CHAPTER ONE

OCCURRENCE OF GROUNDWATER

1.1 Introduction

Groundwater is water that exists in the pore spaces and fractures in rocks and sediments beneath the Earth's surface. It originates as rainfall or snow, and then moves through the soil and rock into the groundwater system, where it eventually makes its way back to the surface streams, lakes, or oceans.

- Groundwater makes up about 1% of the water on the Earth (most water is in oceans)
- But, groundwater makes up to 35 times the amount of water in lakes and streams.
- Groundwater occurs everywhere beneath the Earth's surface, but is usually restricted to depth less than about 750 meters.
- The volume of groundwater is equivalent to a 55-meter thick layer spread out over the entire surface of the Earth.
- **Technical note**: Groundwater scientists typically restrict the use of the term "groundwater" to underground water that can flow freely into a well, tunnel, spring, etc. This definition excludes underground water in the unsaturated zone. The unsaturated zone is the area between the land surface and the top of the groundwater system. The unsaturated zone is made up of earth materials and open spaces that contain some moisture but, for the most part, this zone is not saturated with water. Groundwater is found beneath the unsaturated zone where all the open spaces between sedimentary materials or in fractured rocks are filled with water and the water has a pressure greater than atmospheric pressure.

To understand the ways in which groundwater occurs, it is needed to think about the ground and the water properties.

- Porosity, which is the property of a rock possessing pores or voids.
- Saturated and unsaturated zones.
- Permeability, which is the ease with which water can flow through the rock.
- Aquifer, which is a geologic formation sufficiently porous to store water and permeable enough to allow water to flow through them in economic quantities.
- Storage coefficient, which is the volume of water that an aquifer releases from or takes into storage per unit surface area of aquifer per unit change in the component of area normal to surface.

1.2 Origin of Groundwater

The origin of groundwater is primarily one of the following:

- Groundwater derived from rainfall and infiltration within the normal hydrological cycle. This kind of water is called **meteoric water.** The name implies recent contact with the atmosphere.

- Groundwater encountered at great depths in sedimentary rocks as a result of water having been trapped in marine sediments at the time of their deposition. This type of groundwater is referred to as **connate waters**. These waters are normally saline. It is accepted that connate water is derived mainly or entirely from entrapped sea water as original sea water has moved from its original place. Some trapped water may be brackish.

- **Fossil water** if fresh may be originated from the fact of climate change phenomenon, i.e., some areas used to have wet weather and the aquifers of that area were recharged and then the weather of that area becomes dry.

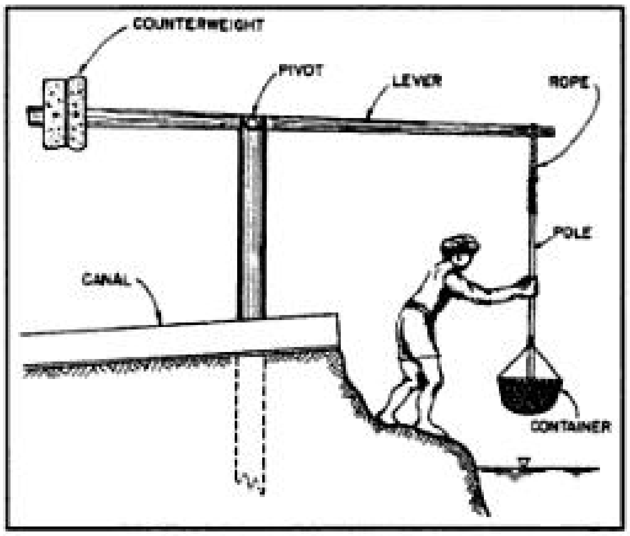


Figure 1.3. Water-lift method used by the Egyptians early as 2000 B.C. (McWhorter and Sunada, 1977).

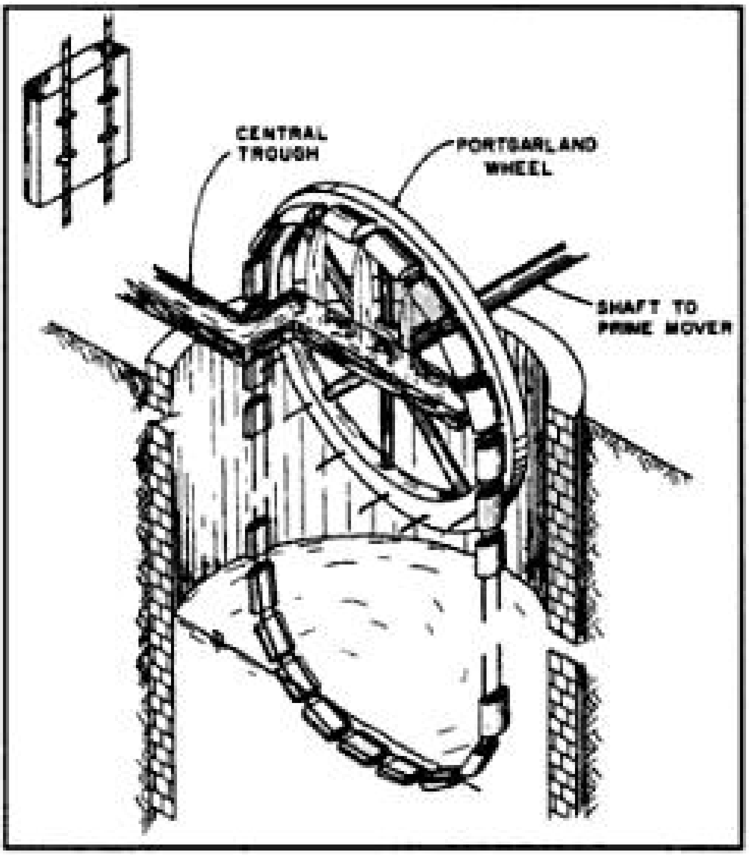


Figure 1.4. Persian wheel (McWhorter and Sunada, 1977).

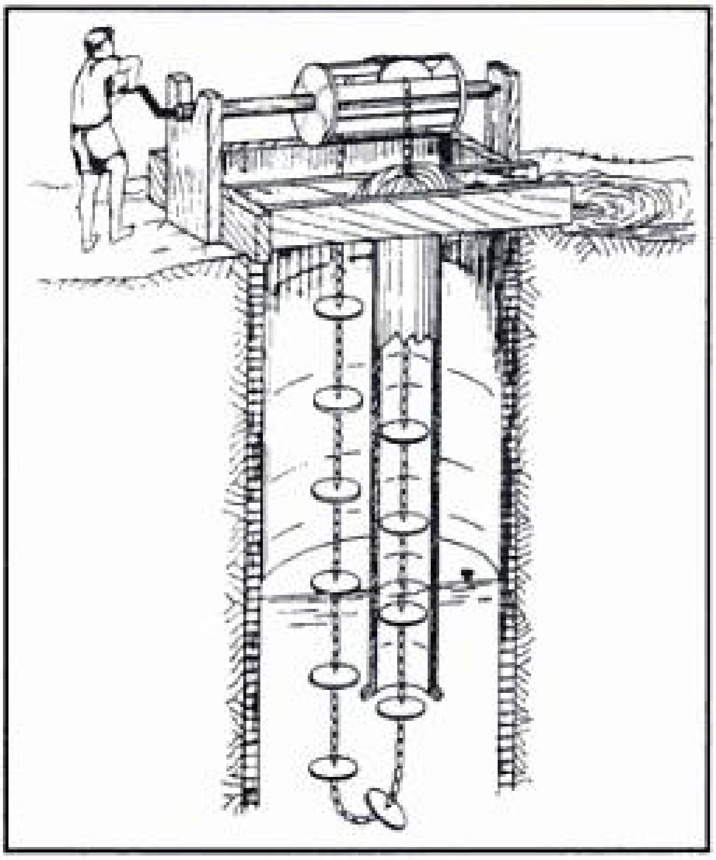


Figure 1.5. Chinese water ladder (McWhorter and Sunada, 1977).

1.3 Groundwater and the Hydrologic Cycle

- The hydrological cycle is the most fundamental principle of groundwater hydrology.
- The driving force of the circulation is derived from the radiant energy received from the sun.

Water **evaporates** and travels into the air and becomes part of a cloud. It falls down to earth as precipitation. Then it evaporates again. This happens repeatedly in a never-ending cycle. This **hydrologic cycle** never stops. Water keeps moving and changing from a solid to a liquid to a gas, repeatedly.

Precipitation creates **runoff** that travels over the ground surface and helps to fill lakes and rivers. It also **percolates** or moves downward through openings in the soil and rock to replenish **aquifers** under the ground. Some places receive more **precipitation** than others do with an overview balance. These areas are usually close to oceans or large bodies of water that allow more water to **evaporate** and form clouds. Other areas receive less. Often these areas are far from seawater or near mountains. As clouds move up and over mountains, the water vapor condenses to form precipitation and freezes. Snow falls on the peaks. **Figure 1.1** shows a schematic representation of the hydrological cycle.

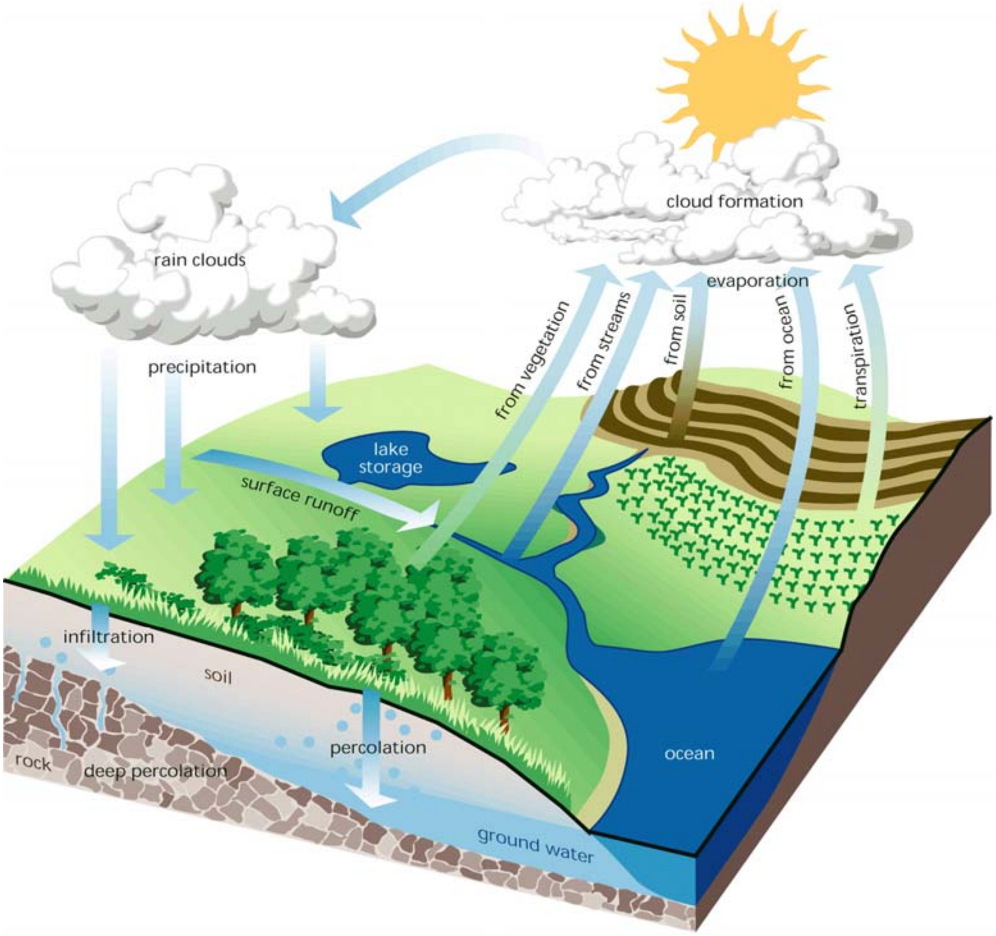


Figure 1.1 Schematic Representation of the Hydrological Cycle

In recent years there has been considerable attention paid to the concept of the **world water balance**, and the most recent estimates of these data emphasize the ubiquitous nature of groundwater in hydrosphere. With reference to **Table 1.1**, if we remove from consideration the 94% of the earth's water that rests in the oceans and seas at high levels of salinity, then groundwater accounts for about two-thirds of the freshwater resources of the world.

Table 1.1 Estimate of the Water Balance of the World

Parameter	Surface area $(Km^2)*10^6$	Volume $(Km^2)*10^6$	Volume (%)	Equivalent depth (m)*	Resident time
Oceans and seas	361	1370	94	2500	~ 4,000 years
Lakes and reservoirs	1.55	0.13	< 0.01	0.25	~ 10 years
Swamps	< 0.1	< 0.01	< 0.01	0.007	1-10 years
River channels	< 0.1	< 0.01	< 0.01	0.003	~ 2 weeks
Soil moisture	130	0.07	< 0.01	0.13	2 weeks – 1 year
Groundwater	130	60	4	120	~ 2 weeks – 10,000
Icecaps and glaciers	17.8	30	2	60	years
Atmospheric water	504	0.01	< 0.01	0.025	10-1000 years
Biospheric water	< 0.1	< 0.01	< 0.01	0.001	~ 10 days
					~ 1 week

* Computed as though storage were uniformly distributed over the entire surface of the earth.

1.4 Vertical Distribution of Groundwater

1.4.1 Volumetric Properties

Flow in soils and rocks takes place through void spaces, such as pores and cracks. The hydraulic properties of soils and rocks therefore depend on the sizes and shapes of the void spaces. These vary over very short distances (e.g. micrometers or millimeters). The idea of defining volumetric or hydraulic properties which apply at a given point in the unsaturated zone therefore has sense only if the properties relate to a finite volume of the soil/rock centered at that point. This volume is usually called the **representative elementary volume** (REV) and the properties defined in this fashion are sometimes called **point-scale** properties.

The point-scale properties vary in space. Part of this variation is associated with variations in the degree of compaction, weathering, cracking, and holing (such as holes left by decayed plant roots). The term **macropore** is often used to describe a feature such as a crack which allows rapid subsurface flow. Macropores and their effects on flow (and chemical transport) lie at the heart of many of the difficult, unresolved, problems in Near-Surface Hydrology.

At many locations, the subsurface flow is dominated by flow through complex networks of macropores. There may even be a few large soil pipes or subsurface channels (for example, subsurface pipes in steep hill slopes and channels in karst areas) which completely dominate the local flow conditions.

At present, there are no reliable techniques for measuring and quantifying macropore networks, and the modelling of macropore flow is in its infancy. The theory given below therefore concentrates on matrix flow (i.e. flow through the pores in media which do not contain macropores).

The point-scale properties can also vary in a systematic manner. There is usually **vertical layering**, resulting from the long-term evolution of the soil/rock profile by deposition processes, weathering, land management, etc. There are also variations associated with gradual horizontal changes (for instance, as shown in geological maps for a hill slope, catchment or region).

The concept of defining large-scale properties (e.g. a single, average, property for an entire hill slope) is controversial, but is being considered by some research workers.

The **porosity** n at a point is defined as:

$$n = \frac{volume\ of\ voids}{total\ volume} \tag{1.1}$$

The **volumetric moisture content** θ is:

$$\theta = \frac{volume\ of\ water}{total\ volume} \tag{1.2}$$

and the **relative moisture content** R is

$$R = \frac{volume\ of\ water}{volume\ of\ voids} \tag{1.3}$$

where, total volume = volume of solids + volume of voids.

In the geotechnical literature, property values are often quoted in mass terms (the moisture content by mass, for example), making use of data for the **bulk dry density** ρ_d of the medium (i.e. the dry mass per unit volume of soil/rock).

Approximate properties such as **field capacity** and **wilting point** are used in the hydrological and agricultural literature. Field capacity is the volumetric moisture content left in the medium after it has drained under gravity from saturation for a period of two days (definitions vary), and the wilting point is the volumetric moisture content which is just low enough so that any plants growing in the medium will fail to transpire, so will wilt and die.

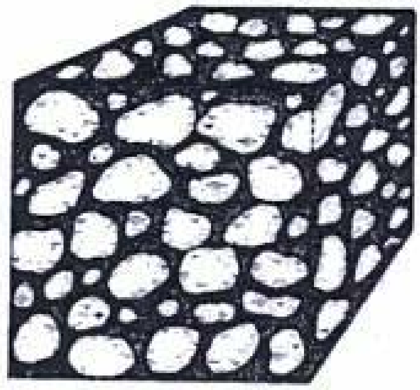


Figure 2.1. Pores in unconsolidated sedimentary rock (Heath, 1984).

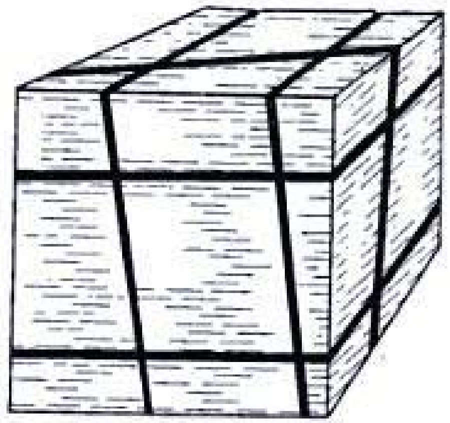


Figure 2.2. Fractures in intrusive igneous rocks (Heath, 1984).

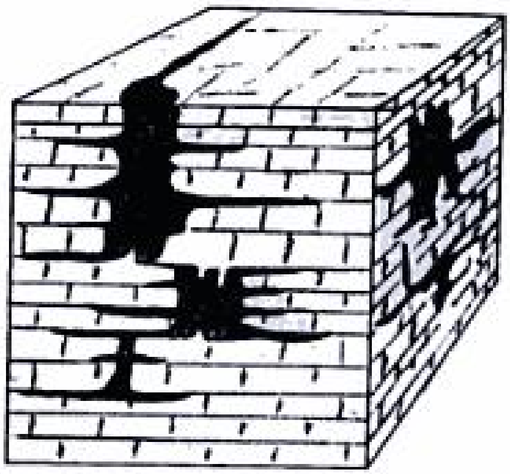


Figure 2.3. Caverns in limestone and dolomite (Heath, 1984).

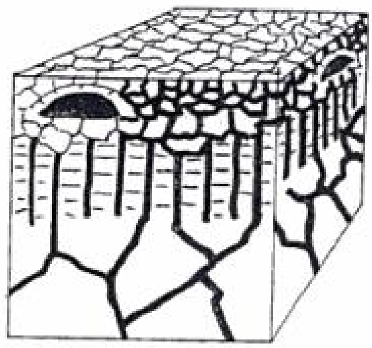


Figure 2.4. Lava tubes and cooling fractures in extrusive igneous rocks (Heath, 1984).

1.4.2 The Occurrence of Subsurface Water

The subsurface occurrence of groundwater may be divided into zones of aeration and saturation. The zone of aeration consists of interstices occupied partially by water and partially by air. In the zone of saturation all interstices are filled with water, under hydrostatic pressure. One most of the land masses of the earth, a single zone of aeration overlies a single zone of saturation and extends upward to the ground surface, as shown in **Figure 1.2**.

In the zone of aeration (unsaturated zone), *Vadose water* occurs. This general zone may be further subdivided into the soil water zone, the intermediate Vadose zone (sub-soil zone), and capillary zone (**Figure 1.2**).

The saturated zone extends from the upper surface of saturation down to underlying impermeable rock. In the absence of overlying impermeable strata, the *water table*, or *phreatic surface*, forms the upper surface of the zone of saturation. This is defined as the surface of atmospheric pressure and appears as the level at which water stands in a well penetrating the aquifer. Actually, saturation extends slightly above the water table due to capillary attraction; however, water is held here at less than atmospheric pressure. Water occurring in the zone of saturation is commonly referred to simply as *groundwater*, but the term *phreatic water* is also employed.

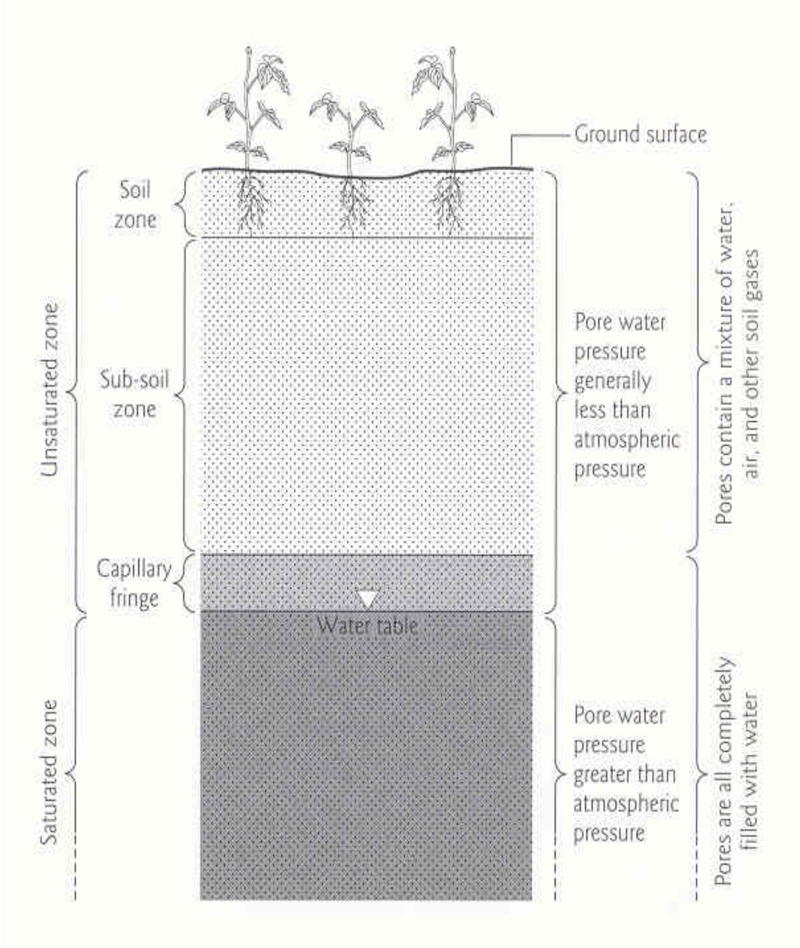


Figure 1.2 A schematic cross-section showing the typical distribution of subsurface waters in a simple "unconfined" aquifer setting, highlighting the three common subdivisions of the unsaturated zone and the saturated zone below the water table.

1.5 Types of Geological Formations and Aquifers

There are basically four types of geological formations (Aquifers, Aquitard, Aquiclude, and Aquifuge)

1.5.1 Aquifer

An aquifer is a ground-water reservoir composed of geologic units that are saturated with water and sufficiently permeable to yield water in a usable quantity to wells and springs. Sand and gravel deposits, sandstone, limestone, and fractured, crystalline rocks are examples of geological units that form aquifers. Aquifers provide two important functions: (1) they transmit ground water from areas of recharge to areas of discharge, and (2) they provide a storage medium for useable quantities of ground water. The amount of water a material can hold depends upon its porosity. The size and degree of interconnection of those openings (permeability) determine the materials' ability to transmit fluid.

Types of Aquifers

Most aquifers are of large areal extent and may be visualized as underground storage reservoirs. Water enters a reservoir from natural or artificial recharge; it flows out under the action of gravity or is extracted by wells. Ordinarily, the annual volume of water removed or replaced represents only a small fraction of the total storage capacity. Aquifers may be classed as unconfined or confined, depending on the presence or absence of a water table, while a leaky aquifer represents a combination of the two types.

Unconfined Aquifer. An unconfined aquifer is one in which a water table varies in undulating form and in slope, depending on areas of recharge and discharge, pumpage from wells, and permeability. Rises and falls in the water table correspond to changes in the volume of water in storage within an aquifer. **Figure 1.2** is an idealized section through an unconfined aquifer; the upper aquifer in **Figure 1.3** is also unconfined. Contour maps and profiles of the water table can be prepared from elevations of water in wells that tap the aquifer to determine the quantities of water available and their distribution and movement.
A special case of an unconfined aquifer involves ***perched water bodies***, as illustrated by **Figure 1.3.** This occurs wherever a groundwater body is separated from the main groundwater by a relatively impermeable stratum of small areal extent and by the zone of aeration above the main body of groundwater. Clay lenses in sedimentary deposits often have shallow perched water bodies overlying them. Wells tapping these sources yield only temporary or small quantities of water.

Confined Aquifers. Confined aquifers, also known as ***artesian or pressure aquifers***, occur where groundwater is confined under pressure greater than atmospheric by overlying relatively impermeable strata. In a well penetrating such an aquifer, the water level will rise above the bottom of the confining bed, as shown by the artesian and flowing wells of **Figure 1.3.** Water enters a confined aquifer in an area where the confining bed rises to the surface; where the confining bed ends underground, the aquifer becomes unconfined. A region supplying water to a confined area is known as a ***recharge area;*** water may also enter by leakage through a confining bed. Rises and falls of water in wells penetrating confined aquifers result primarily from changes in pressure rather than changes in storage volumes. Hence, confined aquifers display only small changes in storage and serve primarily as conduits for conveying water from recharge areas to locations of natural or artificial discharge.

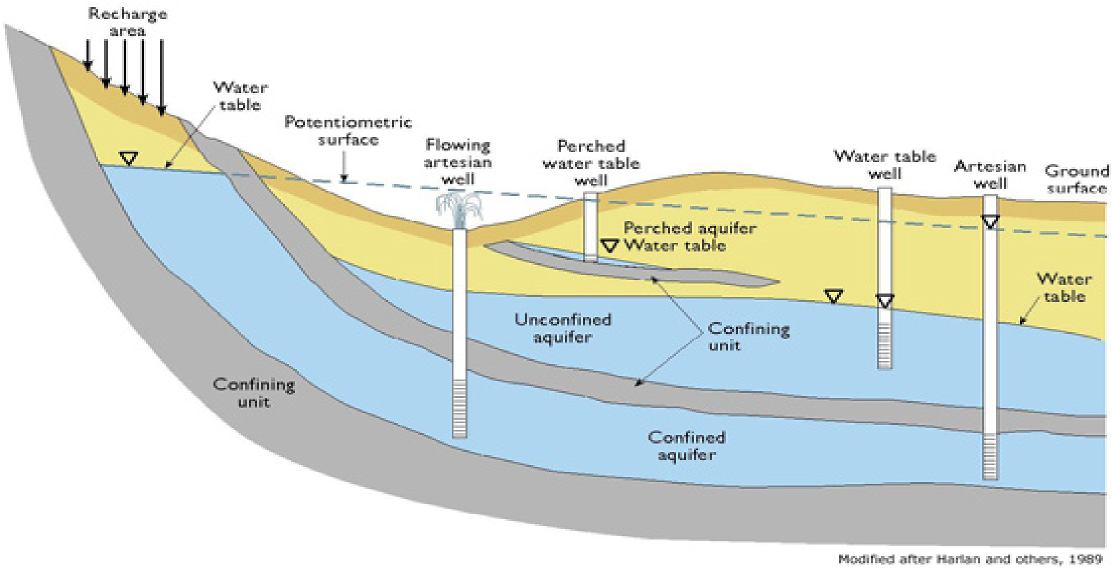


Figure 1.3 Schematic Cross-sections of Aquifer Types

Leaky Aquifer. Aquifers that are completely confined or unconfined occur less frequently than do *leaky,* or *semi-confined*, aquifers. These are a common feature in alluvial valleys, plains, or former lake basins where a permeable stratum is overlain or underlain by a semi-pervious aquitard or semi-confining layer. Pumping from a well in a leaky aquifer removes water in two ways: by horizontal flow within the aquifer and by vertical flow through the aquitard into the aquifer (see **Figure 1.4**).

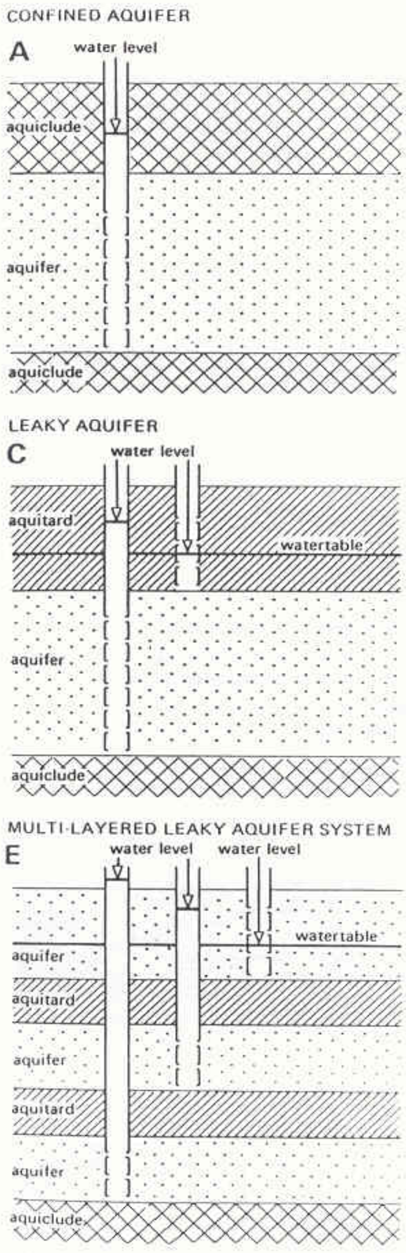


Figure 1.4 Different types of aquifers; A. Confined aquifer, B. Unconfined Aquifer, C. and D. Leaky aquifers, E. Multi-layered leaky aquifer system.

1.5.2 Aquitard

An aquitard is a partly permeable geologic formation. It transmits water at such a slow rate that the yield is insufficient. Pumping by wells is not possible. For example, sand lenses in a clay formation will form an aquitard.

1.5.3 Aquiclude

An aquiclude is composed of rock or sediment that acts as a barrier to groundwater flow. Aquicludes are made up of low porosity and low permeability rock/sediment such as shale or clay. Aquicludes have normally good storage capacity but low transmitting capacity.

1.5.4 Aquifuge

An aquifuge is a geologic formation which doesn't have interconnected pores. It is neither porous nor permeable. Thus, it can neither store water nor transmit it. Examples of aquifuge are rocks like basalt, granite, etc. without fissures.

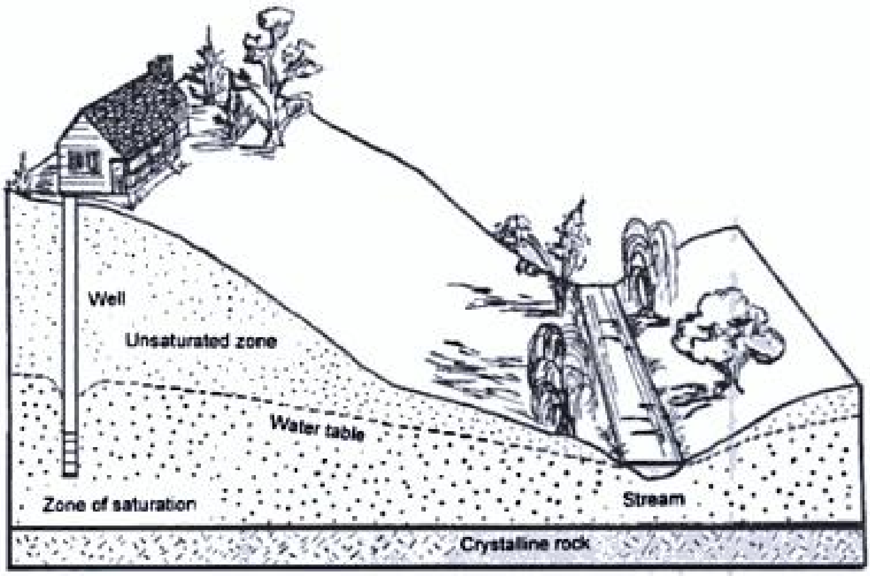


Figure 2.5. Unconfined aquifer. Simple riverine system comprised of sand.

