CHAPTER 1

Geologic Formations as Aquifers

1.1 INTRODUCTION

Water is an essential ingredient for all life forms on earth. Although generally abundant, it is not always available where needed, nor may it be of suitable quality. In many arid lands the only reliable supply lies beneath the earth's surface. In wetter areas, ground water may be superior in quality to surface supplies and also provide a reserve against drought. Because ground water is a hidden resource, knowledge of the conditions of its occurrence is essential in designing programs to utilize it efficiently.

Historical Perspective

Archeological evidence shows that ground water has been used since prehistoric times, especially in arid regions. Springs, for example, provided an easily available supply of freshwater, especially valuable during dry periods, and villages developed around them. Extensive references to the use of springs may be found in Genesis, Chapter 26, thought to be written as early as 2000 B.C.

Throughout the Middle East, tens of thousands of horizontal tunnels (*ghanats*) have been utilized for centuries and presently supply as much as 75% of all water used in Iran (1). *Ghanats* follow water table flow lines for distances as much as 20 mi, reaching depths in excess of 250 ft.

In ancient China water from hand-dug wells was raised using bucket and pulley devices. Cased wells were constructed by driving hollow bamboo rods into shallow saturated soils.

Hand-dug wells, probably the oldest method of developing ground water, are still in general use throughout the world. Production is limited as saturated materials can be penetrated only a few feet. Water is extracted from these shallow wells, using human, animal, or wind power.

The earliest methods of ground water development changed little until the advent of the industrial age. Not until the late 1800s, with development of mechanized techniques for drilling and bringing the water to the surface, could deeper, more productive wells be constructed.

Despite the early extensive use of ground water, little was understood of its origin. As a result of incomplete observation, Aristotle, and other ancient Greek and Roman natural philosophers, concluded that infiltration from precipitation was insufficient to produce observed spring flow. They theorized that ground water was produced in the "roots of mountains," either by transformation of an assumed subterranean atmosphere or by conversion of seawater into freshwater in submarine canyons. This freshwater then rose to the surface discharging as springs.

Marcus Vitruvius (17 B.C.) introduced the present-day concept of ground water originating as precipitation and returning to earth in a continuous cycle. This theory gained little acceptance until propounded independently by Leonardo da Vinci and Bernard Palissy in the sixteenth century.

With the understanding of the "hydrologic cycle," the science of hydrology was born. Basic principles of artesian pressure and hydrostatics were formulated in the seventeenth and eighteenth centuries by Johann Becher, Bernardino Ramazzini, Antonio Vallisnieri, and others. Pierre Perrault, Edme Mariott, and Edmund Halley pioneered quantitative hydrology, measuring and correlating river flow, precipitation, and evaporation.

Development of theories of ground water storage and movement had to await the formulation of basic geological precepts. William Smith (1769–1839) was among the first to recognize the ground water storage potential of sedimentary strata.

The year 1856 marks the birth of ground water hydrology as a quantitative science. Henri Darcy, a French hydraulic engineer with the city of Dijon, published the results of a series of experiments on percolation of water through filter sands. Darcy showed that flow through sand is directly proportional to head difference and cross-sectional area. The constant of proportionality in Darcy's relation later became known as hydraulic conductivity, an important parameter in aquifer evaluation. The Darcy equation remains the basic expression describing flow through a confined aquifer. Jules Dupuit in 1863 developed a similar equation for ground water flow in a gravity drainage system (unconfined aquifers) and identified the curvilinear nature of flow near the well. Phillip Forcheimer, from 1886 through 1898, noted the analogy of ground water flow to heat flow and was the first to apply the LaPlace equation to solve problems in steady-state ground water flow.

Since the beginning of the twentieth century, the science of well hydraulics has been advanced by such investigators as Adolph and Gunther Thiem (father and son), Charles Theis, Hilton Cooper, C. E. Jacob, and Mahdi Hantush. They developed methods of determining aquifer characteristics through use of observations from pumping wells. Their methods are discussed in Chapters 5 and 10. Major contributions to ground water flow hydraulics were made by Oscar Meinzer on compressibility and elasticity of artesian aquifers, C. E. Jacob, who developed the general flow equation from fundamental principles, M. King Hubbert, who clarified the concept of hydraulic potential, and Mahdi Hantush, who, along with C. E. Jacob, developed the theory of leaky aquifers and made major contributions to well hydraulics.

Modern ground water investigators utilize computers to simulate ground water flow and contaminant migration so that ground water resources may be safely exploited. R. Allen Freeze, John Bredehoeft, Shlomo Neuman, Paul Witherspoon, Stavros Papadopoulos, and George Pinder are a few of the those who have contributed to further understanding of the ground water environment.

1.2 THE HYDROLOGIC CYCLE

Seventy percent of the surface of the earth is mantled in water. Over 90% of the world supply is found in the ocean, the source of virtually all water found upon or within the continental land masses. The processes that distribute the world's water supply are described collectively as the hydrologic cycle, shown in Figure 1.1.

At the ocean's surface, evaporation (powered by the energy of solar radiation) converts liquid water into water vapor. This vapor, which contains small amounts of dissolved material, is carried by air masses moving across the ocean. Approximately 3.6 million tons of water are evaporated annually from an average square mile of the ocean surface (2).

Moisture-laden air masses rise as they are heated by the sun's radiation. At higher elevations they expand and cool due to the lower pressure in the upper atmosphere. Minute particulate matter, such as combustion products, salt crystals, dust, and clay particles carried aloft by rising air currents, provide surfaces upon which water vapor is converted to liquid water or ice. These particles are referred to as condensation or freezing nuclei. Through this process, clouds composed of water droplets or ice crystals form. When individual droplets gain sufficient mass they accelerate downward by the force of gravity and fall as precipitation.

Interception and Infiltration

Some precipitation approaching the ground is intercepted and held on the surfaces of foliage, buildings, and so forth, by surface tension forces. After these surfaces have become saturated, interception occurs only when water is removed by evaporation or dripping.

Soil absorbs water because it is porous. The maximum rate at which a particular soil can absorb water is known as its "infiltration capacity." Infiltration capacity decreases as pore space become filled with water and ceases when the soil is fully saturated. Infiltration continues as water is removed from the pores by evaporation, lateral movement (interflow), or downward percolation.

