Elements of Physical Hydrology

SECOND EDITION

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6 Groundwater Hydraulics

6.1 Introduction

Of the total amount of freshwater on this planet, about 30% is contained beneath the surface of the Earth. This figure is even more impressive when one considers that all but a few tenths of a percent of the remaining freshwater is held in ice caps and glaciers (Table 1.1). Most subsurface water is found in rocks and soils that are saturated with water; that is, materials in which water occupies all pores, openings, and fractures. We refer to water in the saturated region of the subsurface as **groundwater**.

Groundwater is an important resource across the globe. Irrigated agriculture in many areas is based on withdrawal of groundwater. For example, Shah et al. (2000) estimate that 60% of irrigated grain production in India depends on the use of groundwater. Half of the world's mega cities (population of 10 million or more) are dependent on groundwater (Giordano 2009). Even in areas where surface water is available, groundwater often is a preferable source because of water temperature, quality, or accessibility. For example, in Dhaka, a city of 10 million in Bangladesh, groundwater is the predominant source of freshwater even though the city is located in the delta region of the Ganges and Brahmaputra rivers and is bordered by major rivers.

The study of groundwater is motivated partly by practical considerations of water supply. In addition, an understanding of the hydrological cycle for a catchment requires a quantitative description of how the groundwater reservoir functions. For example, recall that streamflow (baseflow) is maintained in perennial streams between precipitation events. We can hypothesize that this is in large part the result of the discharge of groundwater into stream channels. In this chapter and in Chapter 7 we discuss subsurface pathways through which precipitation eventually may reach a surface water body. We return to these ideas again in Chapter 10, when we examine mechanisms of runoff generation.

A number of environmental issues involve groundwater, especially the remediation of sites that have been contaminated by poorly controlled dumping practices and the identification of and planning for sites to safely dispose of hazardous wastes. An example is planning for the disposal of radioactive wastes.

The Waste Isolation Pilot Plant (WIPP) in New Mexico, United States, is an underground repository built for certain radioactive wastes that were generated during the construction of nuclear weapons. These wastes, including work gloves and laboratory glassware, have been stored in 55-gallon drums at facilities of the Department of Energy. The wastes are not *high-level*, but they contain isotopes that remain radioactive for very long periods of time (tens of thousands of years). These wastes must be isolated for millennia to ensure that they do not pose risks to human health. In the U.S. these wastes are being buried in a repository in a 600-m thick salt formation (the Salado Formation) some 700 m below the ground surface. Rock salt is thought to be a good host rock for radioactive wastes because it flows and over time will seal the wastes off from the environment. The only path that would lead to the release of radioactivity to parts of the environment where it might adversely affect people or ecosystems is through dissolution of the waste by water and transport of the dissolved constituents by groundwater. The Culebra dolomite, a regional aquifer, overlies the Salado Formation. Assessment of the suitability of the WIPP site for disposal of radioactive waste requires knowledge of how groundwater flows in the aquifer above the repository. Several questions might be raised. What causes or drives the movement of groundwater? What physical characteristics of subsurface fluids and porous media determine the rate of fluid movement?

Because we are again interested in the flow of water in relation to imposed forces, the equations of fluid mechanics provide the basis for the quantitative description of groundwater flow. At the scale of the pores in rocks and soils, however, the paths along which groundwater flows are complex, with many twists, turns, contractions, and expansions. So we recognize immediately that simplifications will have to be made to enable useful equations to be derived.

6.2 A Conceptual Model

Let us try to picture the flow of water in a **porous medium**, for example, like sand. The path a "parcel" of water might follow in moving through a material containing pores or void spaces is convoluted (Figure 6.1). Not only does the water follow tortuous paths, but the geometry of the channels of flow is extremely complex and cannot be specified completely (i.e., the position, size, and shape of all of the sand grains cannot be known). Finally, we recognize that the openings in which water flows are very small. Therefore, we might expect that frictionless flow is totally meaningless in this situation and that head losses will play the predominant role.

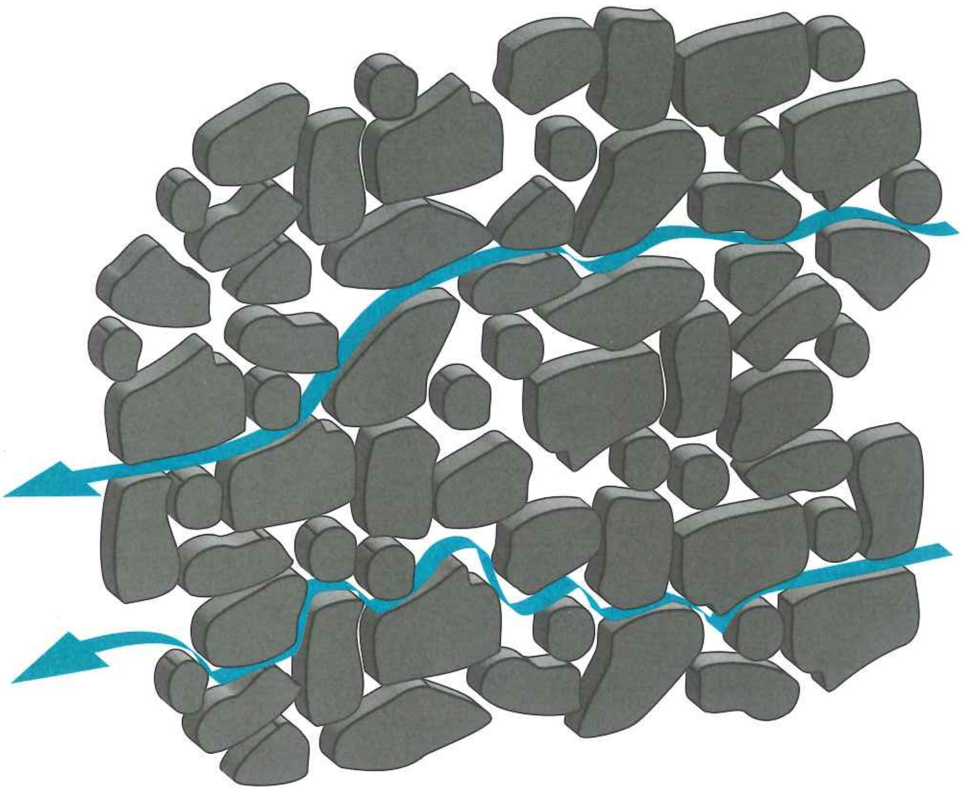

Figure 6.1 Schematic of a thin section of a porous medium and the tortuous flow path of two water "parcels." (Note that a real medium is three-dimensional with flow through the open spaces within the three-dimensional matrix.)

This picture of groundwater flow is exceedingly complex. To try to understand the fluid mechanics of the flow, we must resort to a conceptual model. We first consider the tortuous flow path (Figure 6.1) to be "straightened" by somehow stretching the path. The opening through which flow occurs then might be depicted as a pipe with a continually varying cross section (Figure 6.2). Of course, there are actually many flow paths and the entire system would consist of many different variable-radius tubes. Next we replace the variable-radius tubes with an *equivalent* set of constant-radius tubes. Thus, our conceptual model of flow through a porous medium is flow through a bundle of very small (capillary) tubes of different diameters. To be sure, the model is almost unacceptably oversimplified. Nevertheless, we can derive certain insights into flow in rocks and soils by examining this conceptual model.

Consider flow through a capillary tube. Because the tube is very narrow and the velocities are relatively small, it is reasonable to assume that the flow in the capillary tube is laminar. Moreover, because the velocities are very small, we can neglect the effect of changes in velocity and velocity head through the capillary tube, and assume that the diameter is constant. In Section 3.7.1, we found an expression for the average velocity,

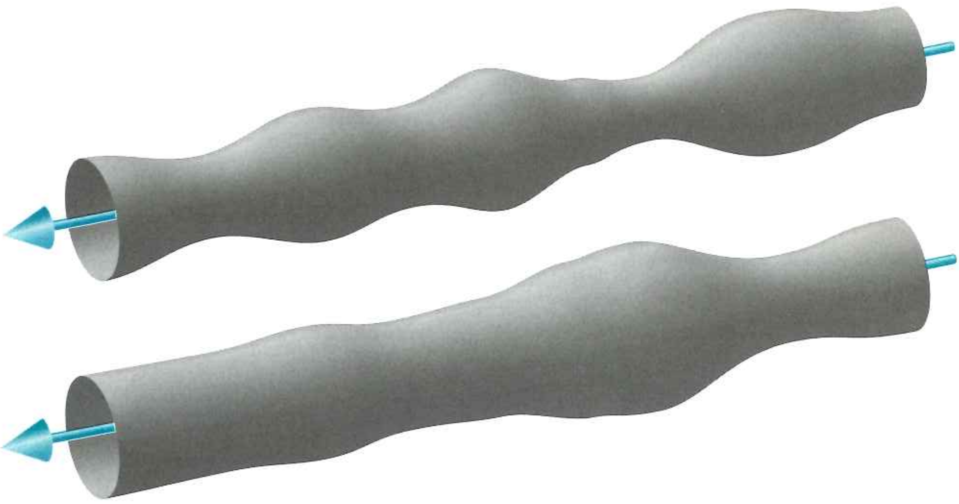

Figure 6.2 Straightened flow "tubes" representing flow in a porous medium.

$$U = -\frac{d}{dl}\left(\frac{p}{\rho g} + z\right)\frac{D^2 \rho g}{32\mu} \qquad \textbf{(3.38)}$$

of laminar flow through a pipe or tube of circular cross section in the direction l; D is the diameter of the tube and l is the flow direction. The quantity between parentheses is the sum of pressure head and gravitational or elevation head and is also known as the **hydraulic head** [L]

$$h = \frac{p}{\rho g} + z. \qquad \textbf{(6.1)}$$

The hydraulic head is a quantity, measurable at every point in a groundwater flow system. Because in these slow flows the velocity head is negligible, h represents the fluid mechanical energy per unit weight.

The negative sign in Equation 3.38 indicates that the flow is down the hydraulic head gradient from high to low hydraulic head. The discharge through the tube is given by the product of average velocity and the cross-sectional area of the tube:

$$Q = AU = \frac{\pi D^2}{4}\left(-\frac{dh}{dl}\frac{D^2 \rho g}{32\mu}\right) = -\frac{\pi D^4 \rho g}{128\mu}\frac{dh}{dl}. \qquad \textbf{(6.2)}$$

Equation 6.2 is a form of Poiseuille's law for the flow of a viscous fluid through a capillary tube. Discharge is directly proportional to the hydraulic head gradient (or hydraulic gradient), inversely proportional to the fluid viscosity, and directly proportional to the fourth power of the radius of the tube.

The equations derived from this conceptual model have several important implications. First, Equation 6.2 indicates that for a given fluid and given hydraulic gradient (*dh*/*dl*), the discharge varies as the fourth power of the radius of the tube. For example, the discharge from a tube of radius 10 mm will be 10^4 times greater than that from a tube of radius 1 mm, if all other conditions are the same. Because the size of the capillary tubes in our conceptual model is related (on an intuitive basis) to the texture or grain size of the porous medium, flow rates will depend on the texture. In addition, Equation 3.38 specifies that the average velocity, *U*, is proportional to the hydraulic gradient, *dh*/*dl*,

$$U = -\frac{D^2}{32}\frac{\rho g}{\mu}\frac{dh}{dl}. \tag{6.3}$$

Equation 6.3 is the form of Poiseuille's law that provides the most useful analogy for flow through a porous medium. As we will see, an equation almost identical to Equation 6.3 is the basis for studies of groundwater flow. Because Equation 6.3 was derived in Chapter 3 using well-known principles and stated assumptions, we can use the analogy between the conceptual model and the actual porous medium to great advantage in the sense that we can apply, in qualitative terms at least, knowledge about the factors that control laminar flow in tubes directly to the flow of groundwater.

6.3 Darcy's Law

In 1856, a French hydraulic engineer named Henry Darcy published an equation for flow through a porous medium that today bears his name. In designing a water treatment system for the city of Dijon, Darcy found that no formulas existed for determining the capacity of a sand filtration system. Consequently, Darcy performed a series of experiments on water flow through columns of sand.

Darcy packed sand into iron pipes and systematically measured parameters that he expected to influence the flow. Consider flow through a cylindrical volume of sand, with cross-sectional area $= A$ and length $= L$ (Figure 6.3). The sum of elevation head (z) and pressure head ($p/\rho g$) is represented as h and varies from h_1 to h_2 along the column. The hydraulic head at each end of the column is measured with an open tube, a simple manometer, as shown in the diagram. By varying L and the hydraulic head difference across the column ($h_1 - h_2 = \Delta h$), Darcy found that the total discharge Q varies in direct proportion to A and to Δh and inversely with L. That is,

$$Q = KA\frac{h_1 - h_2}{L}, \tag{6.4}$$

where K is a constant of proportionality called the **hydraulic conductivity** [L T^{-1}]. Equation 6.4 can be rewritten as:

$$\frac{Q}{A} = -K\frac{h_2 - h_1}{L} = -K\frac{h_2 - h_1}{l_2 - l_1}. \tag{6.5}$$

This can be written more generally as:

$$q = -K\frac{dh}{dl}, \tag{6.6}$$

where $q = Q/A$ is the **specific discharge**. Equation 6.6 is the form of **Darcy's law** that we will use in our studies. Although q has dimensions of velocity [L T^{-1}], keep in mind that we obtained this term by dividing the discharge by the *total area* and that water flows only through a fraction of the area, the spaces between the solid grains of the medium.

In Equation 6.6, dh/dl is the **hydraulic gradient** and the negative sign indicates that positive specific discharge (indicating direction of flow) corresponds with a negative hydraulic gradient. Thus, Darcy's law states that specific discharge in a porous medium is in the direction of decreasing head and directly proportional to the hydraulic gradient (Figure 6.4).

