroachment is a serious problem in some coastal areas. *Upconing of salty water* beneath pumping wells is a more imminent problem in most areas than is *lateral encroachment*. One of the reasons for this is that a much larger volume of fresh water has to be displaced in lateral encroachment than in upconing.

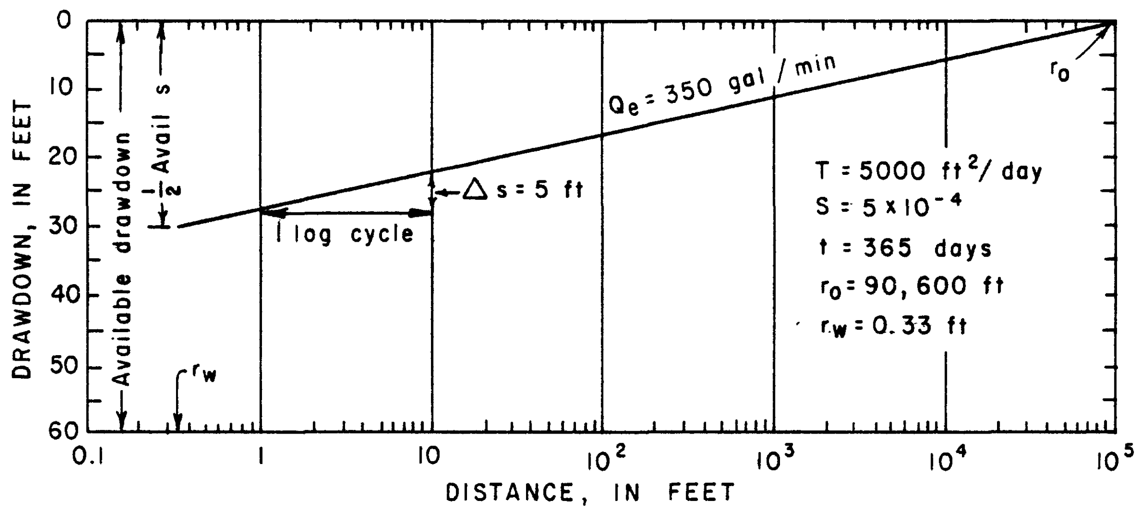
Salty water occurs in the sediments underlying the eastern part of the Coastal Plain at depths controlled by both the fresh-water heads and the stratification of the sediments. As shown on the accompanying sketch of eastern North Carolina, the depth to salty water ranges from less than 200 ft to more than 600 ft. On most of the Outer Banks, the depth to salty water is less than 25 ft.

In the design of supply wells in areas underlain by or adjacent to salty water, consideration must be given to the possibility of salt-water encroachment. This consideration may involve selection of shallow aquifers or small pumping rates to avoid upconing or involve moving wells to more inland locations to avoid lateral encroachment.

References: Cooper, H. H., Jr., Kohout, F. A., Henry, H. R., and Glover, R.E., 1964, Sea water in coastal aquifers: U.S. Geological Survey, Water-Supply Paper 1613-C

Winner, M.D., Jr., 1975, Ground-water resources of the Cape Hatteras National Seashore, North Carolina: U.S. Geological Survey, Hydrologic Atlas HA-540.

Well-Field Design



The development of moderate to large supplies of water from most aquifers requires more than one well; in other words, it requires what is commonly referred to as a *well field.* Consequently, the design of well fields is an important problem in ground-water development. The objective of well-field design is to obtain the required amount of water for the least cost, including both the initial construction cost and the cost of operation and maintenance.

The final product of a design is a plan showing the arrangement and spacing of wells and specifications containing details on well construction and completion, including information on well diameter, depth, and position of screens or open hole, type of casing and screens, and the type, size, and setting of pumps.

The key elements in well-field design are: (1) the total quantity of water to be obtained from the field, (2) the rate at which each well can be pumped (which determines the number of wells that will be required), and (3) the spacing of the wells.

The *pumping rate for each well* can be estimated with Jacob's modification of the Theis equation. (See DISTANCE-DRAWDOWN ANALYSIS.) It depends on the transmissivity and storage coefficient of the aquifer, the available drawdown, and the pumping period. (For a discussion of available drawdown, see WELL INTERFERENCE.) The *pumping period* is normally taken as 1 yr. To determine the pumping rate, Jacob's equations are solved as follows:

$$r_0^2 = \frac{2.25Tt}{S} \qquad (1)$$

$$Q_e = 2.7T\Delta s \qquad (2)$$

where r_o is the distance from the pumping well, in feet, to the point of zero drawdown on a semi-logarithmic graph in which drawdown is on the arithmetic scale and distance is on the logarithmic scale,

T is aquifer transmissivity, in ft^2/day,

t is 365 days (1 yr),

S is the aquifer storage coefficient (dimensionless),

Δs is the drawdown, in feet, across one log cycle along a line connecting point r_o and a point at the proposed radius of the pumping well at which the drawdown equals about half the available drawdown[1].