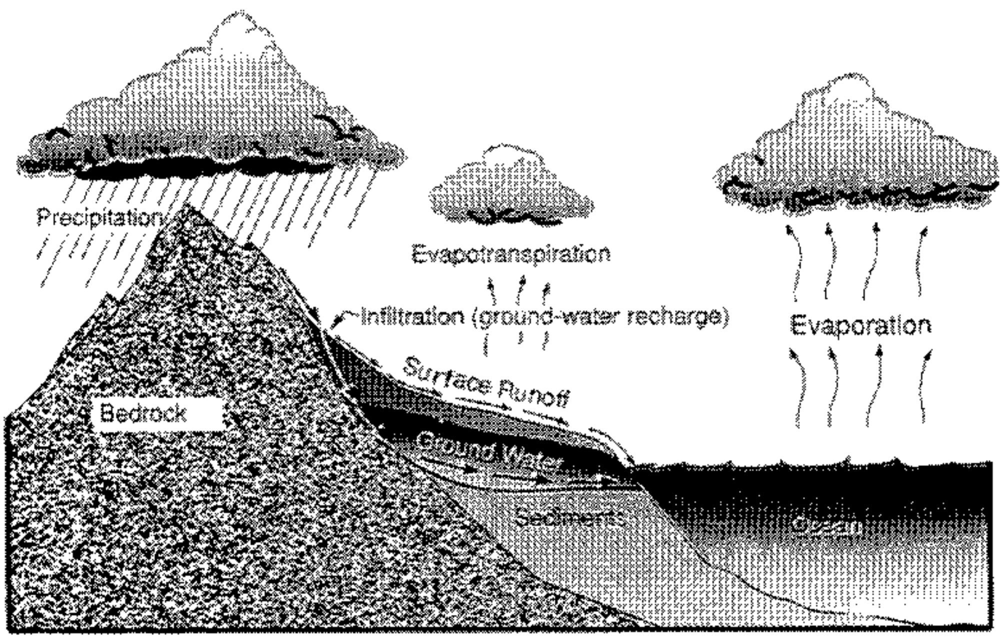

Figure 1.1. The hydrologic cycle.

Runoff, Depression Storage, and Interflow

Water that does not infiltrate into the soil or become stored in depressions flows down-gradient on the surface, initially moving for short distances in no particular channel as "sheet" or "overland flow." However, because flowing water possesses energy, it moves soil particles to form temporary rivulets. Water flowing in these rivulets reaches larger channels that eventually carry the runoff to a large water body (ocean or lake).

Some infiltrated water travels laterally through the shallow soil horizon, discharging where it intercepts a channel. This "interflow" delays the runoff due to the water traveling in longer flow paths around soil grains. Interflow will continue to discharge into channels for many hours after surface runoff has ceased (Figure 1.2). Interflow is distinguished from "base flow," which is the discharge from an aquifer to an adjoining channel with a lower water level.

Evapotranspiration

Depending upon climatic conditions such as temperature, humidity, and wind speed, most precipitation evaporates during and shortly after it occurs. Most of the remaining water percolates to the vegetation root zone and is absorbed through osmosis and discharged or transpired as water vapor through the plant leaves. The processes of evaporation and transpiration are known as "evapotranspiration." Evapotranspiration is also referred to as "consumptive use," because the water returns to the atmosphere as water vapor and is unavailable for further use at the earth's surface.

Recharge and Discharge

Typically, depending upon many factors, only a very small percentage of the infiltrated water percolates deeper. It ranges from possibly 10% to 20% in coarse alluvial deposits to very little in clayey soils. This percolating water moves

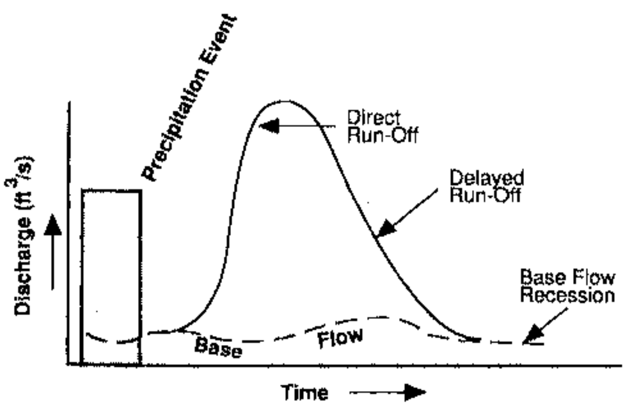

Figure 1.2. Stream hydrograph response to a precipitation event

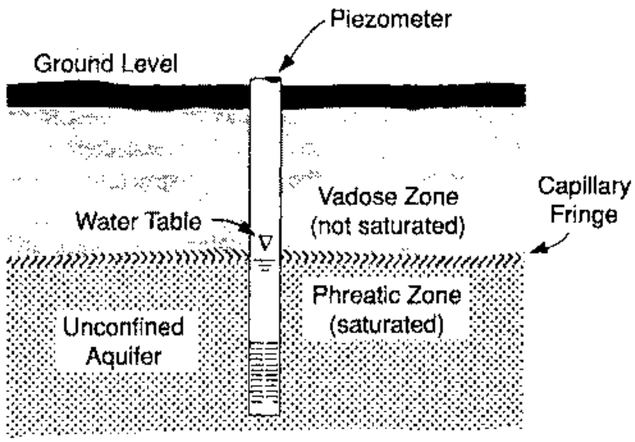

Figure 1.3. Piezometer installed to monitor water table elevation.

downward under the influence of gravity through a region defined as the "vadose zone." In this zone the pores between grains are only partially filled with water, some of which is bound to grains by surface tension and molecular forces.

Unbound water continues downward until it reaches the lower boundary of the vadose zone, known as the "capillary fringe." Here pore spaces are completely filled with water. The thickness of the capillary fringe varies from a few inches to several tens of feet, depending upon the nature of materials forming the zone. Coarse-grained deposits, which have large pore spaces and a low ratio of surface area to volume, have little or no capillary fringe. Materials composed primarily of fine particles have a large surface area to volume ratio and may have capillary fringes of 50 ft or more.

Water molecules are bound to grain surfaces in both the vadose zone and capillary fringe by a process known as surface retention. Here, water within the pore spaces of the capillary fringe is under tension at pressures that are less than atmospheric, and this zone is referred to as the "tension saturated zone" (3). Water held by molecular and surface tension forces does not drain by gravity and is said to be in "dead storage."

If not held in dead storage, the percolating water will eventually reach the zone of saturation, or the "phreatic zone." Water reaching the saturated zone constitutes ground water recharge. Ground water recharge may be natural or artificial, and it replaces ground water removed from storage by pumping or natural discharge such as springs.

A piezometer[1] installed into the saturated zone shows water standing at a level known as the "water table," or phreatic surface (Figure 1.3). The water table is the upper boundary of an unconfined aquifer.

Hydraulic head (potential) consists of both hydrostatic and gravitational potential. Ground water responds to decreasing head (hydraulic gradient) by moving slowly through

1. A small-diameter pipe open or perforated at the bottom.

porous or fractured geologic materials. If not removed by pumping, the ground water may discharge as springs or seeps into channels, eventually reaching a body of surface water. Occasionally, ground water may come to rest in an area with no hydraulic gradient. It will remain there until either tectonic forces modify the geologic structure or it is removed by wells.