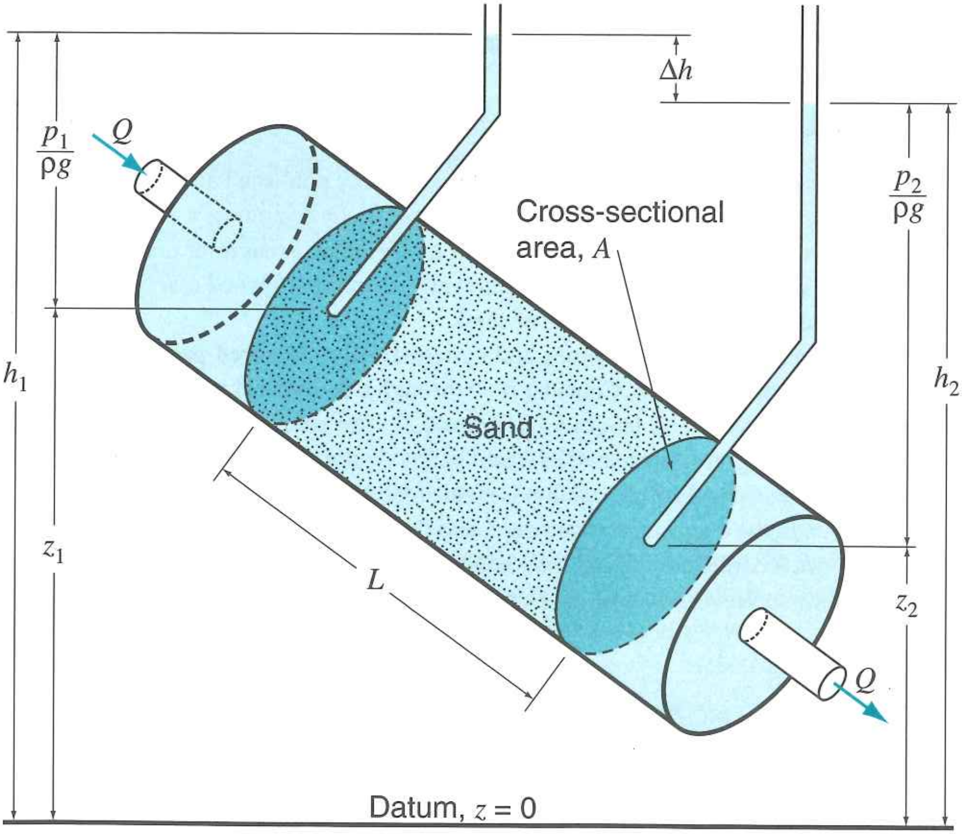

Figure 6.3 Schematic diagram of an apparatus for illustrating Darcy's law.

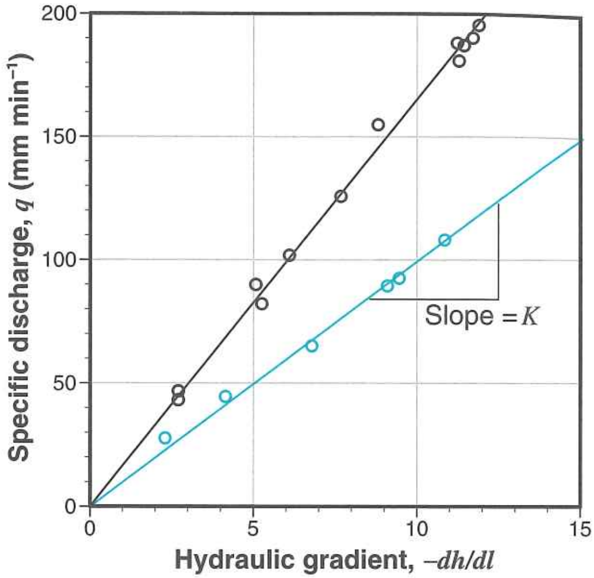

Figure 6.4 Darcy's (1856) original data showing a linear relationship between specific discharge and hydraulic gradient for two different sands.

6.3.1 Hydraulic conductivity, intrinsic permeability, and porosity

We can envision the hydraulic conductivity as the slope of a line relating specific discharge, q, with hydraulic gradient, dh/dl (Equation 6.6). Imagine a set of experiments using a given sample of material and a fluid of constant density and viscosity. By varying the hydraulic gradient and measuring the discharge, q can be plotted against dh/dl. According to Darcy's law a straight line should be the result, the slope of which will be K (Figure 6.4). Using a fluid with a constant density and viscosity, the slope of the relationship between q and dh/dl will depend only on the material and generally will increase with the coarseness of the material. Now imagine repeating the experiments with another fluid having different properties (e.g., a greater viscosity). We would expect the more viscous fluid to move more slowly if everything else remains constant.

The example above suggests that hydraulic conductivity depends on the nature of both the fluid and the porous material. The way in which K depends on these properties can be inferred by reference to the conceptual model discussed earlier. We already have remarked on the similarity between Equations 6.3 and 6.6, which are reproduced below (for one-directional flow in the l direction) to facilitate direct comparison.

$$U = -\frac{D^2}{32}\frac{\rho g}{\mu}\frac{dh}{dl}. \qquad \textbf{(Poiseuille's law)} \qquad (6.3)$$

$$q = -K\frac{dh}{dl}. \qquad \textbf{(Darcy's law)} \qquad (6.6)$$

This analogy suggests that $\rho g/\mu$ should describe the variation of K with fluid density and viscosity, and that $D^2/32$ should describe the variation of K with pore diameter. Of course, the "diameter" of the pores is not measurable (or even well defined), so the analogy is not perfect. However, qualitatively we might expect the grain size of the material to give an indication of the size of the openings (i.e., we would expect larger pores in a boulder field than in a silt deposit) and we can define and measure the average grain size of a granular material such as sand. Experimental evidence supports this analogy, at least for simple granular porous media. Based on experimental results, the following empirical relationship for hydraulic conductivity can be written:

$$K = (Nd^2)\left(\frac{\rho g}{\mu}\right), \tag{6.7}$$

where N = a factor to account for shape of the passages [dimensionless]; d = the mean grain diameter [L]; ρg = the unit weight of the fluid [M L^{-2} T^{-2}]; μ = the viscosity of the fluid [M L^{-1} T^{-1}]. From this we see that K is composed of two factors, one representing fluid properties and the other representing properties of the medium. Darcy's law can be written in a manner that clearly separates these two influences:

$$q = -k\left(\frac{\rho g}{\mu}\right)\left(\frac{dh}{dl}\right) \tag{6.8}$$

where k is referred to as the **intrinsic permeability** [L^2] of the porous medium. The factor Nd^2 in Equation 6.7 is equivalent to k. Once the intrinsic permeability of a certain rock formation is known, Equation 6.8 can be used to describe the flow of any fluid (oil, gas, or water) through that formation.

In addressing issues such as the movement of contaminants in the subsurface, it should be apparent that the hydraulic conductivity or intrinsic permeability of natural materials plays a major role, with higher values of K or k resulting in faster transport. Measuring or estimating these properties is a fundamental step in applying Darcy's law to a natural setting. There are a variety of techniques and methods for either directly or indirectly determining the permeability of a sample of porous material. A small sample can be placed in a device called a **permeameter**, not unlike Darcy's original column. The flow rate through the sample can be measured for a known hydraulic gradient and the permeability can be calculated directly using Darcy's law. In some situations, for example in looking at flow deep underground, samples may be difficult to obtain. There are a variety of methods, known as aquifer tests or "pump" tests, for determining permeability in such cases. By withdrawing or adding water to a well, and measuring the water level in that well or other wells nearby as a function of time, we can calculate the permeability. Finally, there are indirect methods that are based on measuring some other parameter, such as grain size, that is related to the permeability.

Literally thousands of permeability measurements have been made in different materials. The results show that the range of permeability of natural materials is quite large (Figure 6.5). Recalling the discussion of the WIPP site in the introduction, one reported range of hydraulic conductivities for salt deposits, such as the Salado Formation, is 10^{-12} to 10^{-10} m s^{-1} (Domenico and Schwartz, 1990). Relative to the other types of materials shown in Figure 6.5, this range is near the low end and is similar to shales and unfractured crystalline rocks. The resulting slow rate of groundwater flow is one reason that salt formations have been considered for waste repositories. For example, a typical hydraulic gradient of 1/100 in a salt formation with a hydraulic conductivity of 10^{-10} m s^{-1} will produce a specific discharge of 10^{-12} m s^{-1}, or less than 1 mm per 30 years!

Darcy's law indicates that for a given value of the hydraulic gradient (dh/dl), the specific discharge will be greater for a permeable material, such as a sand or gravel, than for a granite. The difference can be several orders of magnitude (Figure 6.5). As mentioned earlier, the specific discharge, despite having dimensions of velocity [L T^{-1}], is not a velocity. We could not use specific discharge to determine how long it will take a parcel of water to move from one point to another. The cross-sectional area available to the water is smaller than the actual cross-sectional area, such that the solid portion of the porous medium acts as a constriction. This constriction means that a tagged parcel of water, or, better, many tagged parcels that are averaged together, will appear to move through a porous medium at a speed that is faster than the specific discharge. The effect is similar to the constriction in a pipe discussed in Chapter 3; the constricted flow has a greater mean velocity for the same value of discharge.

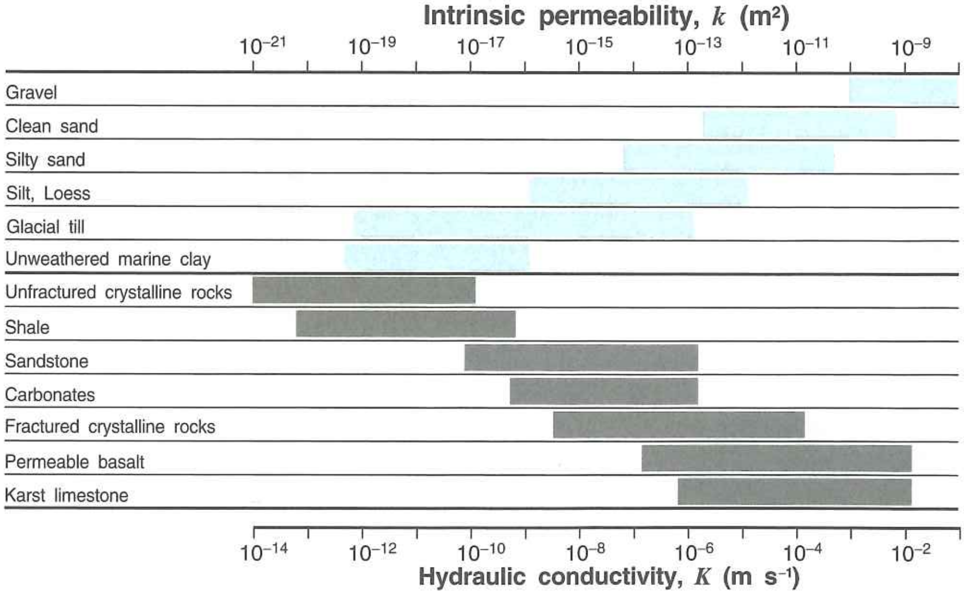

Figure 6.5 Ranges of intrinsic permeability and hydraulic conductivity for a variety of rocks (*gray bars*) and sediments (*blue bars*).
Data from Freeze and Cherry (1979).

We can determine the mean pore water velocity if we have some idea of the amount of "constriction." For porous materials, this property is the **porosity** [dimensionless], which is simply the fraction of a porous material that is void space:

$$\phi = \frac{V_v}{V_t}, \tag{6.9}$$

where V_v is the volume of void space [L^3] and V_t is the total volume [L^3]. If we have measured the porosity, which by definition must be between 0 and 1, we can determine the mean pore water velocity, or **average linear velocity,** $\overline{v}$ [$L\ T^{-1}$]:

$$\overline{v} = \frac{q}{\phi}. \tag{6.10}$$

6.3.2 Restrictions on Darcy's law

Darcy's law is widely used for almost all situations involving motion of a fluid through soil or rock in the natural environment. Although the conceptual model we used to introduce our discussion of Darcy's law can apply (to a limited extent) only to granular materials such as sand, Equation 6.6 is applied in materials ranging from clay to limestone to fractured crystalline and metamorphic rocks. The openings through which fluid flows in most of these materials cannot be envisioned as capillary tubes. Nevertheless, Darcy's law usually can be applied with success.

There are some general limitations on the use of Equation 6.6, and because of the analogy we developed we can infer how these restrictions arise by examining the assumptions implicit in the derivation of Equation 6.3. First, Darcy's law has been found to be invalid for high values of Reynolds number. Experiments using very high flow rates in very permeable materials have found that when values of the Reynolds number exceed the range 1–10, Darcy's law does not describe the relationship between flow rate and hydraulic gradient. For flow through porous materials, the characteristic length in the Reynolds number is the mean grain diameter, *d*:

$$\mathbf{R} = \frac{\rho q d}{\mu}, \tag{6.11}$$

where q, the specific discharge, is used in place of the mean velocity, U. This restriction can be explained by virtue of the fact that Poiseuille's law neglects inertial forces (accelerations). The simplest evidence of this statement is that a velocity head term does not appear in Poiseuille's law (or in the definition of hydraulic head) and that this term derives from consideration of acceleration.

Darcy's law also may fail to hold at very low values of hydraulic gradient in some very low-permeability materials, such as clays. Equation 6.6 implies that any hydraulic gradient, no matter how small, will cause some motion of water. In certain clay materials it is observed that below some threshold value, a small hydraulic gradient will not cause motion; that is, the force due to pressure and gravity must exceed some critical value before motion of water ensues. Poiseuille's law, which is our analogy for Darcy's law, is an expression of Newton's second law, and is based on the idea of balanced forces. However, the only forces considered are pressure and shear. Deviation from the law can be the result of the presence of other forces that we failed to consider (e.g., the inertial terms discussed in the previous paragraph). In the case of clays, there may be important electrostatic forces resulting from charge imbalances within the mineral structure and at low hydraulic gradients these may contribute to a force balance.