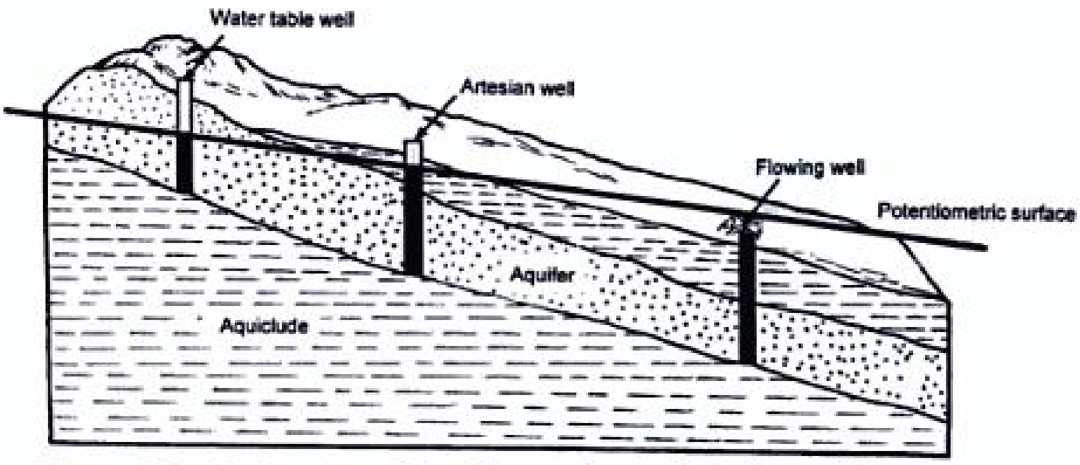


Figure 2.6. Confined aquifer with artesian and flowing wells.

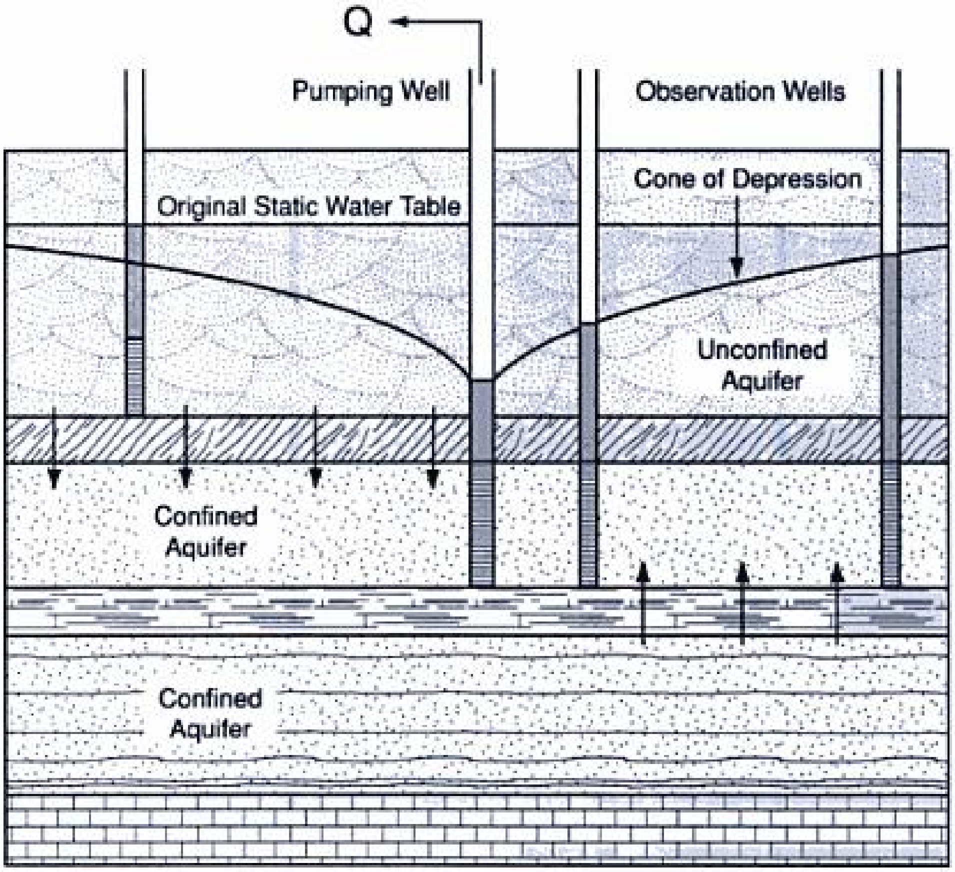


Figure 2.7. Development of a leaky confined aquifer during a pumping period.

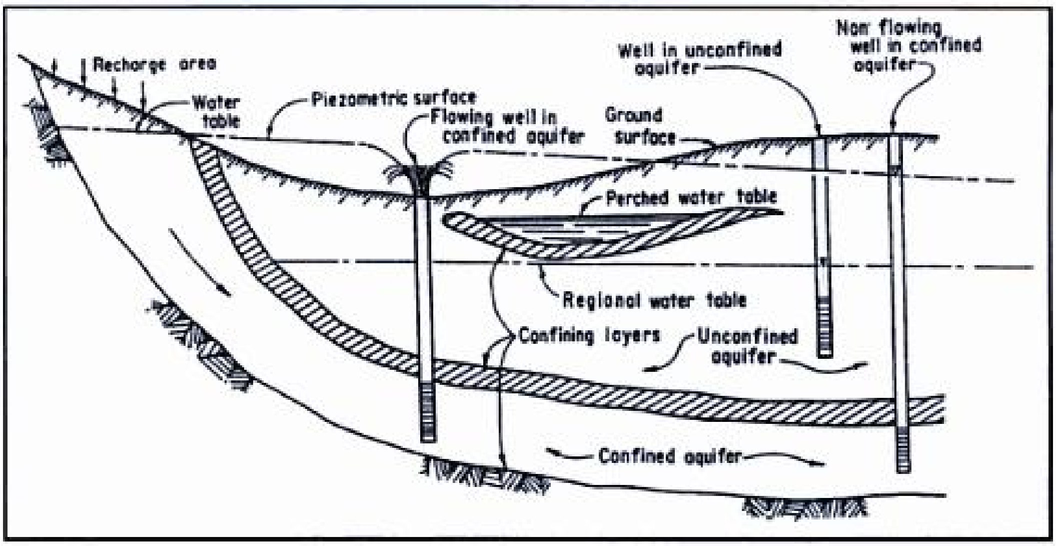


Figure 2.8. Confined, unconfined and perched aquifers (Source: USDI, 1981).

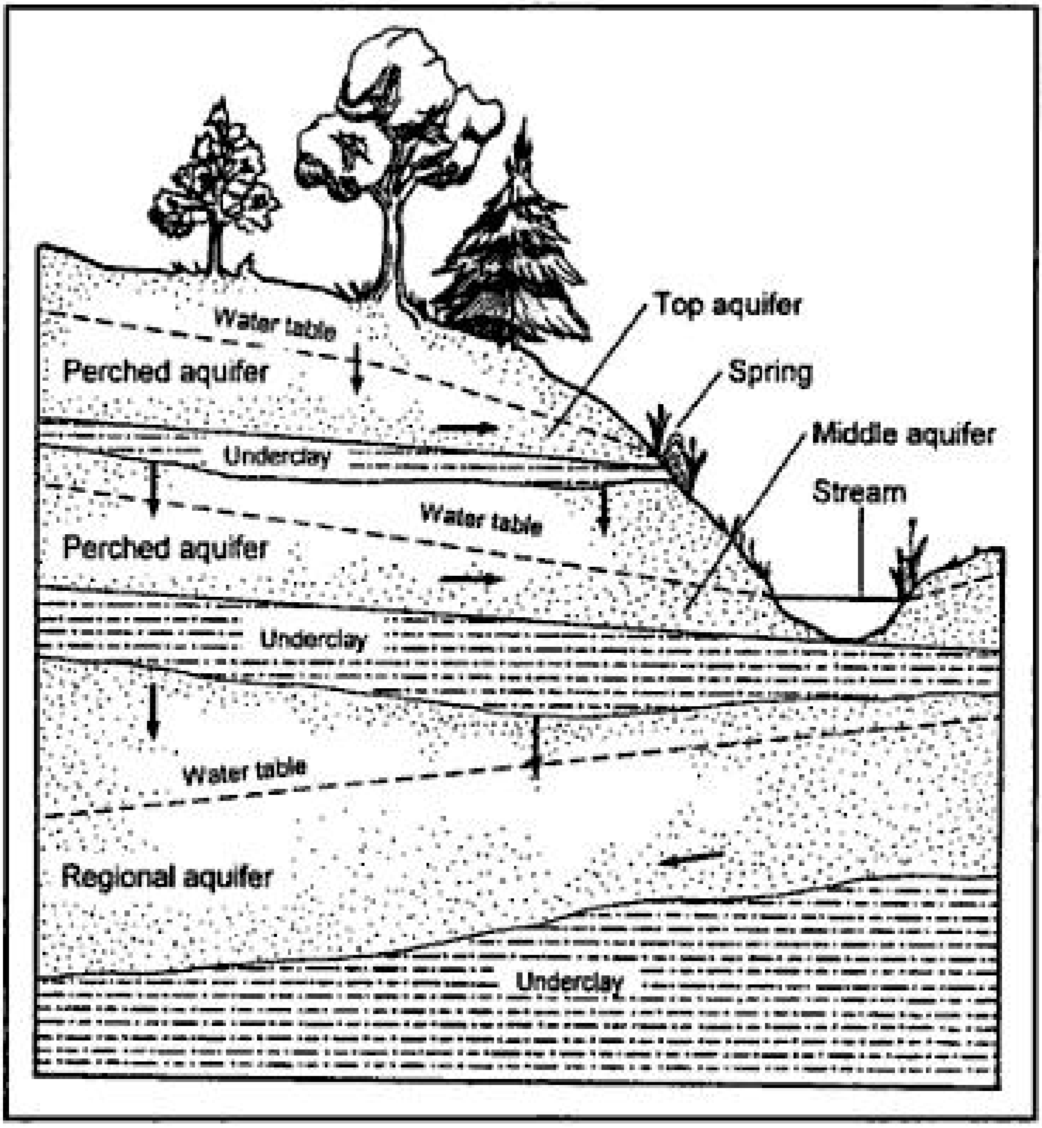


Figure 2.9. Perched aquifers above a regional aquifer system.

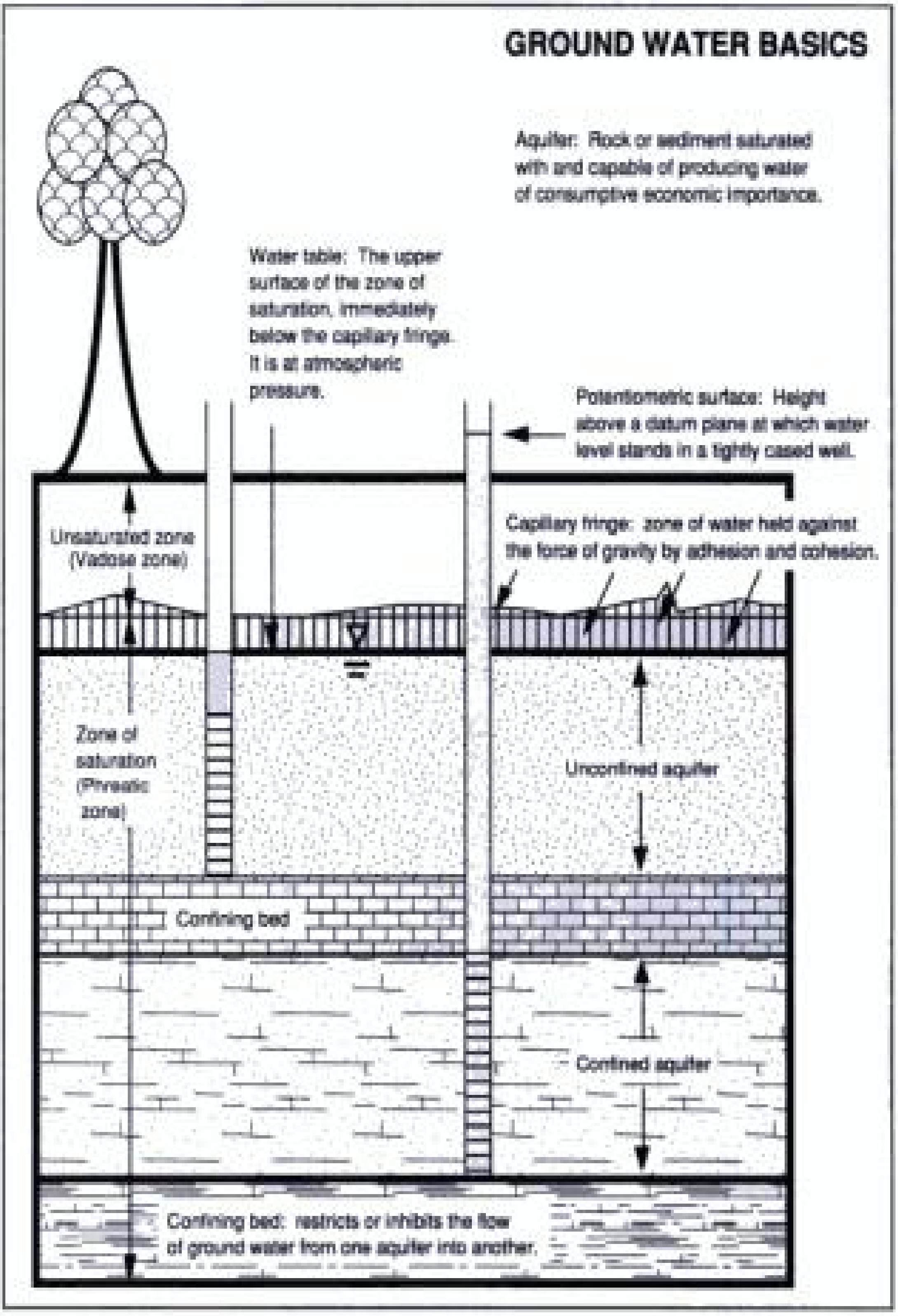


Figure 2.10. Ground-water basics (as defined by Heath, 1983).

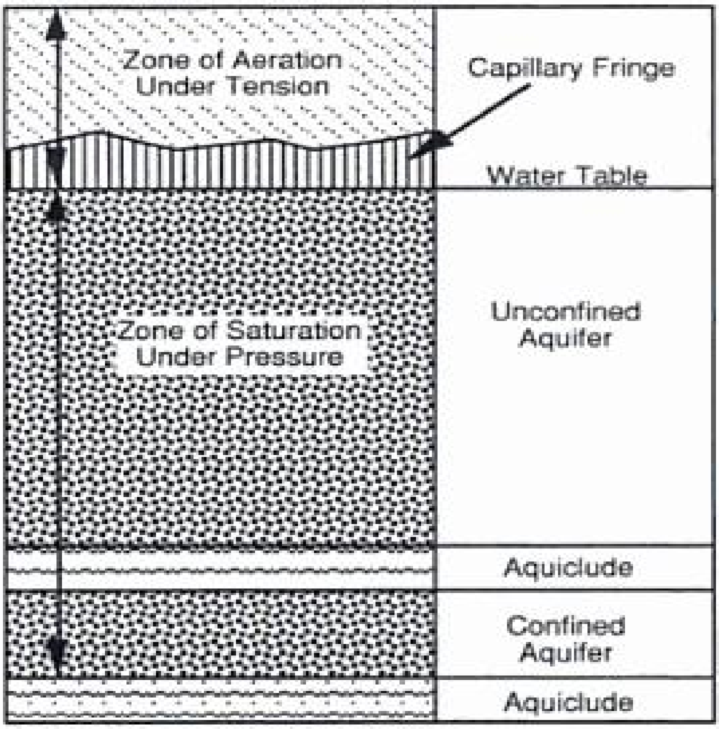
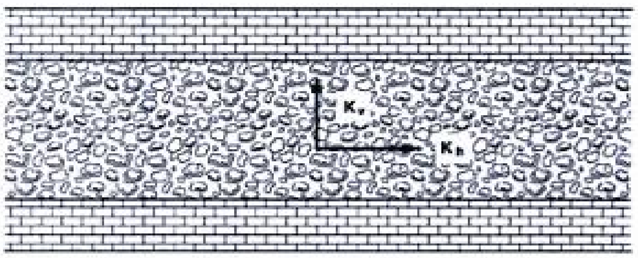
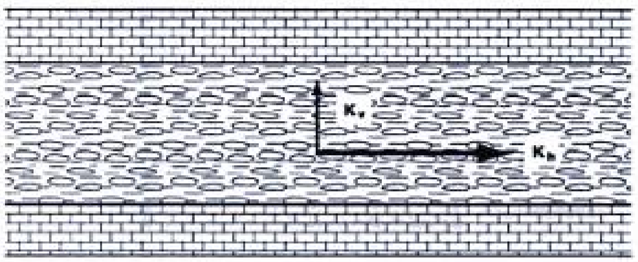


Figure 2.11. Ground-water zones according to Meinzer (1949).



Homogeneous, isotropic aquifer: rounded or sub-rounded sediment.



Homogeneous, anisotropic aquifer: sediment laid in the direction of their long axis.

Figure 2.12. Homogeneous aquifers.

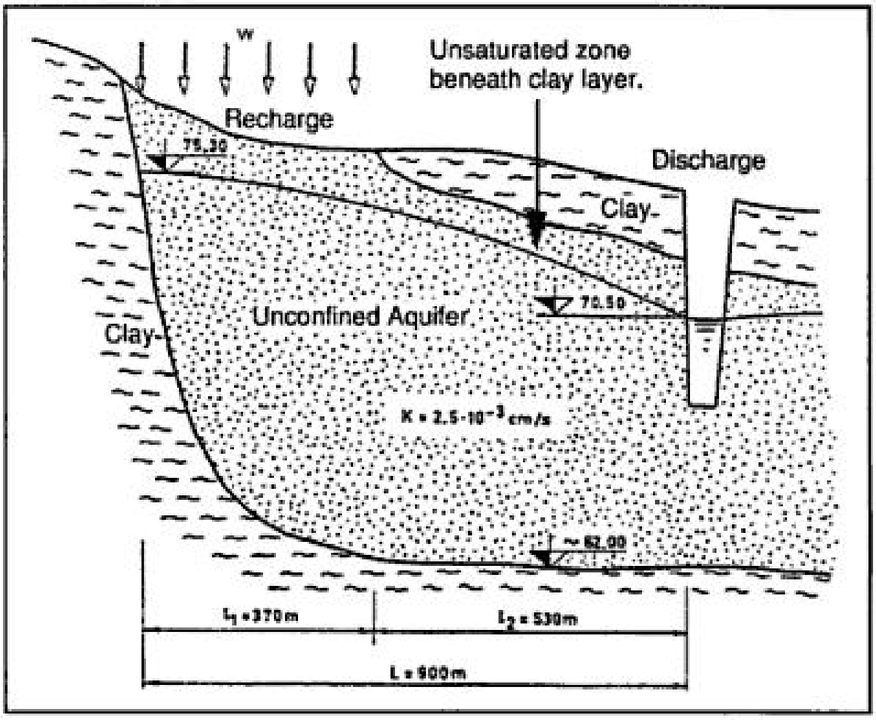


Figure 2.19. **Unconfined aquifer with a thick clay layer above the local water table. Note the areas of recharge and discharge (modified from Vukovic and Soro, 1997).**

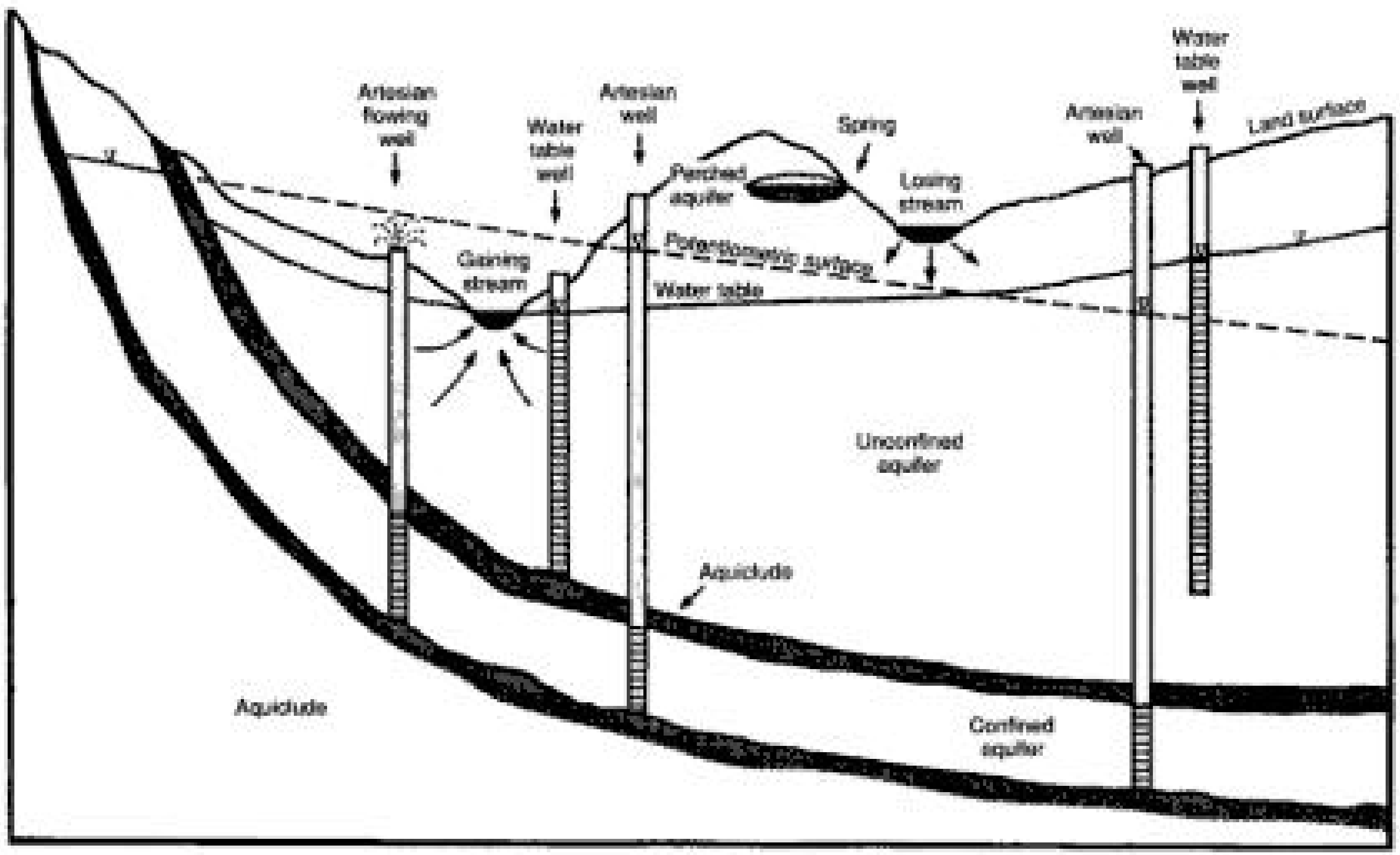


Figure 2.20. Relationship between the unconfined water table and the potentiometric surface (National Engineering Handbook 18, 1978, U.S. Soil Conservation Service).

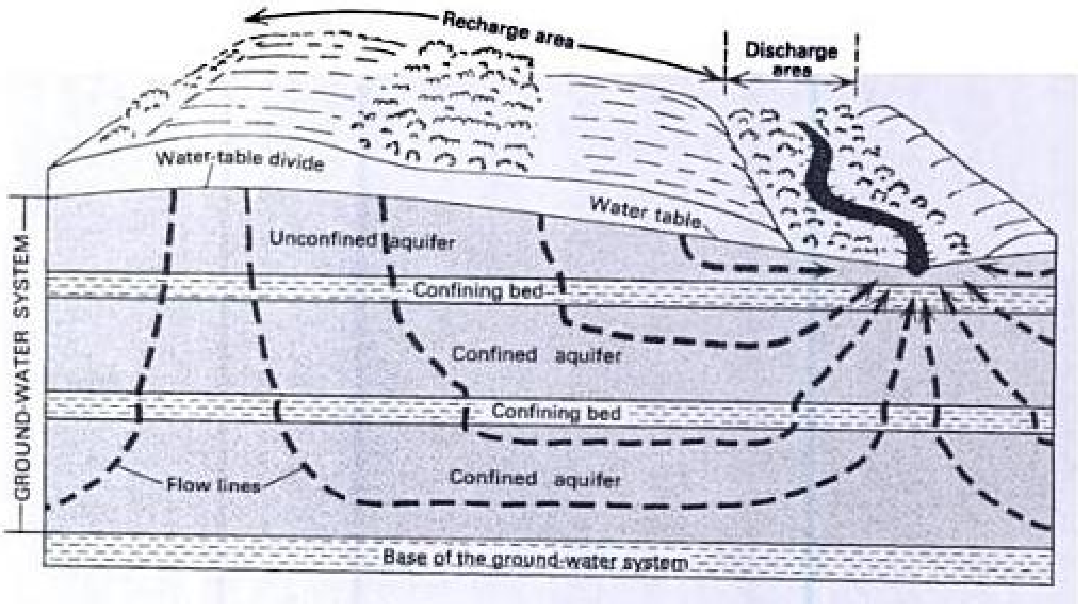


Figure 2.21. Multi-layered aquifer system (Heath, 1983).

1.6 Water Table and Piezometric Surface

1.6.1 Water table

Water table is the surface of water level in an unconfined aquifer at which the pressure is atmospheric. It is the level at which the water will stand in a well drilled in an unconfined aquifer. The water table fluctuates whenever there is a recharge or an outflow from the aquifer. In fact, the water table is constantly in motion adjusting its surface to achieve a balance between the recharge and the out flow. Generally, the water table follows the topographic features and is high below ridges and low below valleys. However, sometimes the topographic ridge and the water table ridge may not coincide and there may be flow from one aquifer to the other aquifer, called **watershed leakage.** Wherever the water table intersects the ground surface, a seepage surface or a spring is formed.

Perched water table when a small water body is separated from the main groundwater body by a relatively small impermeable stratum. Wells drilled below the perched water table up to the small impervious stratum yield very small quantity of water and soon go dry.

1.6.2 Piezometric surface

The water in a confined aquifer is under pressure. When a well is drilled in a confined aquifer, the water level in it will rise above the top of aquifer. The piezometric surface is an imaginary surface to which the water level would rise if a piezometer was inserted in the aquifer. Thus, it indicates the pressure of the water in the aquifer. Hence, a piezometric surface is the water table equivalent of the confined aquifer (see **Figure 1.5**).

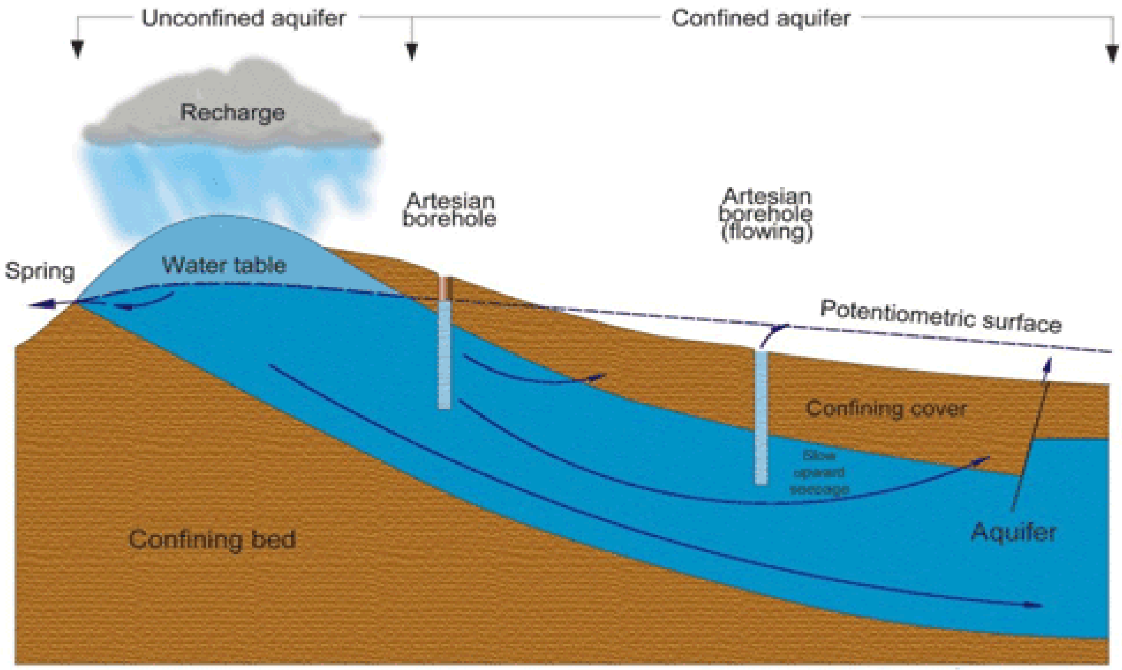


Figure 1.5 Water Table and Piezometric Surface

1.7 Aquifer Properties

The following properties of the aquifer are required for study of groundwater hydrology:

1. Porosity
2. Specific Yield
3. Specific Retention
4. Coefficient of permeability
5. Transmissibility
6. Specific Storage
7. Storage Coefficient

1.7.1 Porosity

Definition of Porosity

Porosity (n) is the percentage of rock or soil that is void of material. The larger the pore space or the greater their number, the higher the porosity and the larger the water-holding capacity. It is defined mathematically by the equation:

$$n = \frac{V_v}{V} \times 100\,\% \qquad \textbf{(1.4)}$$

Where,

n is the porosity (percentage)
V_v is the volume of void space in a unit volume of earth materials (L^3, cm^3 or m^3)
V is the unit volume of earth material, including both voids and solids (L^3, cm^3 or m^3)

In sediments or sedimentary rocks the porosity depends on **grain size**, the **shape of the grains**, the **degree of sorting** and the **degree of cementation.** In rocks, the porosity depends upon the **extent, spacing** and **pattern of cracks** and **fractures.**

- The porosity of well-rounded sediments, which have been sorted so that they are all about the same size, is independent of particle size, depending upon the packing.

- Well-rounded coarse-grained sediments usually have higher porosity than fine-grained sediments, because the grains don't fit together well (see **Figure 1.6**)

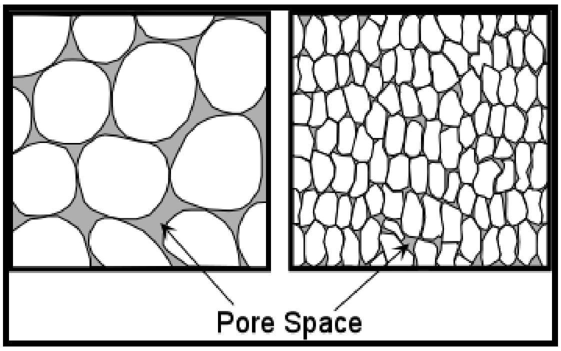


Figure 1.6 Porosity of Well-Rounded Coarse-Sediments vs. Fine Grained Sediments

- In igneous and metamorphic rocks porosity is usually low because the minerals tend to be intergrown, leaving little free space. Higher fractured igneous and metamorphic rocks, however, could have high secondary porosity.