A system of branching streams collects drainage and carries it out of a basin. Such streams may flow perennially or intermittently. During and shortly after rainfall events, stream flow is composed primarily of surface runoff. During high flow some stream water infiltrates into stream banks, recharging the ground water. Such a stream is referred to as a "losing stream," and the process is known as "bank storage." After rainfall ends, stream stage diminishes and bank storage is released into the stream channel. The stream is then known as a "gaining stream."

Discharge of ground water into channels occurs whenever the head potential in the aquifer is higher than the elevation of the water surface in the channel, and the discharge may continue uniformly through the year. As shown in Figure 1.2, it constitutes the "base flow" of a stream. This is the component that the aquifer continues to contribute to stream flow after direct and delayed runoffs have ceased. At the low stream stage, the major component of flow is ground water discharge. Water flowing from an aquifer into a stream is known as "effluent seepage," and when flowing from the stream to an aquifer as "influent seepage."

1.3 GROUND WATER IN STORAGE

In the uppermost portion of the earth's crust, large volumes of porous and fractured materials store ground water. Although only a small fraction of the precipitation from any single storm may reach the saturated zone, the process has continued through geologic time, resulting in large reserves of subsurface water. Ground water storage represents the largest volume of freshwater available for consumption, as seen in the distribution of the earth's water resources (Table 1.1).

TABLE 1.1 Estimate of World Water Inventory

	Volume ($km^3 \times 10^6$)	Volume (%)
Oceans and seas	1370	93.77
Lakes, reservoirs, swamps, river channels	0.18	0.01
Soil moisture	0.07	0.01
Ground water	60	4
Icecaps and glaciers	30	2
Atmospheric water	0.01	<0.01
Biospheric water	0.01	<0.01

Source: Nace (4).

1.4 CONNATE WATER

Although virtually all ground water has been derived from geologically recent precipitation, a small amount may be the result of ancient events. Water trapped in sedimentary strata as they were formed is occasionally encountered. This water, known as "connate water," is normally highly mineralized because of the conditions during its deposition and its long residence in the formation. Connate water may discharge along fractures or faults, due to recent earth movement, or be extracted by wells penetrating these formations.

1.5 AQUIFERS

"Aquifers" are a formation or group of saturated geologic formations capable of storing and yielding freshwater in usable quantities. They are the target for all ground water exploration and development programs. The term aquifer can be ambiguous. In addition to its geological definition, it has an economic connotation. The same formation may be considered an aquifer in one location but not at another where an alternate formation can produce ground water at lower cost.

Saturated formations that store and yield little ground water are known as "aquicludes." This term has been largely supplanted by the term "aquitard." Some geohydrologists restrict the use of "aquitard" to formations whose yield capacities are significantly less than adjacent formations. Although an aquitard may not yield water economically, it can hold appreciable amounts of water.

Aquifer Characteristics

Good aquifers are characterized by relatively large values of storage, yield, and replenishment. The major determinant of aquifer storage is the amount of usable void space, or "effective porosity"; the major determinant of yield is the rate at which the formation allows ground water movement. Replenishment, also known as recharge, depends upon a number of factors such as precipitation and soil characteristics.

1.6 POROSITY

Total Porosity

The storage available in an aquifer is related to the void space that it contains (total porosity). Total porosity as a percentage is expressed as

$$n = \left(\frac{V_v}{V_T}\right)100 ,$$

TABLE 1.2 Soil Classification, Total Porosities, and Effective Porosities of Well-sorted, Unconsolidated Formations

Material	Diameter (mm)	Total Porosity (%)	Effective Porosity (%)
Gravel			
Coarse	64.0–16.0	28	23
Medium	16.0–8.0	32	24
Fine	8.0–2.0	34	25
Sand			
Coarse	2.0–0.5	39	27
Medium	0.5–0.25	39	28
Fine	0.25–0.162	43	23
Silt	0.062–0.004	46	8
Clay	<0.004	42	3

Source: Morris and Johnson (5).

where

n = total porosity [%],
V_v = volume of void space in sample,
V_T = total volume of sample.

Unless individual voids are interconnected, they do not allow transmission of water. Some pyroclastic rocks such as pumice are an example of nonconnected pore spaces.

Effective Porosity

The portion of total interconnected pore space through which flow of water takes place is called the "effective porosity." Effective porosity is always less than total porosity, since not all interconnected pore space in a saturated material is available for flow. A portion of the voids is filled with water held in place against gravity by molecular and surface tension forces. The percentage of water that drains by gravity from a unit volume of material is the "drainable or active porosity," sometimes known as specific yield. Total porosity is therefore the sum of specific yield and surface retention. Table 1.2 lists values for total porosity and effective porosity for various materials.

Surface Retention. Individual water molecules are formed by an oxygen ion bonded between two hydrogen ions with a bond angle of about 105°. This imparts an unbalanced electrical charge to the molecule so that it is polarized with the oxygen ions electrically negative relative to the hydrogen ions. Porous media grains also bear unbalanced electrical charges on their surfaces. Attractive electrical forces, operating over very short distances strongly bind a layer of ground water a few molecules thick to grain surfaces. Because of this process of "surface retention," ground water firmly bound to grain surfaces cannot be removed by gravity. Surface retention is responsible for creating the capillary fringe.

The ratio of surface area to mass increases as particle size decreases. For a material of uniform specific gravity, mass and volume vary directly. The term "specific surface" is the ratio of surface area to volume of uniform spheres. Representative values of specific surface for porous media components are shown in Table 1.3.

Porosity in Aquifers

In aquifers there are three major types of porosity:

1. Spaces between grains in unconsolidated to moderately consolidated media.
2. Fractures in dense rocks.
3. Through-going tubes.

Porous media formations result from the deposition of individual grains eroded from rocks and transported by wind, water, or gravity. The void spaces are formed as the formation is deposited ("primary porosity").

Long after the rock is formed, fractures in rocks result from processes such as structural deformation and erosion ("secondary porosity").

Through-going tubes may represent primary porosity, such as lava tubes formed when a thread of molten lava is extruded from a solidified outer chamber. Tubes may also represent secondary porosity; for example, cavities formed

TABLE 1.3 Relationship of Surface Area per Unit Volume to Particle Size

Particle Diameter (mm)	Classification	Number of Particles	Surface Area per Volume (cm²)	Total Surface Area per cm³ (mm²)
10.	Medium gravel	1	3.14	3.14
1	Coarse sand	10^2	3.14×10^{-2}	31.4
0.1	Fine sand	10^6	3.14×10^{-4}	3.14×10^2
0.01	Silt	10^9	3.14×10^{-6}	3.14×10^3
0.001	Clay	10^{12}	3.14×10^{-8}	3.14×10^4

Sources: Todd (1) and Morris and Johnson (5).