6.4 Water in Natural Formations

The subsurface is a complicated assortment of different materials, some of high permeability and some of low permeability. Hydrologists studying groundwater flow are interested in how the distribution of different materials in the subsurface influences patterns and rates of groundwater movement, and how the water interacts chemically with natural materials. The first topic will be addressed more fully in Chapter 7, but for now we introduce some terms that are applied to different soil or rock units, based on whether they are relatively permeable or not, and where they reside in the subsurface. An **aquifer** is a saturated geological formation that contains and transmits "significant" quantities of water under normal field conditions. "Significant" is a vague term but the implication is that aquifers are formations that can be used for water supply. Obviously, whether the supply is significant or not depends upon whether one is referring to a supply for a single rural dwelling or a large municipality.

Many aquifers are unconsolidated materials, mainly gravel and sand. Examples of this type of aquifer include those in coastal plain settings and intermontane valleys. Limestones, partially cemented sandstones and conglomerates, and permeable volcanic and igneous rocks are also important as aquifers. The limestone aquifers in the Paris basin in Western Europe and those underlying most of the Florida peninsula in the United States are critical resources as are the Deccan basalt groundwater system in India and the basalts of the Hawaiian Islands. Similarly, unconsolidated materials form important aquifers in the North China Plains in China and in the High Plains in the central United States.

Of course, not all formations are aquifers. An **aquiclude** is a formation that may contain water but does not transmit significant quantities. Clays and shales are examples of aquicludes. An **aquifuge** is a formation that neither contains nor transmits significant quantities of water. Unfractured crystalline rocks would fall into this category. The more general term **aquitard** is often used to denote formations that are of relatively low permeability and that may include both aquicludes and aquifuges.

Aquifers are classified according to hydraulic conditions and type of material. One type of aquifer is an **unconfined** or **water-table aquifer**. If an excavation is made in soil, then the near-surface material is usually not saturated (the unsaturated or vadose zone, Chapter 8). Deeper in the soil profile, saturated conditions prevail (**saturated zone**); groundwater by definition refers to water in the saturated zone of the subsurface.

The **water table** or phreatic surface is defined as a surface of zero gage pressure within the subsurface, and separates the saturated and unsaturated zones. Water will flow into an excavation or well up to this level; the water table is equivalent to a free surface. An aquifer with the water table as the bounding surface of the top of the aquifer is an unconfined aquifer.

The second type of aquifer is called **confined** or **artesian**. This type of aquifer is found when permeable material (the aquifer) is overlain by relatively impermeable material (an aquiclude; Figure 6.6). The water in a confined aquifer is under pressure and, in a well penetrating the aquifer, will rise above the top of the aquifer (see Figure 6.6). The height to which water rises in a well defines the **piezometric** or **potentiometric surface**. Note that a well penetrating a confined aquifer can be thought of as a **piezometer**, or single tube manometer. The water level in a piezometer is a measure of water pressure in the aquifer. Thus, the elevation of the potentiometric surface above an arbitrary (horizontal) datum is the sum of pressure head and elevation head, or the hydraulic head, h, of Darcy's law.

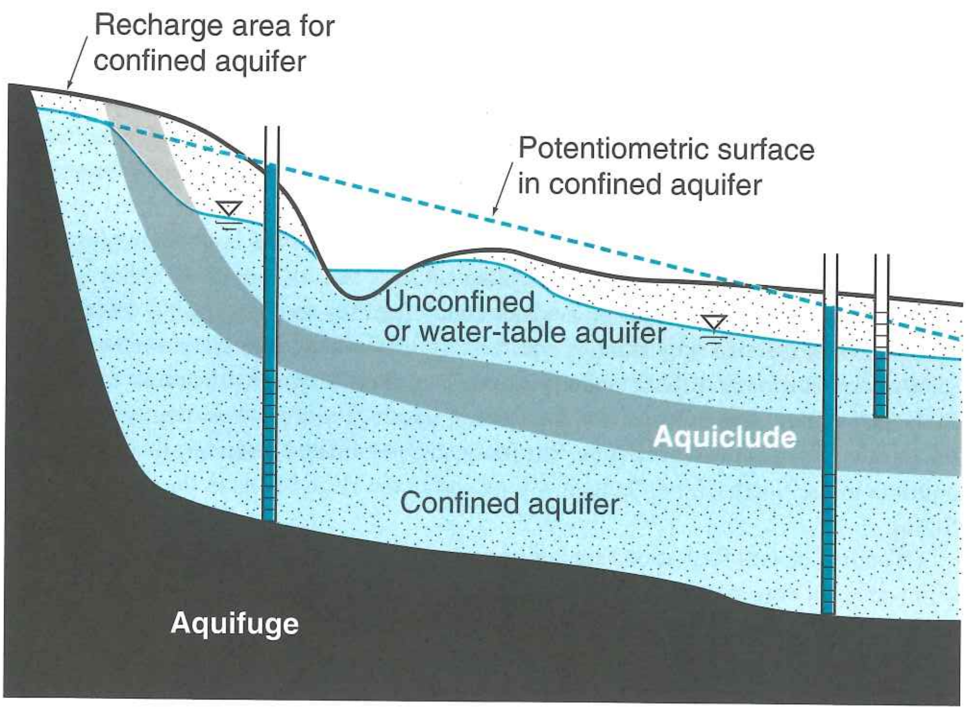

Figure 6.6 Hydrogeological units. Three piezometers are depicted, which are open in either the confined or unconfined aquifer, as indicated by the short horizontal lines. Note that in unconfined aquifers, the water level in the piezometer (*far right*) indicates the height of the water table; in confined aquifers, the water level in the piezometers (*left* and *center*) rises above the top of the aquifer and indicates the position of the potentiometric surface.

6.4.1 Construction of wells and piezometers

We derive information about the subsurface in large part by constructing wells and piezometers. A well is an opening, generally a cylindrical opening, from the ground surface down to a geological formation containing water. The formation is typically an aquifer, at least for wells constructed to supply water or to monitor a water source. In unconsolidated materials such as sand and gravel aquifers, a **well casing** is installed as part of the construction. A casing is made of solid material intended to support the opening and keep the well open. The well is screened over some interval to allow water to enter the otherwise impermeable casing. A **well screen** can be as simple as slots cut into a polyvinylchloride (PVC) pipe that is used as a casing. Wells typically are screened over a considerable depth of the aquifer to allow easy entry of water. A piezometer, by contrast, typically is screened over a narrow interval to reflect groundwater head at one location in an aquifer. If there is little vertical flow of groundwater, the head is sensibly constant over the depth of the aquifer and water levels recorded by wells screened over large vertical intervals will provide the same measurements as piezometers.

There are many methods for constructing wells including hand digging, augering, hammering, and jetting (using water under pressure to "erode" material and excavate a hole). The maximum depth that can be reached in the excavation of a well strongly depends on which of these techniques is used (only a few meters for hand dug wells, several tens of meters for jetted wells). One common method for constructing wells and piezometers in unconsolidated material involves the use of a hollow-stem auger. A **hollow-stem auger** is just what it sounds like (Figure 6.7a): auger blades surround an open core. There is a removable plug at the bottom and once the desired depth is reached, the plug is removed and a casing is inserted into the hollow stem. The auger is withdrawn, leaving the casing in place. The screened area of the well outside of the casing is backfilled with coarse material to allow ready flow of water to the well through the screen but to prevent fine material from clogging the screen. The upper part of the hole outside the casing is backfilled with a relatively impermeable material such as clay to prevent downward leakage of surface water into the well (or upward flow in the case of a confined aquifer). Near the surface a concrete cap is generally installed to complete the seal of the casing and prevent surface water and contaminants from entering the well (Figure 6.7b).

6.4.2 Water-level measurements

Key measurements that we derive from wells and piezometers are the elevations of the potentiometric surface or the water table. It is these measurements that allow us to establish head gradients that can be used with estimates of hydraulic conductivity to calculate groundwater flow rates. Once the elevation of the top of the well casing is established by surveying or the use of satellite positioning, we can determine the elevation of the water surface in a well by measuring the distance from the top of the well casing. One instrument that we routinely use to make these measurements is a **pressure transducer**. This is a device, as the name suggests, that measures water pressure. The transducer is lowered a known distance from the top of the well casing and below the water

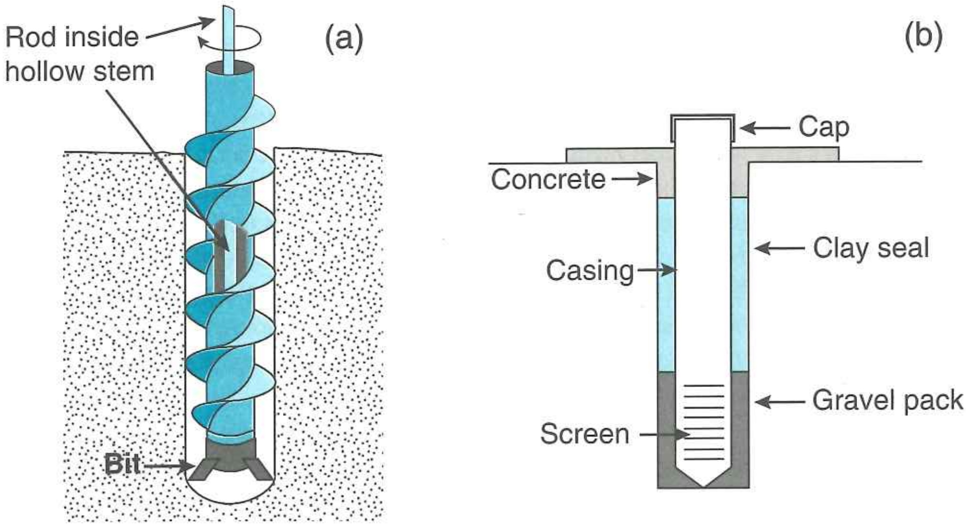

Figure 6.7 Schematic of a hollow-stem auger (*a*), one technique used to drill wells. Once wells are drilled (*b*), construction typically is completed as follows: coarse material is placed around the well screen to facilitate water inflow to the well, backfill is packed around the rest of the casing to seal off the screened region from possible surface contamination, and a grout and concrete cap is constructed to provide surface security.

level in the well and the pressure is measured. The hydrostatic Equation 3.5 can be used to compute the pressure at the depth of the transducer. Pressure transducers can be left in place for extended periods and connected to a data logger to record well hydrographs (Figure 6.8).

6.4.3 Geophysical techniques

There are other, indirect techniques that are used to estimate subsurface conditions that cannot be "seen" in the conventional sense of the word. Geophysics is a field that uses measurements of physical processes and properties to infer information about the structure of the Earth, and in the case of hydrogeology, the relatively shallow solid Earth in particular. That is, we use measurements involving the propagation of sound (seismic vibrations), electricity, or radar, for example, to infer the distribution of Earth properties that affect transmission. These include density for seismic vibrations, electrical resistivity for electricity, and the dielectric constant for radio waves (radar). These physical properties often can be correlated with hydraulic properties because the differences in rock type and structure that affect the physical properties also affect the hydraulic properties such as porosity and hydraulic conductivity. In a sense, use of geophysical techniques can be considered to be a way of "seeing into the Earth" (NRC, 2000). The array of quite sophisticated techniques used in geophysical investigations is extensive and a description of methods is beyond the scope of this book. Suffice it to say that the visualization of subsurface structure can provide significant information for understanding

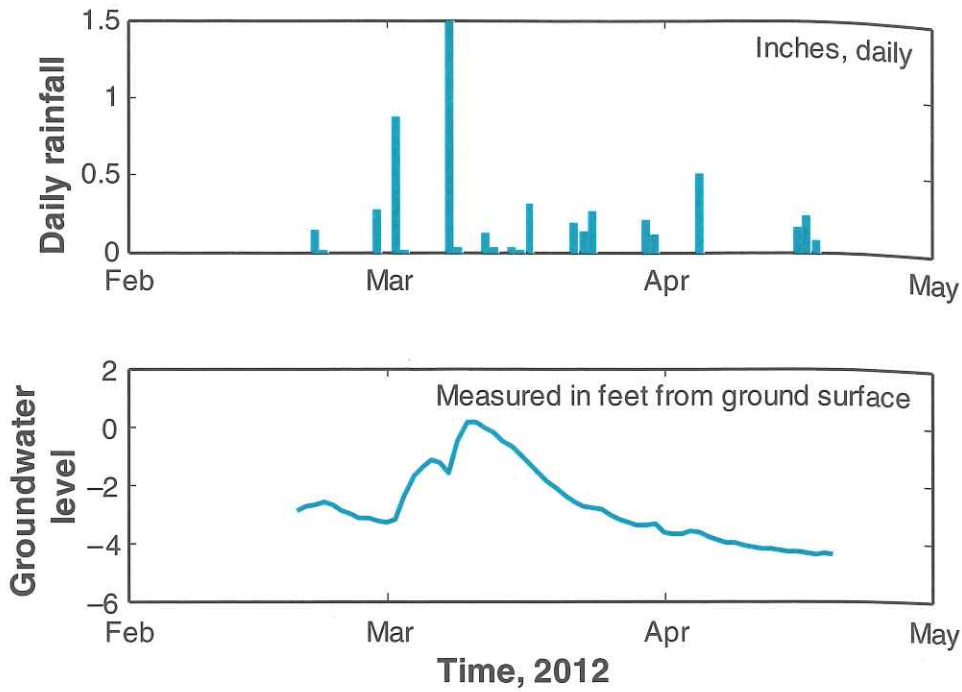

Figure 6.8 Well hydrograph for USGS Well 351428085003600, Hamilton County, Tennessee, USA.
Provisional data from USGS.

hydrogeology. Even a subtle underground feature may have important implications for the flow of water in soils and rocks (Figure 6.9).