- Since cements tend to fill in the pore space, highly cemented sedimentary rocks have lower porosity (see **Figure 1.7**).

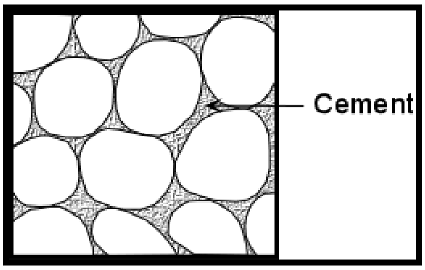


Figure 1.7 Highly Cemented Sedimentary Rock

- Poorly sorted sediments (sediments contains a mixture of grain sizes) usually have lower porosity because the fine-grained fragments tend to fill the open spaces (see **Figure 1.8**).

- The porosity of sediments is affected by the shape of the grains. Well-rounded grains may be almost perfect spheres, but many grains are very irregular. They can be shaped like rods, disks, or books. Sphere-shaped grains will pack more tightly and have less porosity than particles of other shapes. The fabric or orientation of the particles, if they are not spheres, also influences porosity (**Figure 1.8**).

- Porosity can range from zero to more than 60%. Recently deposited sediments have higher porosity. Dense crystalline rock or highly compacted soft rocks such as shale have lower porosity.

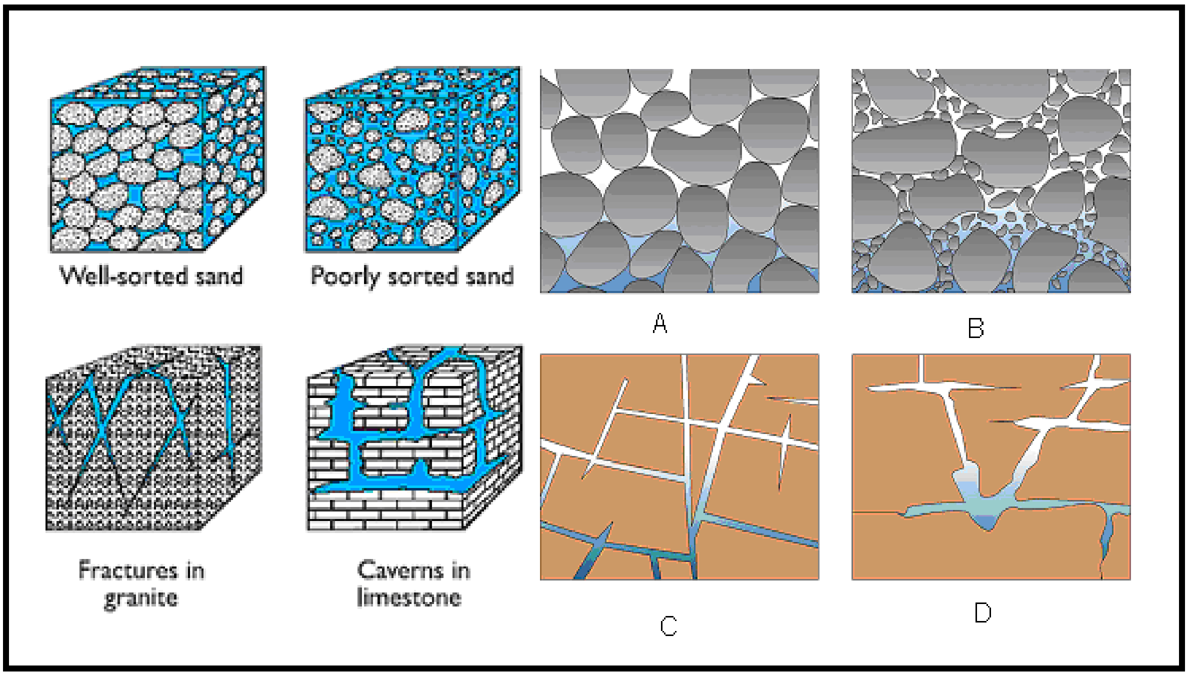


Figure 1.8 Relation Between Texture and Porosity **A.** Well –Sorted Sand Having High Porosity; **B.** Poorly- Sorted Sand Having Low Porosity; **C.** Fractured Crystalline Rocks (Granite); **D.** Soluble Rock-Forming Material (Limestone).

- In porous rock, there may be small pores known as dead end pores which have only one entrance, and so water molecules can diffuse in and out of them, but there can be no hydraulic gradient across them to cause bulk flow of groundwater. In extreme cases, there may be pores containing water that are completely closed so that the water in them is trapped. This may occur during digenetic transformations of the rock. Since we are frequently interested in the movement of groundwater, it is useful to define a porosity that refers only to the movable water in the rock.

This is called **the kinematic or effective porosity n_e [dimensionless]**

$$n_e = \frac{volume\, of\ rock\ occupied\, by\, movable\, water}{total\, volume\, of\ rock} \qquad \textbf{(1.5)}$$

- It is worth distinguishing between **Intergranular** or **matrix** or **primary porosity** as the latter is the porosity provided by small spaces between adjacent grains of the rock, and **secondary porosity** of **fractured rocks** is the porosity provided by discrete rock mass discontinuities (faults, joints and fractures).

➢ **Table 1.2** lists representative porosity ranges from various geologic materials.

Table 1.2 Range of Values of Porosity (after Freeze & Cherry, 1979)

Formation	n (%)
Unconsolidated deposits	
Gravel	25 - 40
Sand	25 - 50
Silt	35 - 50
Clay	40 - 70
Rocks	
Fractured basalt	5 - 50
Karst limestone	5 - 50
Sandstone	5 - 30
Limestone, dolomite	0 - 20
Shale	0 – 10
Fractured crystalline rock	0 - 10
Dense crystalline rock	0 – 5

Classification of Sediments

Sediments are classified on the basis of the size (diameter) of the individual grains. There are many classification systems in use. The engineering classification of sediments is somewhat different that the geological classification. The American Society of Testing Materials defines sediments on the basis of the grain-size distribution shown in **Table 1.3.**

Table 1.3 Engineering grain-size classification (after Fetter, 1994)

Formation	Size Range (mm)	Example
Boulder	> 305	Basketball
Cobbles	76 – 305	Grapefruit
Coarse gravel	19 – 76	Lemon
Fine gravel	4.75 – 19	Pea
Coarse sand	2 – 4.75	Water softener salt
Medium sand	0.42 – 2	Table salt
Fine sand	0.075 – 0.42	Powdered sugar
Fines	< 0.075	Talcum powder

The grain-size distribution of a sediment may be conveniently plotted on semi-log paper. The cumulative percent finer by weight is plotted on the arithmetic scale and the grain size is plotted on the logarithmic scale. The grain size of the sand fraction is determined by shaking the sand through a series of sieves with decreasing mesh openings. The 200 mesh screen, with an opening of 0.075 mm, separates the sand fraction from the fines (see **Figure 1.9**).

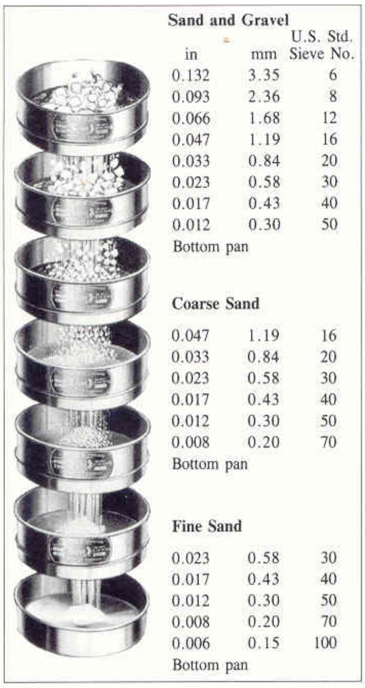


Figure 1.9 Recommended sieve groups suitable for sieving various classes of unconsolidated sediments.

The gradation of the fines is determined by a hydrometer test, which is based on the rate that the sediment settles in water. **Figure 1.10** is a grain size distribution curve for a silty fine to coarse sand. This sample is somewhat poorly sorted as there is a wide range of grain sizes present. **Figure 1.11** is the grain-size distribution curve for well-sorted fine sand. Less than 5% of the sample consisted of fines that pass the 200 mesh sieve.

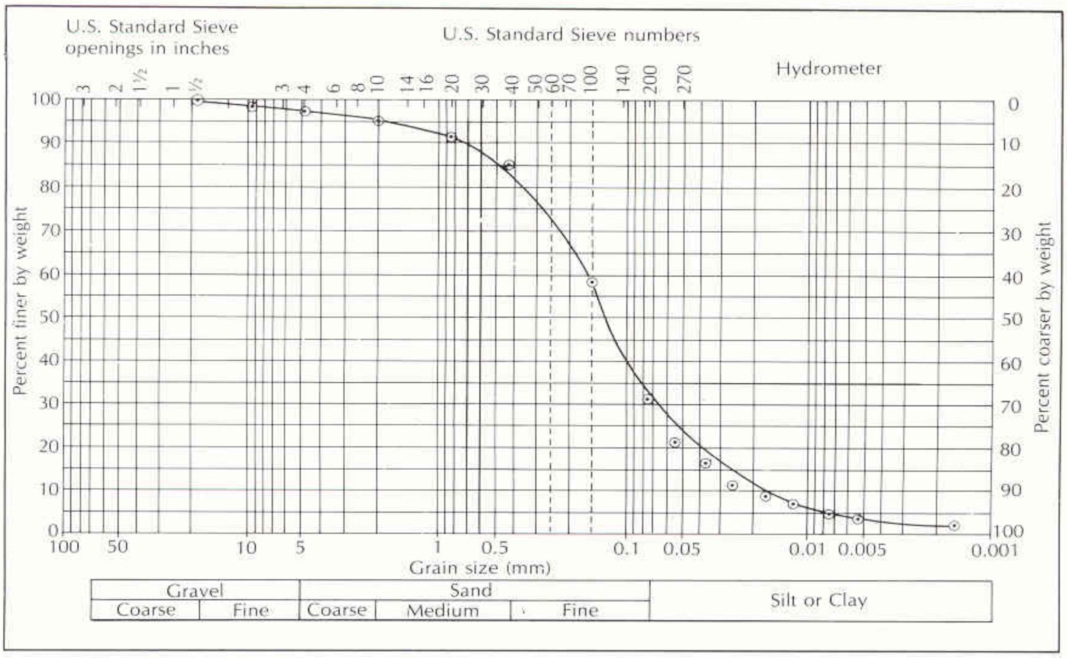


Figure 1.10 Grain-size distribution curve of a silty fine to medium sand

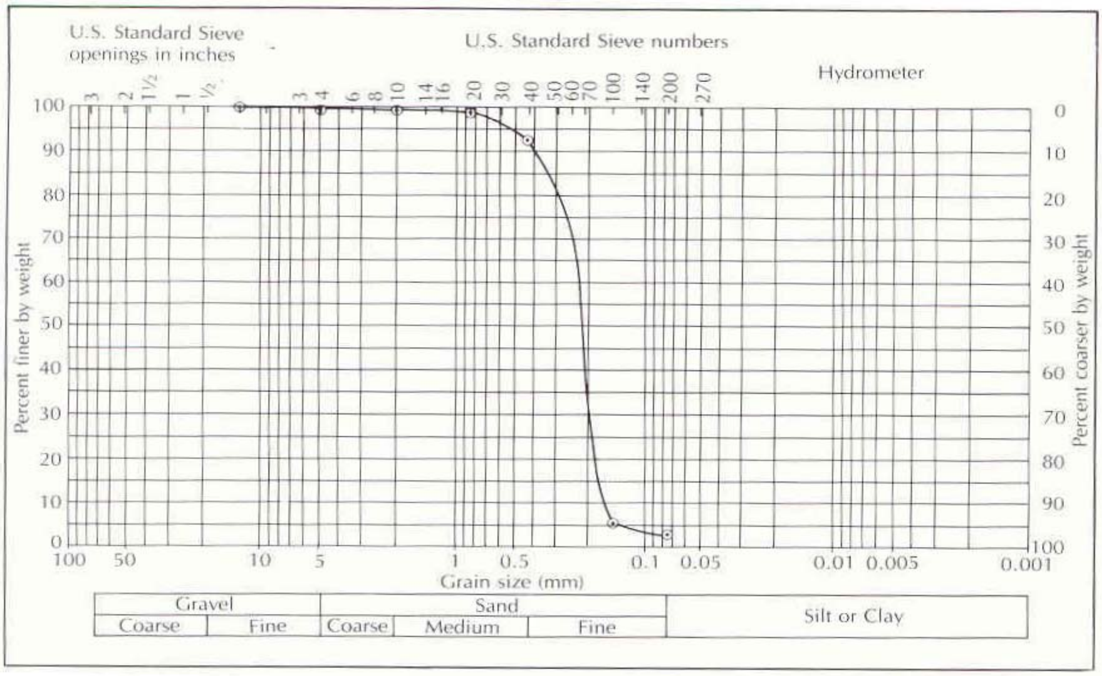


Figure 1.11 Grain-size distribution curve of a fine sand

The **uniformity coefficient** of a sediment is a measure of how well or poorly sorted it is. The uniformity coefficient, U_c, is the ratio of the grain size that is 60% finer by weight, D_{60}, to the grain size that is 10% finer by weight, D_{10}.

$$U_c = \frac{D_{60}}{D_{10}} \qquad \textbf{(1.6)}$$

where,

D_{60} grain size in which 60 percent of sample is passed
D_{10} grain size in which 10 percent of sample is passed (effective diameter)

A sample with an U_c less than 4 is well sorted; if the U_c is more than 6 it is poorly sorted. The poorly sorted silty sand in **Figure 1.10** has a U_c of 8.3, whereas the well-sorted sand of **Figure 1.11** has a U_c of 1.4.

1.7.2 Specific Yield (S_y)

Specific yield (S_y) is the ratio of the volume of water that drains from a saturated rock owing to the attraction of gravity (or by pumping from wells) to the total volume of the saturated aquifer. It is defined mathematically by the equation:

$$S_y = \frac{V_w}{V} \times 100\% \qquad \textbf{(1.7)}$$

where,

V_w is the volume of water in a unit volume of earth materials (L^3, cm^3 or m^3)
V is the unit volume of earth material, including both voids and solids (L^3, cm^3 or m^3

All the water stored in a water bearing stratum cannot be drained out by gravity or by pumping, because a portion of the water is rigidly held in the voids of the aquifer by molecular and surface tension forces (see **Table 1.4**).

Table 1.4 Specific Yield in Percent (after Freeze & Cherry, 1979)

Formation	S_y (range)	S_y (average)
Clay	0 - 5	2
Sandy clay	3 - 12	7
Silt	3 - 19	18
Fine sand	10 - 28	21
Medium sand	15 - 32	26
Coarse sand	20 - 35	27
Gravelly sand	20 - 35	25
Fine gravel	21 - 35	25
Medium gravel	13 - 26	23
Coarse gravel	12 - 26	22
Limestone		14

1.7.3 Specific Retention (S_r)

Specific retention (S_r) is the ratio of the volume of water that cannot be drained out to the total volume of the saturated aquifer. Since the specific yield represents the volume of water that a rock will yield by gravity drainage, hence the specific retention is the remainder. The sum of the two equals porosity.

$$n = S_r + S_y \tag{1.8}$$

- The specific yield and specific retention depend upon the shape and size of particle, distribution of pores (voids), and compaction of the formation.
- The specific retention increases with decreasing grain size.
- It should be noted that it is not necessary that soil with high porosity will have high specific yield because that soil may have low permeability and the water may not easily drain out. For example, clay has a high porosity but low specific yield and its permeability is low.
- **Figure 1.12** illustrates the concept of specific yield.

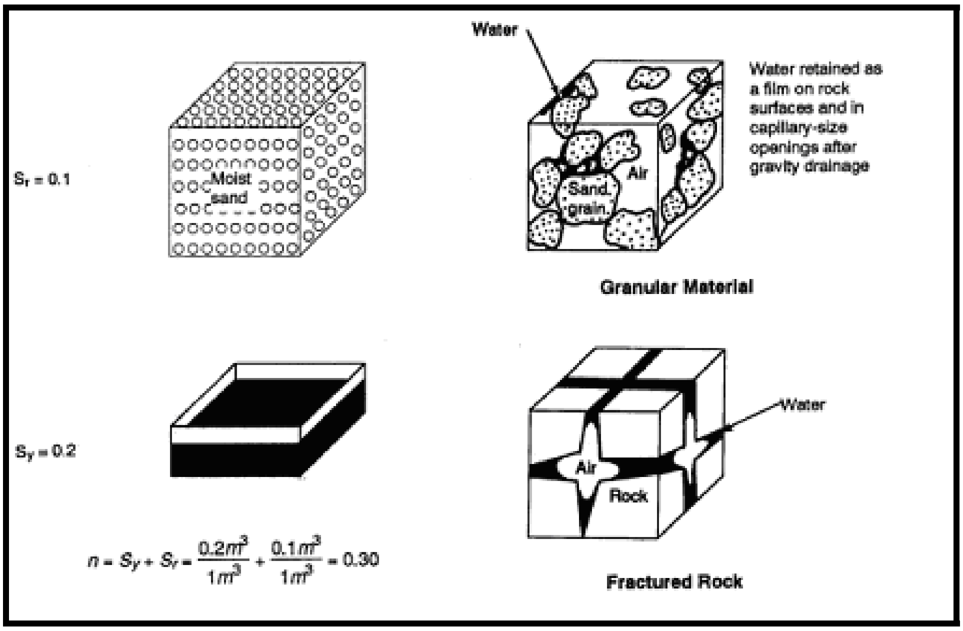


Figure 1.12 Specific Retention

1.7.4 Coefficient of Permeability (Hydraulic conductivity) (K)

Permeability is the ease with which water can flow in a soil mass or a rock. The **coefficient of permeability (K)** is equal to the discharge (m^3/s) per unit area (m^2) of soil mass under unit hydraulic gradient. Because the discharge per unit area equals to the velocity, the coefficient of permeability has the dimension of the velocity [L/T]. It is usually expressed as cm/s, m/s, m/day, etc. The coefficient of permeability is also called **hydraulic conductivity** (see **Figure 1.13**).

Hydraulic Conductivity can be determined and expressed as follows:

Formulas

1 [Hazen method]. The coefficient of permeability (K) depends on the properties of both porous medium and fluid. It can be expressed as,

$$K = \frac{[Cd_m^2]\rho g}{\mu} \qquad \textbf{(1.9)}$$

where,

C	is the shape factor which depends upon the shape, particle size and packing of the porous media
d_m	is the mean particle size (d_{50}) (L, m)
ρ	is the mass density (M/L^3, kg/m^3)
g	is the acceleration due to gravity (L/T^2, m/s^2)
μ	is the viscosity (M/T.L, kg/s.m)

- Another coefficient of permeability, called **intrinsic permeability (k)**, is sometimes used. The intrinsic permeability depends upon the porous medium and is independent of the properties of the fluid. It is usually expressed as,

$$k = Cd_m^2 \qquad \textbf{(1.10)}$$

- The intrinsic permeability k has the dimensions of [L^2] and is usually expressed in cm^2 or **Darcy,** where 1 Darcy = 0.987 * 10^{-8} cm^2.

- The intrinsic permeability is rarely used in groundwater hydrology, but this term is very popular in the petroleum, natural gas industries, and in density-dependent flow problems such as saline water intrusion.

- The intrinsic permeability is also called the absolute permeability.

- The rate of groundwater flow is controlled by the two properties of the rock, porosity and permeability.

- Low porosity usually results in low permeability, but high porosity does not necessarily imply high permeability. It is possible to have a highly porous rock with little or no interconnections between pores. A good example of a rock with high porosity and low permeability is a vesicular volcanic rock, where the bubbles that once contained gas give the rock a high porosity, but since these holes are not connected to one another, the rock has low permeability.

- Typical values of hydraulic conductivity for unconsolidated and hard rocks are given in **Table 1.5** which are taken from Marsily [1986].

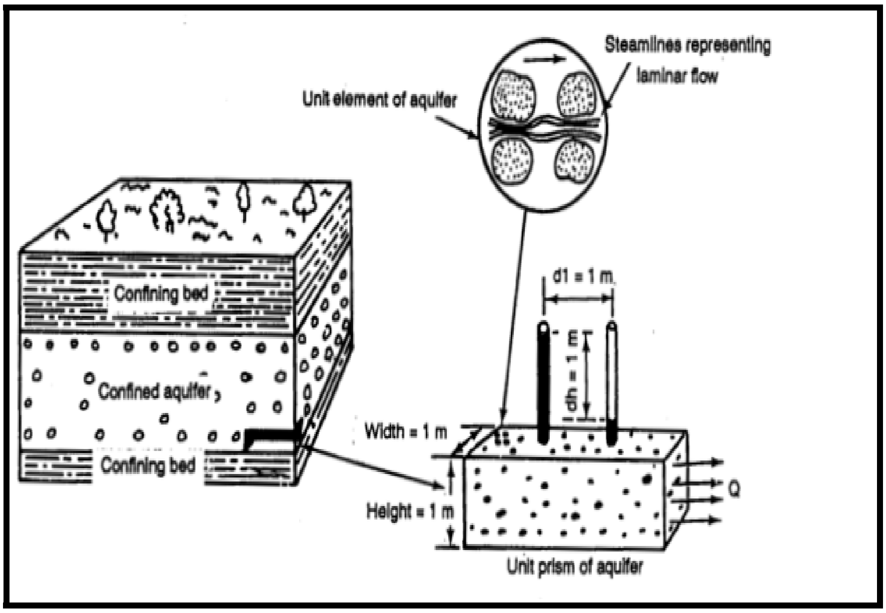


Figure 1.13 Hydraulic Conductivity

Tables 1.5 Hydraulic Conductivity for Unconsolidated and Hard Rocks

Medium	**K (m/day)**
Unconsolidated deposits	
Clay	$10^{-8} – 10^{-2}$
Fine sand	1 - 5
Medium sand	5 - 20
Coarse sand	$20 - 10^{2}$
Gravel	$10^{2} - 10^{3}$
Sand and gravel mixes	$5 - 10^{2}$
Clay, sand, gravel mixes (e.g. till)	$10^{-3} – 10^{-1}$
Hard Rocks	
Chalk (very variable according to fissures if not soft)	30.0
Sandstone	3.1
Limestone	0.94
Dolomite	0.001
Granite, weathered	1.4
Schist	0.2

2 [Kozeny-Carmen]

$$K = \frac{\rho g}{\mu}.\left(\frac{n^3}{(1-n)^2}\right).\left(\frac{d_m^2}{180}\right) \quad \textbf{(1.11)}$$

where, n is porosity,
d_m is representative of grain size (L, m).

3 [Shepherd] – Empirically derived

$$K = c\,d^{1.65\,to\,1.85} \quad \textbf{(1.12)}$$

where c and d exponent values vary with material description

1.7.5 Transmissivity (T)

Transmissivity (T) is the discharge rate at which water is transmitted through a unit width of an aquifer under a unit hydraulic gradient. Thus,

$$\begin{aligned} T &= Kb \quad [confined\ aquifer] \\ T &= Kh \quad [unconfined\ aquifer] \end{aligned} \tag{1.13}$$

where, b is the saturated thickness of the aquifer. b is equal to the depth of a confined aquifer. It is equal to the **average** thickness of the saturated zone of an unconfined aquifer.

- Transmissibility is usually expressed as m^2/s, or $m^3/day/m$ or l/day/m.
- Transmissibility of most formations lies between $1*10^4$ -$1*10^6$ l/d/m, with an average value of $1*10^5$ l/d/m.
- **Figure 1.19** illustrates the concepts of hydraulic conductivity and transmissivity.

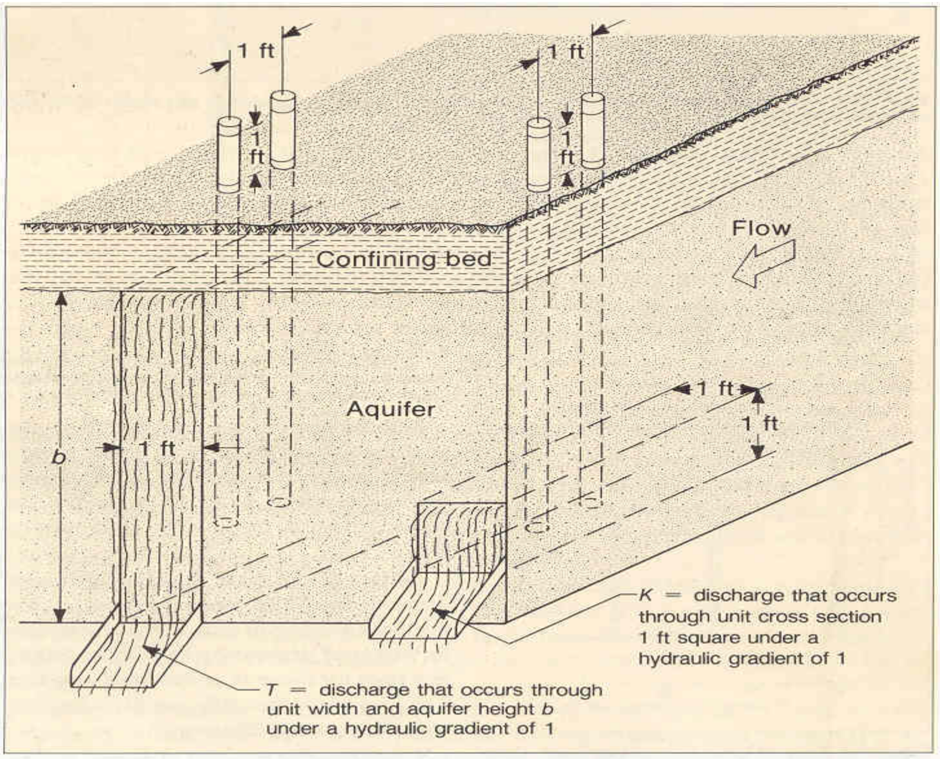


Figure 1.19 Illustration of the Coefficients of hydraulic conductivity and transmissivity. Hydraulic conductivity multiplied by the aquifer thickness equals coefficient of transmissivity.

Table 1.8 Classification of Transmissivity

Magnitude (m^2/day)	Class	Designation	Specific Capacity (m^2/day)	Groundwater supply potential	Expected Q (m^3/day) if s=5m
> 1000	I	Very high	> 864	Regional Importance	> 4320
100-1000	II	High	86.4 – 864	Lesser regional importance	432 – 4320
10-100	III	Intermediate	8.64 – 86.4	Local water	43.2 – 432

				supply	
1-10	IV	Low	0.864 – 8.64	Private consumption	4.32 – 43.2
0.1-1	V	Very low	0.0864 – 0.864	Limited consumption	0.423 – 4.32
<0.1	VI	Imperceptible	< 0.0864	Very difficult to utilize for local water supply	< 0.432

1.7.6 Specific Storage (S_s)

Specific Storage (S_s) is the amount of water per unit volume of a saturated formation that is stored or expelled from storage owing to compressibility of the mineral skeleton and the pore water per unit change in head. This is also called the *elastic storage coefficient.* The concept can be applied to both aquifers and confining units.

The specific storage is given by the expression (Jacob 1940, 1950; cooper 1966):

$$S_s = \rho_w g(\alpha + n\beta) \qquad \textbf{(1.14)}$$

where
- ρ_w is the density of the water (M/L^3; Kg/m^3)
- g is the acceleration of gravity (L/T^2; m/s^2)
- α is the compressibility of the aquifer skeleton (1/(M/LT^2); m^2/N)
- n is the porosity
- β is the compressibility of water (1/(M/LT^2); m^2/N)

The specific storage is usually expressed as cm^{-1} or m^{-1}. For most aquifers, the specific storage is about 3*10^{-7} m^{-1} (see **Table 1.9**).

Table 1.9 Values of Specific Storage Assuming Porosity Equal to 15 % (after Younger, 1993)

Typical Lithologies	Specific Storage (m^{-1})
Clay	9.81 x 10^{-3}
Silt, fine sand	9.82 x 10^{-4}
Medium sand, fine	9.87 x 10^{-5}
Coarse sand, medium gravel, highly fissured	1.05 x 10^{-5}
Coarse gravel, moderately fissured rock	1.63 x 10^{-6}
Unfissured rock	7.46 x 10^{-7}

SOURCE: (Younger, 1993)

In a confined aquifer, the head may decline-yet the potentiometric surface remains above the unit. Although water is released from storage, the aquifer remains saturated. Specific storage (S_s) of a confined aquifer is the storage coefficient per unit-saturated thickness of the aquifer. Thus,

$$S_s = \frac{S}{b} \qquad \textbf{(1.15)}$$

where, **b** is the thickness of aquifer.

1.7.7 Storage Coefficient (S)

Storage coefficient (S) is the volume of water released from storage, or taken into storage, per unit of aquifer storage area per unit change in head.

- The storage coefficient is also called **Storativity**.
- The storage coefficient is a **dimensionless** as it is the ratio of the volume of water released from original unit volume.
- The water-yielding capacity of an aquifer can be expressed in terms of its storage coefficient.
- In unconfined aquifers, Storativity is the same as the specific yield of the aquifer.
- In confined aquifer, Storativity is the result of compression of the aquifer and expansion of the confined water when the head (pressure) is reduced during pumping.

For a vertical column of unit area extending through a confined aquifer, as in **Figure 1.20a**, the storage coefficient equals the volume of water released from the aquifer when the piezometric surface declines a unit distance. In most **confined aquifers**, values fall in the range **$0.00005 < S < 0.005$**, indicating that large pressure changes over extensive areas are required to produce substantial water yields. Storage coefficients can best be determined from pumping tests of wells or from groundwater fluctuation in response to atmospheric pressure or ocean tide variation.