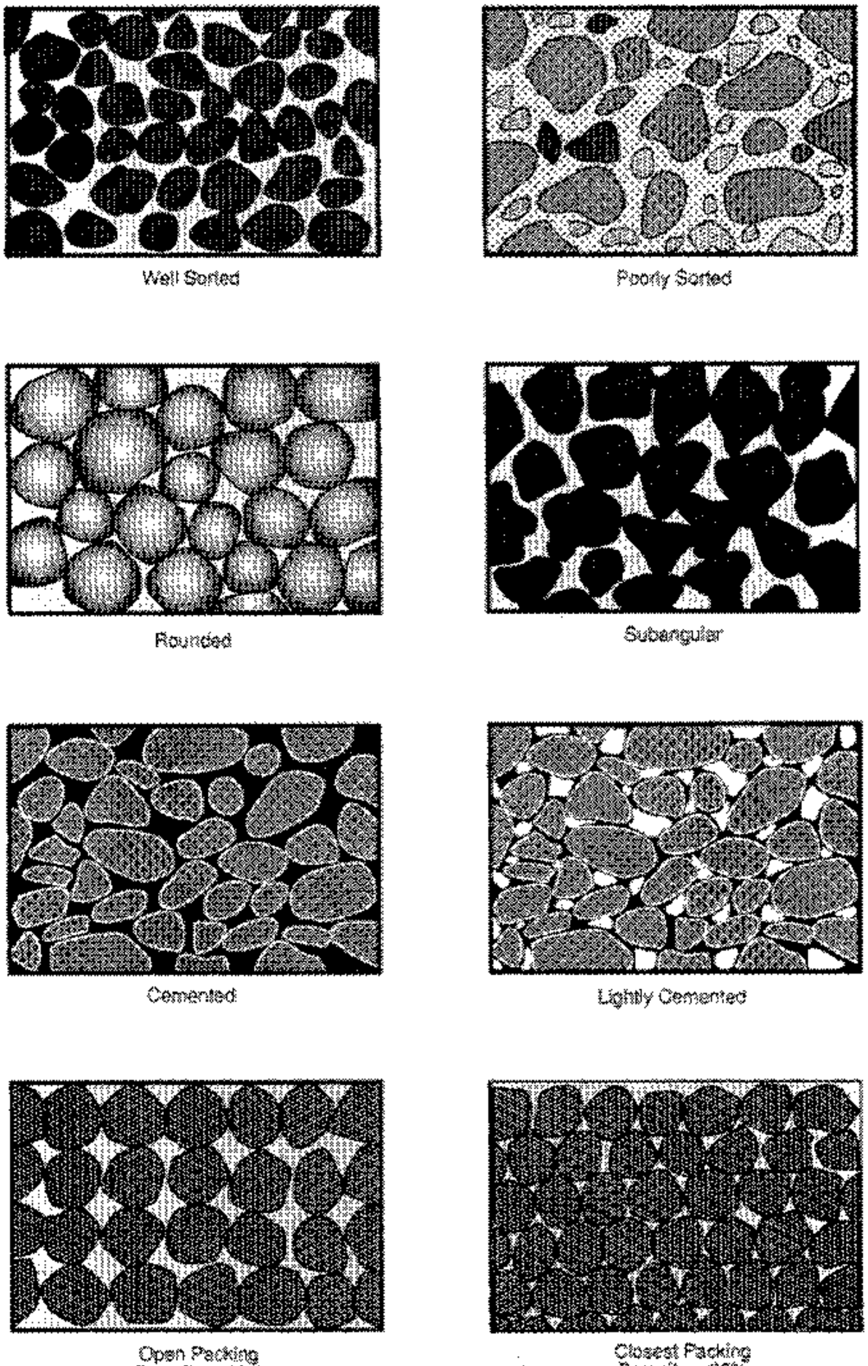

Figure 1.4. Porosity variations.

in limestone from the solution of carbonates by ground water.

Porous Media. The factors controlling the amount of void space in a porous medium vary according to its type. In unconsolidated formations the primary determinants are shape, sorting, and packing of the grains (Figure 1.4). In a consolidated (or semiconsolidated) formation of porous media, an additional factor is the extent to which pore spaces have become filled with cementing material. Formations of geologically older[2] porous media are usually cemented to

2. Older than Pleistocene or late Pliocene, or about a million to a few million years.

some degree. The cementing material may be calcium carbonate, silica, iron oxide, or sometimes clay. Since cementing greatly reduces porosity, older formations usually provide less ground water storage than younger ones.

Grain shape is measured in terms of sphericity and roundness. Sphericity is a three-dimensional aspect and describes how closely a particle approaches the shape of a sphere. Roundness is a two-dimensional aspect and relates to the presence or absence of angular corners. A cube has high sphericity and low roundness; a coin has low sphericity and high roundness.

Quartz and feldspar grains, which predominate in sand-size porous media, have high sphericity. Their roundness depends largely upon the distance they were transported prior to deposition; the longer the distance, the more corner abrasion and the higher the roundness. Clay particles, which predominate in shales, are platelike and have low sphericity.

"Grain packing" is the term describing the arrangement of individual grains. Spheres have the lowest ratio of surface to volume, whereas a circle has the lowest ratio of circumference to area. The most open type of packing of uniform spheres will result in a porosity of 48%, whereas the closest packing (rhombohedral) exhibits a porosity of 26%. Therefore, depending on packing, an aquifer composed of uniform-sized, spherical, well-rounded grains will have higher porosity than one composed of angular or flat particles.

Geologists refer to a porous medium whose grains are uniform in size as "well sorted." They refer to a porous medium exhibiting a wide range of grain sizes as "poorly sorted." If grain size varies, the smaller grains will fit into the pore spaces between larger grains, significantly reducing porosity. Paradoxically, in engineering practice a sample containing a wide range of grain sizes is termed "well graded."

Very fine-grained porous media may exhibit high porosity. However, because pore spaces are very tiny, specific surface is large relative to volume. Surface retention and frictional forces bind virtually all pore water, and ground water cannot move freely through the formation.

Clay particles are the smallest grains found in porous media formations. By definition, individual grains are less than 0.00015 in. in diameter. Clays have high porosity but, because of their minute size, very low permeability. The ratio of surface area to volume is very high, and pore spaces between grains are small. Dry clay will quickly absorb water (by capillary action) that binds to the grain surfaces, and the clay-water system becomes impermeable. Clays, which have a total porosity as high as 60% to 70%, surprisingly represent the most effective geological aquitards.

Fractured Rocks. Fracture and tubal porosities are characteristic of rocks sufficiently hard to maintain these openings. With few exceptions fractures occur long after the rock has been formed. In fractured media the determinants of porosity are the number and width of fractures per unit volume of rock, the degree to which they are interconnected, and the extent to which they may be filled with eroded or injected materials (Figure 1.5).