6.5 Steady Groundwater Flow

The concepts developed in Section 6.3 can be applied readily to flow in natural formations. Consider first a simple example in which a confined aquifer is bounded by two channels (Figure 6.10). The height of the water in the channel on the left defines the hydraulic head at that boundary of the aquifer and the same is true on the right boundary. If the flow is steady and the confined aquifer is of constant thickness, b, then the specific discharge through the aquifer from $x = 0$ to $x = L$ is constant and we can write Darcy's law for flow in the x-direction as:

$$q = -K\frac{dh}{dx}. \tag{6.12}$$

We can rearrange this expression and integrate over the length of the confined aquifer to obtain an expression for hydraulic head, h:

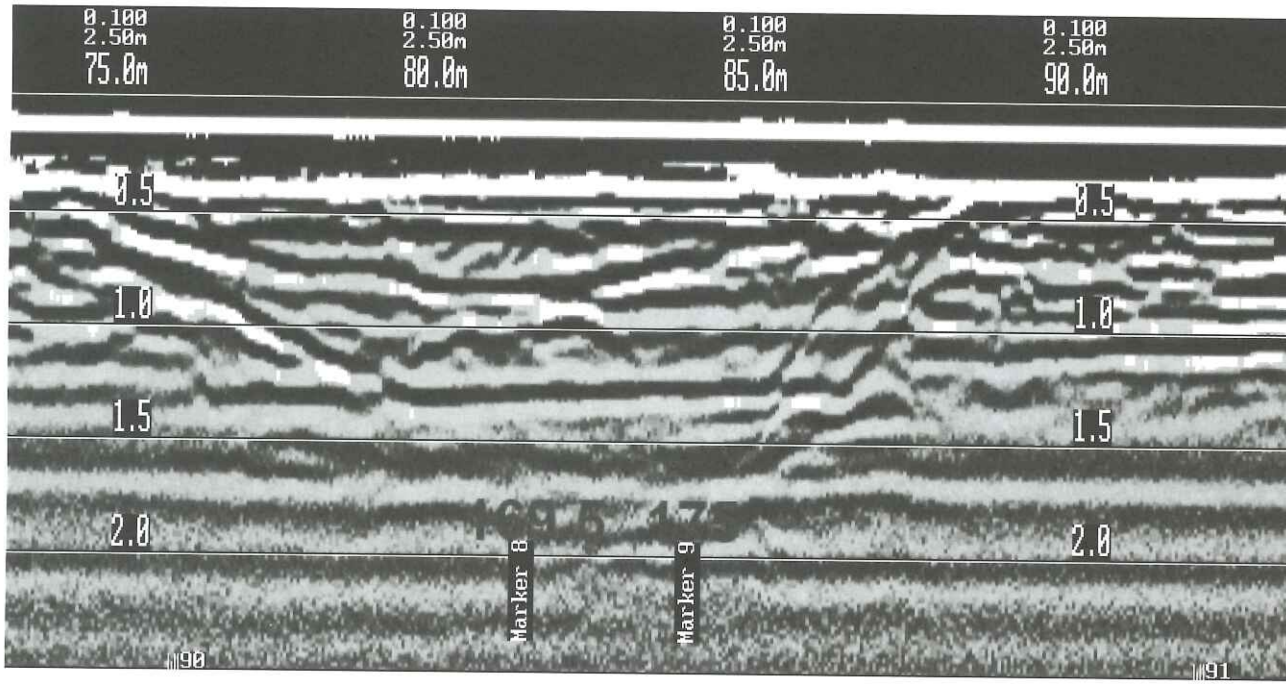

Figure 6.9 Trace from a ground-penetrating radar transect showing a buried channel. Such features often may be important for guiding groundwater flow.

Image courtesy of Alan D. Howard.

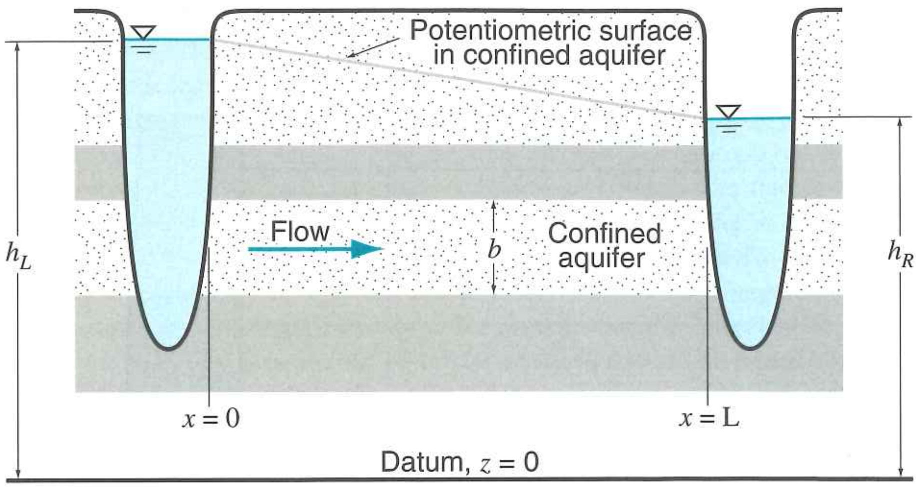

Figure 6.10 Horizontal flow in a confined aquifer.

$$dh = -\frac{q}{K}dx$$

$$\int_{h_L}^{h} dh = -\frac{q}{K}\int_{0}^{x} dx$$

$$h - h_L = -\frac{q}{K}x$$

$$h = h_L - \frac{q}{K}x. \tag{6.13}$$

Equation 6.13 specifies that head decreases linearly with distance from the left boundary; that is, the potentiometric surface is a plane sloping from left to right. This is shown as a solid gray line in Figure 6.10. Water flows in the downhill direction of this surface.

On a plan view of the channel-aquifer system considered above, the contour lines of the potentiometric surface are parallel to the channels (Figure 6.11). The spacing of the contours indicates the slope of this surface, just as do the contour lines of a standard topographic map. Once these lines of equal hydraulic head, or **equipotentials**, have been established, lines that indicate the direction of flow can be sketched in by constructing *perpendiculars* to the equipotentials because this represents the downhill direction of the potentiometric surface, the direction of flow specified by Darcy's law. These lines are called **streamlines**. Together, the equipotentials and the streamlines constitute a **flow net**. Flow nets can be applied to great advantage in actual field problems where groundwater flow patterns are to be established based on the measured water levels in a series of wells (see Fetter, 2000, for examples).

We will discuss the use of flow nets in greater detail in a later section of this chapter, but at this point we need to consider the physical reasoning behind the idea that streamlines

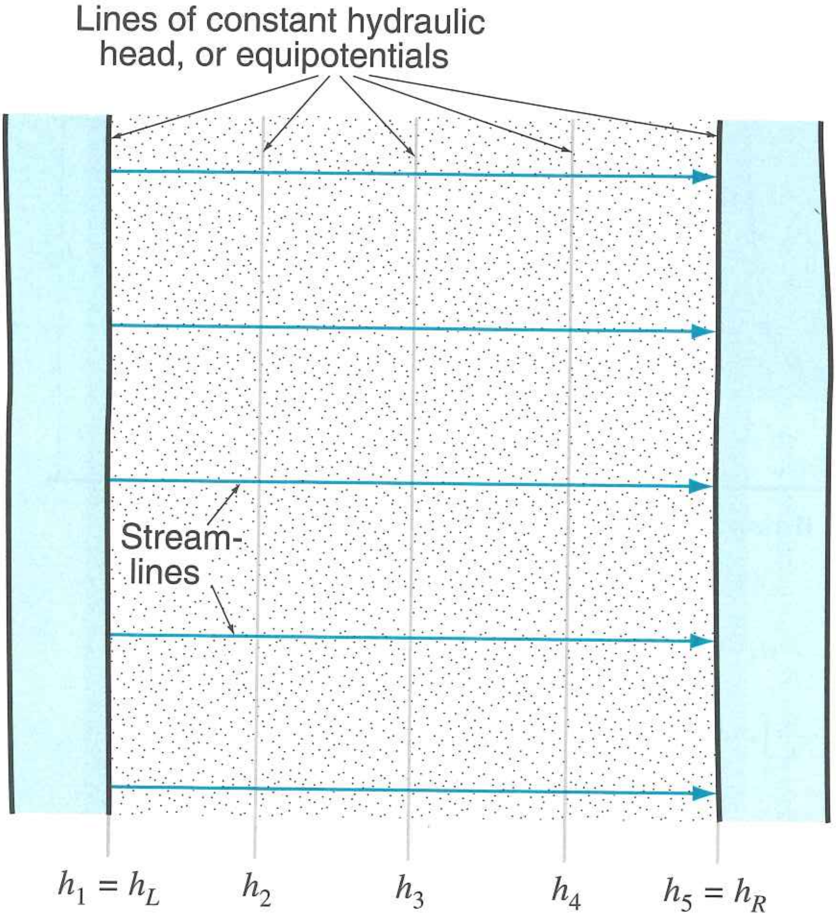

Figure 6.11 Plan view of the aquifer shown in Figure 6.10: a simple flow net. Note that the streamlines are perpendicular to the equipotentials.

must be perpendicular to equipotentials. Darcy's law indicates that flow should always be from high values of the hydraulic head to low values. In fact, flow must follow the *path of steepest descent*. Picture the potentiometric surface or the water table as a topographic surface, with hills (high h) and valleys (low h). Now imagine a drop of water moving along this surface in response to the pull of gravity. We have to imagine this water drop moving rather slowly, without the momentum we might expect a ball bearing or other object to have on a steep surface. The water drop moves down the slope and perpendicular to the "topographic" contours rather than following some arbitrary path down the slope; that is, it follows the path of steepest descent. The rate of descent is proportional to the slope, just as the specific discharge is proportional to the hydraulic gradient.

An example of the use of flow nets to obtain an understanding of hydrological phenomena is presented by Hubbert (1940). Consider an idealized valley in which the groundwater flow from the hillslopes is perpendicular to the axis of the valley (Figure 6.12). A flow net for a cross section through the hillslope shows flow along "U-shaped" paths from the ridge to the valley bottom. Although the homogeneity of the material and two-dimensional nature of the flow would never exist in nature, the general picture of flow is instructive. Note that water flow near the stream is *not* horizontal (as one might expect intuitively) but has a significant upward component. Groundwater flow is not concentrated near the water table with a large volume of stagnant water at depth—the flow patterns are actually such that flow occurs throughout the saturated zone. The fact that water flows along "U-shaped" paths is of practical and theoretical importance. The U.S. Geological Survey conducted a study on the presence and the movement of agricultural chemicals in shallow groundwater on the Delmarva (Delaware-Maryland-Virginia) Peninsula (Hamilton and Shedlock, 1992). They found nitrate (partly from applied fertilizers) at almost all depths sampled in the groundwater and attributed the patterns of contamination to land-use practices and to groundwater flow paths.

The construction of accurate flow nets by hand is a task requiring considerable practice. More commonly today, flow nets are constructed using numerical solutions to the equations governing groundwater flow. Regardless of how a flow net is constructed, it

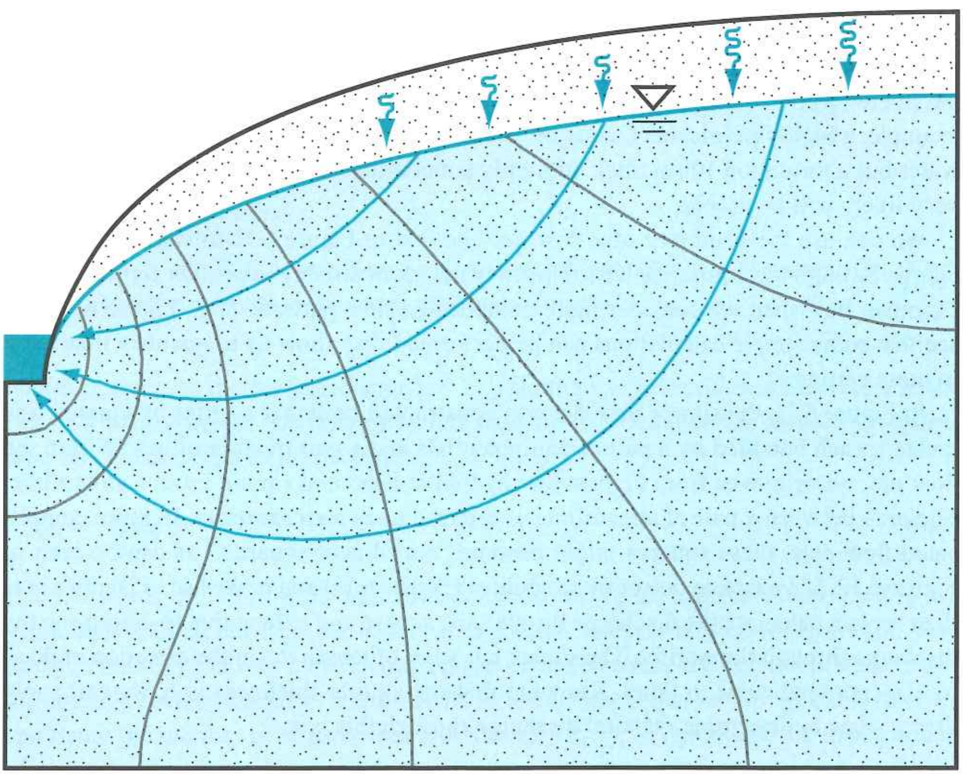

Figure 6.12 Flow net for two-dimensional groundwater flow through a hillslope.
Redrawn from Hubbert, 1940, fig. 45.

provides not only a visualization of the groundwater flow paths, but also information on the rate of groundwater flow in a particular region.