Q_e is the first estimate of the pumping rate in ft^3/day. To convert to gallons per minute, divide Qe by 192.

[1]At this point, we use 1/2 the available drawdown in order to get a first estimate of well loss and well interference. If we determine that at a pumping rate of Q_e the drawdown in the formation is less than the available drawdown and the drawdown in the well is above the top of the screen, we can assume a larger Δs and recompute Q_e. It is important also to note that in the initial determination of available drawdown, the seasonal fluctuation of static water level must be considered.

The estimated pumping rate is divided into the total quantity of water needed from the well field in order to determine the number of wells that will be needed. The next step is to calculate the *well spacing*. Theis devised the following equation for this purpose:

$$r_s = \frac{3.2 \times 10^7 Q_e^2 c}{kT} \qquad (3)$$

where r_s is the distance, in feet, between wells,

Q_e is the estimated pumping rate of each well,

c is the power charge for lifting the water,

k is the initial cost of construction, depreciation, and maintenance of the interconnecting water and power lines, expressed in units of dollars per year per foot, and

T is transmissivity in ft^2/day.

To illustrate the use of equation 3, we will assume electrical power costs 3 cents per kilowatt hour and pump efficiency is 50 percent. The power charge, c, will then be about 2×10^{-7} dollars per gallon per ft. Assuming the cost of the water line and power line is $20 per ft and capitalizing this at 10 percent gives a cost, k, of $2 per ft per yr for capital charges, depreciation, and maintenance. If the transmissivity of the aquifer is 5,000 ft^2/day and the pumping rate, Q_e, is 350 gal/min, then

$$r_s = \frac{(3.2 \times 10^7)\ (350)^2\ (2 \times 10^{-7})}{(2)\ (5{,}000)}$$

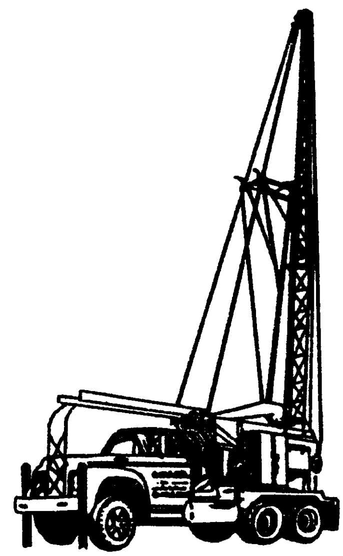
$$= \frac{7.8 \times 10^5}{1 \times 10^4} = 78 \text{ ft}$$

According to equation 3, if the cost factors, c and K, are constant, the optimum spacing, r_s, of two discharging wells varies directly with the square of the pumping rate, Q_e, and inversely with transmissivity. Thus, the larger the pumping rate and the smaller the transmissivity, the greater the spacing.

Obviously, equation 3 does not cover all factors involved in well-field design. Among the additional factors that should be considered are:

1. The minimum distance between pumping wells should be at least twice the aquifer thickness if the wells are open to less than about half the aquifer thickness.
2. Wells near recharging boundaries should be located along a line parallel to the boundary and as close to the boundary as possible, and
3. Wells near impermeable boundaries should be located along a line perpendicular to the boundary and as far from the boundary as possible.

Reference: Theis, C. V., 1963, Spacing of wells: U.S. Geological Survey, Water-Supply Paper 1545-C, p. C113-C117.