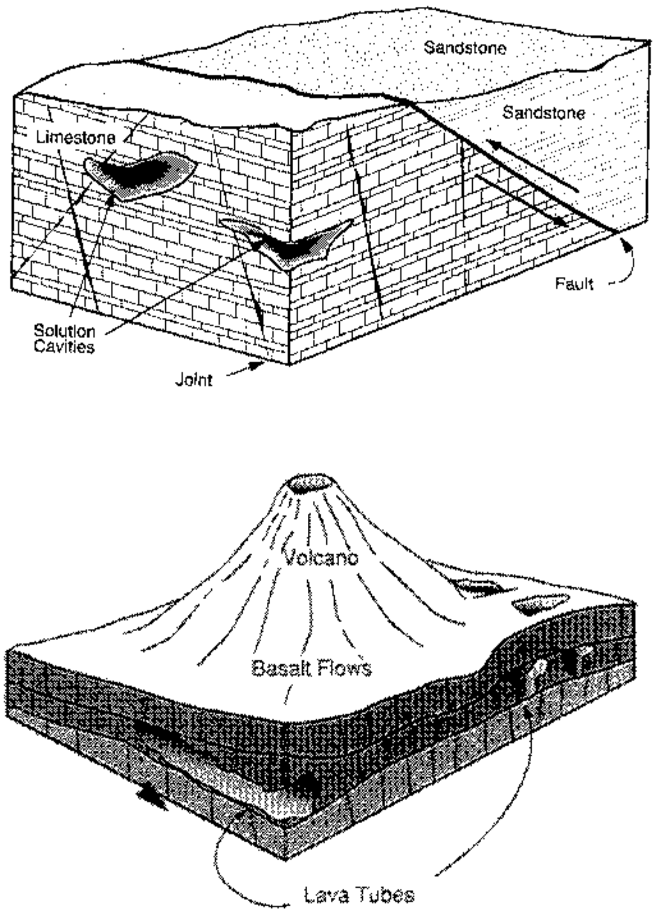

Figure 1.5. Fractures and tubes in rocks.

Fracture and tubal porosities tend to decline rapidly with increasing depth. As the pressure of the overburden increases, fractures close and tubes collapse. Nevertheless, in competent rock of low primary porosity, fractures may operate as conduits for ground water movement at great depth.

Permeability

Permeability is an intrinsic geometric property of the formation and is related to the ease with which a fluid can move through it. It is defined by the size and connections of the voids in the porous medium. For sand and silts, it is proportional to the square of material grain size.

1.7 HETEROGENEITY AND ANISOTROPY

An aquifer usually varies spatially in grain size and porosity due to different modes of deposition and stresses imparted after formation. Permeability therefore may differ significantly within an aquifer. Such a condition is referred to as "heterogeneity." If an aquifer shows little spatial variation in permeability, it is "homogeneous."

The term "isotropy" defines a condition in which ground water is able to move in any direction within the aquifer with equal ease. If the water may move with greater ease in one direction than another, the aquifer is "anisotropic." Subangular or subrounded grains are often deposited in a preferred orientation, having been transported by a moving fluid. This will cause horizontal anisotropy. A porous formation may be deposited over geologic time under varying conditions, resulting in a layered formation with permeability differing from layer to layer. Even if each layer is isotropic, the formation as a whole will be anisotropic. Some degree of heterogeneity and anisotropy characterize all aquifers.

Compressibility and Elasticity

All aquifers are elastic. Under an imposed stress of sufficient magnitude, they compress. If the stress is removed, they tend to return to their original volume. The modulus of elasticity expresses the elasticity of a material as the ratio of change in stress to change in strain. The ratio of the change in volume to change in stress is compressibility, the inverse of the modulus of elasticity.

A porous medium skeleton may compress through two mechanisms: the individual grains may decrease in volume, or the grains may be rearranged to a closer packing. Since individual grains are relatively incompressible, the result of an imposed stress is usually closer packing accompanied by a decrease of porosity. In a fractured rock aquifer, porosity decreases under stress through closure of fractures. In tubal aquifers, an imposed stress may cause collapse.

In many aquifers the response to a decrease in stress is not equal to the response to increase in stress. If an aquifer has compressed, it may not return to its original volume when stress is removed. For clays, the ratio of expansibility to compressibility is on the order of 1:10 (1). Extensive removal of ground water may lead to subsidence, formation of sinkholes, and structural damage to wells and surface installations.

1.8 FRACTURES: JOINTS AND FAULTS

All hard rocks contain fractures caused by tectonic or erosional processes after the rocks were formed. Joints are the most common fractures. They are simply cracks caused by the parting of rock masses. The only motion is perpendicular to the fracture plane. Faults are fractures along which the rocks have moved parallel to the fracture plane. Joints vary in size from submicroscopic to hundreds of feet in length; faults may range up to hundreds of miles.

The capacity of joints and faults to store and transmit water is highly variable and depends upon the type of stress, nature of the rock, depth of burial, and other factors. Rocks in which interconnected joints are only a short distance apart make reasonably good aquifers. Lava flows, in which joints are formed by extensive shattering of solidified lava from movement of underlying molten lava, may make excellent aquifers. This is especially true of recent lavas in which the joints have not yet collapsed or been filled with weathered material.

Joints are usually not random directionally but tend to occur in preferred orientations. This imparts anisotropy. They are more open on or close to the surface but become tightly closed at depth by the overburden pressure. Surface or near-surface open joints may become filled with products of weathering, restricting the movement of ground water. Consequently the interval at which jointed rock is effective in transmitting ground water generally occurs from a few tens of feet to a few thousand feet below the surface.

Faults, as shown in Figure 1.6, although similar in some respects to joint-type fractures, are less numerous and can extend for greater distances. Blocks of the earth's crust move laterally against each other, shattering rocks on each side of the fault. The wider the zone of shattering in consolidated formations, the more effectively the fault transmits ground water.