6.5.1 Quantifying groundwater flow using flow nets

Returning to the example of horizontal flow through an aquifer bounded by two channels separated by distance L (Figure 6.10), we find that

$$q = K\frac{h_L - h_R}{L}. \tag{6.14}$$

Because the hydraulic gradient, hydraulic conductivity, and aquifer thickness are constant, the value calculated using Equation 6.14 represents the specific discharge at any point in the aquifer. Quite often the total discharge in an aquifer, rather than specific discharge at a point, is of primary concern. Discharge per unit length of stream can be calculated from Equation 6.14 by multiplying by the thickness of the aquifer, b:

$$\textbf{Discharge per length} = Kb\frac{h_L - h_R}{L} = T\frac{h_L - h_R}{L}, \tag{6.15}$$

where $T = Kb$ [$L^2\ T^{-1}$] is called the **transmissivity** of the aquifer. This is an important parameter when considering the development of a water supply from an aquifer. For example, a highly permeable formation 10 mm thick may not provide a usable supply of water but a formation 100 m thick with only a moderate value of hydraulic conductivity very likely will be usable.

We can use the geometry of a flow net like that in Figure 6.11 to calculate the discharge through the aquifer as well. This may not seem terribly important for this simple example, but when we consider more complicated flow patterns (e.g., Figure 6.12) we see that some simple calculations can be performed that allow us to quantify flow even in these settings.

In Figure 6.11, the simple flow net was constructed such that a series of squares was created, each bounded by a pair of equipotentials and a pair of streamlines. Within the flow net, water moves from high to low hydraulic head and cannot cross a streamline. The area between a pair of streamlines is referred to as a **streamtube**. In more complicated flow nets, these squares might become "curvilinear squares," as can be seen in Figure 6.12. If we isolate one of these squares (Figure 6.13) and make use of Darcy's law, then we can calculate the discharge through the square and extend this to determine the discharge through the aquifer. The square has sides of length ds (in the direction of flow) by dm (perpendicular to flow). Knowing that the aquifer thickness is b, we can apply Darcy's law to determine the total discharge through this box:

$$Q_S = qA = K(dm\, b)\frac{dh}{ds}, \tag{6.16}$$

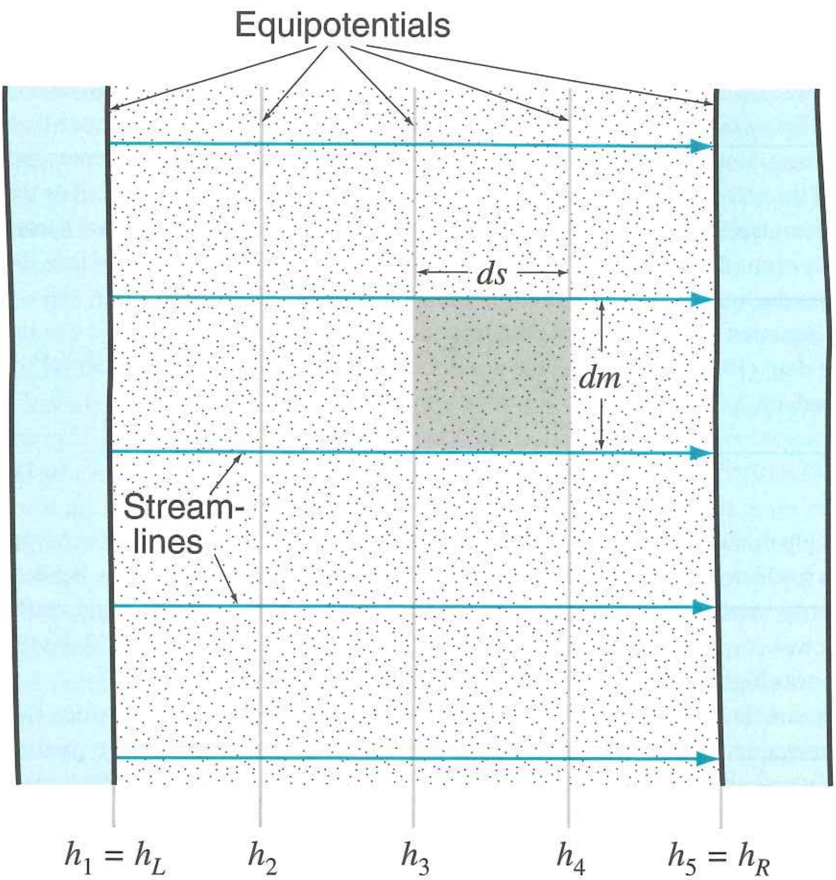

Figure 6.13 A simple flow net showing terms used to quantify flow in the aquifer. The gray square of dimension *ds* by *dm* is analyzed in the text.

where Q_s refers to the total discharge [$L^3\ T^{-1}$] through a streamtube and *dh* refers to the head difference across the box, or in this case $h_4 - h_3$. Because the domain is square, $dm = ds$, and equation (6.16) becomes:

$$Q_S = Kbdh. \tag{6.17}$$

In other words, if we know the hydraulic conductivity (or the transmissivity, $T = Kb$), we can simply look at our flow net to see what contour interval is used for hydraulic head (*dh*) and multiply that value by the transmissivity to determine the amount of water moving through each streamtube. If the aquifer is bounded at upper and lower ends, then we can count the number of streamtubes and multiply by Q_s to determine the total amount of water flowing through the aquifer.

We now look at a somewhat more complicated example. One effect of placing a dam in a stream or river is that a hydraulic gradient is created beneath the dam. At the upstream end, the hydraulic head (relative to the base of the dam) at the bottom of the reservoir is equal to the depth of water in the reservoir (Figure 6.14). On the downstream side, the hydraulic head is equal to the height of water in the river. A flow net can be constructed, as shown in Figure 6.14, to determine the pattern and rate of steady groundwater flow beneath the dam. Note that a low-permeability layer exists at depth, which represents the bottom of the aquifer beneath the dam. We can envision this boundary, as well as the base of the dam itself, as streamlines, because there will be relatively little flow across them. For this example, we assume that the dam is 100 m wide (in the direction into the page), and that the hydraulic conductivity of the material beneath the dam is 10^{-10} m s^{-1}. We can use Equation 6.17 to calculate the total discharge beneath the dam. We use the length of the dam (100 m) in place of the aquifer thickness (b). The contour interval for hydraulic head, dh, is 2 m. Then,

$$Q_S = Kbdh = 10^{-10}\ \text{m s}^{-1} \times 100\ \text{m} \times 2\ \text{m} = 2 \times 10^{-8}\ \text{m}^3\ \text{s}^{-1}. \tag{6.18}$$

We can multiply this value by the number of streamtubes (3) to obtain the total discharge, 6×10^{-8} m^3 s^{-1}, which is approximately equal to 1.9 m^3 yr^{-1}. In this case, the flow beneath the dam is fairly small, which is what we would hope for; the dam would be fairly inefficient if water was constantly leaking around it! A larger value would result if the hydraulic conductivity was higher, or the difference in head across the dam was greater.

Before we conclude this chapter with a discussion of some of the complexities that enter into our evaluation of groundwater flow, we mention several other points regarding the use of flow nets as we have described it. First, we have considered only *steady groundwater flow*. It is not possible to construct a flow net for unsteady or transient groundwater

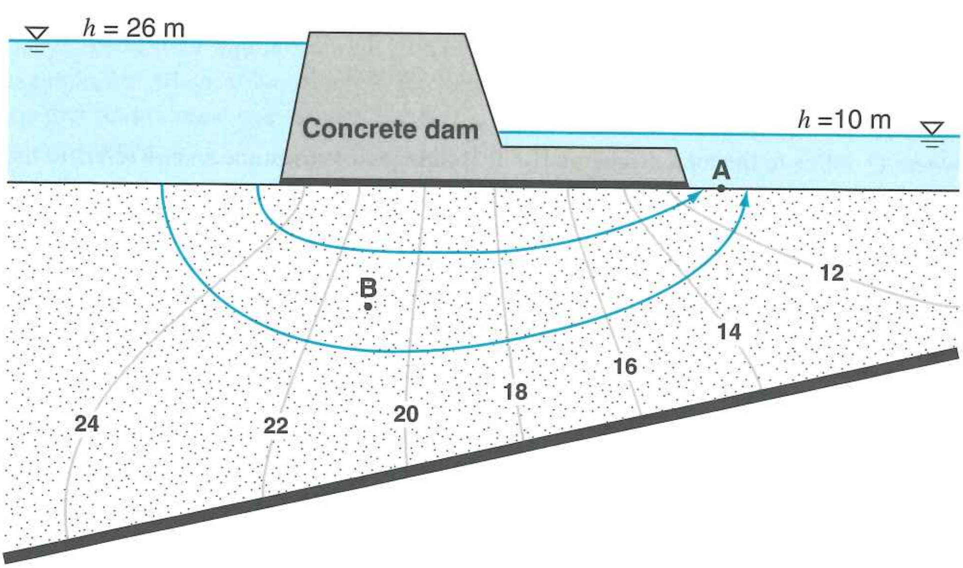

Figure 6.14 Flow net for groundwater movement beneath a concrete dam. The equipotentials (*gray lines*) are labeled with values of hydraulic head.

flow. Also, we can use a flow net to describe the variation in specific discharge, and, therefore, flow velocity, in an aquifer. You can imagine each streamtube as a "pipe," because water cannot cross a streamline. Therefore, from the principle of conservation of mass described in Chapters 3 and 4, we can state that within a streamtube, the specific discharge will be greatest where the streamtube is narrowest. The total discharge through the streamtube must be the same at any cross section. Referring to Figure 6.14, the flow velocity will be greater at point A than at point B. Finally, we have considered only homogeneous porous media, that is, materials with a spatially constant k. In the next section, we discuss the implications for flow through materials that are not homogeneous.

6.5.2 Heterogeneity and anisotropy

Virtually all natural materials through which groundwater flows display variations in intrinsic permeability from point to point. This is referred to as **heterogeneity**. The natural processes that create and modify rocks, sediments, and soils give rise to heterogeneity at all scales—from minor variations in grain size and the small holes called vugs in carbonate rocks (mm scale) to sedimentological and soil features (m scale) to variations in fracture spacing and lithological layering as depicted in Figure 6.6 (m to km scale). These variations in permeability from point to point complicate flow net construction. Permeable zones tend to focus groundwater flow, while, conversely, flow tends to avoid less permeable zones. It is possible to construct flow nets for some simple cases, such as layered aquifers and aquitards. However, when the permeability distribution is more complicated, hydrogeologists often rely on numerical simulation to depict the pattern and rate of groundwater flow.

Another complication often arises when measuring the permeability of natural materials. Rocks, sediments, and soils often have textural features that cause the permeability at a point to *vary with the direction of measurement*. Materials that display this trait are referred to as **anisotropic**, whereas **isotropic** refers to the condition in which the permeability does not depend on the direction of measurement. Consider a fractured rock aquifer, in which the fractures are predominantly horizontal. In this case the permeability will be higher when measured in the horizontal direction than in the vertical direction. You might picture this situation by referring to Figure 6.2 and imagining that the capillary tubes are aligned in the horizontal direction with relatively little hydraulic communication between them. Other features that produce anisotropy include (but are not limited to) the orientation of: platy minerals and small-scale layering in sedimentary rocks, such as clays; cooling cracks and lava flow tubes in basalts; large pores due to animal burrowing and plant roots in soils; and schistosity and fractures in metamorphic and igneous rocks.

Anisotropy, like heterogeneity, makes the job of constructing flow nets difficult, so that once again we must rely on numerical models that are capable of including this aspect of natural porous media. In the case of anisotropy, because water will tend to flow in a "preferred" direction, that is, in the direction of maximum permeability, Darcy's law must be modified to include this preference. As a result, in an anisotropic medium, the streamlines and equipotentials may not be perpendicular to one another at all points.

In introducing the concept of a flow net in Section 6.5, we described streamlines as paths of steepest descent in the "topographic landscape" of hydraulic head. We can extend this analogy to anisotropic media if we now imagine a series of ridges running at some angle with respect to the equipotentials. These ridges cause the water drop to move at some angle other than straight "downhill," since the water wants to follow the ruts between the ridges at least part of the time. (The flow will be straight downhill if the ridges are either straight downhill or follow the equipotentials.) These ridges in the hydraulic "topography" have the same effect as anisotropy in porous media. The flow direction will be altered from the normal direction parallel to the hydraulic gradient toward the direction of maximum intrinsic permeability.

6.6 Concluding Remarks

Darcy's law (and the law of conservation of mass implied in the construction and use of flow nets) provides the basis for computing steady groundwater flow patterns and rates. In some cases a simple "back-of-the-envelope" calculation is sufficient to gain a rough estimate of flow rates. In other cases, the aquifer may have a complex geometry or be heterogeneous and anisotropic and simple calculations will not suffice. Today hydrologists routinely use groundwater models (by which the groundwater flow equations are solved with the assistance of computers) to address environmental issues (e.g., Konikow et al., 2006). The computational techniques used in groundwater models are much more sophisticated than those we have used in this chapter. However, the ideas behind the computation are essentially the same. The geometry, conditions at the boundaries, and hydraulic parameters (intrinsic permeability, porosity) of the aquifer must be specified before predictions can be made with the models.

One use of groundwater models is in making assessments of compliance with environmental regulations. For example, before the WIPP site was licensed to begin operation, a set of regulations issued by the U.S. Environmental Protection Agency (EPA) had to be met. For WIPP, the U.S. Department of Energy had to show that the probability of significant releases to the accessible environment over 10,000 years into the future will be very small. The "accessible environment" for WIPP means groundwater in the Culebra formation "down gradient" of WIPP, that is, in the direction of decreasing hydraulic head. The demonstration of compliance with the EPA standard is done using a performance assessment analysis (Helton et al., 1997). In short, performance assessment uses a set of *scenarios*—sequences of hypothetical events that might occur in the future—and models to predict the impact of these scenarios. For WIPP, one scenario that is of concern involves someone in the distant future drilling a well into the repository and allowing some of the radioactive wastes to flow up the well and into the Culebra dolomite. A groundwater model must be used to "route" the hypothetical contaminant through the aquifer to decide whether significant quantities of contaminant might reach a human population in this scenario. These and many other environmental problems require the kind of knowledge of groundwater hydraulics that we have introduced in this chapter.