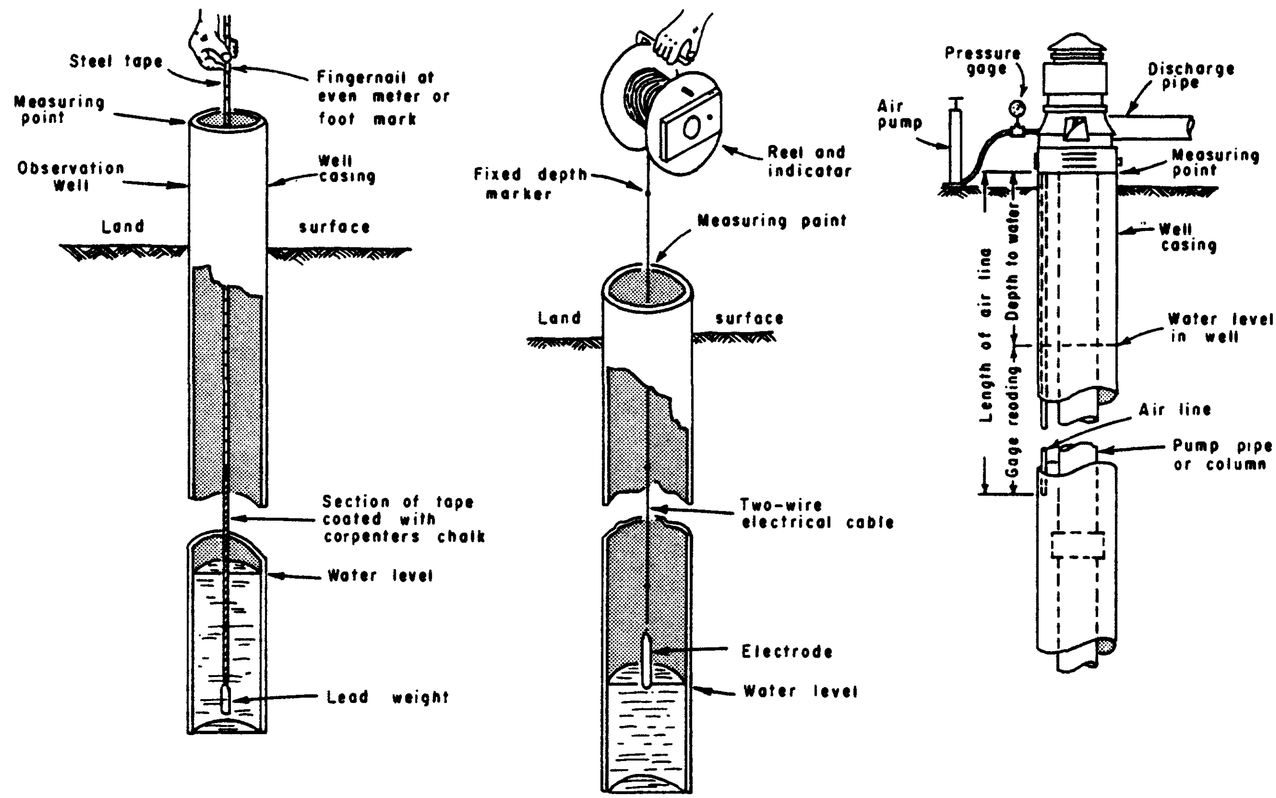


PART V. OPERATION OF GROUND-WATER SUPPLIES

"there is increasing emphasis on the development of water supplies from streams and reservoirs in preference to the development of ground-water supplies. Various reasons are given for this, including fear of salt-water contamination of ground-water supplies. The possibility of salt-water contamination is certainly a real problem in some coastal areas, but it is important to realize that no significant contamination has yet occurred any place in the State. There are several advantages in the use of ground-water supplies. These may include a considerable saving in the cost of development and operation, relative immunity of the supplies to droughts, natural and manmade disasters, and bacterial and chemical pollutants. Another and possibly even more important advantage concerns the effect of ground-water use on conservation of the State's water resources. The development of ground-water supplies often does not require large-scale surface-storage facilities with the resulting losses to evaporation. Second, where ground-water withdrawals result in a managed decline in the water table the loss of water by evaporation from the land surface and by the transpiration of plants is reduced. Reductions in these losses represents potential increase in the available water supply. Thus, the extensive development of ground-water supplies, especially from the shallowest aquifers, may have the same beneficial effect as a desalting plant.

-Review of water-resources problems in North Carolina: *News*; Water Resources Research Inst., Univ. of North Carolina, No. 110, March 1975.

Measurements of Water Levels and Pumping Rates



Each supply well should be provided with a means for measuring the position of the water level in the well and the pumping rate. (The use of these measurements is discussed in SUPPLY-WELL PROBLEM - *Decline in Yield.*)

The three most common methods used in measuring the depth to water in wells are (1) wetted tape, (2) electric tape, and (3) air line. The *wetted-tape method* is probably the most common and most accurate of all methods. This method utilizes a graduated-steel tape with a weight affixed to its end. The graduations on the lower 3 to 4 ft. of the tape are coated with blue-colored carpenter's chalk and the tape is lowered into the well until the lower part of the tape is submerged and an even foot mark is at the measuring point. The tape is then quickly withdrawn and the value held at the measuring point and the length of tape that was submerged are entered on the record form. The length of tape that was submerged is obvious from the change in color of the chalk coating. The difference between the two readings is the depth to water below the measuring point.

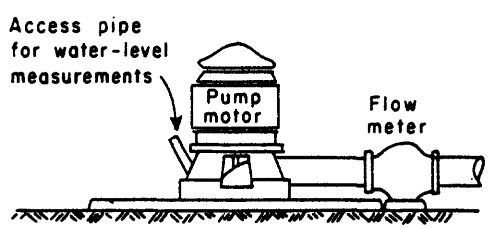
The *electric-tape method* involves an ammeter connected across a pair of insulated wires whose exposed ends are separated by an air gap in an electrode and containing, in the circuit, a source of power such as flashlight batteries. When the electrode contacts the water surface, a current flows through the system circuit as indicated by a deflection of the ammeter needle. The insulated wires are marked at five foot intervals. The nail of the index finger is placed on the insulated wires at the measuring point when the ammeter indicates the circuit is closed. A steel tape or carpenter's rule is used to measure the distance from the point indicated by the fingernail to the next highest five foot mark. This distance is subtracted from the value of the five foot mark to determine the depth to water. One difference between the wetted-tape and the electric-tape method is that in the wetted-tape method the subtraction involves the length of the submerged tape whereas in the electric-tape method the subtraction involves the distance between the measuring point and the next highest mark.