The intense grinding along a fault plane produces a fine-grained product called "gouge." This is often composed of clay-sized particles and is very impermeable. Therefore a fault is a permeable zone containing a central axis of impermeable gouge. The zone can vary in width from a few

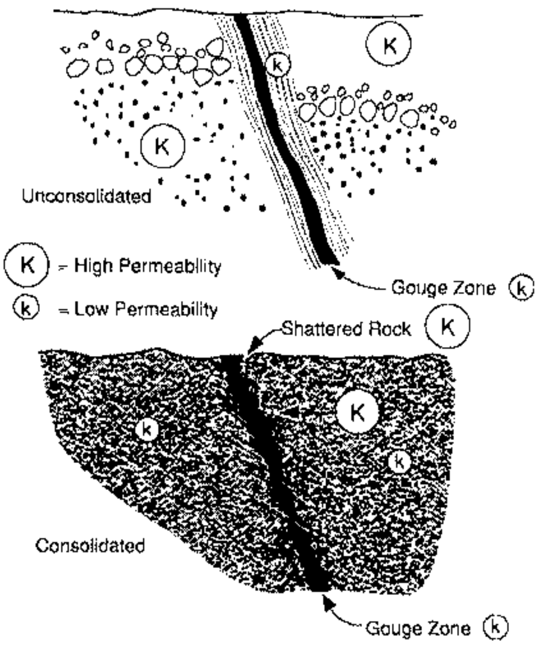

Figure 1.6. Effects of faulting in unconsolidated and consolidated formations.

TABLE 1.4 Representative Properties of Geologic Materials

Formation	Total Porosity (%)	Intrinsic Permeability (in.^{2})	Vertical Compressibility (ft^2/lb)
Permeable basalt	17	10^{-8}	
Fractured igneous and metamorphic rocks	5	10^{-9}	10^{-7}–10^{-9}
Limestones	30	10^{-9}	
Sandstones	35	10^{-9}	
Unfractured igneous and metamorphic rocks	0–10	10^{-10}	10^{-8}–10^{-10}
Shale	6	10^{-13}	
Unweathered clay	42	10^{-12}	10^{-5}–10^{-7}
Glacial till	6–34	10^{-9}	
Silt, loess	49	10^{-8}	10^{-6}–10^{-8}
Dune sand	45	10^{-8}	10^{-6}–10^{-8}
Clean sand	39–43	10^{-8}	10^{-6}–10^{-8}
Gravel	28–34	10^{-7}	10^{-7}–10^{-9}

feet to a few thousand feet in the case of the great plate boundary faults.

Although faults may operate as conduits for ground water flow in hard rocks, in unconsolidated formations they may act as barriers. Springs, oases, and marshes may occur on the up-gradient side of a fault that operates as a barrier. Although unconsolidated deposits cannot hold open fractures, they may be displaced on opposite sides of a fault, abutting a permeable formation against an impervious one.

1.9 DESCRIPTIVE PROPERTIES OF AQUIFERS

An aquifer as a water-bearing formation can be described by three basic physical properties: porosity, permeability, and compressibility. Table 1.4 lists representative values for each property for common formations. It must be remembered that properties vary considerably within any given formation. However, the table is useful in comparing values among formations.

Aquifer Conditions

There are two major types of aquifers: confined and unconfined (Figure 1.7). Unconfined aquifers receive recharge directly from the overlying surface through percolation from precipitation or surface water. They usually exhibit a shallow water level. At the water table, water is at atmospheric pressure. Potential aquifer material extends to the ground surface, and the water table rises or falls in response to

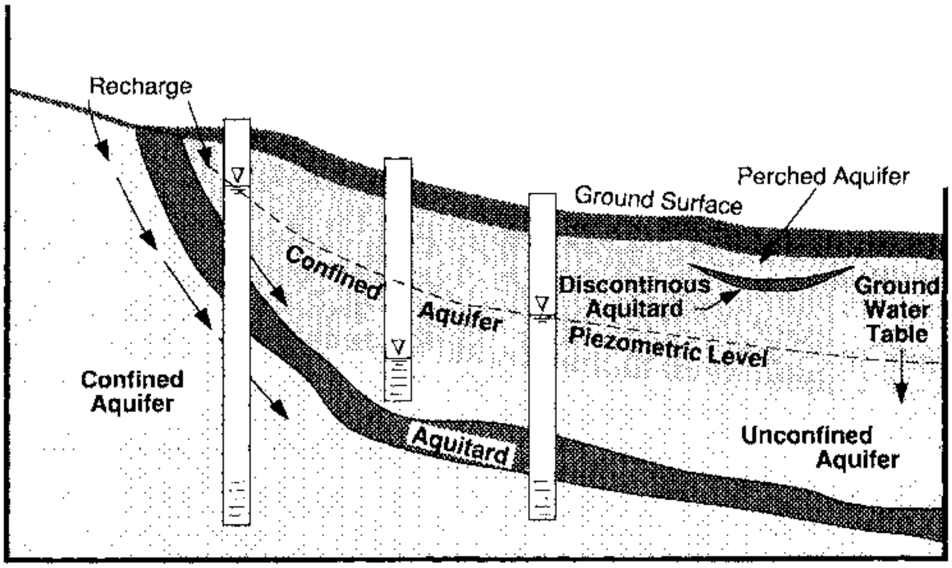

Figure 1.7. Types of aquifers.

changing patterns of precipitation, recharge, and discharge. When a well in an unconfined aquifer is pumped, the water level is lowered and gravity causes water to flow to the well. Formation materials in proximity to the well above the pumping water level are dewatered.

A confined aquifer is bounded both above and below by aquitards. It does not receive significant amounts of percolation from the overlying surface. The ground water within a confined aquifer is under a pressure that includes the sum of the weight of the atmosphere and the overburden. When a well penetrates a confined aquifer, the water level rises above the upper boundary of the aquifer. Such wells are artesian. In some instances the water may rise to the surface and flow (flowing artesian well).

Most confined aquifers are unconfined at their exposed edges and are recharged from percolation into this unconfined portion. Confined aquifers may also receive recharge through their bounding aquitards (semiconfined), especially when pressure changes are induced through pumping or injection. Recharge to an aquifer from or through an aquitard is termed "leakage." An aquifer system in which there is ground water communication through an aquitard is known as a "leaky aquifer system."

Because the water level in an artesian well stands above the upper boundary of the aquifer, pumping does not dewater the aquifer (unless the head is drawn below the top of the aquifer) but will reduce its pressure. This reduction increases the overburden load on the aquifer skeleton, resulting in compression of the aquifer.

A "perched aquifer" results when a lower permeability layer impedes the downward movement of water above it. Perched aquifers may occur above a confined or unconfined aquifer. They are usually very limited areally and may not receive sufficient recharge to support significant well production.

1.10 APPLICATION TO GEOLOGIC FORMATIONS

A formation can be assessed for its value as an aquifer by considering its storage, transmitting, and recharge potentials. Knowledge of the aquifer characteristics of porous media and other formations is of great importance in the study of ground water and its exploitation.

Porous Medium Aquifers

Porous medium aquifers are the most important sources of ground water due to their availability, accessibility, and productivity. Since most wells are drilled in porous media, their aquifer properties are best understood. The classic theories of ground water hydrology have been developed from observations and experimentation on porous media.

Weathering, tectonic, and other processes continually loosen rock. Fragments not carried aloft by winds are moved down-gradient by water or ice or occasionally gravitational force. Sediment deposited on land by running water is called "alluvium"; that deposited by wind is "aeolian."