7 Groundwater Hydrology

7.1 Introduction

Concerns about groundwater arise in many settings, including, for example, developing plans for sustainable water uses in urban areas with limited surface water. The question of how to deal with the resource problems associated with urbanization—including water resources—is one that taxes responsible governments in almost every part of the world. In the United States, for example, more than 40% of the water used by people in Orange County south of Los Angeles comes from groundwater. Because precipitation and surface runoff vary with the seasons, Orange County uses spreading basins (gravel-lined areas) over which they spread water so it can infiltrate to recharge the groundwater that is in high demand. During most years there is not enough natural flow in the rivers to accommodate the demand for recharge. In those years, reclaimed wastewater may be used for recharge. For example, in the city of Anaheim the Santa Ana River is diverted into spreading basins to allow the water to infiltrate and recharge the underlying aquifer. In the summer over 90% of the flow in the river is treated wastewater that is returned to the river upstream of the spreading basins (NRC, 2012c). Trade-offs between amount and quality of water are inevitable and careful plans must be made to sustain the resource.

One example of a difficult problem in the use and protection of groundwater resources can be found in Mexico City, which is now one of the world's "megacities." Water must be supplied—and wastewater treated—for about 20 million people. Usable surface water is scarce in the Mexico City Basin, so groundwater from the Mexico City Aquifer is a primary source of freshwater. Near Mexico City, the aquifer is a sequence of alluvial fill sediments interstratified with basalt deposits and overlain by clays. The principal aquifer is from 100 to 500 m thick and the overlying clay aquitard is approximately 100 m thick. The water table below Mexico City has been declining at a rate of 1 to 1.5 m per year; the rate at which water is being withdrawn by pumping wells is much greater than the natural rate of replenishment. The practice of withdrawing groundwater faster than it can be replenished, referred to as "overdraft," has been common since about 1900. Furthermore, poor waste-disposal practices have adversely affected water quality in the aquifer. Clearly, if the water resource for Mexico City is to be made sustainable, changes in the pattern of use based on an understanding of the basin hydrology will be required.

In the United States, the High Plains Aquifer (often referred to as the Ogallala Aquifer) has experienced similar problems (McGuire, 2011). The High Plains Aquifer is an important source of water for much of the central United States, including parts of Colorado, Kansas, Nebraska, New Mexico, Oklahoma, South Dakota, Texas, and Wyoming. About 20 percent of the irrigated land in the United States is in the High Plains, and about 30 percent of the groundwater used for irrigation comes from the High Plains Aquifer. Overdrafts of groundwater have resulted in a decline in water levels throughout much of the aquifer. Between 1950 and 2009, the decline was more than 4 m (on average) and exceeded 45 m in some places. Declining water levels (which are expected to continue into the future) increase the cost of water, because pumps require more energy to lift the water a greater distance.

In places such as the High Plains, Mexico City, northern India, the North China Plain, and the southeast of Spain, groundwater is a limited resource and is being depleted (Wada et al., 2010). In other words, the rate of groundwater withdrawal exceeds the rate at which it is naturally replenished. As a result there is a net reduction in groundwater storage. In some cases, modern societies are using groundwater that accumulated in aquifers over geological time scales and under different climate conditions. This unsustainable use of non-renewable groundwater resources is sometimes called **groundwater mining**. The water removed from the ground is ultimately released to the ocean, and contributes—in addition to the effects of climate change—to sea-level rise (Wada et al., 2010; Konikow, 2011). Global estimates of groundwater depletion vary between 27 $km^3\ yr^{-1}$ (Margat et al., 2006) and 283 $km^3\ yr^{-1}$ (Wada et al., 2010). Table 7.1 shows an intermediate estimate (145 $km^3\ yr^{-1}$) along with the major hotspots of groundwater depletion in the world. As a term of comparison, globally, the irrigation water used by crops is about 545 $km^3\ yr^{-1}$ (Siebert et al., 2010).

To understand limitations on the sustainable use of groundwater, we must examine the water balance for the subsurface. How and at what rate is groundwater replenished or recharged? Hydrological basins (the rocks, sediments, and soil underlying catchments) exhibit natural rates of groundwater flow as a result of the infiltration of precipi-

Table 7.1. Total net groundwater depletion and average groundwater depletion in 2001–2008

Aquifer	**Net depletion in 2001–2008 (km^3)**	**Average annual rate ($km^3\ yr^{-1}$)**
In the USA		
Atlantic Coastal Plain	2.8	0.3
Gulf Coastal Plain	67.4	8.4
High Plains (Ogallala) Aquifer	94.7	11.8
Central Valley, California	31.4	3.9
Western Alluvial Basins	2.1	0.3
Western Volcanic Systems	2.9	0.4
Deep Confined Bedrock Aquifers	2.6	0.3
Agricultural and Land Drainage	0	0
Total (all USA systems)	203.9	25.5
Non-USA Aquifer Systems		
Nubian Aquifer System	18.9	2.4
North Western Sahara Aquifer System	17.6	2.2
Saudi Arabia Aquifers	109.1	13.6
North China Plain	40.0	5.0
Northern India and Adjacent Areas	423.5	52.9
Indirect Estimates for Other Areas	350.0	43.7
Total Global	1163.0	145.4

Source: According to Konikow (2011).

tation, exfiltration or discharge of groundwater, and evapotranspiration. Flow nets provide a tool for evaluating the natural rate of groundwater flow. How is that flow altered by perturbations such as pumping? Groundwater pumping in the Mexico City Basin has changed the pattern of flow as well as the water balance. Because the flow pattern determines how pollutants move within the subsurface, flow nets are also important in evaluating the risk of contamination.

Our best information about the water balance of an aquifer system, information that is essential for water resources planning, usually comes from records of water levels and groundwater pumping. This is one of the reasons why groundwater hydrology, as interpreted from flow nets and water-level records, is an important area of study.

7.2 Flow Nets and Natural Basin Yield

Flow nets can be constructed for any setting where the approximations of steady and two-dimensional flow are valid, as long as the conditions at the boundaries and the distribution

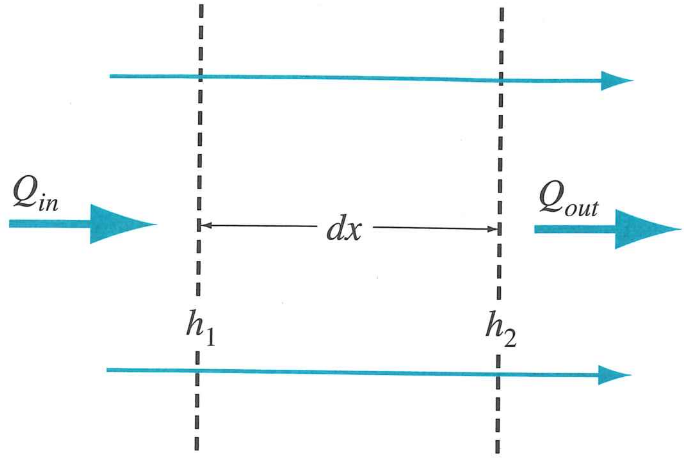

Figure 7.1 Definition sketch for the derivation of the steady groundwater flow equation.

of hydraulic conductivity are known. Later in this chapter we explore cases where flow is not steady. For the present discussion, we consider steady flow to be an approximation to the real system, such that temporal variations in the height of the water table are small relative to the thickness of the flow system.

Flow net construction is based on a mathematical model of groundwater flow. The form of this model is a differential equation that describes the physical process of groundwater motion. We can develop such an equation using conservation of mass and Darcy's law. Consider a single curvilinear square within a flow net as a control volume (Figure 7.1). This volume is bounded by two equipotentials and two streamlines. The streamlines can be thought of as impermeable boundaries, because no water crosses them. For steady flow, the inflow (Q_{in}) must equal the outflow (Q_{out}), or the change in storage must be zero. For the streamtube (the area bounded by a pair of streamlines), this may be written:

$$\frac{Q_{out} - Q_{in}}{dx} = \frac{dQ}{dx} = 0,$$

where Q [$L^3\ T^{-1}$] refers to the total discharge at any section through the streamtube. For uniform flow in the x-direction (both streamlines are horizontal), the specific discharge, q_x, also must be a constant:

$$\frac{dq_x}{dx} = 0. \tag{7.1}$$

Darcy's law may be written for the specific discharge as:

$$q_x = -K\frac{dh}{dx}, \tag{7.2}$$

where dh is equal to $h_2 - h_1$. Substituting Equation 7.2 into Equation 7.1 and assuming that K does not vary in space (i.e., the material is homogeneous):

$$\frac{dq_x}{dx} = -K\frac{d}{dx}\left(\frac{dh}{dx}\right) = 0$$

or

$$\frac{d^2h}{dx^2} = 0. \tag{7.3}$$

In two spatial dimensions, this would become:

$$\frac{\partial^2 h}{\partial x^2} + \frac{\partial^2 h}{\partial y^2} = 0. \tag{7.4}$$

Equation 7.4 is referred to as the **Laplace equation** and may be solved to determine the distribution of hydraulic head (the equipotentials) in two dimensions, provided that the region is homogeneous (constant K). Of course in the two-dimensional case, the equipotentials (and streamlines) are not straight parallel lines as in the one-dimensional case in Figure 7.1. The conditions at the boundaries of a domain are required to solve this equation. A flow net may be thought of as a graphical solution to the Laplace equation. The Laplace equation is simply a mathematical statement of the law of conservation of mass (Equation 7.1) combined with Darcy's law (Equation 7.2). It is now common for the Laplace equation to be solved using computers. Computer-based groundwater models have become important tools for solving problems of groundwater hydrology and contaminant transport.

For steady flow in two dimensions through a hillslope, conservation of mass requires the inflow of water to be balanced by the outflow, because there can be no change in storage. In other words, the same number of streamtubes must leave the hillslope as enter it. This idea gives rise to naturally occurring **recharge** and **discharge areas**. A recharge area occurs where water is crossing the water table downward, hence, recharging the groundwater system. A discharge area occurs where groundwater is moving upward across the water table, thereby discharging into the unsaturated zone above, or to the land surface or a surface-water body such as a lake or stream. Consider a hillslope bounded by an impermeable bottom (for example, the bedrock beneath an unconfined aquifer) and two impermeable sides, called **groundwater divides**. The upper boundary is the water table. Often, the water table is a subdued replica of the land surface topography, such that the water table is higher beneath hilltops than beneath a valley. A flow net shows a balance between recharge in the upland portion and discharge in the valley (Figure 7.2a).

Flow nets must obey certain rules at the boundaries. For example, it is probably apparent that streamlines cannot cross impermeable, or "no-flow," boundaries. Equipotentials must be perpendicular to impermeable boundaries, because otherwise some flow across the boundary would be implied. A constant-head boundary is represented by an

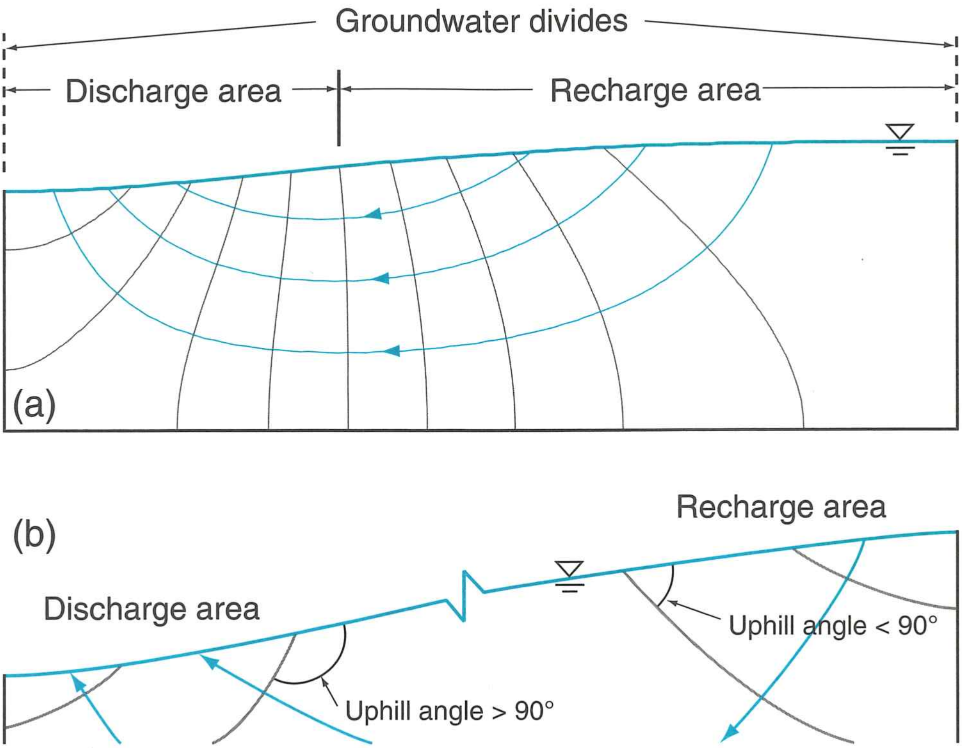

Figure 7.2 The natural pattern of groundwater flow in a simple basin (*a*). Equipotentials are gray and streamlines are blue. Recharge and discharge areas may be distinguished by looking at the angle that an equipotential makes with the water table (*b*).

equipotential, and streamlines must be perpendicular to a constant-head boundary. At the water table, the hydraulic head is everywhere equal to the elevation, because the gage pressure is zero. Equipotentials and streamlines may both intersect a non-horizontal water table, and the orientation of these lines provides a means of delineating recharge and discharge areas (Figure 7.2b).

The hydraulic head along any equipotential is a constant by definition. One way of labeling an equipotential is by noting where it intersects the water table (Figure 7.3). Water in a piezometer that is open at a particular depth will rise to an elevation equal to the hydraulic head at that point. For example, in Figure 7.3 the piezometer is open at the depth depicted. . The elevation head (z) at that point is 20 m. Because the open interval intersects the 40-m equipotential, the hydraulic head (h) at that point is 40 m. The pressure head ($p/\rho g$), therefore, is equal to $h-z$, or 20 m.