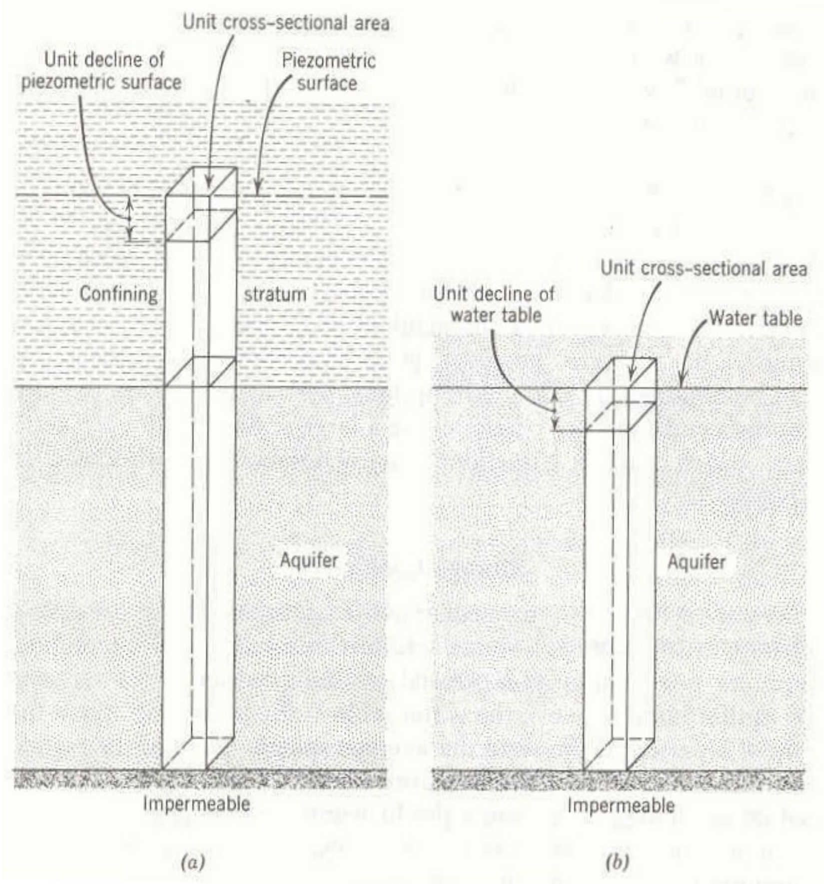


Figure 1.20 Illustrative sketches for defining storage coefficient of (a) confined and (b) unconfined aquifers

Storage coefficient normally varies directly with aquifer thickness

$$S = S_s \,.\, b \quad but \quad S_s = \rho g[\alpha + n\beta]$$
$$\Rightarrow \quad S = \rho g b[\alpha + n\beta] \qquad \textbf{(1.16)}$$

where *b* is the saturated aquifer thickness in meters to be applied for estimating purposes.

The storage coefficient for **unconfined aquifer** corresponds to its **specific yield**, as shown in **Figure 1.20b.**

In an unconfined unit, the level of saturation rises or falls with changes in the amount of water in storage. As the water level falls, water drains from the pore spaces. This storage or release is due to the specific yield (S_y) of the unit. Water is also stored or expelled depending on the specific storage of the unit. For an unconfined unit, the storativity is found by the formula

$$S = S_y + hS_s \qquad \textbf{(1.17)}$$

where h is the thickness of the saturated zone.

The value of S_y is several orders of magnitude greater than hS_s for an unconfined aquifer, and the storativity is usually taken to be equal to the specific yield. For a fine-grained unit, the specific yield may be very small, approaching the same order of magnitude as hS_s . Storativity of unconfined aquifers ranges from **0.02 to 0.30**.

The volume of the water drained from an aquifer as the head is lowered may be found from the formula

$$V_w = S.A.\Delta h \tag{1.18}$$

where V_w is the volume of the water drained (L^3; m^3)
S is the storativity (dimensionless)
A is the surface area overlying the drained aquifer (L^2; m^2)
Δh is the average decline in head (L; m)

The **transmissivity** and **storage coefficients** are especially important because they define the hydraulic characteristics of a water-bearing formation. The coefficient of transmissivity indicates how much water will move through the formation, and the coefficient of storage indicates how much can be removed by pumping or draining. If these two coefficients can be determined for a particular aquifer, predictions of great significance can usually be made. Some of these are:

1 Drawdown in the aquifer at various distances from a pumped well.
2 Drawdown in a well at any time after pumping starts.
3 How multiple wells in a small area will affect one another?
4 Efficiency of the intake portion of the well.
5 Drawdown in the aquifer at various pumping rates.

1.8 Springs

A spring is a concentrated discharge of groundwater appearing at the ground surface as a current of flowing water. To be distinguished from springs are *seepage areas*, which indicate a slower movement pond and evaporate or flow, depending on the magnitude of the seepage, the climate, and the topography.

Springs occur in many forms and have been classified as to cause, rock structure, discharge, temperature, and variability. Springs can be divided into (1) those resulting from mongravitational forces, and (2) those resulting from gravitational forces. Under the former category are included volcanic springs, associated with volcanic rocks, and fissure springs, resulting from fractures extending to great depths in the earth's crust.

Gravity springs result from water flowing under hydrostatic pressure; the following general types are recognized, (see **Figure 1.21**)

1. **Depression Springs** – formed where the ground surface intersects the water table.

2. **Contact Springs** – created by permeable water-bearing formation overlying a less permeable formation that intersects the ground surface.

3. **Artesian Springs** – resulting from releases of water under pressure from confined aquifers either at an outcrop of the aquifer or through an opening in the confining bed.

4. **Impervious Rock Springs** – occurring in tubular channels or fractures of impervious rock.

5. **Tubular or fracture Springs** – issuing from rounded channels, such as lava tubes or solution channels, or fractures in impermeable rock connecting with groundwater.

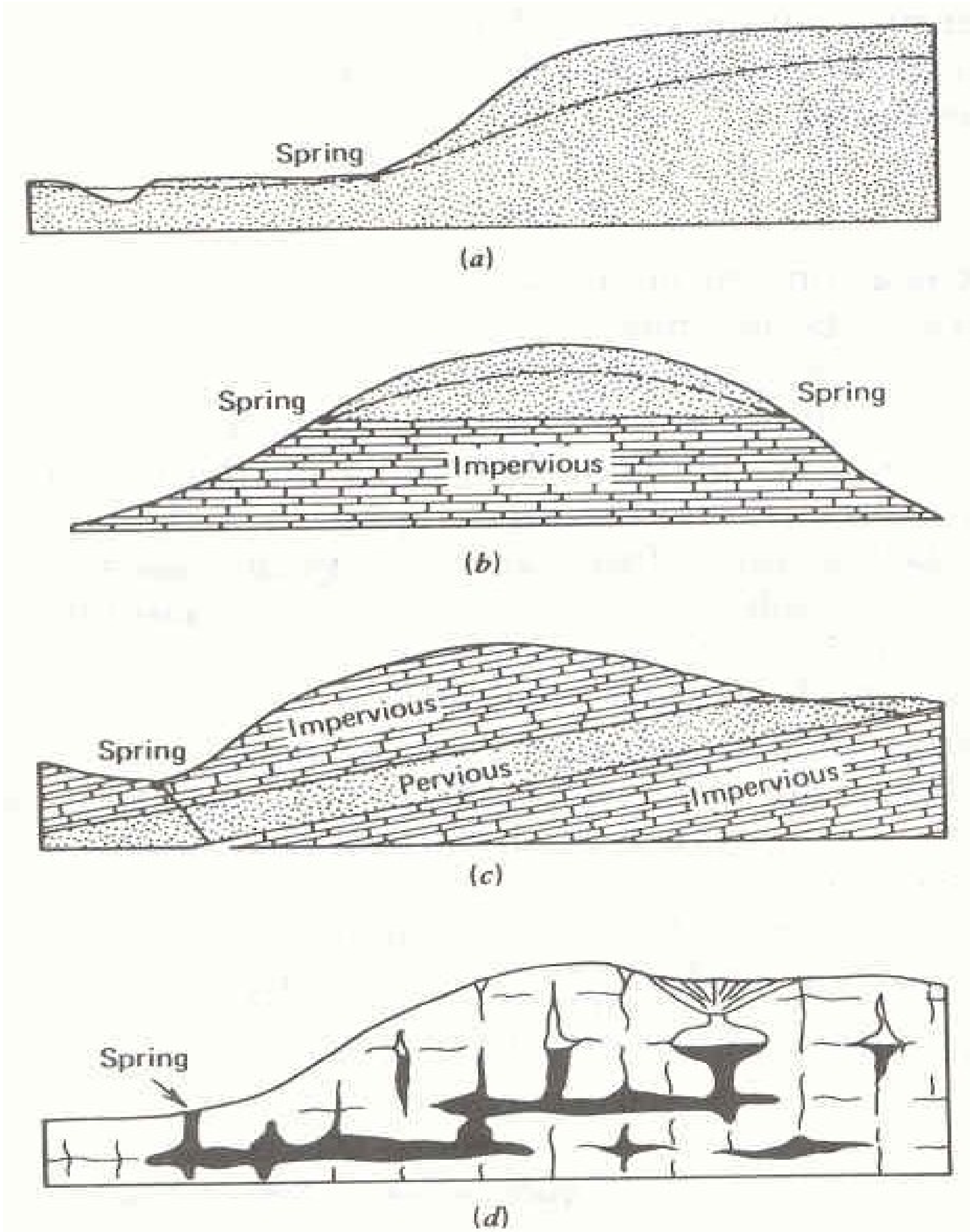


Figure 1.21 Diagrams illustrating types of gravity springs. (a) Depression spring. (b) Contact springs. (c) Fracture artesian spring. (d) Solution tubular spring (after Bryan, 1919).

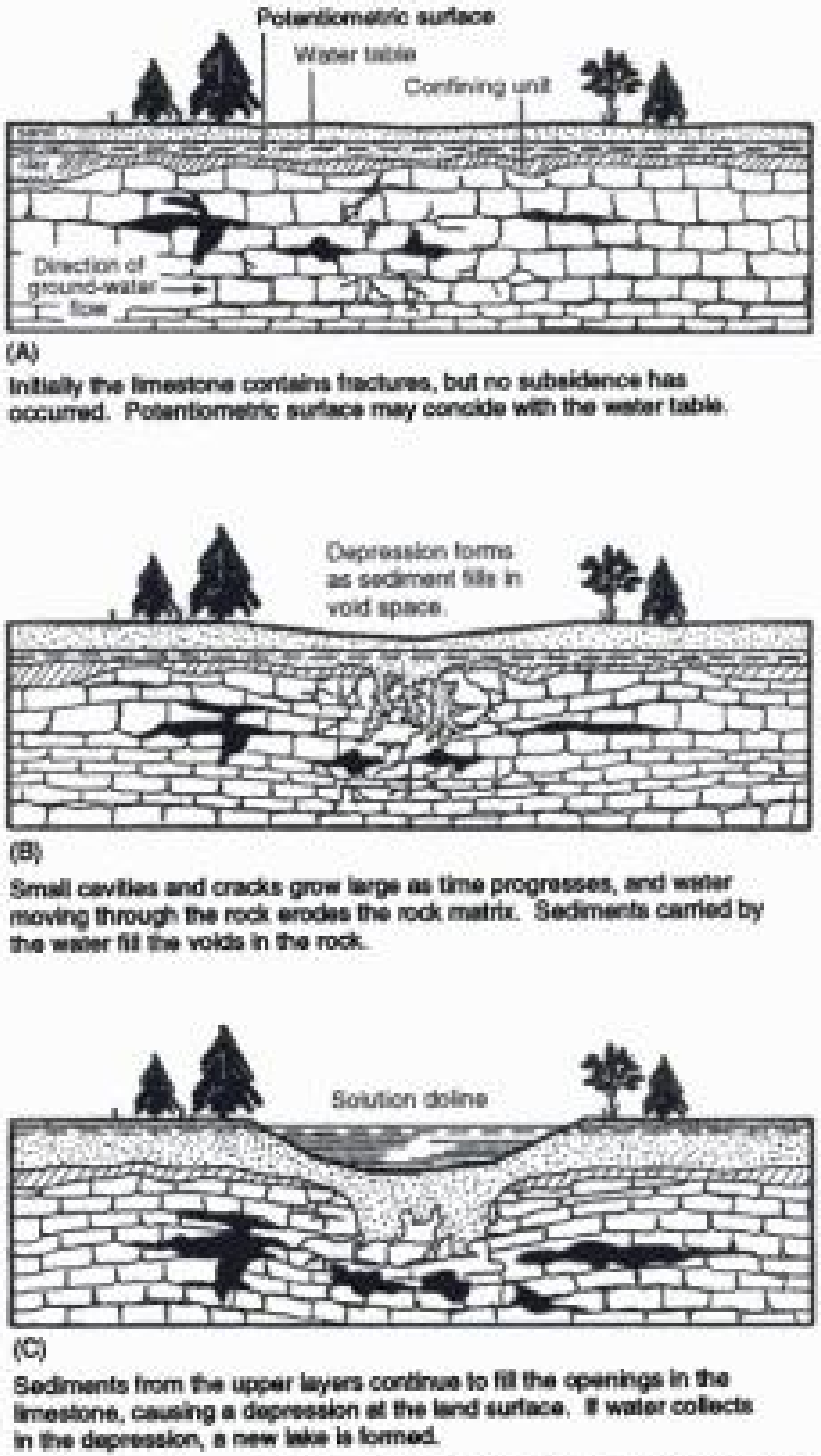
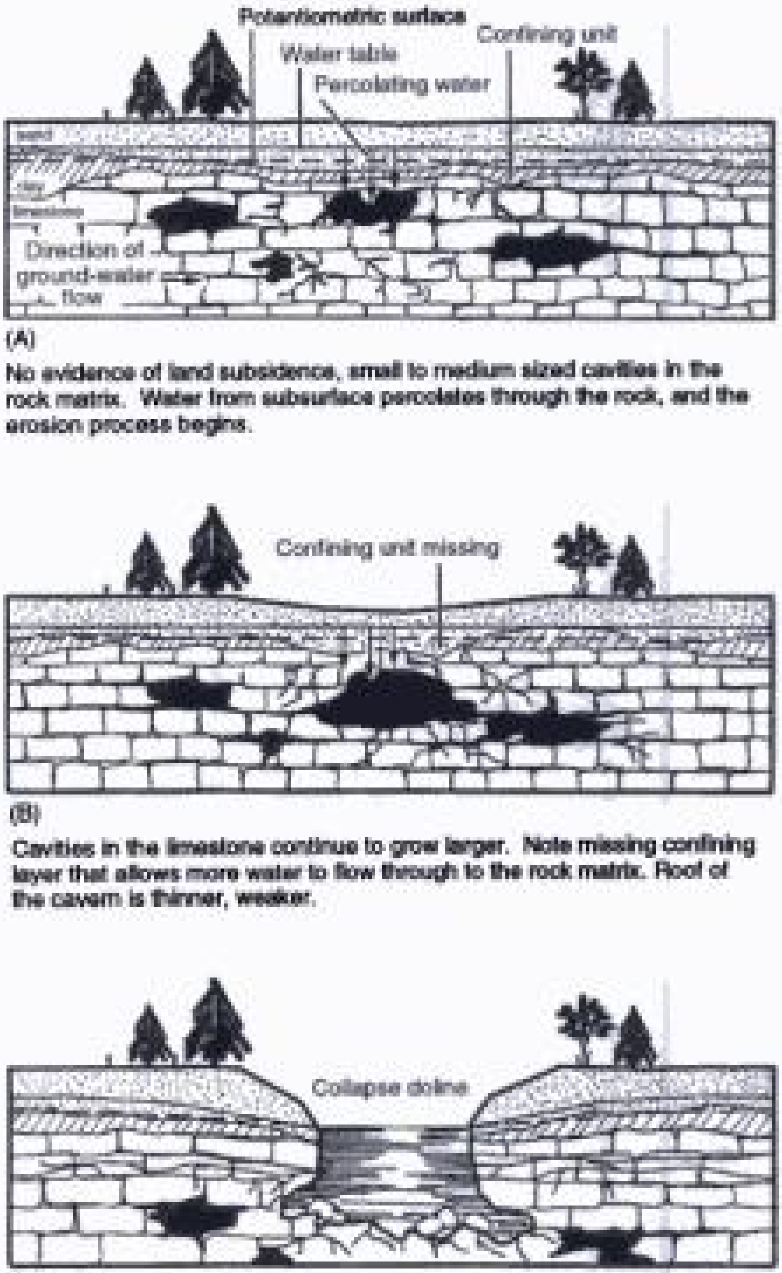


Figure 2.23. **Slow development of a solution doline (sinkhole) (source: USGS Circular 1137).**



(A)
No evidence of land subsidence, small to medium sized cavities in the rock matrix. Water from subsurface percolates through the rock, and the erosion process begins.

(B)
Cavities in the limestone continue to grow larger. Note missing confining layer that allows more water to flow through to the rock matrix. Roof of the cavern is thinner, weaker.

(C)
As ground-water levels drop during the dry season, the weight of the overburden exceeds the strength of the cavern roof, and the overburden collapses into the cavern, forming a sinkhole.

Figure 2.24. Rapid development of a collapse doline (sinkhole) (source: USGS Circular 1137).

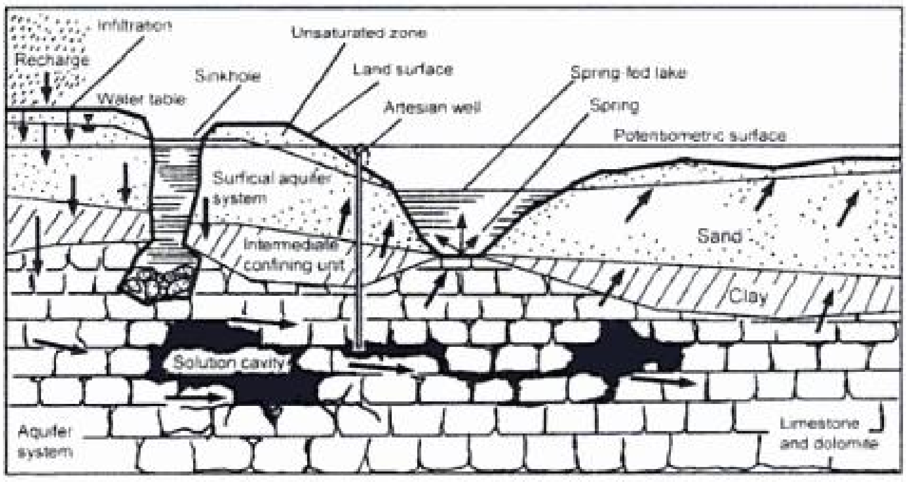
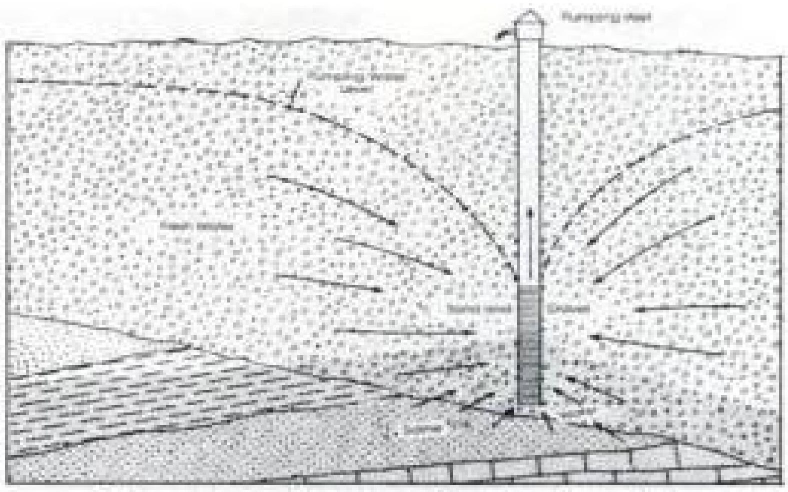
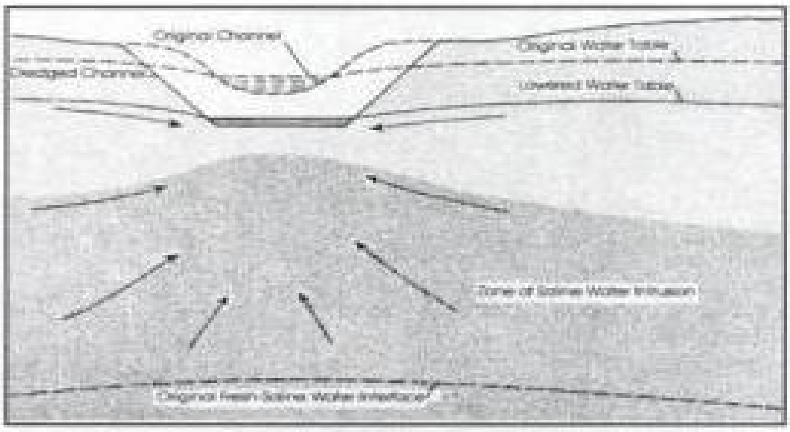


Figure 2.25. Karst ground-water flow system (source: USGS Circular 1137).



Saltwater rising into an aquifer due to pumping.



Migration of saline water caused by lowering of water levels in a gaining stream.

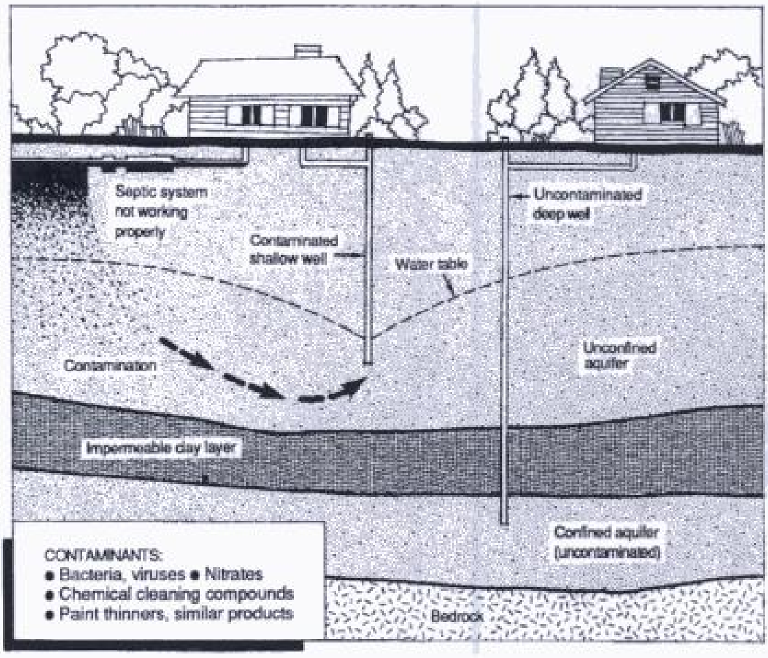
Figure 2.27. Freshwater aquifers contaminated by saline water from underlying rocks (Deutsch, 1963).

SEPTIC SYSTEMS

CONTAMINATION EVIDENCE:
- Wastewater shows above ground
- Detection of excessive bacteria, chemicals in well water tests

CAUSES:
- Poor installation and/or maintenance
- Disposal of household chemicals, such as paint thinners, into the system
- Overloading the system with a garbage disposal unit
- Use of septic tank cleaning additives
- Too many closely-spaced septic systems in a limited area

PREVENTION:
- Proper installation
- Inspection and cleaning every 2-4 years, annually if garbage disposal unit is used
- Do not dispose of household chemicals into the system
- Ban hazardous cleaning additives for septic systems
- Develop local septic system codes
- Public sewers when feasible
- Public information/education



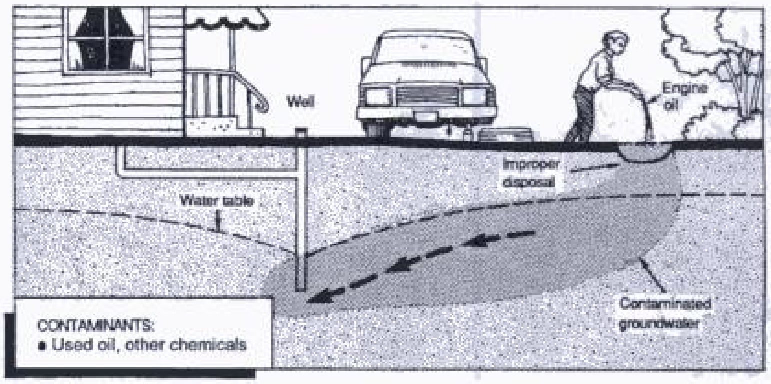
SMALL DISPOSAL PITS

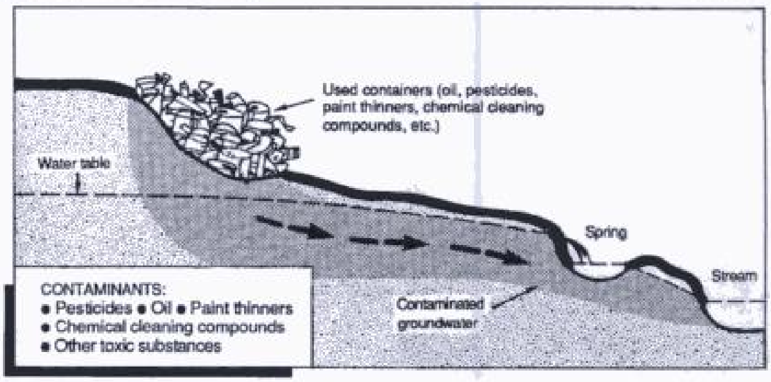
Used for dumping or burning wastes by businesses and households

CONTAMINATION EVIDENCE:
- Petroleum odor in well water
- Other chemical odors
- Detection of chemicals in well water tests

CAUSES:
- Improper disposal of chemicals, oil, pesticides, other wastes and used containers
- Lack of disposal facilities for small amounts of hazardous wastes

PREVENTION:
- Public information/education
- Disposal facilities for small hazardous wastes generators
- Enforcement against improper waste disposal



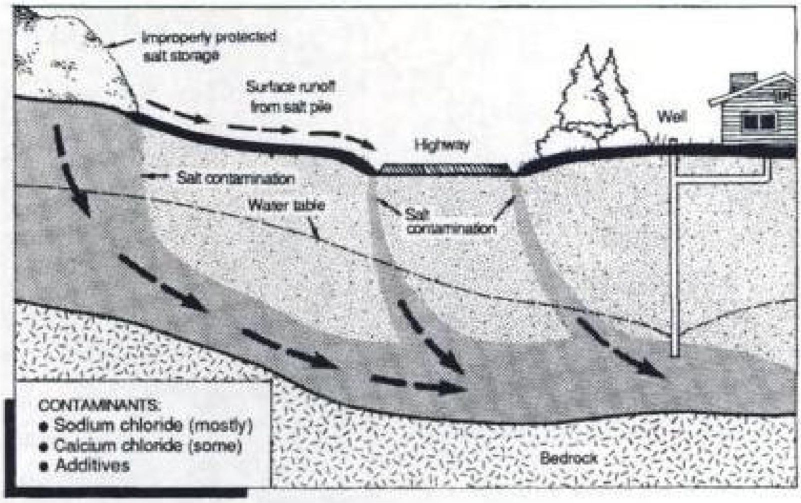


DEICING SALTS

CONTAMINATION EVIDENCE:
- **Salty taste in well water**
- **High chloride level in well water tests**

CAUSES:
- **Runoff from salt storage piles and highways**

PREVENTION:
- **Proper protection of salt storage piles**
- **Minimize use**
- **Use alternative deicing materials**



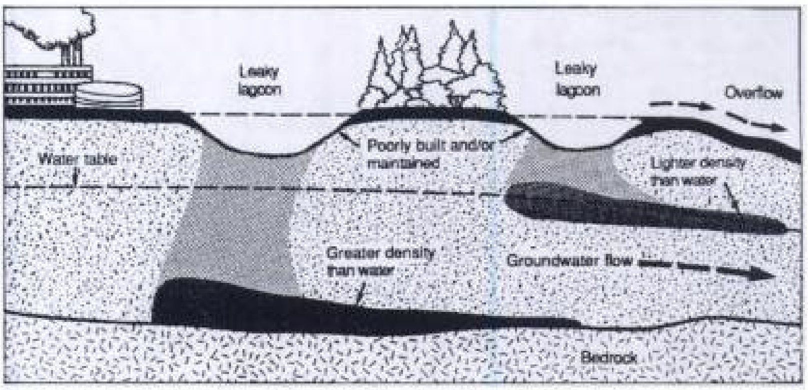
STORAGE LAGOONS

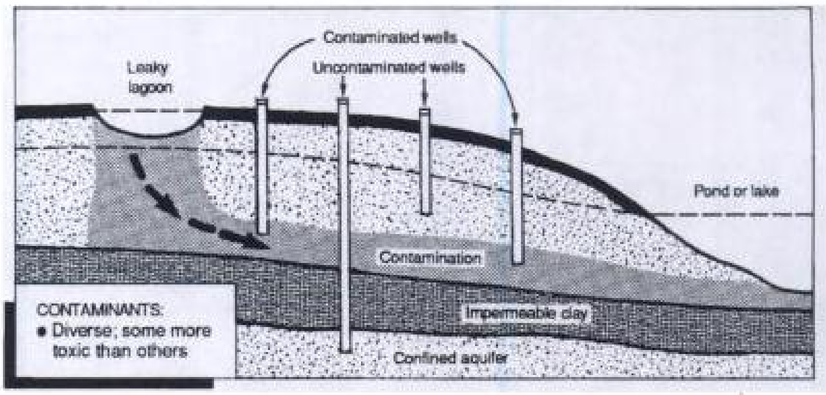
Used by industries, farms, municipalities, mining operations, oil/gas producers

CONTAMINATION EVIDENCE:
- Spills
- Changes in color, taste, odor of well water
- Unhealthy or dead vegetation near lagoon
- Greener and more vigorous plant growth near lagoon
- Detection of excessive bacteria, chemicals in well water tests

CAUSES:
- Poor installation and maintenance
- Overflows
- Seepage
- Liner failure
- Structural collapse
- Location in sensitive groundwater area

PREVENTION:
- Proper installation and maintenance
- Locate away from sensitive groundwater areas



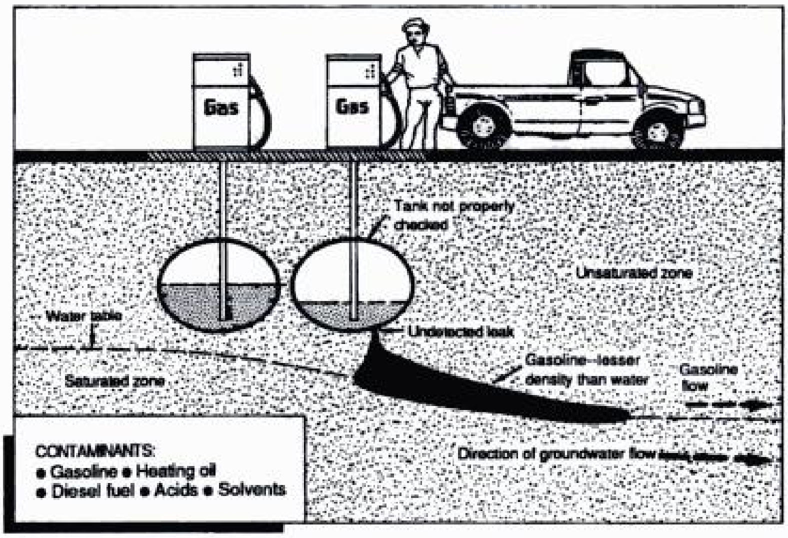


UNDERGROUND STORAGE TANKS

CONTAMINATION EVIDENCE:
- Petroleum odor in wells or basements
- Tank inventory losses
- Spills
- Detection of leaks

CAUSES:
- Corroded tanks
- Poor installation and/or maintenance
- No testing for tank leaks
- Poor inventory control
- No leak backup containment
- Deterioration of abandoned tanks

PREVENTION:
- Proper installation, maintenance, leak testing and inventory control
- Permit compliance
- Leak backup containment
- Removal of abandoned tanks or filling with inert material