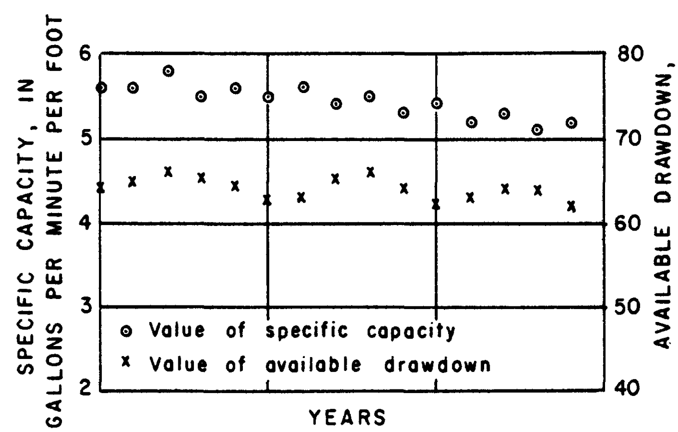
The *air-line method* is generally used only in wells on which pumps are installed. This method involves the installation of a small diameter pipe or tube (the air line) from the top of the well to a point about 10 ft below the lowest anticipated position of the water level during extended pumping periods. The water level in this pipe is the same as in the well. To determine the depth to water an air pump and a pressure gage are attached to the top of the air line. Air is pumped into the line to force the water out of the lower end. As the water level in the air line is depressed, the pressure indicated by the gage increases. When all the water has been forced out of the line, the pressure-gage reading stabilizes and indicates the length of the water column originally in the air line. If we subtract the pressure-gage reading from the length of the air line below the measuring point, which was carefully determined when the air line was installed, the answer is the depth to water below the measuring point.

The *measurement of pumping rate* requires the installation of a flow meter in the pump-discharge line. Either of two types of meters may be used, depending on the pumping rate. Up to a rate of about 200 gal/min an "active-element" type meter may be used. These meters utilize either a propeller or a disk that is turned by the moving water. For pumping rates larger than about 200 gal/min meters that utilize a constriction in the discharge pipe are commonly used. These include venturi meters, flow nozzles, and orifices.

Flow meters have dials that either show the total amount of water that has passed the meter or the rate at which the water is passing. With the former type (the totalizing dial), the rate of discharge is determined by timing with a stopwatch the period required for a certain volume of water to be pumped.

Supply-Well Problems -
Decline in Yield





The yield of any water-supply well depends on three elements; (1) the aquifer, (2) the well, and (3) the pump. A decline in yield is due to a change in one of these elements and correction of the problem depends on identification of the element that is involved. This identification in many cases can be made only if data are available on (1) the depth to the water level in the well, and (2) the pumping rate. Inability to identify reasons for a decline in yield frequently results in discontinuing the use of ground water and the development of more expensive supplies from surface-water sources.

The *depth to the water level* in a well equipped with a pump may be determined by (1) use of a steel tape, (2) use of an electric-tape, or (3) use of an air line and pressure gage. The *pumping rate of a supply well* can be determined with any one of several different types of metering devices. (See MEASUREMENT OF WATER LEVELS AND PUMPING RATES.)

The yield of a well depends on the drawdown and on the specific capacity. The *specific capacity* is the yield per unit of drawdown and, in nearly all pumping wells, it varies with the pumping rate. Therefore, a discussion of decline in yield is meaningful only in terms of the maximum yield. The *maximum yield* of a well is controlled by the available drawdown and the specific capacity when the drawdown in the well equals the available drawdown.

The *available drawdown* is determined at the time of construction of a supply well and consists of the difference between the static (non-pumping) water level and the lowest practical water level. The *lowest practical water level* depends on the type of well. In screened wells it is at the top of the uppermost screen. In open-hole, fractured-rock wells it is at the position of the lowest water-bearing fracture, or at the lowest level at which the pump intake can be placed.

The specific capacity and the "yield" of supply wells is determined at the time of well construction. (See WELL-ACCEPTANCE TESTS.) If the pumping level during the well-acceptance test is relatively close (within several feet) to the lowest practical level, the specific capacity determined during the test can be used to accurately estimate the maximum yield. However, it is important to note that apparent declines in yield after wells are placed in production reflect, in many cases, over estimation of the yields at the time of construction. Actual declines in yield after wells are placed in operation result from declines in the static water level, or the specific capacity, or both.

The yield of a ground-water system is the sum of the yields of the individual wells. Successful operation, therefore, requires periodic measurements of both the specific capacity and the available drawdown for each well. Changes in these values are used to predict the yield of the system at different times in the future and, when used in conjunction with predictions of needs, to plan the construction of new wells.

Measurements of specific capacity and available drawdown are neither difficult nor time consuming. The determination of both requires only the three measurements listed below:

1. *Static (nonpumping) water level* - measured weekly near the end of the longest non-pumping period which, in most systems, is near the end of the weekend.
2. *Maximum pumping water level* - measured weekly near the end of the longest period of continuous use which, in most systems, is near the end of the workweek.
3. *Pumping rate* - measured at the same time as the maximum pumping water level.