Flow nets have a "dimensionless" quality. They are a picture of the *pattern* of steady groundwater flow and depend only on the physical features of the basin. For example, the pattern shown in Figure 7.2a does not depend on the actual size of the basin or the (homogeneous) permeability of the basin rocks or sediments. However, quantities determined from the flow net *do* have particular dimensions (and units). Recalling the discus-

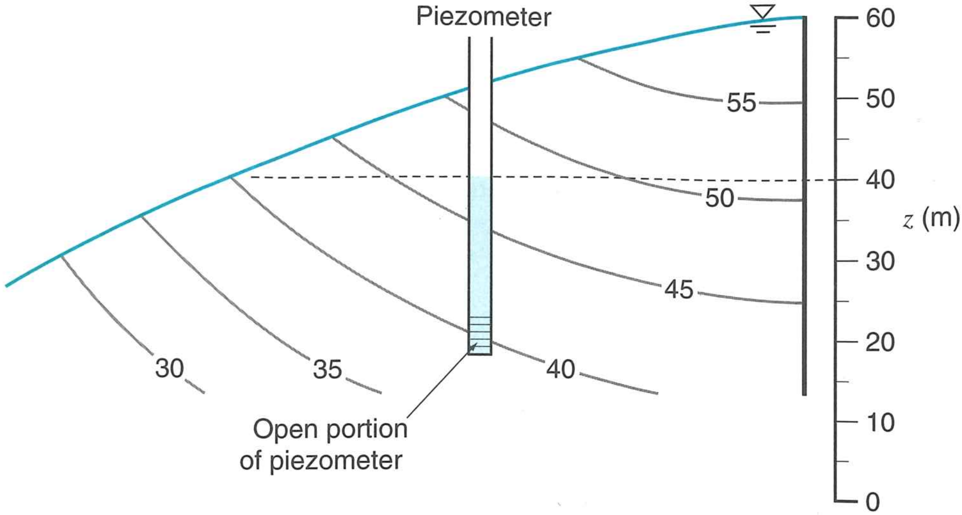

Figure 7.3 The hydraulic head along any equipotential is equal to the elevation of its intersection with the water table.

sion in Chapter 6 on quantifying flow using flow nets, the discharge through each streamtube may be calculated knowing only the hydraulic conductivity and the difference between adjacent equipotentials (*dh*):

$$Q_S = Kbdh. \tag{6.17}$$

Equation 6.17 is valid as long as the flow net has been constructed using curvilinear squares. For a horizontal flow net, *b* is the aquifer thickness (see Chapter 6). The total volume of flow through a vertical cross section depends on the cross section width, however, and so *b* refers to that width. If we let $b = 1$ unit [L], the discharge calculated from Equation 6.17 is simply the discharge *per unit basin width.*

Consider how we might use the flow net shown in Figure 7.2a. We would have to know the length of the basin, for example $L = 100$ m, in which case the hill is 30 m ($0.3L$) high on the right side. The difference in hydraulic head between adjacent equipotentials is equal to $0.005L$, or 0.5 m. If the hydraulic conductivity is 0.5 m day^{-1} (approximately 5.8×10^{-6} m s^{-1}), the discharge through each streamtube is:

$$Q_S = Kbdh = (0.5 \text{ m day}^{-1})(1 \text{ m})(0.5 \text{ m}) = 0.25 \text{ m}^3 \text{ day}^{-1}. \tag{7.5}$$

There are four streamtubes in the flow net. Therefore, the total discharge through the section is 4×0.25 or 1.0 m^3 day^{-1}. This is the discharge per meter basin width ($b = 1$ m). If the basin is 500 m wide, then we could simply multiply 1.0 m^3 day^{-1} by the 500-m width to find the total discharge from the basin (500 m^3 day^{-1}).

The above calculation shows that the basin discharge is proportional to the hydraulic conductivity of the basin material. Knowledge of the permeability of basin materials is essential to determining the water balance for a groundwater system. Freeze and Witherspoon (1968) referred to a calculation such as that above as the **natural basin yield**. Under natural or undisturbed conditions (i.e., in the absence of anthropogenic groundwater withdrawals or changes in climate or vegetation), this is the average rate of discharge from a hillslope or basin.

7.3 Regional Groundwater Flow

The simple flow net above (Figure 7.2a) provides a template for understanding regional groundwater flow. However, flow patterns are conditioned by variation in the shape of the basin and the water table, and spatial patterns of hydraulic conductivity. This section explores the primary controls on the pattern and rate (natural basin yield) of steady groundwater flow in a basin bounded by divides at the sides, the water table at the top, and a low-permeability unit at the base. The terms introduced in the previous section (recharge and discharge areas, natural basin yield) provide some keys to exploring the importance of each of these controls.

7.3.1 The effect of basin aspect ratio

Hydrological basins occur in a variety of shapes, and we would expect the basin shape to exert some influence over the groundwater flow pattern. Tóth (1962, 1963) examined the influence of **basin aspect ratio** (length to depth) on the pattern of groundwater flow in a homogeneous two-dimensional basin with a gradual water-table slope. Imagine a basin of constant length (L), but with a depth that might vary, depending on the depth to a low-permeability unit, such as crystalline bedrock. The pattern of flow will be similar to that shown in Figure 7.2a, with single recharge and discharge areas. Our experience with Darcy's law should tell us that the natural basin yield will be large for deeper basins (smaller aspect ratio), because the volume through which groundwater is flowing is larger. This assumes that we are comparing basins with identical water-table profiles. An analogy might be constructed using columns of different cross-sectional area. For a given hydraulic gradient (analogous to the slope of the water table in a basin), the discharge will be greater through a larger column. Recall that Darcy found that, for the simple case of one-dimensional flow, the total discharge was proportional to the cross-sectional area.

Compare Figure 7.4a and Figure 7.4b. The natural basin yield in each case is directly proportional to the number of streamtubes and to the hydraulic conductivity of the basin materials. For the shallow basin (large aspect ratio, Figure 7.4a), there are only two complete streamtubes and a small fraction of a third. For the deep basin (small aspect ratio, Figure 7.4b), almost eight complete streamtubes occur. This observation matches our expectation, although the total discharge is not perfectly proportional to the basin depth. In a catchment setting, we might expect the thickness of permeable material near the surface to have a big influence on the rate of groundwater flow. Where thin soils are found, the rate of groundwater flow might be relatively small.

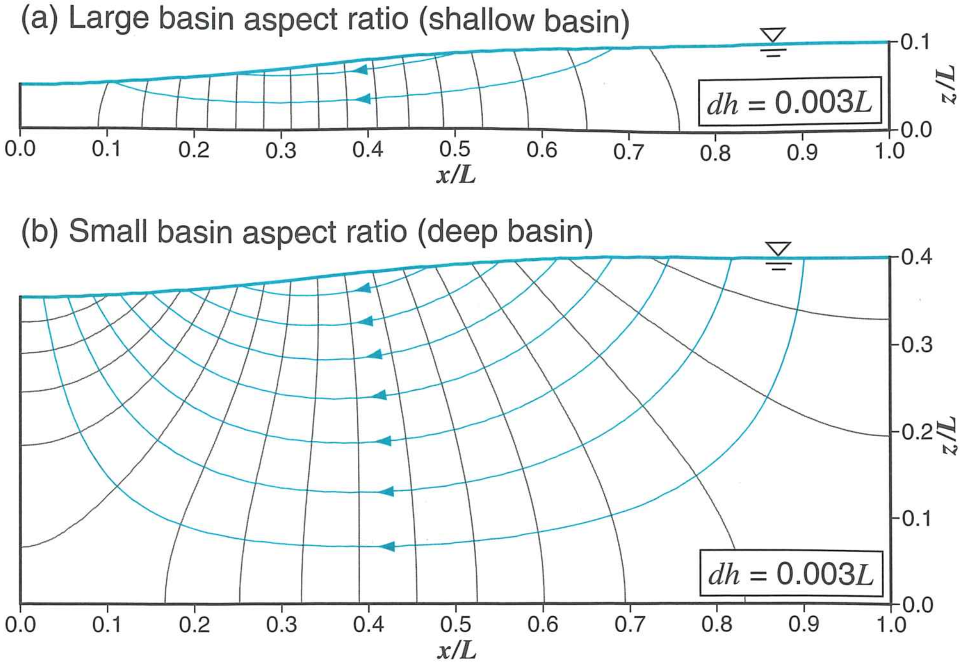

Figure 7.4 The effect of basin aspect ratio (length to depth) on natural patterns of groundwater flow. The water-table profile is the same for the shallow (*a*) and deep (*b*) basins.

There is some difference between the patterns of flow in the two basins in Figure 7.4. For the deep basin (Figure 7.4b), vertical hydraulic gradients (and, therefore, vertical flow) exist over a large portion of the basin. In the shallow basin (Figure 7.4a), the flow is essentially horizontal over most of the basin.

7.3.2 The effect of water-table topography

Variation in hydraulic head resulting from topographic relief of the water table is in most instances the driving force for groundwater flow. Complex land-surface topography should produce similarly complex water-table topography, and so we need to consider the flow systems produced by such topography. Although real topography can be quite complex, we might consider a general picture in which "local" topography (small-scale undulations or hills and valleys) is superimposed on a "regional" slope. The end-member cases are shown in Figures 7.4 and 7.5, for regional and local topography, respectively. These patterns are similar in some respects, because flow occurs from highs to adjacent lows. A single hill-valley flow system (dashed boxes in Figure 7.5) resembles the larger flow systems produced by regional water-table topography (Figure 7.4). These flow systems appear to have a lower depth limit, below which flow is very slow or non-existent, as may be seen particularly well in Figure 7.5b. However, the flow doesn't actually become zero, but only slows as streamtubes become wider and the hydraulic gradient along the

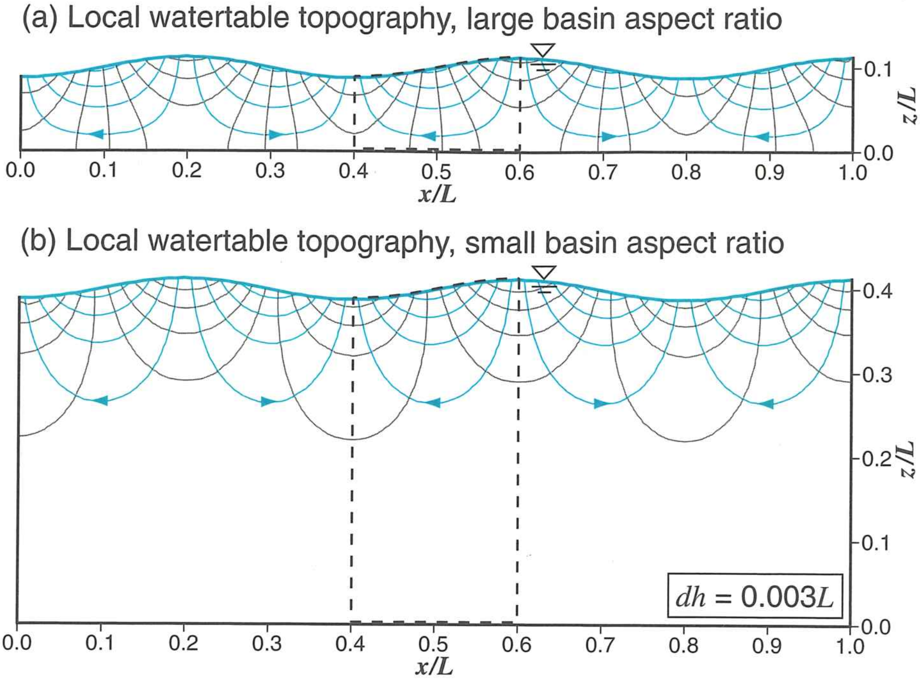

Figure 7.5 Regional groundwater flow patterns for the case of local water-table topography. The dashed boxes indicate an individual local flow system. The same equipotential spacing (dh) of $0.003L$ is used for both the large (a) and small (b) basin aspect ratios.

streamlines decreases. It could also be shown that the apparent "depth" of these local flow systems depends on the local hydraulic gradient.

Tóth, in the same studies cited earlier, referred to these flow cells (dashed boxes in Figure 7.5) in which water flows from topographic high to adjacent low as **local flow systems**. Larger flow systems, from regional high to low, he referred to as **regional flow systems**. The similarity in appearance between these types of flow systems suggests that the use of different terms to describe them depends only on one's definition of "regional." However, the distinction becomes clearer when we superimpose local hill-and-valley topography on top of a regional slope (Figure 7.6). For the case of a deep basin (Figure 7.6b), both local and regional systems develop (streamtubes labeled "L" and "R" in the figure), as well as what Tóth referred to as an **intermediate flow system** (streamtube labeled "I" in the figure). If the basin is relatively shallow (Figure 7.6a), a regional system may still exist, but may be attenuated as a result of the dominant influence of the local flow systems.

The conclusions that we can draw from these flow nets are as follows:

1. If local relief is negligible, but a regional water-table slope exists, only a regional flow system will develop.
2. Conversely, if local hill-and-valley topography exists, but no regional slope, only local flow systems will develop.