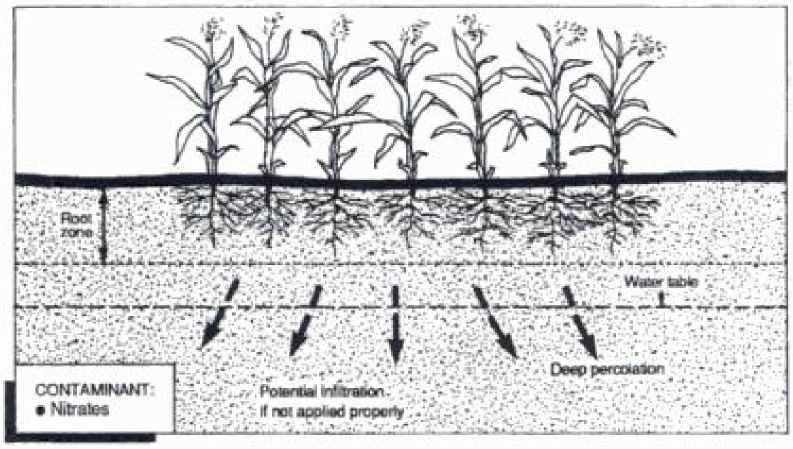


FERTILIZERS

CONTAMINATION EVIDENCE:
- High nitrate level in well water tests

CAUSES:
- Overfertilization
- Ill-timed application

PREVENTION:
- Careful adjustment of fertilizer application to plant needs and timing for maximum growth benefit
- Storage of animal manure to facilitate land spreading at appropriate times

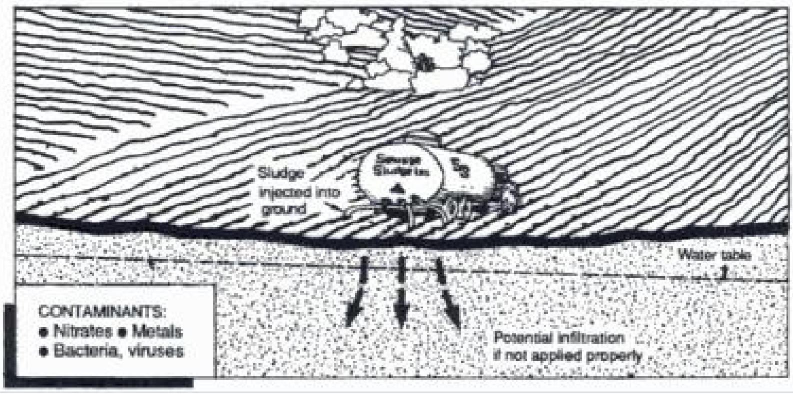


LAND APPLICATION
Sludges and Wastewater

CONTAMINATION EVIDENCE:
- High bacteria, nitrate levels in well water tests

CAUSES:
- Improper application methods
- Inappropriate soils for application

PREVENTION:
- Compliance with permit requirements

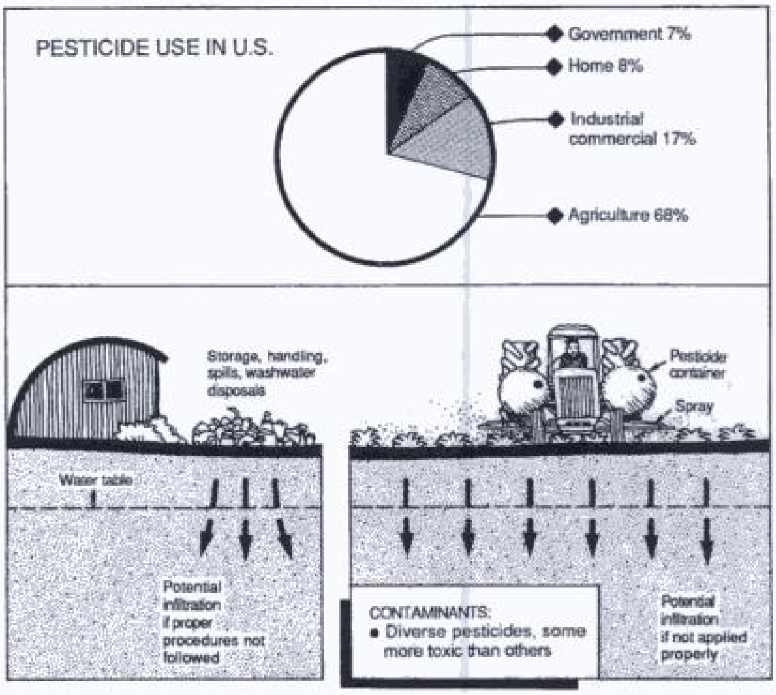


PESTICIDES

CONTAMINATION EVIDENCE:
- Detection of pesticides in well water tests
- Ill effects on animals drinking water from nearby wells, springs or surface water
- Ill effects on plants watered with nearby well water
- Ill effects on aquatic life

CAUSES:
- Excessive or ill-timed application
- Improper storage
- Leaching through the soil
- Improper disposal of excess pesticides and rinsewater

PREVENTION:
- Follow use instructions
- Compliance with pesticide certification requirements
- Reduce pesticide use in recharge areas for water wells
- Encourage alternative pest control methods
- Public information/education

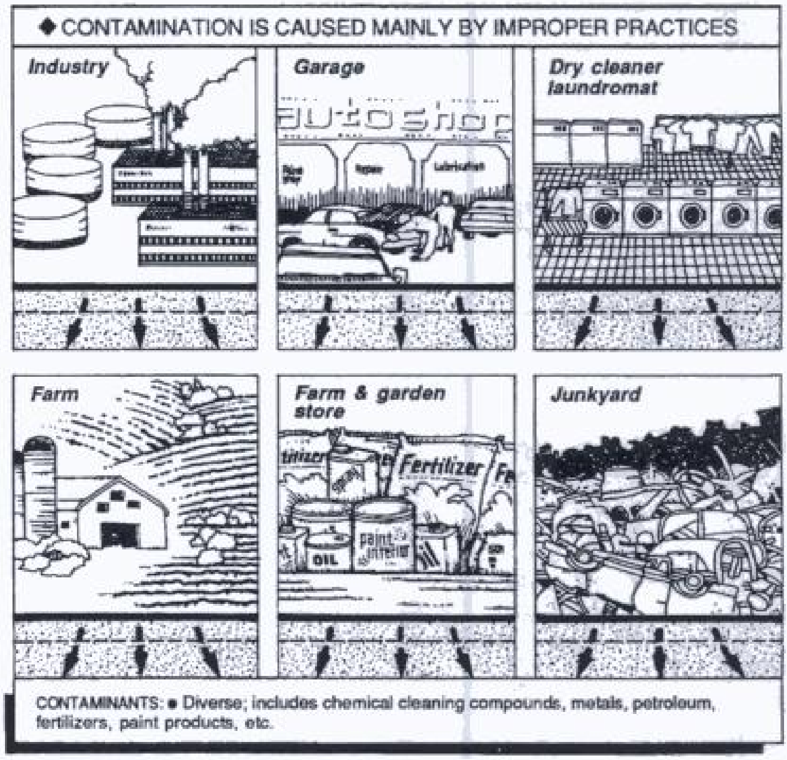


HAZARDOUS MATERIALS

CONTAMINATION EVIDENCE:
- Spills
- Detection of chemical solvents, metals, nitrates, other chemicals in well water tests

CAUSES:
- Improper storage, handling, use, and disposal
- Spills
- Leaks

PREVENTION:
- Proper storage, handling, use and disposal
- Spill prevention and containment measures
- Compliance with laws and regulations
- Zoning to locate heavy users of hazardous materials away from sensitive groundwater areas
- Public information/education



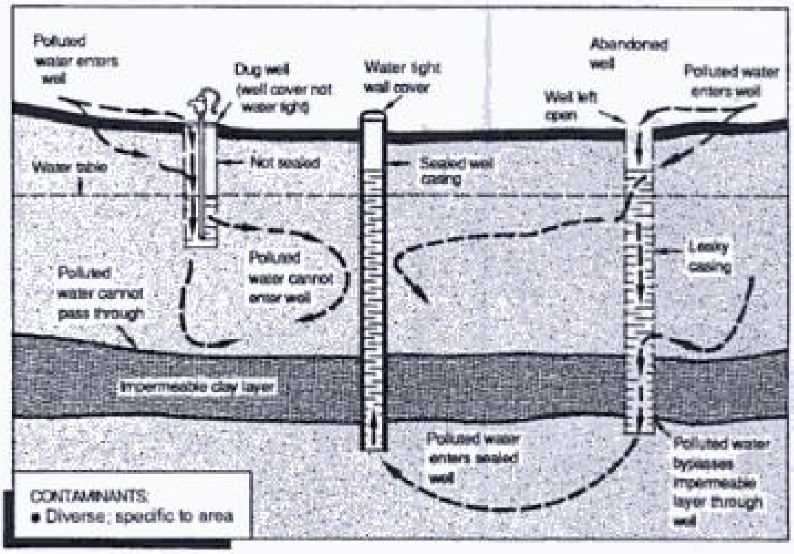
WELLS

Wells are potential pathways for contaminants to enter groundwater

CONTAMINATION EVIDENCE:
- Detection of high bacteria levels in well water tests
- Well water turbidity
- Detection of other contaminants in well water

CAUSES:
- No well casing or leaky casing
- Well cover not watertight
- Open abandoned wells
- Groundwater movement from contaminated to uncontaminated wells

PREVENTION:
- Watertight well cover
- Tight well casing
- Tight plumbing connections
- Identify and seal open abandoned wells

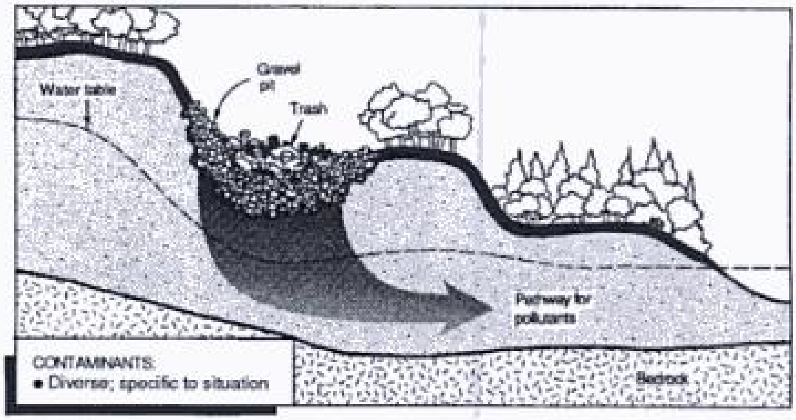


INACTIVE MINING SITES

CONTAMINATION EVIDENCE:
- (Potential) Dumping of wastes in inactive mining pits

CAUSES:
- Rapid infiltration of contaminants due to loss of topsoil filtering capacity

PREVENTION:
- Close unused mining pits by restoring topsoil cover
- Vigilance against waste dumping in inactive mining pits

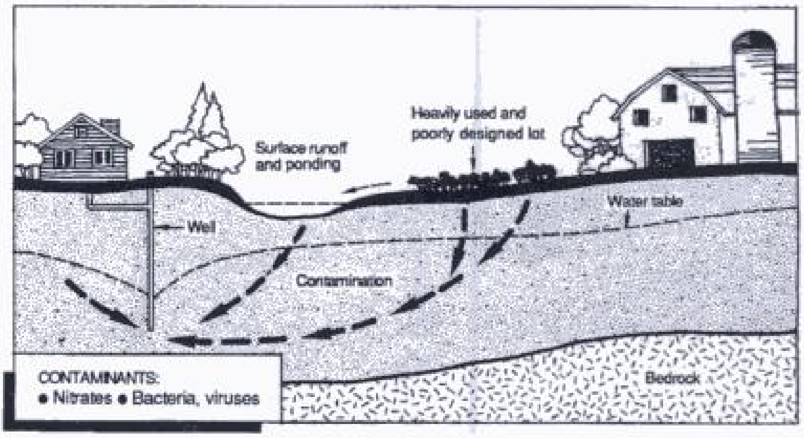


ANIMAL LOTS

CONTAMINATION EVIDENCE:
- High bacteria, nitrate levels in well water

CAUSES:
- High animal density
- Shallow depth to water table
- Poor lot drainage
- Failure to regularly clean lot

PREVENTION:
- Proper siting and design
- Control animal density
- Regular cleaning of lot

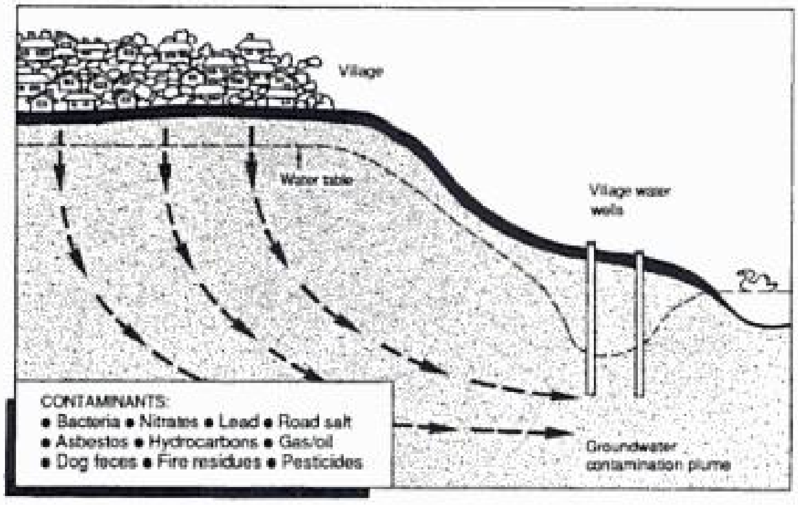


URBAN RUNOFF

CONTAMINATION EVIDENCE:
- Detection of chemicals, metals, nitrates, petroleum, etc. in well water

CAUSES:
- Spills
- Random waste disposal
- Abandoned commercial/industrial sites
- Motor vehicle emissions
- Fires

PREVENTION:
- Public information/education
- Street sweeping
- Anti-dumping codes
- Vegetated collection and infiltration basins for street runoff
- Clean up abandoned commercial/industrial sites
- Proper cleanup of fire sites

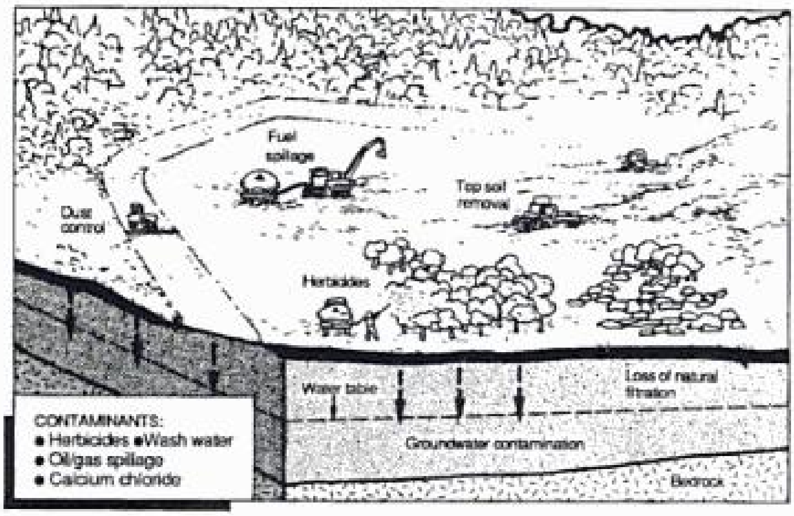


CONSTRUCTION EXCAVATION

CONTAMINATION EVIDENCE:
- Spills
- Changes in color, taste, odor, turbidity of water in nearby wells

CAUSES:
- Fuel, chemical spills
- Road dust control runoff
- Excessive and/or improper use of chemicals

PREVENTION:
- Spill containment and cleanup procedures
- Follow recommended practices for safe use of fuels and other hazardous substances

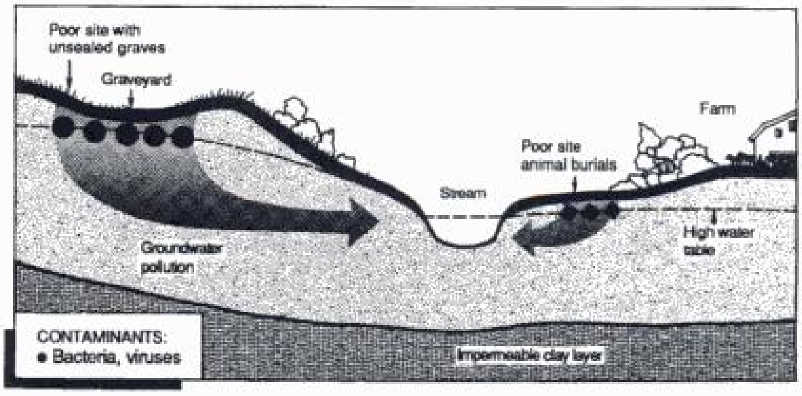


CEMETERIES and ANIMAL BURIALS

CONTAMINATION EVIDENCE:
- Detection of high bacteria levels in nearby well water tests

CAUSES:
- High water table

PREVENTION:
- Avoid high water tables for burial sites
- Use watertight caskets in cemeteries with high water tables

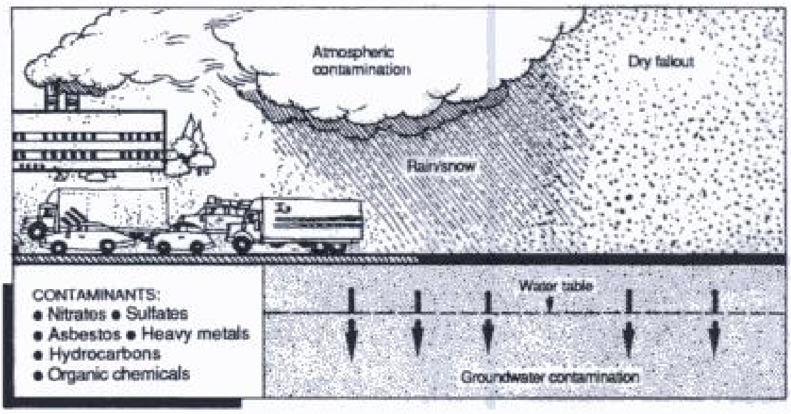


ATMOSPHERIC POLLUTANTS

CONTAMINATION EVIDENCE:
- Detection of elevated levels of sulfates, nitrates, heavy metals, asbestos, hydrocarbons, other chemical compounds in well water tests

CAUSES:
- Emissions from motor vehicles, power plants, industries

PREVENTION:
- Federal and state emission controls

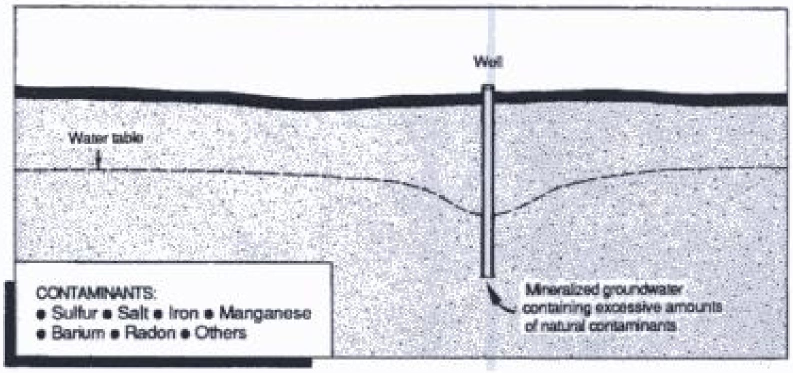


NATURAL SUBSTANCES

CONTAMINATION EVIDENCE:
- Bad taste or odor in well water
- Stains on water fixtures
- Detection in well water tests

CAUSES:
- Natural origin

PREVENTION:
- Avoid areas where natural groundwater problems exist, if feasible
- Use water treatment devices
- Change to public water supply, if feasible



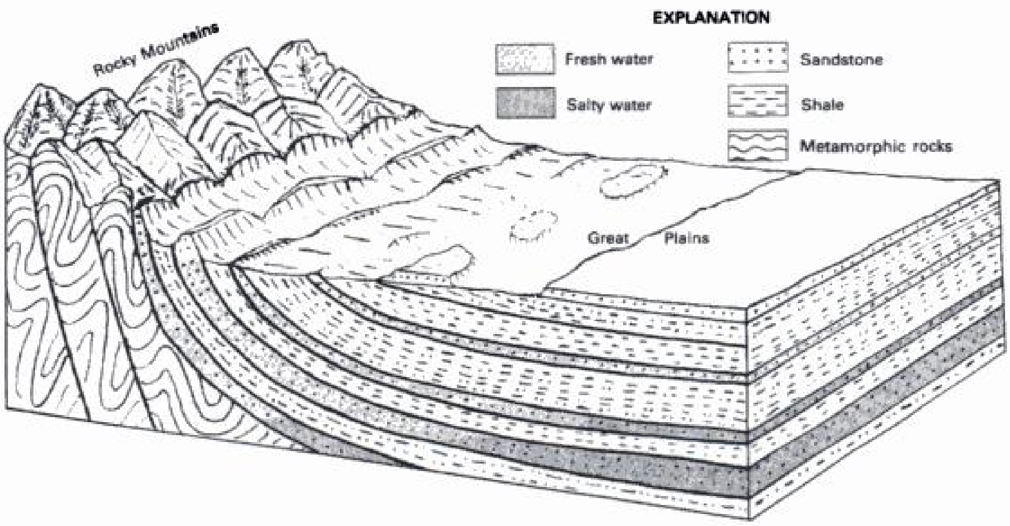


Figure 3.1. Confined aquifer where sedimentary rocks are uplifted (Heath, 1984).

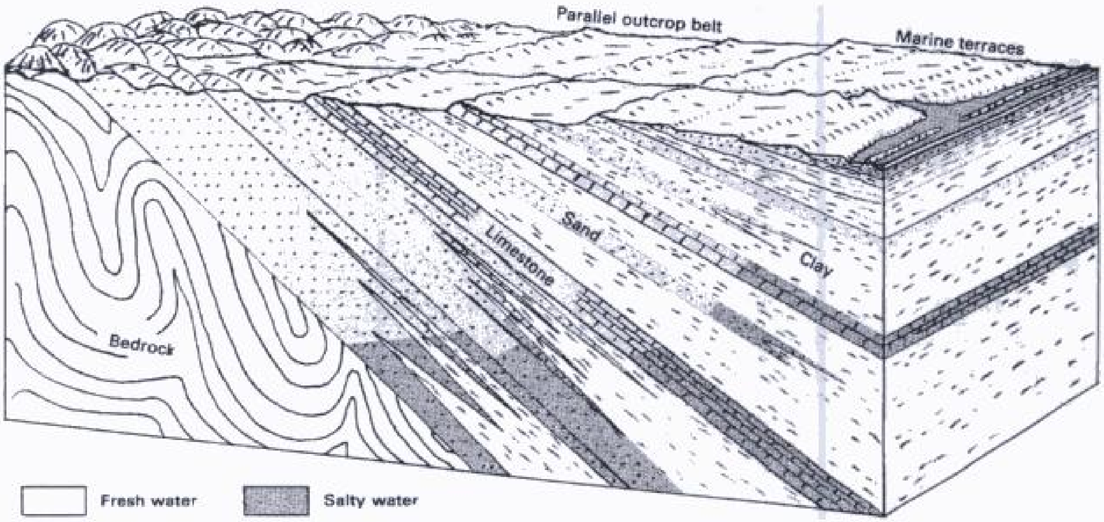


Figure 3.2. Confined aquifer development on a regional dip (Heath, 1984).

Figure 3.3. Alluvial fan development.

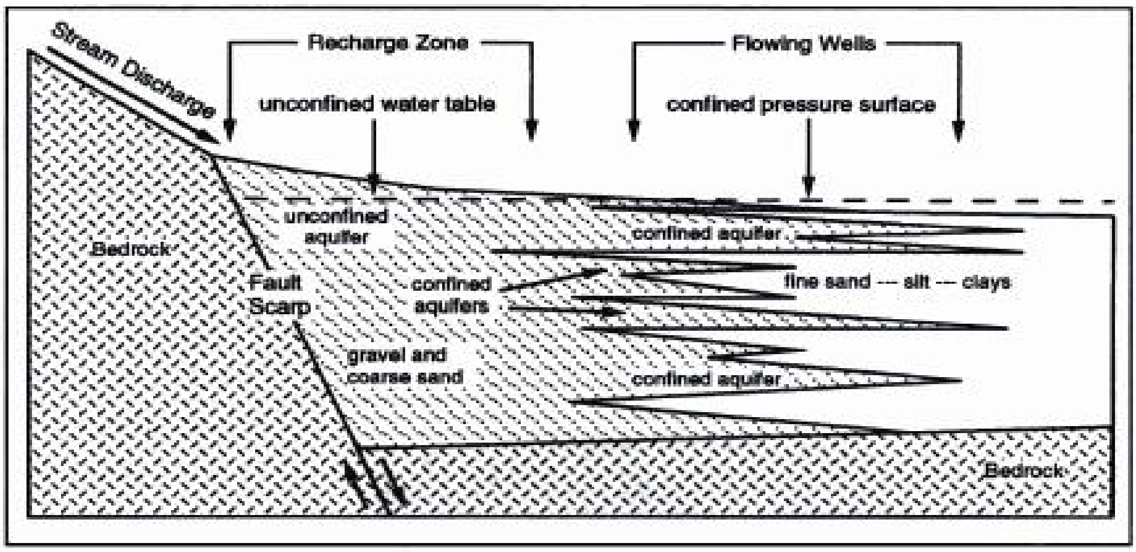
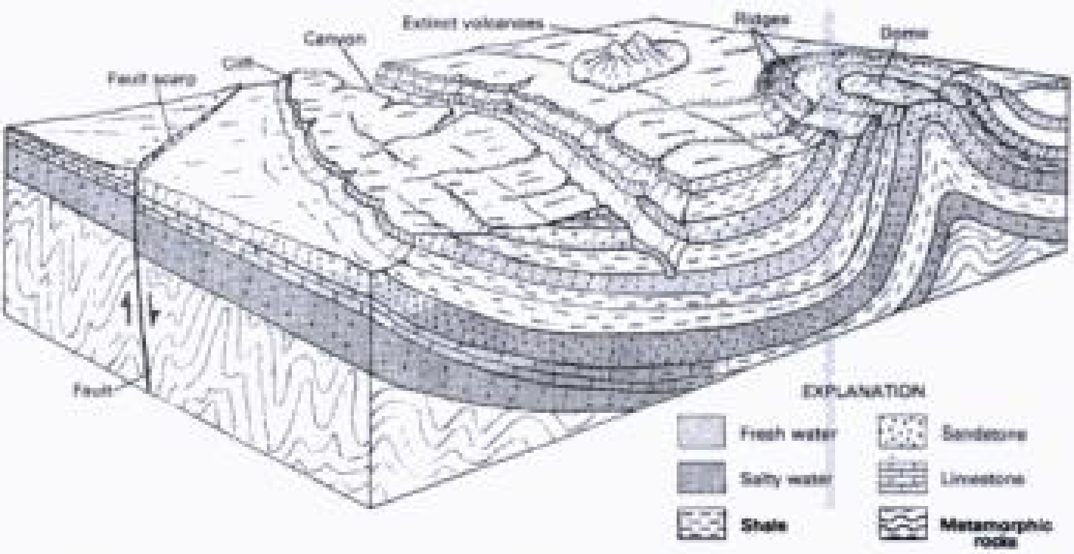
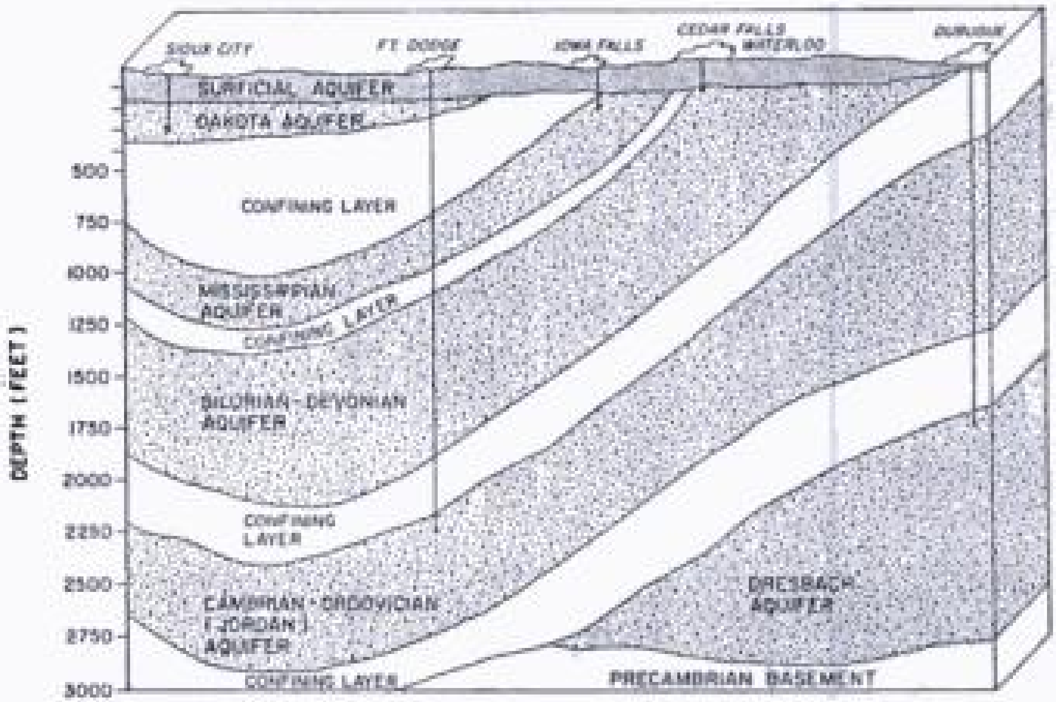


Figure 3.4. Development of confined aquifers in an alluvial valley at the foot of a mountain.



Typical geologic features of the Colorado Plateau and Wyoming Basin (Heath, 1984).



Synclinal aquifers beneath Iowa (Iowa Department of Natural Resources).

Figure 3.5. Development of confined aquifers in folded rock.