Alluvium is by far the most common of the nonmarine porous media. Where channels are steep, flowing water has great energy and carrying capacity. Rock fragments collide with each other and are tumbled along the channel bottom, becoming more rounded and spherical through abrasion. Deposition of increasingly finer materials occurs as the gradient diminishes.

Stream velocity depends directly on flow volume. Variation of stage causes deposition of a succession of layers, each of which is well sorted, but with great variation of grain size from layer to layer. Individual layers range in thickness from a few inches to many feet.

Buried river sands and gravels are very prolific porous media aquifers. The layered deposits do not extend very far laterally from the original course of the river that deposited them. The extent of river deposits beyond the channel is the flood plain, deposited by the river at a stage high enough to overflow its banks. Flood plain materials are coarsest in the channel, becoming increasingly finer with distance from it. However, rivers that flow across softer materials tend to change course frequently because they erode their banks, resulting in more extensive channel and flood plain deposits. Because of the meandering and changing nature of the river channel through geologic time, it is often not possible to correlate formation samples among boreholes in alluvium, even when close together.

Buried river deposits have the shapes of valleys, long in the up-and-down stream direction and narrow across. They are trough-shaped at the bottom, relatively flat at the surface and pinched out at the edges. Coarse gravels are often found as the basal deposit.

It is common to find fine-grained layers (silts and clays) interbedded with coarser deposits, especially in the downstream area where the fine-grained materials originated as swamp, pond, or delta deposits. If they are extensive laterally, they may function as aquitards, confining some aquifers. River-deposited sediments therefore generally exhibit heterogeneity and anisotropy.

Aquifers deposited by a constant flowing river are in contact with and receive recharge from it. Although storage volume may be limited laterally, it is continuously replenished. River-deposited aquifers in arid regions often are not in direct contact with a constant flow, and recharge opportunities may be limited.

Alluvial Fan Sands and Gravels

In arid and semiarid regions, eroded rock is washed from the mountains during infrequent but intense precipitation events. The coarser materials are deposited against the mountains, with the finer grained extending outward. As erosion occurs within the range and deposition occurs beyond the range front, the gradient lessens. Succeeding flows swing back and forth creating a fan-shaped deposit below the range front, with the handle extending up into a canyon.

Streams that deposit alluvial fans have short total reaches and extremely variable flow. They are dry except during and after storms. As the fans build up they cover the retreating range front. Layers closer to the mountain front consequently can be expected to become coarser with increasing depth. During major storms slurries of clay, silt, and water may transport large rocks, many feet in diameter. Termed "debris flows," these are poorly sorted, tight, and may function as aquitards.

The coarse-grained upper edges of the fan are in an excellent position to receive orographically controlled precipitation, while the fine-grained deposits of clays and precipitated salts at the lower edge retard ground water outward flow. The ground water is usually unconfined near the mountain front and confined down-gradient. As a result alluvial fans operate as vast and efficient storage reservoirs in arid regions.

Alluvial fan deposits are heterogeneous, anisotropic, and composed of subrounded to subangular grains of a wide range of sizes. Wells may penetrate a number of aquifers separated by aquitards. Correlation of static water levels and lithologies from well to well may be difficult because of variation of deposition over short distances. It may be difficult to determine when a borehole has entered bedrock because large boulders may be buried within the fan.

Dune Sands

Dune sands are aeolian deposits that form in arid regions where topographic conditions cause a rapid decrease in the velocity of sediment-carrying winds. They consist of well-sorted, well-rounded, spherical sand grains and have high porosity and permeability. They are, however, limited in areal extent. Their high permeability prevents significant water storage since, as surface deposits, they quickly drain.

1.11 GLACIAL DEPOSITS

Glacial deposits formed during the Pleistocene Epoch are common in northern temperate regions throughout a wide range of elevations and at moderate-to-high elevations in the subtropics and tropics. These relatively young deposits are still unconsolidated, and some of them function as porous media aquifers. Ancient pre-Pleistocene glacial deposits are consolidated. If they provide water, they do so as bedrock-type aquifers.

Glacial deposits include all material deposited directly by ice or by meltwater streams derived from ice. The term "glacial drift" refers to all types of glacial deposits, regardless of manner of deposition. The materials deposited directly by ice are called "glacial till." Ridgelike or sheetlike deposits of glacial till are termed "moraines." Sediments deposited by glacial meltwater are referred to as "glacial outwash," "glacifluvial deposits," or similar terms.

Glacial drift is widespread through large regions of North America and Europe. They were deposited by continental ice sheets centered over southeastern Canada and central Scandinavia. From these centers the ice extended radially, reaching southern New England and the central plains states in North America, and covering England and the northern European countries. As these continental glaciers expanded and contracted several times during the Pleistocene Epoch, they deposited vast sheets of rock material. Concurrently, alpine or valley glaciers formed in mountainous regions, following and deepening river valleys already in existence.

Glacial drift may consist of alternating layers of till and outwash, forming a series of aquitards and aquifers. Where several ages of glaciation have occurred, the intervals between episodes may have been sufficient to allow deep weathering of the till. The result is "gumbotil," a thick clay-rich layer of low permeability.

Layered silts and clays of low permeability are often found in glaciated areas. These deposits were created in the abundant lakes formed where water dammed against moraines or in hollows gouged out by ice. Glacial lake deposits are identical to those formed in nonglacial lakes.

Till

Moraines are till deposits that occur laterally along the valley-wall sides of valley glaciers. Terminal moraines were formed at the snouts of glaciers during times of little advance or retreat. In moraines where fine materials have not formed a significant part of the original deposit, or have been washed out by meltwater, local aquifers are created. In general, however, glacial till forms aquitards. The till is less permeable than outwash because it is poorly sorted and rich in silt and clay formed by the grinding of ice-borne sediment against bedrock. Examples are the ground moraines formed as discontinuous sheets of till left in the wake of retreating glaciers or plastered on underlying formations during ice advances.

Outwash

Outwash deposits may form good aquifers. Since they were deposited by running water, they have the hydraulic characteristics of stream sediments. Coarse texture is common because the meltwater from glaciers is abundant, and the climate that supports glacial development is usually wetter than in nonglacial periods. Because deposition is by running water, outwash deposits are well sorted and stratified.

The outwash deposits that spread from the edges of continental glaciers tend to be broad; those that were derived from valley glaciers are long and narrow, resembling river deposits. Outwash deposits are usually coarser grained than stream deposits because glaciers provide large volumes of water continuously.

"Eskers" are a special type of outwash formed by the filling of a crevasse or tunnel on or within the ice. Because they were deposited by high-velocity meltwater, they are generally very coarse grained, well sorted, and highly permeable. Eskers are shaped like railroad embankments and are distinctly linear. Most are several tens of feet wide, sinuous, and can extend for miles. They may not store significant amounts of water because of their high permeability, but if encased within less permeable deposits, they form excellent local aquifers.