The specific capacity is determined by dividing the pumping rate by the difference in the static and pumping water levels. Thus

Specific capacity

$$= \frac{\text{pumping rate (gal/min)}}{\text{static w.1. (ft)-pumping w.1. (ft)}} = \frac{\text{gallons}}{\text{min ft}}$$

Available drawdown (ft)

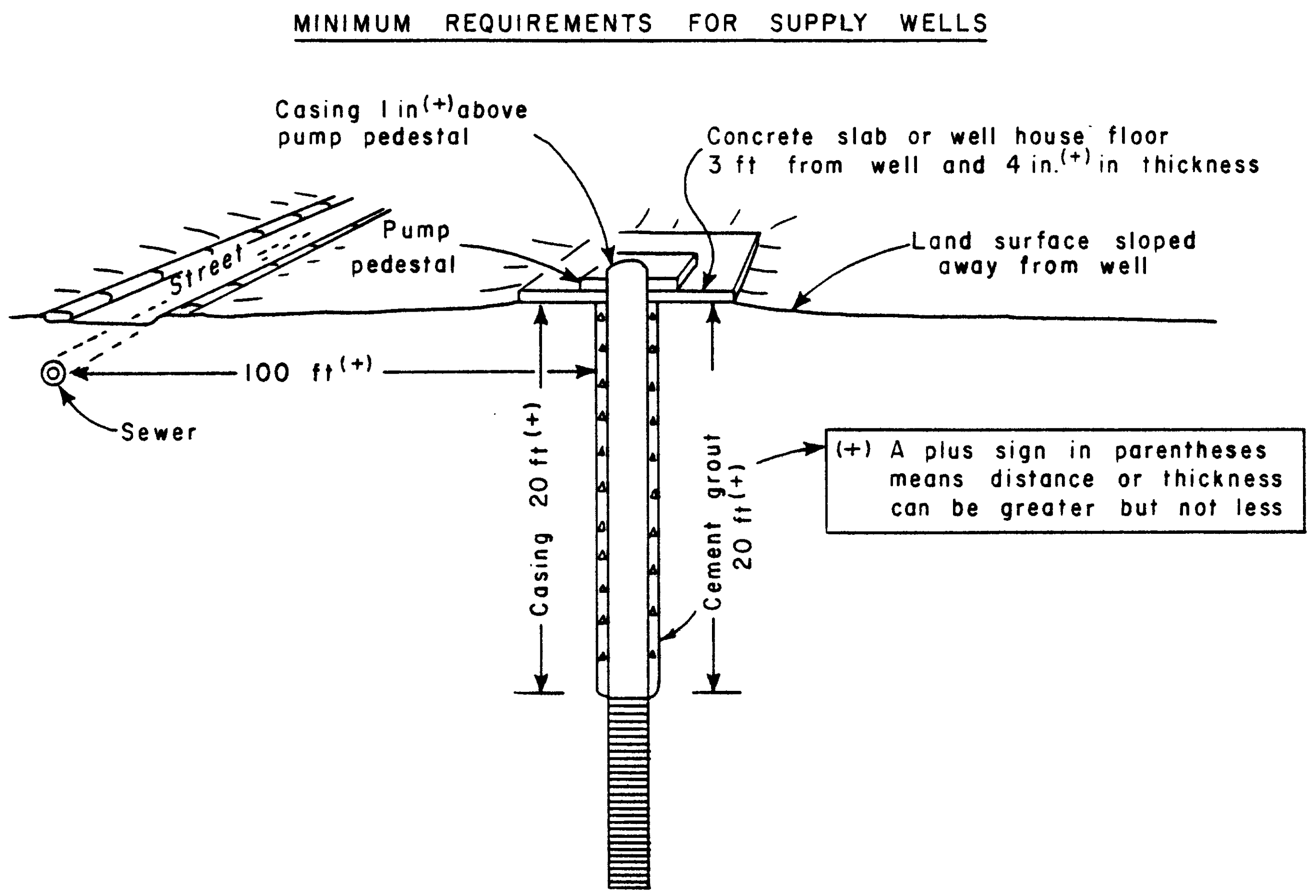
$$= \begin{matrix}\text{(static water} \\ \text{level, ft)}\end{matrix} - \begin{matrix}\text{(lowest practical} \\ \text{water level, ft)}\end{matrix}$$

The determinations of specific capacity and available drawdown should be carefully preserved as a part of the permanent file on each well. (See WELL RECORDS AND FILES.) They should be analyzed at least quarterly to determine if changes in either are occurring. This analysis can be done most conveniently if the values are plotted on graph paper versus the time of the determination. Changes in available drawdown and/or specific capacity and suggested causes and corrective action are listed in the accompanying table.