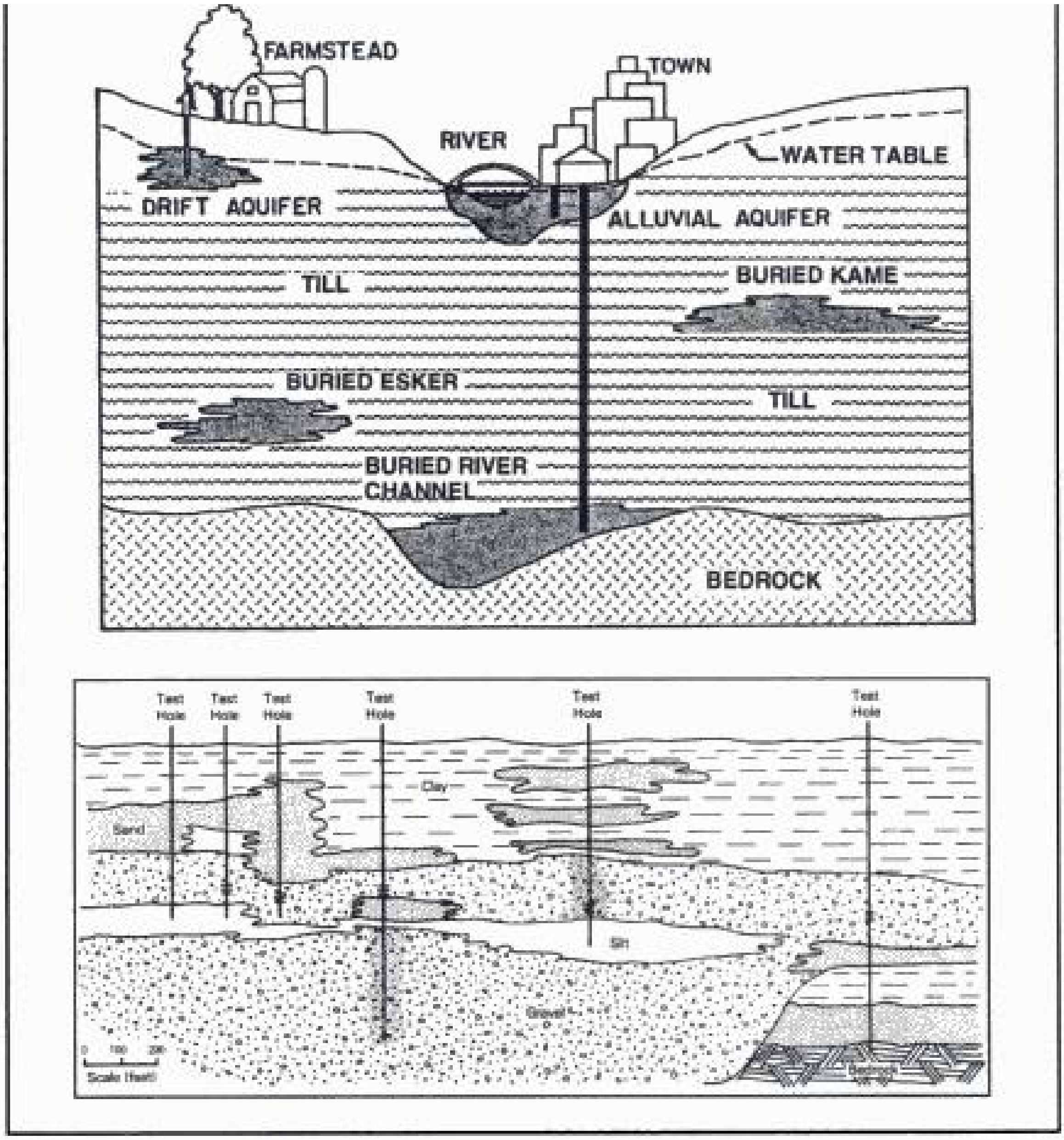


Figure 3.6. Confined aquifers in glacial terrain.

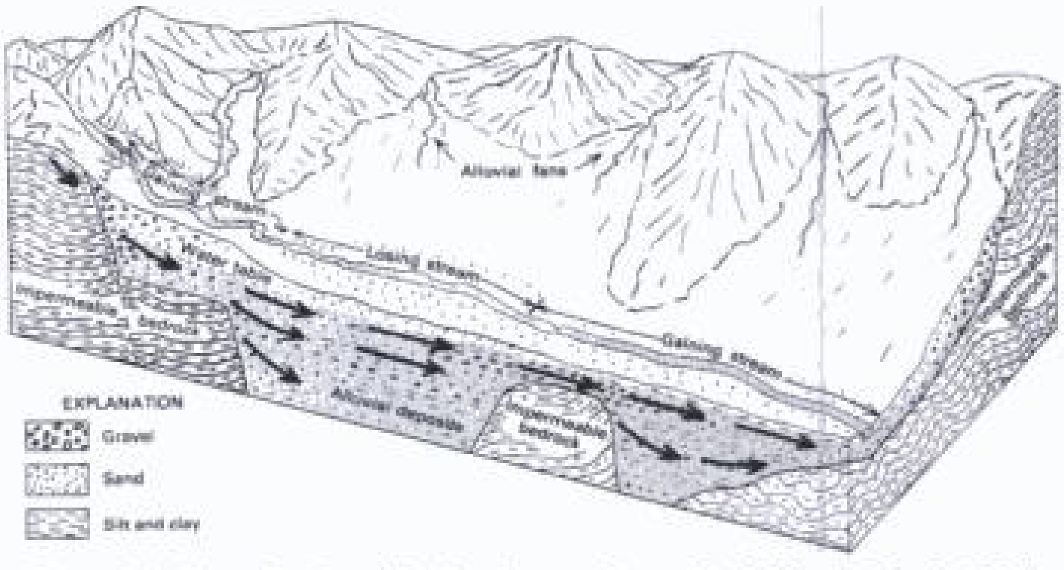


Figure 3.9. Unconfined aquifer development in an alluvial basin (Heath, 1984).

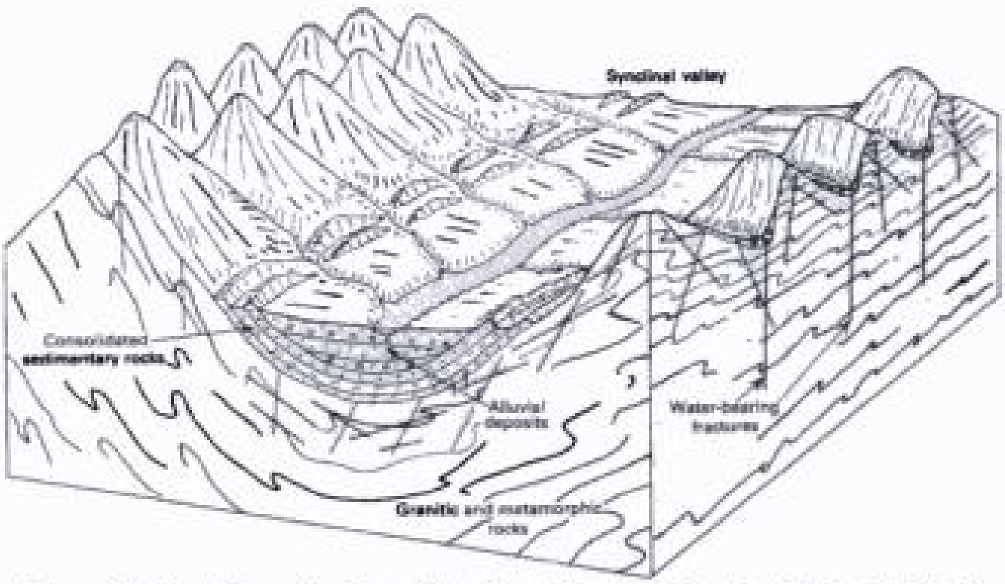
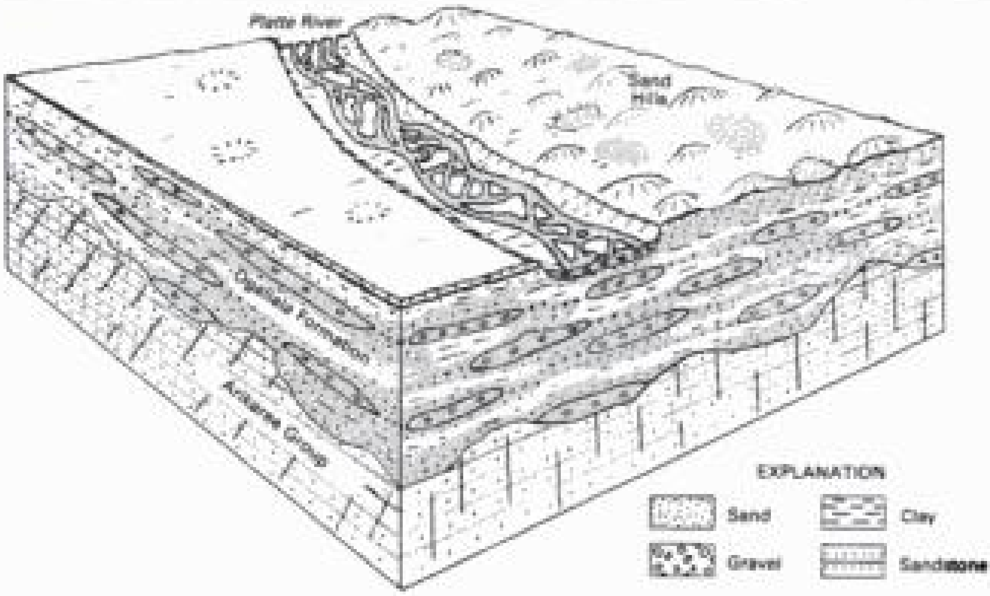


Figure 3.10. Unconfined aquifer above fractured bedrock (Heath, 1984).



Unconfined aquifer development of the High Plains region (Top).
Unconfined aquifers along the flow direction of the Mississippi River (Bottom).

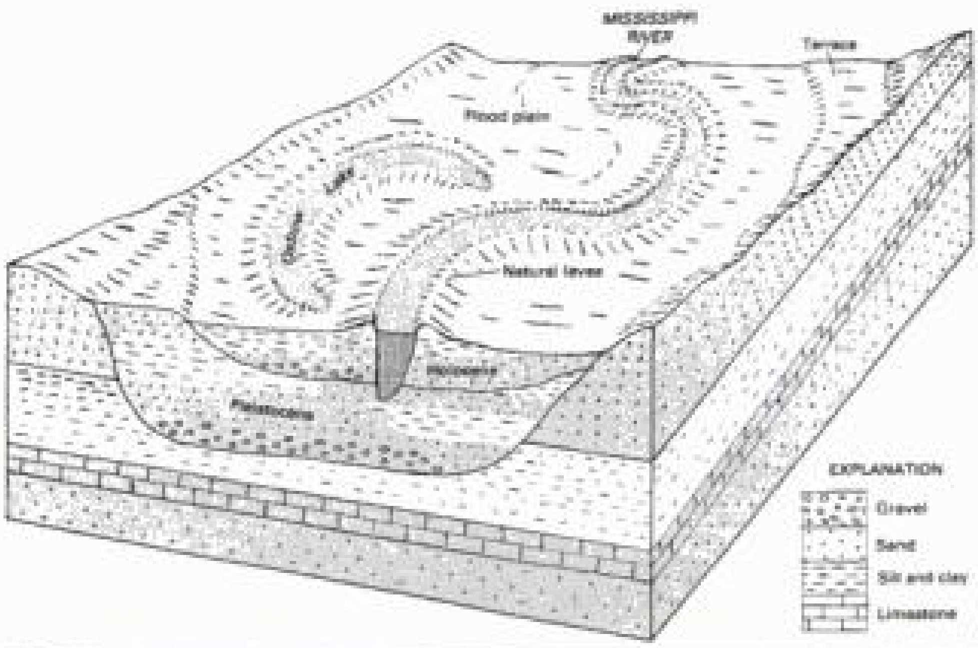


Figure 3.11. Unconfined aquifers in unglaciated river valleys (Heath, 1984).

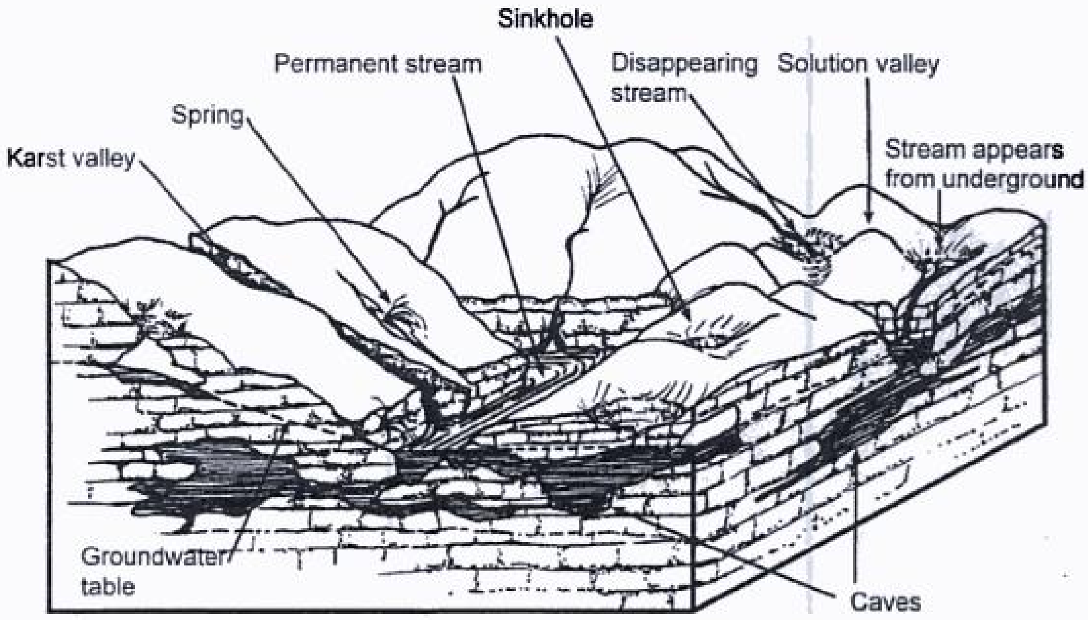


Figure 3.13. Karst system.

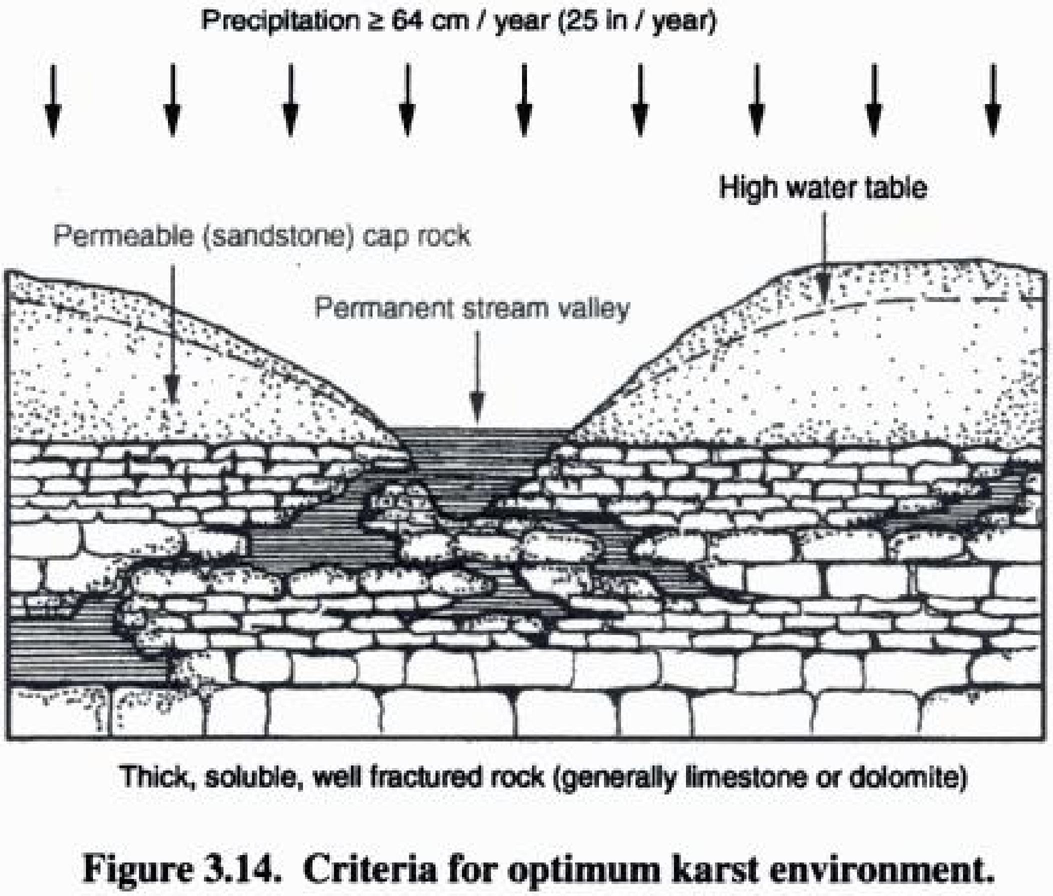


Figure 3.14. Criteria for optimum karst environment.

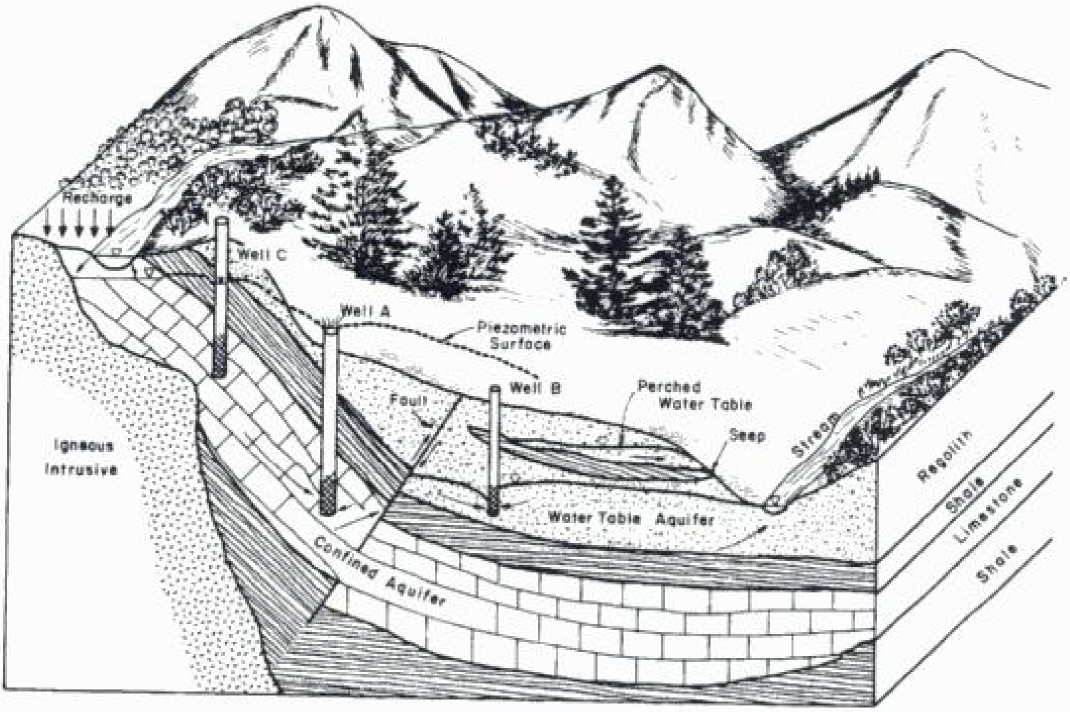


Figure 3.15. Confined aquifer in limestone (McWhorter and Sunada, 1977).

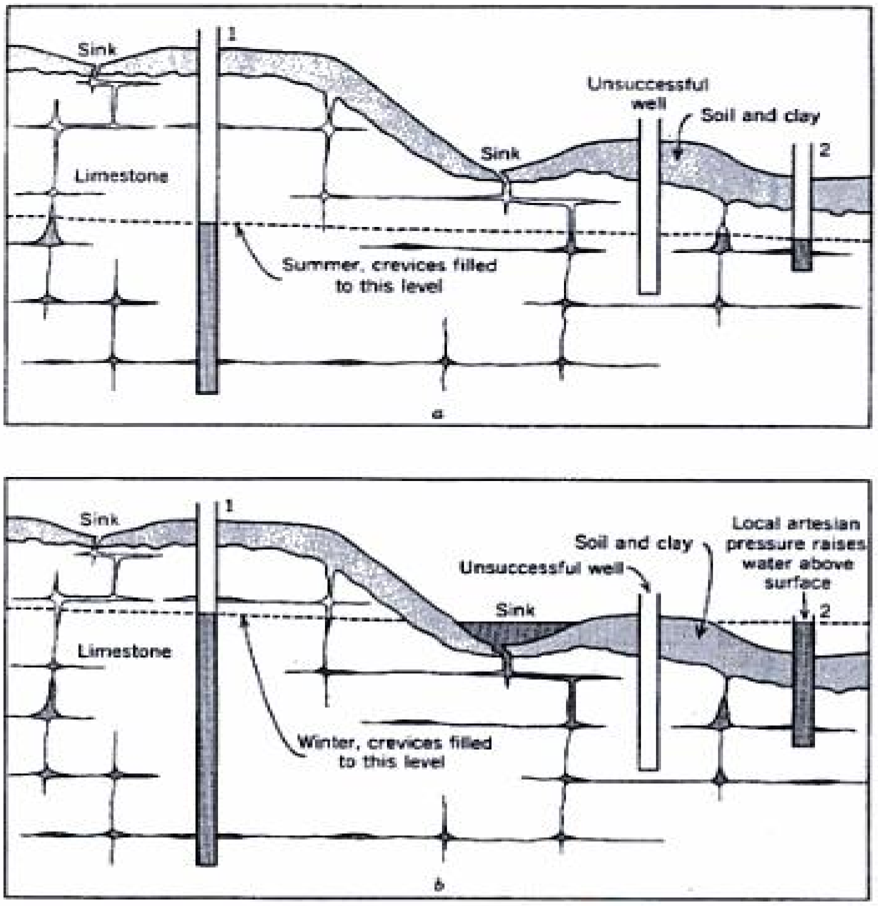


Figure 3.16. Wells constructed in karst fractures (Walker, 1956).

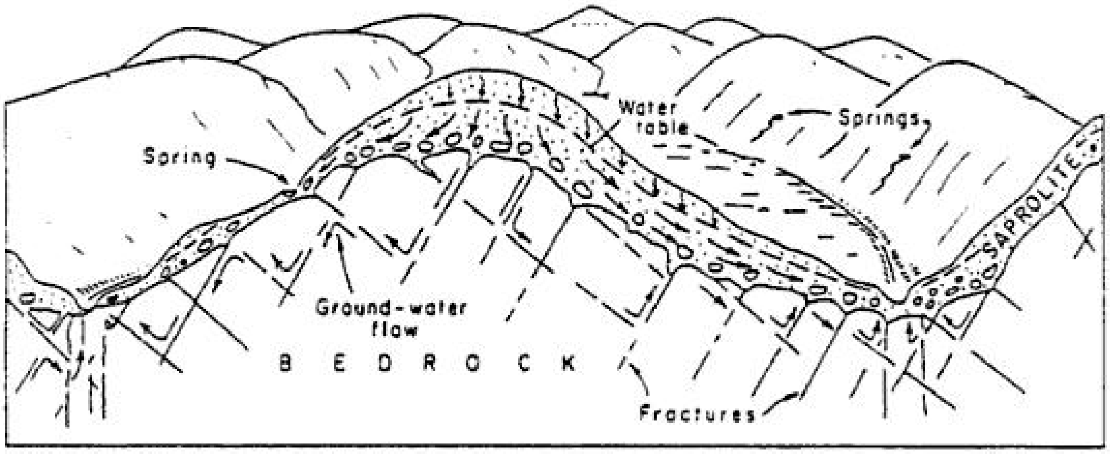
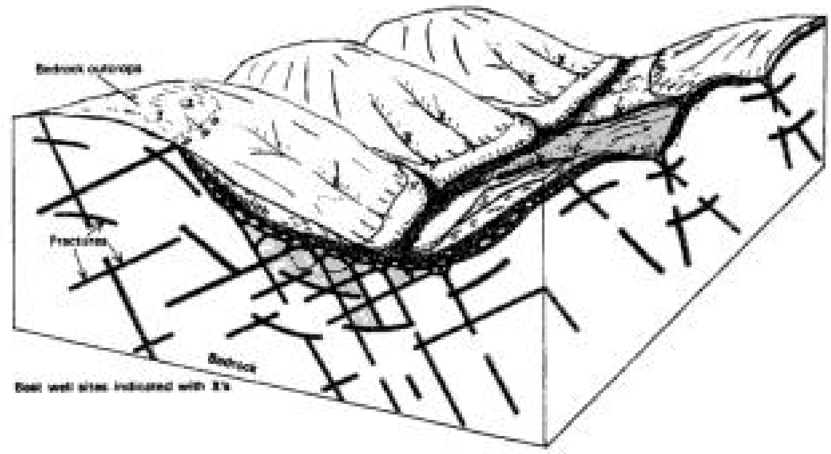
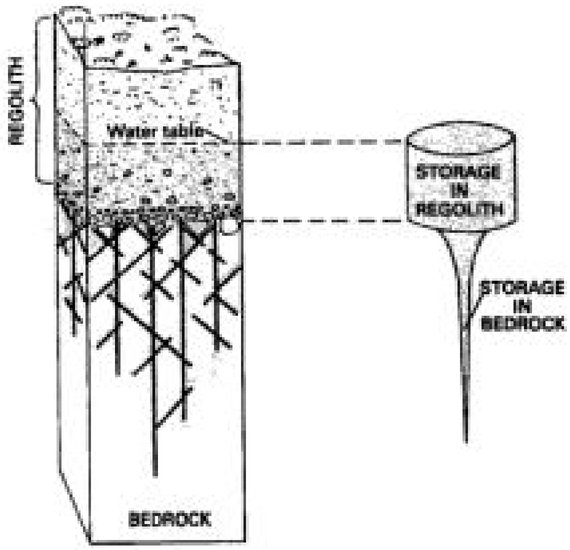


Figure 3.17. Fractures in bedrock acting as "pipelines" (Heath, 1980).

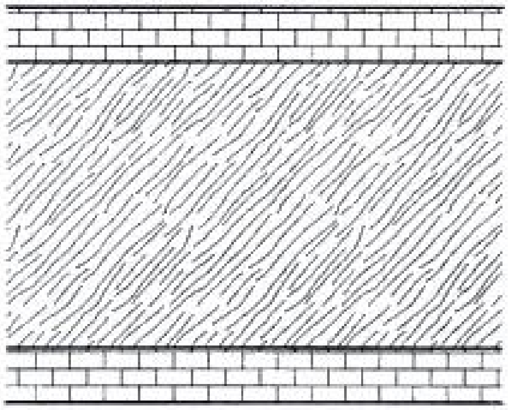


Thin regolith overlying fractured bedrock of the Piedmont Blue Ridge region.

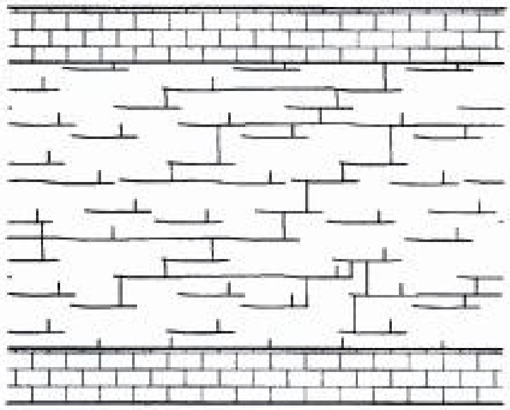


Difference in storage capacity of regolith and bedrock.

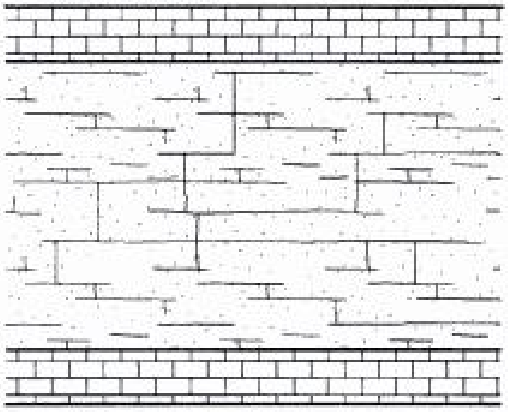
Figure 3.18. Fractured rock aquifer and storage capacity (Heath, 1984).



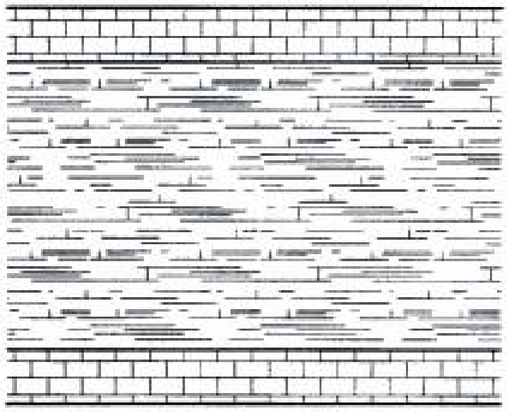
A. Minute multiple fractures closely spaced should produce poor well yields. This is typical of shale and schist.



B. Large multiple fractures closely spaced should produce adequate well yields. This is dependent on the connectivity of the fractures.



C. Multiple fractures in a porous sedimentary rock should produce good well yields. This is called a double-porosity system.



D. Multiples of large, closely spaced, and connected fractures, should produce good, sustained well yields.

Figure 3.19. Fractures in rock relative to ground-water supply.

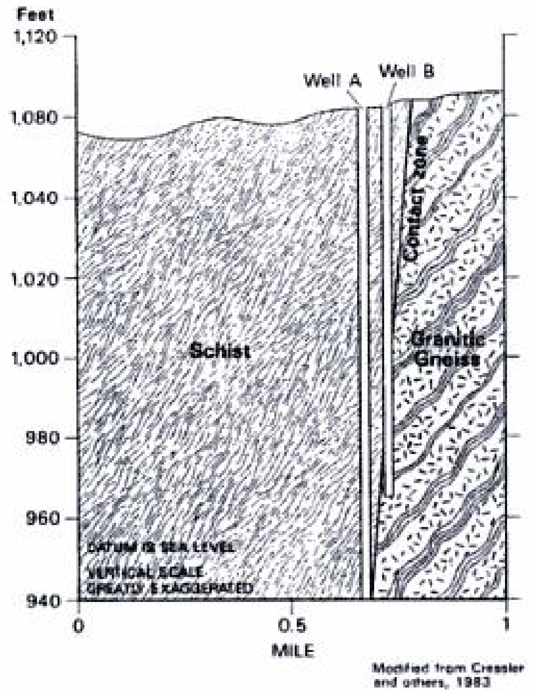


Figure 3.20. Two wells completed within a crystalline environment. Well-A, completed in schist, yielded only about 5.5 m³/day, while well-B, which is screened within the granitic gneiss and the contact zone, produced about 550 m³/day (USGS Hydrologic Investigations Atlas 730-G).

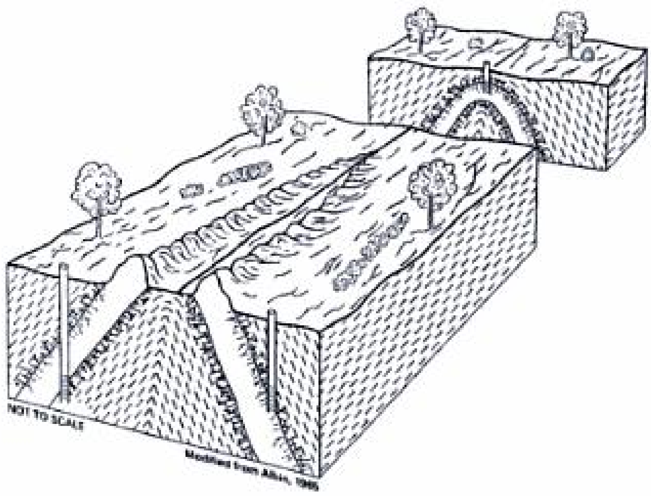


Figure 3.21. Contact zones between fractured sandstone (anticline) and shale (matrix) are favorable places to complete water wells (USGS Hydrologic Investigations Atlas 730-F).