"Loess" deposits are blankets of wind-blown silt associated with continental glaciation. They were formed outside the glaciated areas or in areas recently uncovered by retreating glaciers. Strong winds, abundant fine materials, and lack of vegetation to stabilize these materials have promoted the deposition of loess in vast regions. The deposits may be 50 to 100 feet thick, and they tend to stand in near-vertical bluffs when eroded. Although not good aquifers because of their fine-grained nature, they have formed rich soils throughout much of the American midwest and elsewhere.

1.12 BEDROCK AQUIFERS

Because most rock contains some water, the recognition and designation of bedrock as an aquifer is as much a local economic as a geohydrological decision. The term bedrock is relative. It usually refers to hard formations such as granite. In areas where hard formations do not occur, the term may be applied to weakly consolidated, early Pleistocene sediments when they are overlain by late Pleistocene or Holocene unconsolidated formations.

The competence or degree of consolidation of the bedrock controls the manner in which it stores and transmits water. Weakly consolidated rocks cannot support open fractures. Their value as an aquifer depends upon how closely they resemble porous media. As the degree of consolidation increases, permeability decreases. Fractured and cavernous rocks usually form aquifers only at relatively shallow depths, perhaps less than 2000 ft because, with increasing overburden pressure, fractures close and tunnels collapse.

The processes describing ground water flow in fractured rock have not been well studied until recently. It is usually assumed that if fractures are abundant, open, and interconnected, analytic methods developed for porous media can be applied. The assumption is also made that the aquifer is highly heterogeneous and anisotropic.

Sandstone, Siltstone, and Shale

Sandstone, siltstone, and shale are formed by the cementation of unconsolidated sands, silts, and clays. These porous media deposits have been compressed through burial and impregnated with cementing material.

Semiconsolidated sandstones may have sufficient primary porosity to transmit water as a porous media and may also be sufficiently competent to support open fractures. Sandstone aquifers are widespread, persist for long distances, and are the principle aquifers for many regions from the Colorado Plateau to Appalachia. Siltstone and shales rarely form good aquifers, even when extensively fractured.

Because deposits may grade into other deposits, aquifers may change from confined to unconfined over very short distances and stratigraphy may be complex. Siltstone and shales will function as aquitards in a sequence of lithified sedimentary deposits. Folding of lithified deposits may cause extensive fracturing along the axis of the folds, whereas faulting may offset beds, causing barriers to, or conduits for, ground water flow.

Carbonates, Limestone, and Dolomite

Much limestone and dolomite are chemically and biologically precipitated sedimentary rock formed in marine environments. Both are soluble in slightly acidic water. Limestone is composed primarily of calcium carbonate and is slightly more soluble than dolomite, a mixture of calcium and magnesium carbonate. The term "carbonate" is used for both types of formation. If a limestone incorporates considerable clastic material, it may have some primary porosity. If crystalline (composed of interlocking grains), little primary porosity is likely. Secondary porosity may result from the solution of calcite (calcium carbonate) along bedding planes. Some of these solution planes may enlarge over time to extensive caverns. Where thick limestone deposits prevail, such tunnels may be numerous, interconnected, and extend for several miles. If ground water withdrawals exceed recharge, tunnels may collapse, forming deep sinkholes.

Groundwater flow in limestone tunnels resembles open-channel flow. The aquifer may be confined, with discharge to the surface via springs issuing from solution cavities. Recharge to the system is rapid, so that potentiometric levels may rise and fall dramatically in response to climatic events. The aquifer may appear to be fairly homogeneous and isotropic over long distances if solution planes spread out in all directions.

Carbonates form major aquifers throughout the United States, especially in the southeast.

Plutonic and Metamorphic Rocks

Plutonic and some of the metamorphic rocks can support open fractures. Where fractures are numerous and interconnected, these rock types can function as aquifers. Plutonic rocks include granites, intermediate granitic types such as monzonite and granodiorite, and mafic and ultramafic types. Metamorphic rocks that function as aquifers are gneiss, massive quartzite, and some metavolcanics. Schist and slate are characteristically closely fractured, but these rocks are weak and usually the fractures have been closed. The weathered surface appearance of such rocks can be misleading because fractures that appear open at the surface are closed at shallow depth. Granite, which weathers easily in some climates, may exhibit a surficial zone in which fractures are filled with weathered detritus. At depth, fractures may be clean and open.

Because fractures in rocks, especially those caused by tectonic stresses, have preferred orientations, they are highly anisotropic. Stress-caused fractures may also vary in intensity and interconnectedness over short distances and exhibit heterogeneity. Recharge can be quite rapid if fractures are exposed at the surface. Granitic rocks have not been extensively utilized as aquifers but can contribute local water supplies in mountainous terrain.

1.13 VOLCANICS

Volcanic rocks range from extremely poor to some of the most productive aquifers known. Composition and age are the controlling factors. Rhyolitic and related lavas are the surface flow equivalents of the granitic intrusives. They may be extensively fractured but exhibit low permeability.

Basaltic lavas that form poor-to-moderate aquifers are geologically old or were chemically altered during deposition. In old basalts, fractures and lava tubes that might have formed during deposition have been closed by collapse and consolidation over time. Chemical alteration often occurs in marine basalts and andesites by the interaction of seawater during deposition. Fractures are filled with decomposition products and permeability is very low. Many marine lavas contain pillow structures, an important clue to underwater deposition.

It is important to distinguish pillow structure from cores and rinds developed in nonmarine lavas by spheroidal weathering. Nonmarine lavas that appear tight at the surface may contain abundant open fractures at shallow depth, below the zone of extensive weathering.

Young Quaternary basalts that still maintain open lava tubes and extensive, interconnected fractures formed during deposition are excellent aquifers. The *Aa* structure, which is best described as a deposit of rubble, occurs as congealed flow surfaces are broken up by the movement of liquid lava beneath, making it an excellent aquifer. Individual flows are tongue-shaped and are highly anisotropic and heterogeneous. Recharge into young lavas is very rapid.

Quaternary lavas form the principle aquifers of the Hawaiian Islands. Precipitation is orographically controlled, being in excess of 100 inches annually at higher elevations and 20 inches at sea level. Percolation through the volcanics is vertical until an ash bed is encountered. Ash beds formed of extruded exploded products form aquitards and perching structures for the percolating water. Below the ocean surface the less permeable lavas are saturated with seawater. The fresh percolating rainwater rests upon the saline water in a double convex lens known as the Ghyben-Herzberg lens.

In the Columbia Plateau of the northwestern United States, recent volcanics also provide extensive supplies of ground water.

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