Analysis of Declines in Well Yield

Identifying Criteria	Cause	Corrective action
Decline in available drawdown No change in specific capacity	*The aquifer* - due to a decline in ground-water level resulting from depletion of storage caused by decline in recharge or excessive withdrawals.	Increase spacing of new supply wells. Institute measures for artificial recharge.
No change in available drawdown Decline in specific capacity	*The well* — due to increase in well loss resulting from a blockage of screen by rock particles or by deposition of carbonate or iron compounds; or reduction in length of the open hole by movement of sediment into the well.	Redevelop the well through the use of a surge block or other means. Use acid to dissolve incrustations.
No change in available drawdown No change in specific capacity	*The pump* — due to wear of impellers and other moving parts, or loss of power from the motor.	Recondition or replace motor or pull pump and replace worn and damaged parts.

Protection of Supply Wells



Laws of North Carolina related to water-supply wells include the Well Construction Act of 1967, administered by the Dept. of Natural Resources and Community Development, and those governing public water supplies, administered by the Dept. of Human Resources.

These laws, and the rules and regulations developed for their administration and enforcement, are concerned, among other things, with the protection of supply wells from pollution. Pollution of the environment results from man's activities and, consequently, except where wells are used for waste disposal, it primarily affects the land surface, soil zone, and upper part of the saturated (ground-water) zone. Therefore, the protection of supply wells includes (1) avoiding areas that are presently polluted, and (2) sealing the wells in such a way as to prevent pollution in the future.

Fortunately, most ground-water pollution at the present time affects only relatively small areas that can be readily avoided in the selection of well sites. Among the areas in which at least shallow ground-water pollution should be expected are:

1. *Industrial districts*-that include chemical, metal-working, petroleum refining, and other industries that involve the use of fluids other than cooling water.
2. *Residentail areas* - in which domestic wastes are disposed of through septic tanks and cesspools.
3. *Animal feed lots* - and other areas in which large numbers of animals are kept in close confinement.
4. *Liquid and solid waste disposal sites* - including sanitary landfills, "evaporation ponds," sewage lagoons, and sites used for the disposal of sewage plant effluent and solid wastes.
5. *Chemical stockpiles* - including those for salt used to deice streets and highways and for other chemical substances soluble in water.

In the selection of a well site, not only should the areas listed above be avoided but also the zone surrounding them that may be polluted by movement of wastes in response to both the natural hydraulic gradient and the artificial gradient developed by the supply well.

Rules and regulations intended to prevent future pollution include (1) provision of "exclusion" zones around supply wells, (2) requirements for casing and for sealing of the annular space, and (3) sealing of the upper end of the wells.

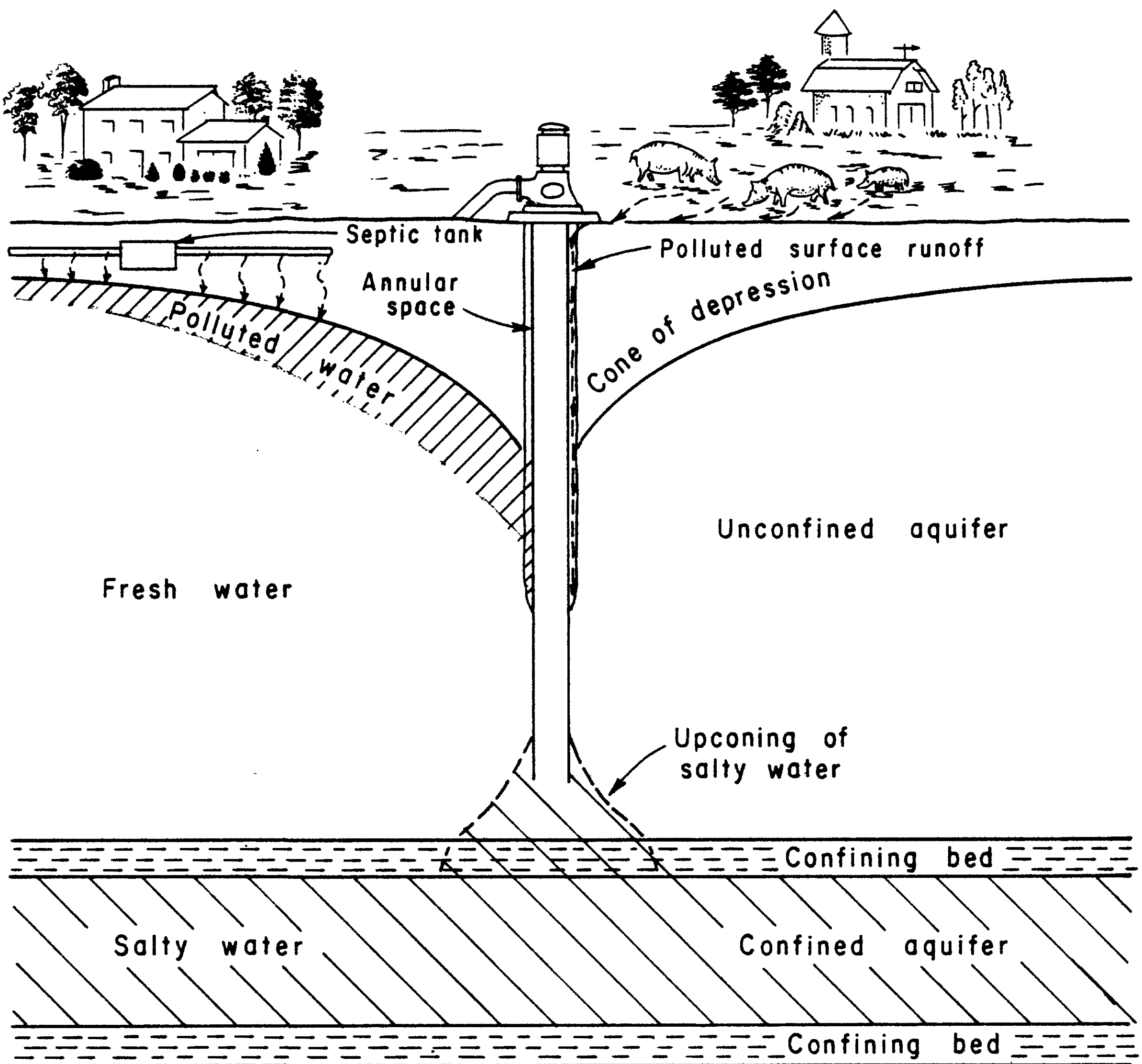
State regulations require that supply wells be located at least 100 ft from any sources or potential sources of pollution. In the case of public supply wells, the well owner must either own or control the land within 100 ft of the well. A public supply well may be located as close as 50 ft to a sewer if the joints in the sewer line meet water-main standards.