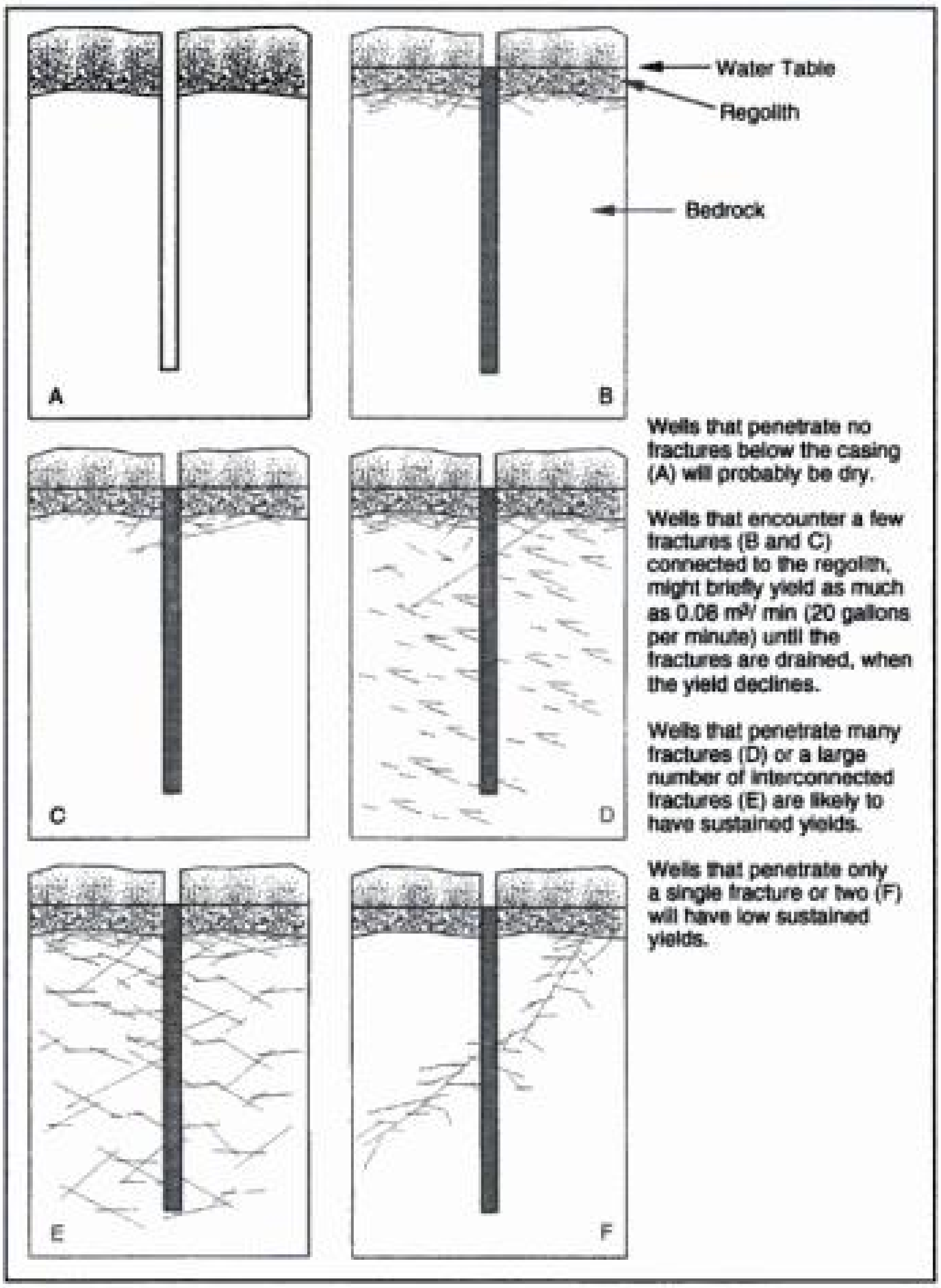


Figure 3.22. Wells constructed in crystalline rocks will have greater yields if they penetrate many fractures or fractures that are interconnected and encounter the surface (After LeGrand, 1967).

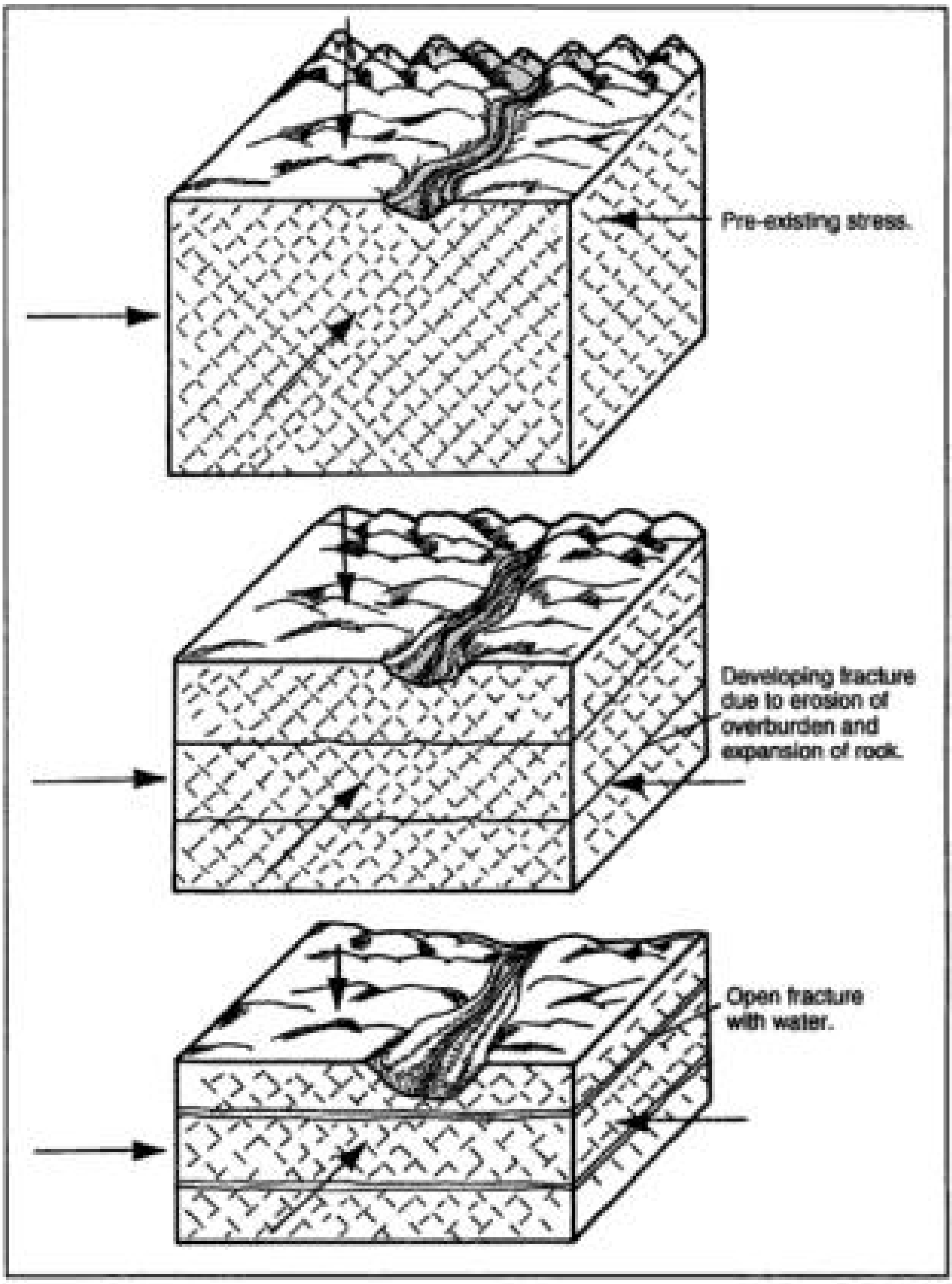


Figure 3.23. Development of horizontal fractures.

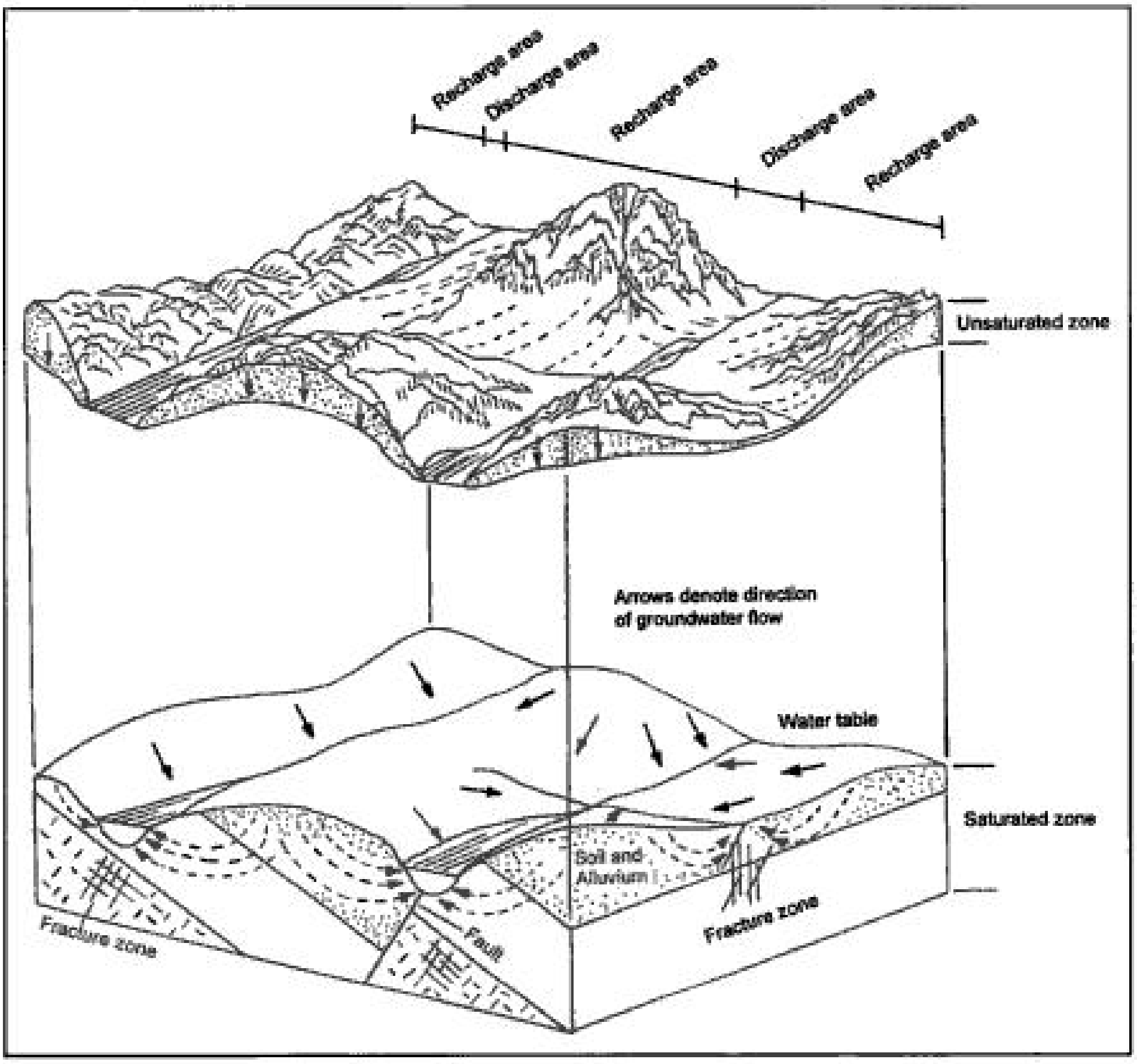


Figure 3.24. Interaction between a fault zone, fractures and permeable soil.

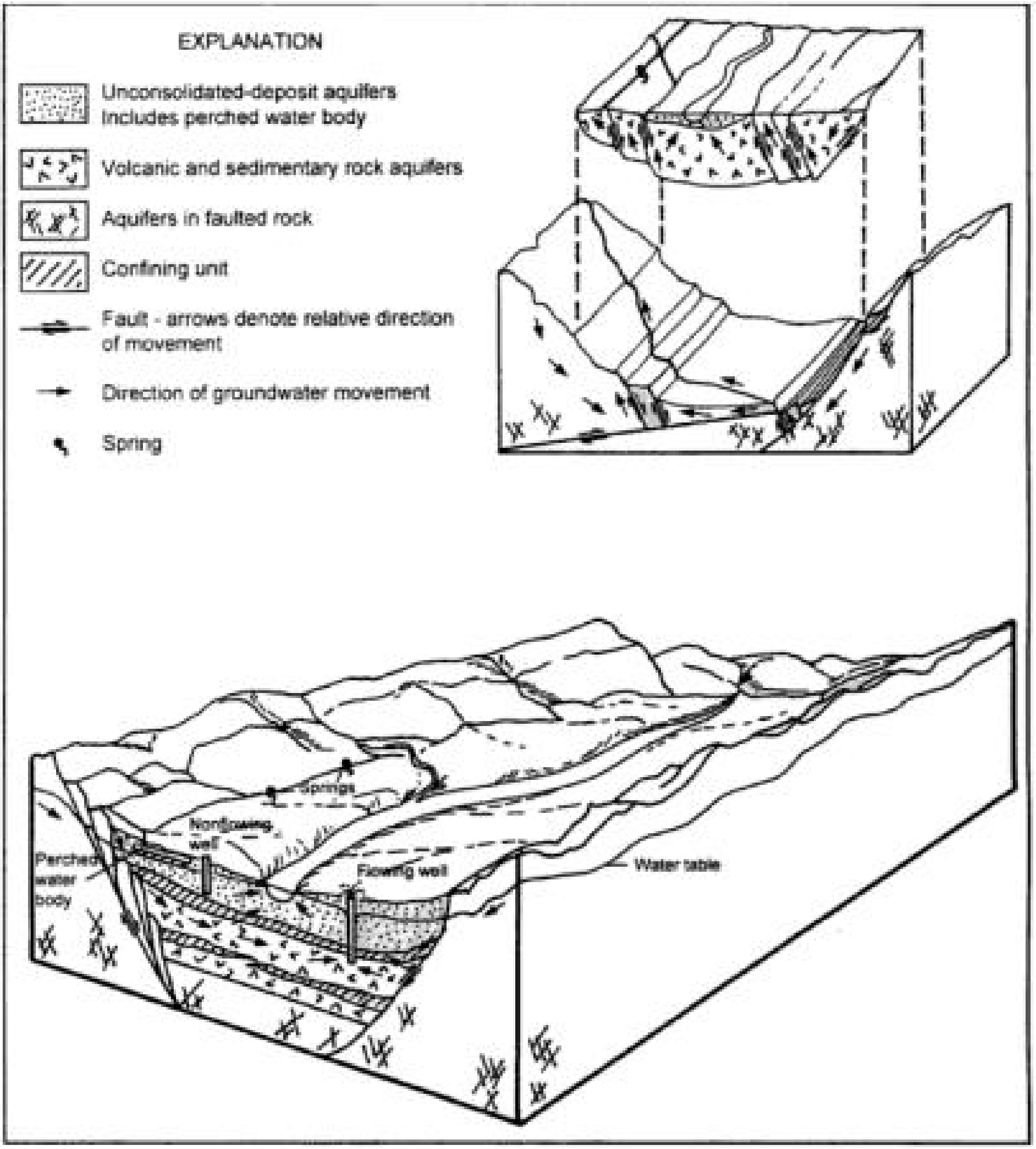


Figure 3.25. Fault zones acting as conduits in horst and graben structures. Unconfined aquifer system (top); complex system with confined aquifer (bottom).

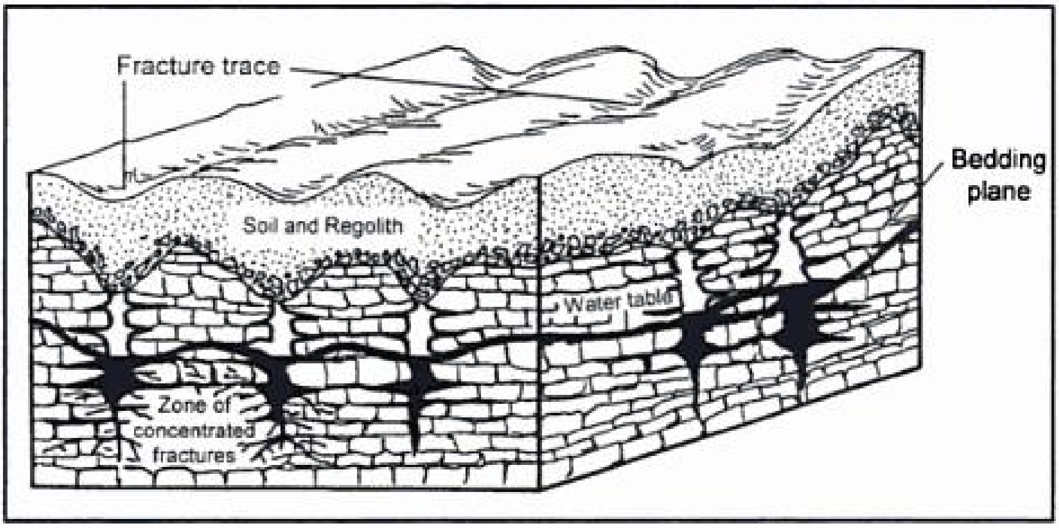


Figure 3.26. Fracture trace analysis is often used to identify zones of high permeabilty, which are identified by locating linear zones of subsidence at the surface.

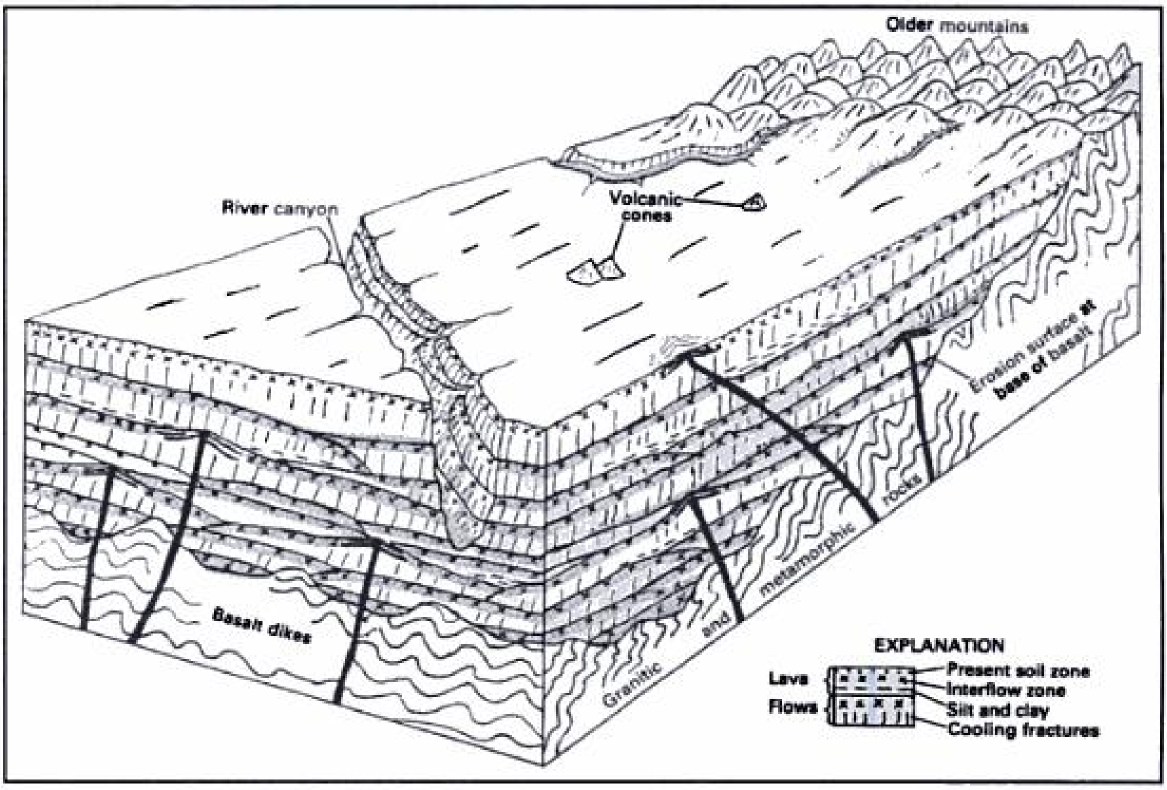


Figure 3.27. Topographic and geologic features of the Columbia Lava Plateau region (Heath, 1984). The interflow zones are represented in gray.

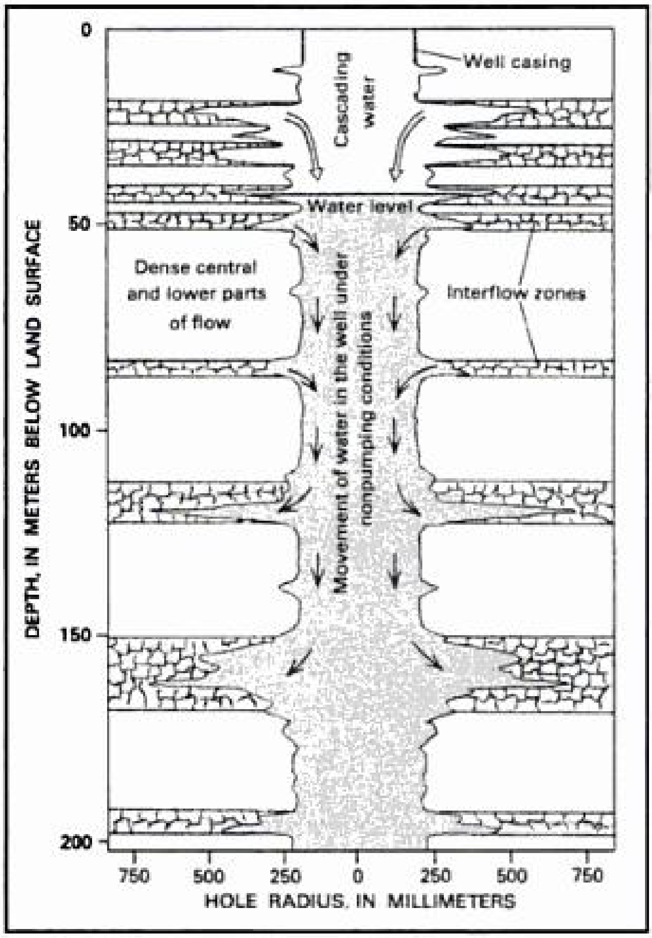


Figure 3.28. Well constructed within interflow zones in the Columbia River Group (Heath, 1984; modified from Luzier and Burt, 1974).

GROUND-WATER STORAGE

4.1 THE IDEAL AQUIFER

The aquifer system has two basic functions: it stores and transmits water; therefore, it is a reservoir and a conduit system. Water is stored between pore spaces of sediment, or in faults, fractures or solution cavities of rock. Ground water is transmitted from areas of recharge to areas of discharge when these void spaces are connected. Quantification of aquifers is concerned with these storage and transmissive properties. The **ideal aquifer** is often assumed for ease of mathematical calculations. Of course, the ideal aquifer does not exist, but this assumption generally results in good approximations, especially over a large area when average values are of concern. The ideal aquifer is rectangular and infinite, and comprised of homogeneous and isotropic sedimentary material.

4.2 HOMOGENEITY AND ISOTROPY

To review, geologic material that is **homogeneous** is of the same size and shape; therefore, it has the same hydraulic properties at all locations. If these properties are independent of direction the formation is said to be **isotropic**. Well rounded quartz beach sand is an example of homogeneous and isotropic sediment. Most geologic formations are **heterogeneous** and **anisotropic**, that is, the hydraulic properties change spatially and are more extreme in a given direction. Glacial till and alluvium are examples of such material.

The hydraulic property of general concern when considering the above conditions is **hydraulic conductivity (K)**. In simple terms, hydraulic conductivity is the rate at which a unit cube of geologic material will transmit a liquid under a hydraulic gradient. This important aquifer property will be discussed later in greater detail, for now, understand this, the hydraulic conductivity of a porous medium made of perfect spheres with the same diameter is equal in all directions. Anisotropy occurs as the shape of the material in the porous medium deviates from the geometry of a sphere, and increases as the material becomes progressively "flatter" or "oblong". Most sediment are laid on their flat sides or in the direction of their long axis; therefore, **horizontal hydraulic conductivity (K_h)** is generally greater when compared to **vertical hydraulic conductivity (K_v)** (Fig. 4.1).

Most mathematical equations that describe storage and flow of ground water assume homogeneous and isotropic conditions. As previously indicated, good quantitative values can be obtained when using these assumptions, because heterogeneities often cancel each other on a large scale where average values can be considered.

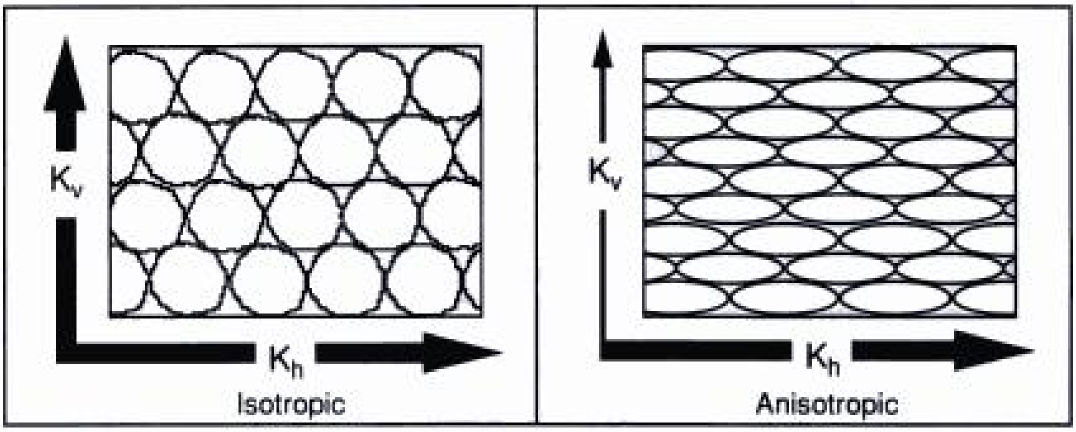


Figure 4.1. Isotropy and anisotropy resulting from grain shape and orientation.

Conditions that create anisotropy include (Fig. 4.2):

- facies changes • variations in sorting • sedimentary structures • fabric or grain orientations
- graded bedding • joints and fractures • layering of units with different K values • clay lenses
- thinning and thickening of adjacent units • inclusions • variations in cementation • faults

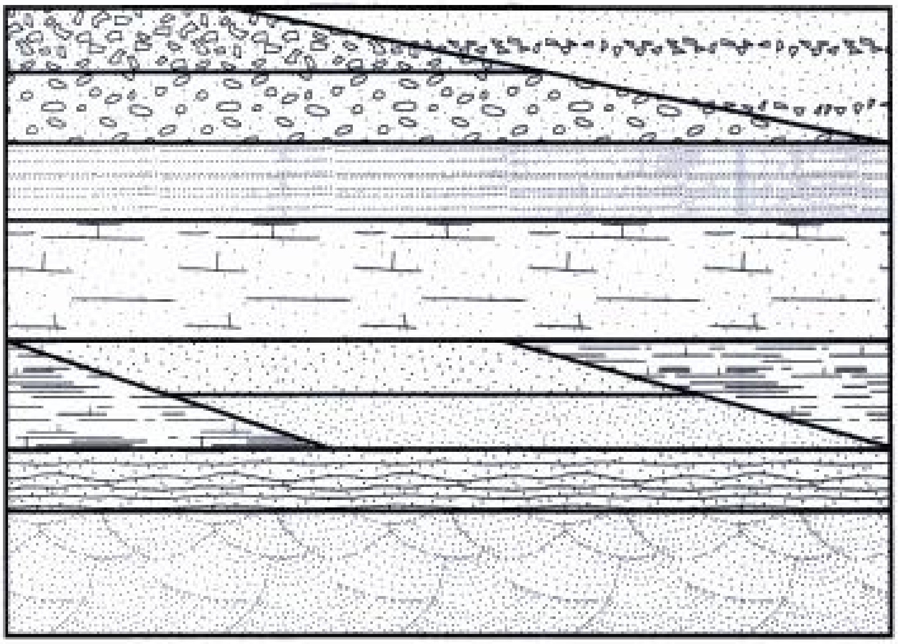


Figure 4.2. Horizontal and vertical spatial changes in sedimentary material produce variability in hydraulic conductivity, resulting in isotropic conditions.

4.3 POROSITY

Ground water is stored and moves through the pore spaces of rock or sediment. The volume of void space in geologic material is called **porosity** (Fig. 4.3).

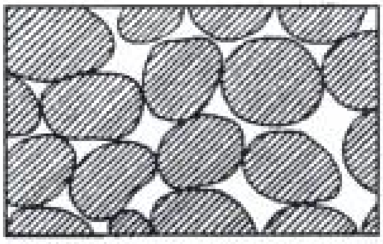


Figure 4.3 (Meinzer, 1942).

Porosity defines how much water a rock or sediment can hold when saturated. It can be expressed quantitatively as the ratio of the volume of voids to the total volume of material:

$$n = \frac{\text{volume of voids}}{\text{total volume of material}} = \frac{V_T - V_s}{V_T}(100) = \frac{V_v}{V_T}(100) = \text{percentage} \tag{4.1}$$

where, for consistent units,

n = porosity = unitless

V_T = total volume of material = L^3

V_s = total volume of solids = L^3

V_v = total volume of voids = L^3

Primary porosity is the original void space created when the rock or soil was formed. In soil and sedimentary rock, the primary voids are the spaces between the mineral grains or pebbles. Sediment shape (sphericity and roundness), orientation, sorting and packing, generally determine primary porosity. Angularity of sediment may increase or decrease porosity, depending if the particles bridge openings or are packed together like pieces of a mosaic (Lohman, 1972). Well sorted sediment has greater porosity compared to poorly sorted sediment, because in poorly sorted sediment the smaller grains take up space between the larger grains, which reduces total porosity (Fig.4.4).

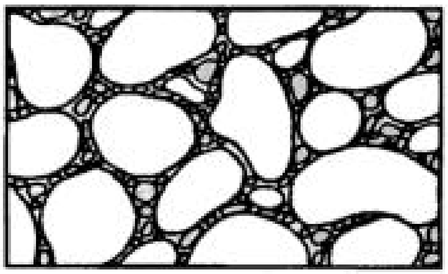


Figure 4.4.

Fine-grained materials tend to be well sorted and generally have large porosities. Clay often contains a porosity of 50%, silt 40%, sand 30% and gravel 25%; therefore, it can be generalized, that porosity by volume in clastic material increases as the actual grain size decreases. **Be careful of such generalities.** Porosity of clastic material of nearly uniform particle size is nonsensitive to actual particle size (McWhorter and Sunada, 1977). It can be shown, that for perfect spheres, porosity is independent of the size of the spheres. A volume of uniformly packed bowling balls has the same porosity as a volume of baseballs or marbles or beebees. In general, primary porosity increases with sphericity and roundness, but the highest degree of primary porosity often occurs in unconsolidated, subangular, well sorted sediment. Primary porosity in igneous and metamorphic rocks is usually negligible. If it exist, as in the extrusive rock pumice, the pore spaces are not connected; therefore, ground water cannot flow.

Secondary porosity is much more important in regard to igneous and metamorphic rocks. Secondary porosity develops after the material has formed. The cementing agent in a sedimentary rock reduces primary porosity, because it occupies part of the voids (Fig. 4.5). Fractures, joints and faults can develop in all rocks after formation, which can increase porosity (Fig. 4.6).

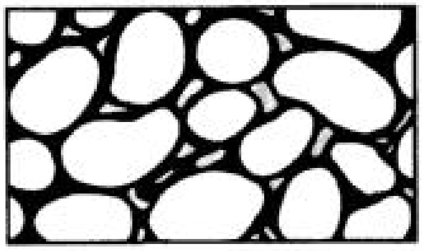


Figure 4.5.

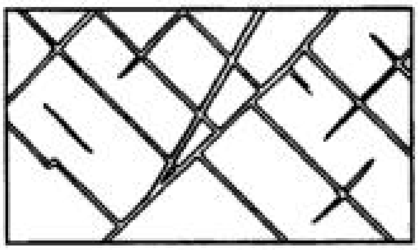


Figure 4.6.

Solution cavities that occur along joints and bedding planes in limestone, dolomite and other carbonate rocks greatly increase porosity (Fig. 4.7). The solution and transport of easily dissolvable mineral salts, such as halite, anhydrite and gypsum, from within sedimentary beds, can also increase porosity.

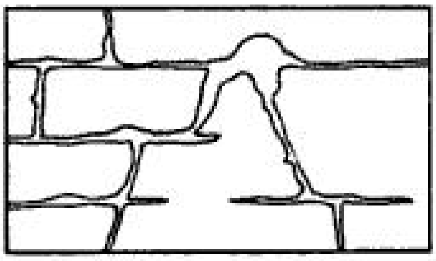


Figure 4.7.

Repacking of earth material due to stresses applied by overburden, earthquakes, landslides and subsidence can reduce primary porosity. When considering spheres, the least compact arrangement produces the greatest porosity -- about 47.6% -- this is called **cubic packing** (Fig. 4.8). The most compact arrangement, **rhombohedral packing**, produces the least porosity --- about 26% (Slichter, 1899). Therefore, the shape of the grains can increase or decrease porosity depending on how they are packed. Angular and irregular shaped, unconsolidated particles, tend to have larger porosities, although the difference is not always substantial (McWhorter and Sunada, 1977).

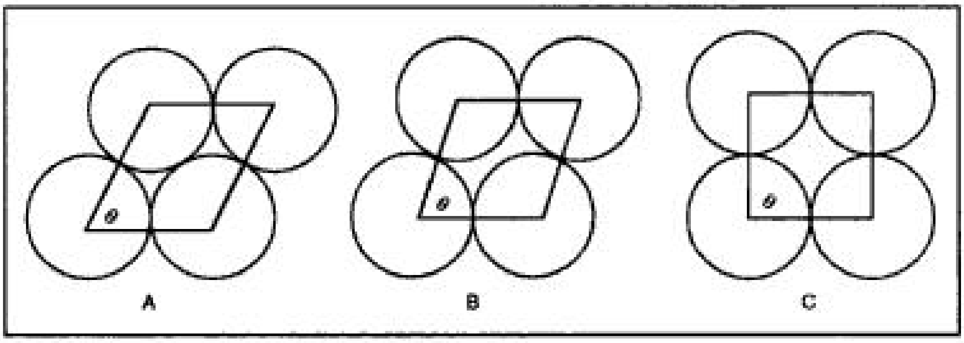


Figure 4.8. Porosity values resulting from different "packing" arrangements. "A" is the most compact arrangement, which has the lowest porosity. "B" has a higher porosity, because its compact arrangement is less than "A". "C" has the least compact arrangement and the highest porosity (after Slichter, 1899).

Porosity can range from 0 in crystalline rocks up to 60% for recent sedimentary deposits and dried, powdery clay. Values generally range between 20 to 50% for most clastic material (Table 4.1). On a relative scale clay has the highest porosity by volume (45-60%), even though clay impedes and often confines ground-water movement. Clays are comprised of extremely tiny plate-like particles with relatively large surface areas, which cause high-molecular attraction between the clay and water. Even though clays by volume have a high percent of voids, these pore spaces are microscopic, which prevents the easy flow of water through clays. In addition, water is often hydrated or absorbed by clay particles, filling the tiny pores, and becoming part of

Table 4.1. Representative values of porosity and specific yield.

SEDIMENTARY MATERIAL	POROSITY PERCENT	SPECIFIC YIELD PERCENT
GRAVEL, COARSE	23.8 - 36.5 - R	13.2 - 25.2
GRAVEL, MEDIUM	23.7 - 44.1 - R	16.9 - 43.5
GRAVEL, FINE	25.1 - 38.5 - R	12.6 - 39.9
SAND, COARSE	30.9 - 46.4 - U	18.4 - 42.9
SAND, MEDIUM	28.5 - 48.9 - U	16.2 - 46.2
SAND, FINE	26.0 - 53.3 - U	1.0 - 45.9
SILT	33.9 - 61.1- U	1.1 - 38.6
CLAY	34.2 - 56.9 - U	1.1 - 17.6
SANDSTONE, FINE-GRAINED	13.7 - 49.3 - U	2.1 - 39.6
SANDSTONE, MEDIUM-GRAINED	29.7 - 43.6 - U	11.9 - 41.1
LIMESTONE	6.60 - 55.7 - U	0.2 - 35.8
DOLOMITE	19.1 - 32.7- U	---
DUNE SAND	39.9 - 50.7 - U	32.3 - 46.7
LOESS	44.0 - 57.2 - U	14.1 - 22.0
PEAT	92 - U	44
SILTSTONE	21.2 - 41.0 - U	0.9 - 32.7
CLAYSTONE	41.2 - 45.2 - U	--
TILL, HIGH CLAY CONTENT	--	--
TILL, HIGH SILT CONTENT	29.5 - 40.6 - U	0.5 - 13.0
TILL, HIGH SAND CONTENT	22.1 - 36.7 - U	1.9 - 31.2
TILL, HIGH GRAVEL CONTENT	22.1 - 30.3 - R	5.1 - 34.2
WASHED DRIFT WITH CLAY	38.4 - 59.3 - U	33.2 - 48.1
WASHED DRIFT WITH SILT	36.2 - 47.6 - U	29.0 - 48.2
WASHED DRIFT WITH SAND	34.6 - 41.5 - U	---

R = repacked sample; U = undisturbed.
Source: USGS Water Supply Papers 1662-D and 1839-D.

the structure of clay, in essence, impeding the movement of new water through the clay. This is especially true of expanding or swelling clays such as montmorillonite.

4.4 EFFECTIVE POROSITY

The discussion in Section 4.3 assumes that the pore spaces between sediment or rock are interconnected, which would allow ground water to flow from one location to another. This is referred to as **effective porosity** (n_e). Volcanic rocks like pumice and scoria have an abundance of voids that formed when hot gasses escaped; but the spaces are generally not connected; therefore, when considering ground-water flow, such rocks are useless.

Effective porosity applies to all types of rocks. In order for igneous and metamorphic rocks to have effective porosity the joints, faults or fractures must link together to form a natural conduit system. The fundamental importance of effective porosity, for any earth material, is to act as a natural pipeline through which ground water can move. For most practical field applications $n_e = n$, although some investigators contend that effective porosity is equal to specific yield.

4.5 SPECIFIC YIELD

The volume of water that will drain from soil or rock under the influence of gravity is called **specific yield** (S_y). It can be expressed by the following ratio (Meinzer, 1923):

$$S_y = \frac{V_g}{V_T} \tag{4.2}$$

where, for consistent units,

S_y = specific yield as a decimal fraction = unitless
V_g = volume of water drained by gravity = L^3
V_T = total volume = L^3.

Specific yield represents the amount of water that can be available for supply and consumption. It is maximum in sediment ranging from medium to coarse-grained sand (Cohen, 1965). Values of specific yield depend upon grain size, shape, sorting, compaction and time of drainage. Angular and well rounded particles tend to have higher specific yields when compared to subangular particles, especially when subangular particles are compacted in a dense arrangement. After drainage is complete, a thin film of water will remain on the surface of sediment due to surface tension. This is referred to as **specific retention** (S_r) by hydrogeologists and civil engineers, and as **field capacity** by soil scientists. Material dominated by finer grain sizes will have more surface area on which the thin film of water can cling to; therefore, specific retention increases with decreasing grain size. In general, poorly sorted sediment has a higher specific retention, when compared to well sorted samples, due to the increase in surface area. Specific retention can be expressed by the following ratio (Meinzer, 1923):

$$S_r = \frac{V_r}{V_T} \tag{4.3}$$

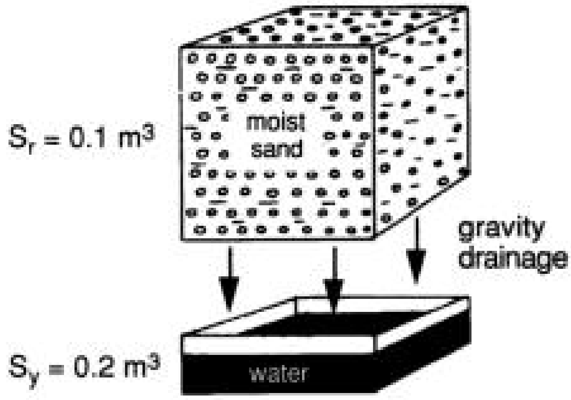
where, for consistent units,

S_r = specific retention as a decimal fraction = unitless
V_r = volume of water retained against gravity = L^3
V_T = total volume = L^3.

Porosity can be expressed as the sum of specific yield and specific retention:

$$n = \frac{V_g}{V_T} + \frac{V_r}{V_T} = S_y + S_r \tag{4.4}$$

The relationship between porosity, specific yield and specific retention is depicted in Figure 4.9.



$$n = S_y + S_r = \frac{0.2\ m^3}{1.0\ m^3} + \frac{0.1\ m^3}{1.0\ m^3} = 0.3$$

Figure 4.9. Relationship between n, S_y, and S_r, (modified from Heath, 1983).

4.6 MEASURING POROSITY AND SPECIFIC YIELD

Volume methods can be used to measure porosity and specific yield. A sample is dried to a constant weight and then saturated. Porosity is equal to the difference between the saturated volume and the original volume. If this sample is allowed to drain to completion, so that evaporation is not a factor, then specific yield is the drained volume of water. Specific retention is the difference between porosity and specific yield.

Example 4.1

A container is filled with 42.7 cm^3 of loose sand. This volume of sand is poured into a graduated cylinder partially filled with water. It is recorded that 28.4 cm^3 of water are displaced. The displaced volume is the volume of solids. Use equation (4.1) to determine porosity:

$$n = \frac{V_T - V_s}{V_T}(100) = \frac{42.7\ cm^3 - 28.4\ cm^3}{42.7\ cm^3} = 0.33 \qquad 4.1$$

Example 4.2

One cubic meter of uniform sand with a porosity of 0.30 is mixed with two cubic meters of uniform gravel that has a porosity of 0.25. What is the resulting porosity?

1) Pore space of gravel = 2(0.25) = 0.50 m^3
   Particles of gravel = 2(0.75) = 1.5 m^3

2) Particles of sand = 1(0.70) = 0.70 m^3
   Remaining volume of sand after the voids in the gravel are filled with sand:

   $1.00\ m^3 - 0.50\ m^3 = 0.50\ m^3$

3) $$n = \frac{V_T - V_s}{V_T}(100) = \qquad 4.1$$

$$\frac{(2.0\ m^3\ \text{gravel} + 0.50\ m^3\ \text{sand}) - (1.5\ m^3\ \text{gravel} + 0.70\ m^3\ \text{sand})}{(2.0\ m^3\ \text{gravel} + 0.50\ m^3\ \text{sand})} = 0.12$$

Example 4.3

The water table at the upper surface of an unconfined sand aquifer declines 7.6 meters (Δh = 7.6 m). The areal extent of the aquifer is about 10 km^2. The porosity of the sand is 40 percent. Specific retention is 12 percent. Rearrange equation (4.4) to calculate specific yield.

$$S_y = n - S_r = 0.40 - 0.12 = 0.28 \qquad (4.4.1)$$

What is the change in storage in m^3?

$$V_w = (S_y)(\Delta h)(A) = (0.28)(0.0076\ km)(10\ km^2) = 0.02128\ km^3$$
$$= 2.1 \times 10^7\ m^3$$

4.7 THE STORAGE COEFFICIENT

The **storage coefficient** (S) is defined as the volume of water that an aquifer releases from or takes into storage per unit surface area of the aquifer when the head is lowered a unit distance (Heath, 1983):

$$S = \frac{\text{volume of water}}{(\text{unit area})(\text{unit change in head})} = \frac{(L^3)}{(L^2)(L)} = \frac{L^3}{L^3} = \text{unitless} \tag{4.5}$$

Head is the height of a column of water above a datum plane. In practical application, head is the elevation of water in a well. The storage coefficient for most unconsolidated and many loosely consolidated aquifers can be expressed as:

$$S = S_y + bS_s \tag{4.6}$$

where, for consistent units,

S_s = **specific storage** (Jacob, 1940,1950 and Cooper, 1966)

and

$$S_s = \rho_w g(\phi + n\beta) = n\gamma\beta + \phi\gamma = 1 / L \tag{4.7}$$

- β = compressibility of water = LT^2 / M: m^2 / N
- ϕ = compressibility of aquifer skeleton = LT^2 / M: m^2 / N
- g = acceleration due to gravity = L / T^2: m / s^2
- ρ_w = density of water = M / L^3: kg / m^3
- n = porosity = unitless
- γ = specific weight of water = M / L^2T^2: N / m^3
- b = aquifer thickness = L: m.

Specific storage (S_s) is the volume of water a unit volume of saturated aquifer stores or releases from storage per unit decline in hydraulic head. It is considered to be a constant with units of $1/L$ generally expressed as $1/m$ or $1/ft$. The volume of water derived from specific storage in an unconfined aquifer is often negligible; therefore, when considering storage in a water table aquifer, the third term in equation (4.6) is neglected. Units cancel by the following relationship:

$$S_y = \frac{\text{volume of water}}{(\text{unit area})(\text{unit change in head})} = \frac{(L^3)}{(L^2)(L)} = \frac{\text{volume of water drained}}{\text{total volume of water}} = \frac{L^3}{L^3} = \text{unitless.}$$

Ground water drains under the force of gravity in an unconfined aquifer, and when drainage occurs the water table declines. The storage coefficient for a water table or unconfined aquifer generally ranges between 0.1 to 0.3.

4.8 STORAGE IN ELASTIC CONFINED AQUIFERS

The storage coefficient for an elastic confined aquifer, **storativity** is the product of specific storage (S_s) and the aquifer thickness (b): $S = S_s b$. It is not as intuitive as specific yield. Water released from storage in an elastic confined aquifer results from compressionof the aquifer and expansion of water. The overburden on top of a confined aquifer is supported by both sediment comprising the aquifer and hydraulic pressure exerted by the water in the aquifer. As the head declines in a confined aquifer water is released, but the aquifer remains filled with water. Hydraulic pressure is reduced as ground water is forced from the pores. Increased support of the overburden is transferred to the aquifer skeleton, the aquifer compacts, and the water forced from the pores represents that part of the storage coefficient due to compression. The aquifer remains filled with water because the pore spaces have been reduced (Fig. 4.10).

Hooke's Law states that strain is proportional to stress within the elastic limit; therefore, compressibility of the aquifer can be expressed by equation (4.8)(Table 4.2):

$$\phi = \frac{\Delta b / b}{\Delta P} \tag{4.8}$$

where, for consistent units,

ϕ = compressibility of the elastic confined aquifer = LT^2 / M: m^2 / N
b = original aquifer thickness = L: m
Δb = change in aquifer thickness = L: m
ΔP = change in hydraulic pressure = M / LT^2: N / m^2.

Table 4.2. Compressibility range of some earth material.

Material	Compressibility Range
Clay	10^{-6} to 10^{-8} m^2/N
Sand	10^{-7} to 10^{-9} m^2/N
Gravel	10^{-8} to 10^{-10} m^2/N
Jointed Rock	10^{-8} to 10^{-10} m^2/N
Competent Rock	10^{-9} to 10^{-11} m^2/N
Water	4.4×10^{-10} to 4.6×10^{-10} m^2/N

The amount of compaction or land subsidence for an elastic confined aquifer can be determined using equation (4.9):

$$\Delta b = \Delta P\left(\frac{S}{\gamma} - nb\beta\right) \tag{4.9}$$

where, for consistent units,

Δb = change in aquifer thickness = L: m

ΔP = change in hydraulic pressure = M / LT^2: N / m^2.

S = storage coefficient = $S_s b$ = unitless

γ = specific weight = M / L^2T^2: N / m^3

n = porosity = unitless

b = original aquifer thickness = L: m

β = compressibility of water = LT^2 / M: m^2 / N.

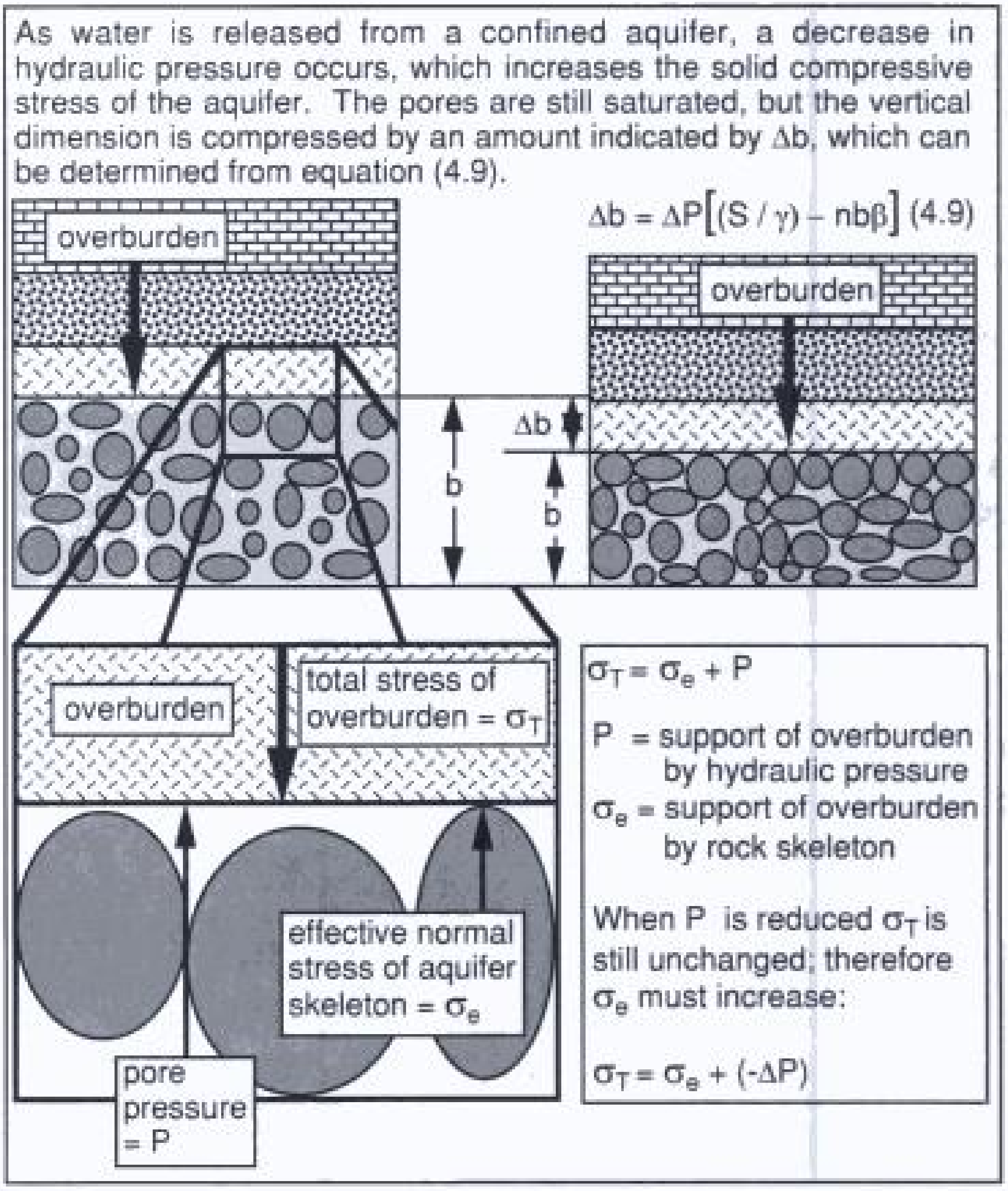


Figure 4.10. Water released from compressive storage in an elastic confined aquifer.

That part of the storage coefficient that results from the expansion of water (S_w) is small and is given by the following equation (Walton, 1962):

$$S_w = 4.6 \times 10^{-6}(nb), \text{ where } n = \text{porosity and } b = \text{aquifer thickness in meters.} \quad (4.10)$$

The English equivalent of equation (4.10) is

$$S_w = 1.4 \times 10^{-6}(nb), \text{ where } b = \text{feet.} \quad (4.11)$$

For consistent units

$$S_w = bn\gamma\beta \quad (4.12)$$

where

γ = ρg = specific weight of water = $M / L^2 / T^2$: N / m^3 (Table 1.4)
β = compressibility of water = LT^2 / M: m^2 / N (Table 1.4).

That part of the storage coefficient due to compression can be expressed as

$$S_c = b\phi\gamma \quad (4.13)$$

where, for consistent units,

ϕ = compressibility of the aquifer skeleton = LT^2 / M: m^2 / N

If the storage coefficient is known, equations (4.14) and (4.15) can be used to calculate that part of the storage coefficient that results from compression of the aquifer (S_c):

$$S_c = S - (4.6 \times 10^{-6}(nb)) = S - S_w \quad \text{(SI)} \quad (4.14)$$

$$S_c = S - (1.4 \times 10^{-6}(nb)) = S - S_w \quad \text{(English)} \quad (4.15)$$

Storativity can be expressed as:

$$S = b[\rho_w g(\phi + n\beta)] = b(n\gamma\beta + \phi\gamma) = bS_s \quad (4.16)$$

where

b = saturated aquifer thickness in a consistent unit.

An alternative expression for equation (4.16) is (Lohman, 1972):

$$S = n\gamma b\left(\beta + \frac{\phi}{n}\right). \quad (4.17)$$

Lohman (1972) used the relationship described by equation (4.11) to predict storativity using only the confined aquifer thickness (b = ft):

$$S = b(1.0 \times 10^{-6}). \tag{4.18}$$

The metric equivalent for equation (4.18) is

$$S = b(3.281 \times 10^{-6}) \tag{4.19}$$

where

b = aquifer thickness in meters.

For a confined aquifer, the second term in equation (4.6) is neglected. Units cancel by the following relationship:

$$S = bS_s = (\text{ft})\left(\frac{1}{\text{ft}}\right) \tag{4.20}$$

or by definition

$$S = \frac{\text{volume of water}}{(\text{unit area})(\text{unit change in head})} = \frac{(L^3)}{(L^2)(L)} = \frac{L^3}{L^3} = \text{unitless} \qquad 4.5$$

The storage coefficient for a confined aquifer generally ranges between 0.00001 and 0.001. The storage coefficient for a leaky confined aquifer often ranges between 0.001 and 0.1.

4.9 UTILIZING THE STORAGE COEFFICIENT

A simple equation can be used to estimate the amount of water released from storage:

$$V_w = (S)(\Delta h)(A) \tag{4.21}$$

where, for consistent units,

S = storage coefficient = unitless
V_w = volume of water released from storage = L^3
Δh = change or decline in the potentiometric surface or water table = L
A = area of aquifer = L^2.

Example 4.4

Calculate the change in storage (V_w) for a confined aquifer over 5.0 km^2. Assume a typical confined storage value of 0.0002. The Δh = 1.7 m. Determine V_w in both m^3 and gallons.

$$V_w = (0.0002)(1.7\ \text{m})(5.0 \times 10^6\ \text{m}^2) = 1700\ \text{m}^3 \qquad 4.21$$

$$1700\ \text{m}^3 = 4.50 \times 10^5\ \text{gallons}$$

Clearly, large volumes of water can be released from confined aquifers even though the storage coefficient is small.

Example 4.5

Equation (4.10) can be used to calculate that part of the storage coefficient that results only from the expansion of water (S_w).

$$S_w = 4.6 \times 10^{-6}(\text{nb}) \qquad 4.10$$

For example: if n = 0.1 and b = 91 m, then S_w = 0.000042. Equation (4.10) offers a check in regard to the storage coefficient. When $S_w \geq S$ then the determined storage coefficient is incorrect.

Example 4.6

The storage coefficient can also be used in equation (4.9) to predict land subsidence for an elastic confined aquifer (Lohman, 1972):

$$\Delta b = \Delta P\left(\frac{S}{\gamma} - nb\beta\right) \qquad 4.9$$

For example, when

ΔP = 6,895,000 N / m^2
S = 0.00035
γ = 9800 N / m^3 (from Table 4)
n = 0.1
b = 91.44 m, and
β = 4.6×10^{-10} m^2 / N (from Table 1.4)

$$\Delta b = 6895000\ \text{N / m}^2\left(\frac{0.00035}{9800\ \text{N / m}^2} - (0.1)(91.44\ \text{m})(4.6 \times 10^{-10}\text{m}^2/\ \text{N})\right) = 0.22\ \text{m}.$$

Example 4.7

The storage coefficient can be calculated using equation (4.17)

$$S = n\gamma b\left(\beta + \frac{\phi}{n}\right); \qquad 4.17$$

For example (data from Ferris et al., 1962),

γ = $9800\ N / m^3$ (from Table 4)
n = 0.3
b = 15.24 m
β = $4.6 \times 10^{-10}\ m^2 / N$ (from Table 1.4)
ϕ = $2.93 \times 10^{-10}\ m^2 / N$.

$$S = (0.3)(9800\ N / m^2)(15.24\ m)\left((4.6 \times 10^{-10} m^2 / N) + \frac{2.93 \times 10^{-10} m^2 / N}{0.3}\right) = 0.000064.$$

Example 4.8

The storage coefficient can also be used to calculate, ϕ, the compressibility of the aquifer skeleton, by rearranging equation (4.17):

$$S = n\gamma b\left(\beta + \frac{\phi}{n}\right); \qquad 4.17$$

$$\phi = \left(\frac{S}{n\gamma b} - \beta\right)n \qquad (4.22)$$

For example, using equation (4.22) (data from Ferris et al., 1962),

S = 0.00006437
γ = $9800\ N / m^3$ (from Table 4)
n = 0.3
b = 15.24 m
β = $4.6 \times 10^{-10}\ m^2 / N$ (from Table 1.4)

$$\phi = \left(\frac{0.00006437}{(0.3)(9800\ N / m^2)(15.24\ m)} - 4.6 \times 10^{-10} m^2 / N\right)0.3 = 2.93 \times 10^{-10} m^2 / N.$$

Hooke's Law can also be used to solve for ϕ,

$$\phi = \frac{\Delta b / b}{\Delta P}. \qquad 4.8$$

Example 4.9

The original thickness of an elastic confined aquifer is 38.10 m. The aquifer compacts 0.152 m after the head is lowered 19.81 m. Calculate the vertical compressibility of the aquifer using equation (4.8). Once this has been completed, calculated S_s and S.

$\Delta P = \Delta h$ in equation (4.8), therefore, Δh must be converted to hydraulic pressure, which can be expressed as

$$\Delta P = \Delta h(\rho g) = \Delta h(\gamma)$$

then

$$\Delta P = \Delta h(\gamma) = (19.81\ \text{m})(9800\ \text{N/m}^3) = 194{,}138\ \text{N / m}^2.$$

Using equation (4.8) with the given data results in

$$\phi = \frac{\Delta b / b}{\Delta P} = \frac{(0.152\ \text{m}) / (38.10\ \text{m})}{194{,}138\ \text{N / m}^2} = 2.0 \times 10^{-8}\ \text{m}^2 / \text{N}. \qquad 4.8$$

Now use the resulting and given data to calculate specific storage (S_s) and the storage coefficient (S). Porosity = n = 0.1; $\gamma = \rho g$.

From Table 1.4: $\beta = 4.6 \times 10^{-10}\ \text{m}^2 / \text{N}$

$$S_s = \rho_w g(\phi + n\beta) = 1 / L \qquad 4.7$$

$$S_s = \left(\frac{9800\ \text{N}}{\text{m}^3}\right)\left(2.0 \times 10^{-8}\frac{\text{m}^2}{\text{N}} + (0.1)\left(4.6 \times 10^{-10}\frac{\text{m}^2}{\text{N}}\right)\right) = 0.000196\frac{1}{\text{m}}$$

or

$$S_s = n\gamma\beta + \phi\gamma = 1 / L$$

$$S_s = \left[(0.1)\left(9800\ \frac{\text{N}}{\text{m}^3}\right)\left(4.6 \times 10^{-10}\frac{\text{m}^2}{\text{N}}\right)\right] + \left[\left(2.0 \times 10^{-8}\frac{\text{m}^2}{\text{N}}\right)\left(9800\ \frac{\text{N}}{\text{m}^3}\right)\right] = 0.000196\frac{1}{\text{m}}$$

Example 4.10 Water Balance

A groundwater basin in a coastal area has an area of 510 km^2. The land area is 500 km^2 and the area of the river is 10 km^2. There is no stream flow or groundwater flow into the basin. A water budget for the basin has the following long-term average annual values.

Precipitation	Evapotranspiration	Overland flow	Baseflow	Runoff	Sub-sea outflow
875 mm/yr	575 mm/yr	75 mm/yr	150 mm/yr	225 mm/yr	75 mm/yr

Notes:

- In order to prepare a water budget, identify all parameters in and out for each component assuming steady conditions.
- River flow or runoff is a combination of land flow and Baseflow.

1 Prepare an annual water budget for the basin as a whole.
2 Prepare an annual water budget for the river.
3 Prepare an annual water budget for the groundwater reservoir.
4 What is the annual river flow from the basin in m^3/sec?
5 What is the average rate of groundwater recharge in million m^3 per day per km^2 of surface area?

Answer 4.10

1 Water Budget for **the Whole Basin**

	In	Out
	Precipitation = 875 mm	Evaporation = 575 mm
		Runoff = 225
		Sub-sea flow = 75 mm
Sum	**875**	**875**

2 Water Budget for **the River**

	In	Out
	Over land flow = 75 mm	Runoff = 225
	Base flow = 150 mm	
Sum	**225**	**225**

3 Water Budget for the **Groundwater Reservoir**

	In	**Out**
	Prec. - Overland flow - Evap. = 875 - 75 - 575 (mm)	Baseflow = 150 mm
		Sub-sea flow = 75 mm
Sum	**225**	**225**

4 **Runoff** = 225 mm/yr

$= (0.225 \text{ m} \times 510\times10^{6} \text{ m}^2) / (365\times24\times60\times60)$

$\mathbf{= 3.64 \text{ m}^3/\text{sec}}$.

5 **Recharge** = 225 mm/yr

$= 0.225 \text{ m/yr} = 0.225 \text{ m}/ 365 = 6.164\times10^{-4} \text{ m/day}$.

$= 6.164\times10^{-4} \text{ m}^3/\text{m}^2 \text{ per day} = 6.164\times10^{-4} \text{ m}^3/ (\text{m}^2\times10^{-6}\text{km}^2/\text{m}^2\times\text{day})$

$= 6.164 \times 10^{-4} \times 10^{6} \text{ m}^3 \text{ per day per km}^2$

$\mathbf{= 1.164\times10^{-4} \text{ million m}^3 \text{ per day per km}^2}$.