State regulations require that all supply wells be cased at least to a depth of 20 ft and that the annular space between the land surface and a depth of 20 ft be completely filled with cement grout. The casing of supply wells drawing water from bedrock must be seated and sealed into the top of the rock.

State regulations require that the casing of all supply wells terminate above land surface and the land surface at the site must be graded or sloped so that surface water is diverted away from the well. Every public supply well must have a continuous bond concrete slab or concrete well house floor at least 4 in. thick and extending at least 3 ft horizontally around the outside of the well casing. The top of the well casing must project not less than 6 in. above the concrete slab or well-house floor. The top of the well casing must also project at least 1 in. above the pump pedestal. The top of the well casing must be sealed water tight except for a vent pipe or vent tube having a downward-diverted, screened opening.

The regulations cited above provide, at best, only minimal protection for supply wells. There are numerous situations in which both the size of the exclusion zone and the depth of casing are inadequate. Relative to the radius of the exclusion zone, there are no arbitrary limits, except the physical boundaries of an aquifer, past which ground water cannot move. Relative to the minimum required casing, there are no vertical limits, except for the impermeable base of the ground-water system, past which polluted water cannot move.

Supply-Well Problems
Changes in Water Quality



The problems most frequently encountered in the operation of supply wells relate either to declines in yield or to deterioration in quality of the water. Declines in yield are discussed in SUPPLY-WELL PROBLEMS - *Decline in Yield.*

Deterioration in water quality may result either from (1) changes in the aquifer or (2) changes in the well. These changes may affect either (1) the biological quality, (2) the chemical quality, or (3) the physical quality. Deterioration in biological and chemical quality generally result from conditions in the aquifer, whereas changes in physical quality result from changes in the well.

Deterioration in biological quality refers to the appearance in the water of bacteria and or viruses associated with human or animal wastes. Such deterioration is referred to under the general term *pollution* and indicates, in nearly all cases, a connection between the land surface or a near surface zone and the open section of the well. The connection most frequently exists in the annular space between the casing and the formation. *To avoid pollution of wells*, the N. C. Well Construction Act of 1967 requires that the annular space be completely filled with cement grout from the land surface to a depth of at least 20 ft.

Analysis of Changes in Water Quality

Change in quality	Cause of the change	Corrective Action
Biological	Movement of polluted water from the surface or near surface layers through the annular space.	Seal annular space with cement grout or other impermeable material and mound dirt around the well to deflect surface runoff.
Chemical	(1) Movement of polluted water into the well from the land surface or from shallow aquifers.	(1) Seal the annular space. If this does not eliminate pollution, extend the casing to a deeper level (by telescoping and grouting a smaller diameter casing inside the original casing).
	(2) Upward movement of water from salty-water zones.	(2) Reduce the pumping rate and/or seal the lower part of the well.
Physical	(1) Migration of rock particles into the well through the screen or from water-bearing fractures penetrated by open-hole wells.	(1) Remove pump and redevelop the well.
	(2) Collapse of the well screen or rupture of the well casing.	(2) Remove screen, if possible, and install new screen. Install smaller diameter casing inside the original casing.

Deterioration in chemical quality refers to the arrival at a supply well of water containing dissolved chemicals in an undesirably large concentration. Withdrawals of water from a well causes water to converge on the well from different directions. If this convergence involves water containing a large concentration of any substance, the concentration of that substance will, after some period of time, begin to increase. The most commonly observed changes in chemical quality involve NaCl (sodium chloride or common salt) and NO_3 (nitrate).

Nitrate is an important constituent in fertilizers and is present in relatively large concentrations in human and animal wastes. Therefore, nitrate concentrations in excess of a few mg/L (milligrams per liter) indicate water is arriving at the well from shallow aquifers that are polluted by septic tanks or animal feed lots or that are contaminated by excess nitrates used in farming operations.

Sodium chloride or, more simply, chloride, is the principal constituent of sea water and is also present in significant concentrations in human and animal wastes and in some industrial wastes. An increase in chloride content in well water most commonly indicates upward movement of water from an underlying salty-water zone. Other increases are due to pollution by sources at or near the land surface.

Deterioration in physical quality most commonly involves either the gradual or the sudden appearance of rock particles in the water. These particles can range in size from clay, which gives the water a turbid or "bluish" appearance, to sand. The size of the particles is indicated by the rate at which the particles settle. If exceedingly slowly, or not at all, the particles are clay-size. If the particles settle immediately, they are sand-size.

The gradual appearance of particles generally indicates the finer-grained material was not adequately removed from the zone adjacent to the well during well development. (See WELL-CONSTRUCTION METHODS.) During the process of well use, these particles slowly migrate to and into the well.

The sudden appearance of particles - that is, the concentration of particles is large (very obvious) from the beginning—generally indicates the failure (collapse) of the screen or a rupture of the well casing.

Well Records and Files

The collection and preservation of records on the construction, operation, maintenance, and abandonment of supply wells is an essential, but largely neglected activity. The consequence of this neglect is that it is not possible to identify and to economically correct problems of declining yield or deterioration in water quality and the design of new wells cannot incorporate past operational experience.

A file should be established on each supply well at the time plans for its construction are initiated. From the initial planning to the final abandonment of the well the following records should be generated and carefully preserved in this file:

1. *Initial design* - including drawings or written specifications on diameter, proposed total depth, position of screens or open hole, method of construction, and materials to be used in construction. (See WATER-WELL DESIGN.)
2. *Construction record* - including the method of construction and the driller's log and a geophysical log of the materials penetrated during construction, the diameter of casings and screens, slot size and metallic composition of screens, depths of casing and screens, total depth of the well, and weight of casing.
3. *Well-acceptance test* - including a copy of the water-level measurements made prior to, during, and after the drawdown (pumping) test, a record of the pumping rate or rates, copies of any graphs of the data, and a copy of the hydrologist's report on the results of the test. (See WELL-ACCEPTANCE TESTS.)
4. *Pump and installation data* - including type of pump, horsepower of motor, depth to the pump intake, and a copy of the pump manufacturer's performance and efficiency data, and data on length of the air line or description of facilities provided for water-level measurements including a description of the measuring point. (See MEASUREMENT OF WATER LEVELS AND PUMPING RATES.)
5. *Operating record* - including data on type of meter used to measure the flow rate, weekly readings of the flow meter dial and weekly measurements of the static and pumping water levels. (See SUPPLY-WELL PROBLEMS - *Decline in Yield.)*
6. *Record of well maintenance* - including the dates and the activities instituted to increase the yield or to improve the water quality, and data showing the results achieved.
7. *Record of well abandonment* - including the date use of the well was discontinued and a description of the methods and materials used to seal or plug the well.

The type of forms used for the records described above are not of critical importance. It is more important that the records be collected, regardless of the type of form that is used. It is important, however, that the date and watch time be noted with each measurement of pumping rate and depth to water.

INTERNATIONAL SYSTEM UNITS

The following factors may be used to convert the U.S. customary units published in this report to the International System of Units. (SI).

Multiply U.S. Customary unit	By	To obtain SI (metric) unit
	Length	
inches (in)	25.4	millimeters (mm)
feet (ft)	.3048	meters (m)
miles (mi)	1.609	kilometers (km)
	Area	
square feet (ft^2)	.0929	square meters (m^2)
square miles (mi^2)	2.590	square kilometers (km^2)
	Volume	
Cubic feet (ft^3)	0.02832	cubic meters (m^3)
	Flow	
cubic feet per second (ft^3/s)	.02832	cubic meters per second (m^3/s)
gallons per minute (gal/min)	0.00379	cubic meters per minute (m^3/min)
	Velocity	
feet per day (ft/day)	.3048	meters per day (m/day)
	Mass	
pounds (lb avoirdupois)	.4536	kilograms (kg)

RELATION OF UNITS OF HYDRAULIC CONDUCTIVITY AND TRANSMISSIVITY

A, Hydraulic conductivity (K)

Feet per day (ft/day)	Meters per day (m/day)	Gallons per day per square foot (gal/day ft^2)
ONE	0.305	7.48
3.28	ONE	24.5
.134	.041	ONE

B, Transmissivity (T)

Square feet per day (ft^2/day)	Square meters per day (m^2/day)	Gallons per day per foot (gal/day ft)
ONE	0.0929	7.48
10.76	ONE	80.5
.134	.0124	ONE