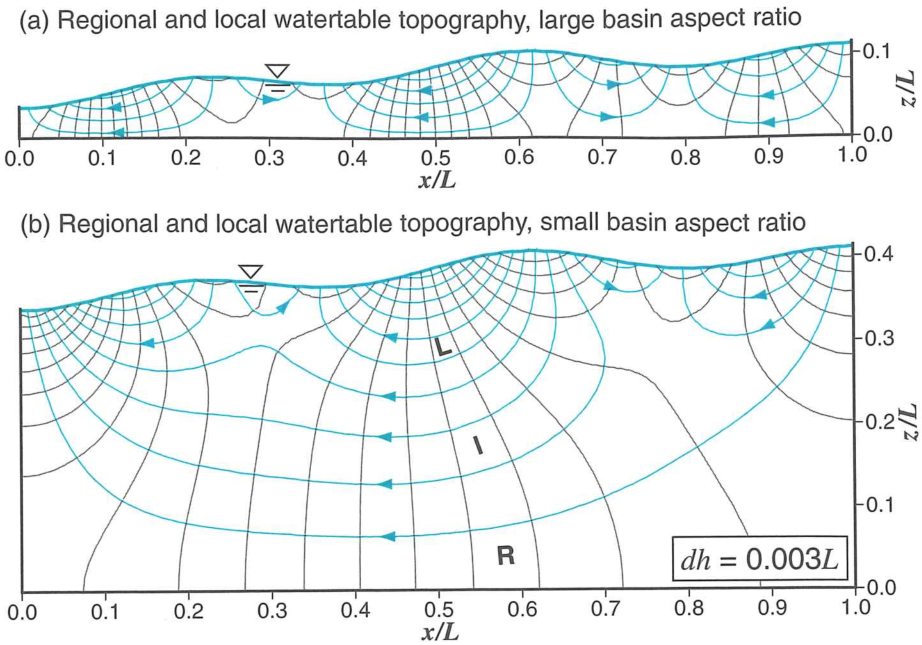

Figure 7.6 Groundwater flow patterns for the case of combined regional and local water-table topography. Streamtubes labeled "L," "I," and "R" indicate the local, intermediate, and regional flow systems, respectively. The same equipotential spacing (*dh*) of 0.003*L* is used for both the large (*a*) and small (*b*) basin aspect ratios.

3. If both local and regional topography exists in a basin, all three types of flow systems (local, intermediate, and regional) will develop. As a result, precipitation infiltrating on a hilltop may eventually discharge at an adjacent low or follow a longer flow path toward the regional low point.

Complex water-table topography will necessarily produce complex patterns of groundwater flow, but the general observations discussed above can be applied to other settings. Referring to Figure 7.7, can you identify local, intermediate, and regional flow systems?

7.3.3 The effect of geological heterogeneity

Geological materials are always heterogeneous, a fact that complicates analysis. Aquifers and other permeable zones in the subsurface are capable of capturing and focusing groundwater flow. In some cases, recharge and discharge areas may develop in locations that are not predicted from the configuration of the water table. Hillside springs are one good example. Freeze and Witherspoon (1966, 1967) performed a series of numerical "experiments" in which they calculated flow nets for a variety of configurations of the water table and subsurface geology. These calculations demonstrated the ability of subsurface aquifers to alter the flow system (Figure 7.8).

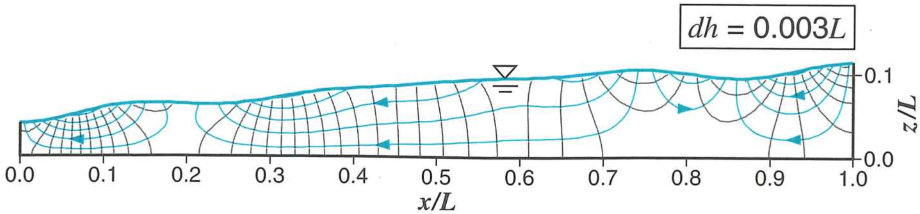

Figure 7.7 A flow net for a shallow basin with complex water-table topography.

A highly permeable unit such as an aquifer causes flow to become more nearly vertical within the overlying unit, and may narrow the discharge area (Figure 7.8b). Aquifers act as conduits that are capable of rapidly transmitting water to the principal discharge area. Because the mean hydraulic conductivity of the basin is greater, the natural basin yield is increased (compare Figure 7.8a with Figure 7.8b). A large portion of this yield is transmitted through the aquifer. An aquitard at depth effectively reduces the basin depth and natural basin yield (Figure 7.8c). Discontinuous zones of high permeability, such as the truncated aquifer shown in Figure 7.8d, may alter the areas of recharge and discharge. The discharge produced at the end of the aquifer may appear as a surface spring or seep.

Hsieh (2001) produced a computer model that is useful for demonstrating the influence of the topography of the water table and heterogeneity on flow nets. Once the model is downloaded (http://water.usgs.gov/nrp/gwsoftware/tdpf/tdpf.html), it can easily be used to explore a variety of different flow conditions to gain insights about recharge and discharge areas of groundwater flow systems.

Mexico City lies within the discharge area for the Basin of Mexico, which is a "closed basin." This means that there are no natural surface water outflows for the basin, and all outflow occurs through evapotranspiration. The Basin is ringed by volcanic mountains that form the recharge areas for the Basin. Because of the heterogeneity of the materials filling the basin, an aquifer-aquitard system exists, and the pattern of flow from the mountains to the center of the basin looks something like Figure 7.8b. Prior to heavy pumping during the past 100 years, groundwater flowed from the recharge areas (mountains) into discharge areas in the Basin center through the aquifers, and discharged upward across the overlying aquitards. Surface springs and lakes resulted from this natural pattern of groundwater flow.

Potable groundwater under artesian pressures (i.e., heads that caused water flow to the surface in wells without pumping) was discovered in the Basin of Mexico in 1846, promoting the rapid development of this resource (NRC, 1995b). Since that time, the large overdrafts of groundwater have reduced the pressures within the confined aquifers and altered the natural pattern of groundwater flow. Now most of the water within the Basin is moving downward toward the heavily pumped aquifers; surface springs and lakes have dried up. In 1983, systematic monitoring of the water levels in wells began. These records of water levels in wells have provided important information on the water balance in the basin.

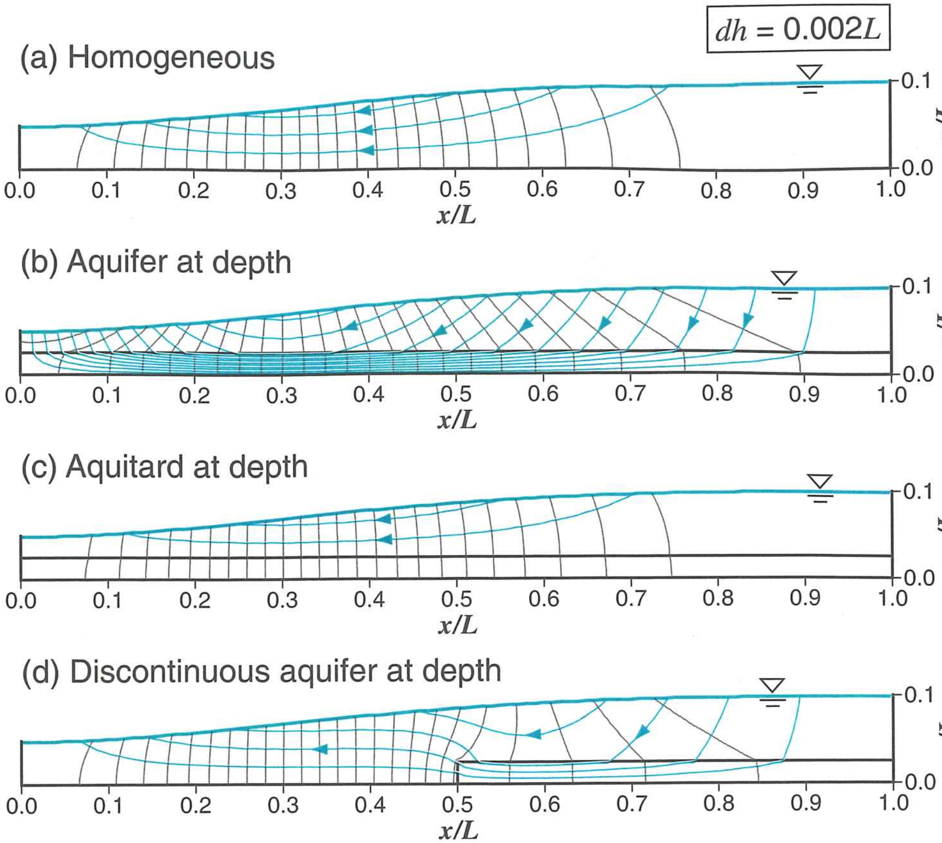

Figure 7.8 The influence of geological heterogeneity on patterns of groundwater flow. The homogeneous case (*a*). An aquifer at depth (*b*; hydraulic conductivity of the aquifer is an order of magnitude greater than the overlying unit). An aquitard at depth (*c*; hydraulic conductivity an order of magnitude less than the overlying unit). A discontinuous aquifer at depth (*d*). The equipotential spacing (*dh*) is the same for all, 0.002*L*.

7.4 Well Hydrographs

The three major controls on patterns of groundwater flow described in the previous section concern *natural* groundwater flow. The natural system is assumed to be at steady state, such that recharge and discharge balance one another and the water table is approximately constant. Groundwater systems, however, are both dynamic (although sometimes sluggish) and subject to human alteration, as we have seen for the Mexico City region. The task of depicting patterns of flow in these circumstances is a difficult one. We often have observations, in the form of well hydrographs, that allow us to gain some understanding of these effects. As we will see in the next sections, confined and unconfined aquifers behave somewhat differently, and we consider each separately.

7.4.1 Unconfined aquifers

A **well hydrograph** shows the variation in water level in a well through time. In an unconfined aquifer, the water level in a well generally indicates the position of the water table. Well hydrographs may show variations over different time scales. For example, daily fluctuations may be observed in carefully monitored wells, and very gradual changes over several years of observation may also be evident. The most obvious cause of water-level variations is fluctuation in the inputs and outputs. In Coffee County, Tennessee, the water-table elevation is observed to vary with the season (Figure 7.9). Precipitation in Tennessee is not very seasonal (monthly precipitation averages are relatively constant). However, recharge to groundwater does vary seasonally because it is influenced by factors other than

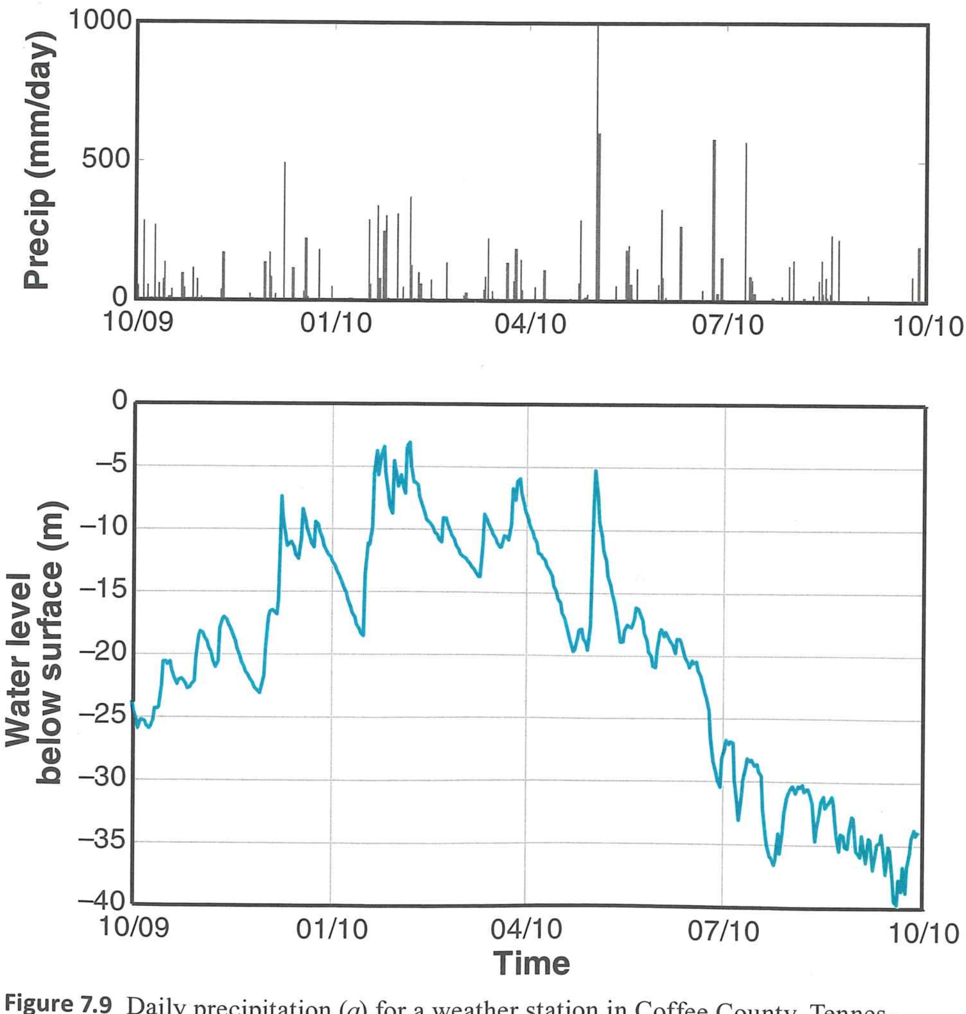

Figure 7.9 Daily precipitation (*a*) for a weather station in Coffee County, Tennessee, USA and water levels (*b*) for water year 2010 in a nearby well in an unconfined aquifer.

the amount of rainfall, the most important of these being evapotranspiration. The variation in the water table in Coffee County can be explained by noting that evapotranspiration is at a peak in the summer months and greatly reduced in the winter months. Consequently, because the ground remains unfrozen during the winter in southern Tennessee, more water is available for recharge in the winter and groundwater levels rise. The converse is true in the summer—little recharge occurs and levels decline as the groundwater discharges to provide the baseflow in streams.

In areas where the water table is close to the ground surface, groundwater levels are influenced directly by the transpirational demands of plants. During the day when transpiration is high, water movement is *upward* from the water table and the level declines. At night, transpiration is reduced, groundwater flows laterally from locations upslope that are relatively unaffected by direct transpiration effects, and the water table recovers (Figure 7.10).

Recharge to aquifers does not occur solely from direct infiltration of rainfall. For example, water can seep from surface-water bodies, such as rivers, ponds and lakes, into the ground. Artificial recharge (recharge induced by activities of people as opposed to that which occurs naturally) can be implemented by introducing water into wells (recharge

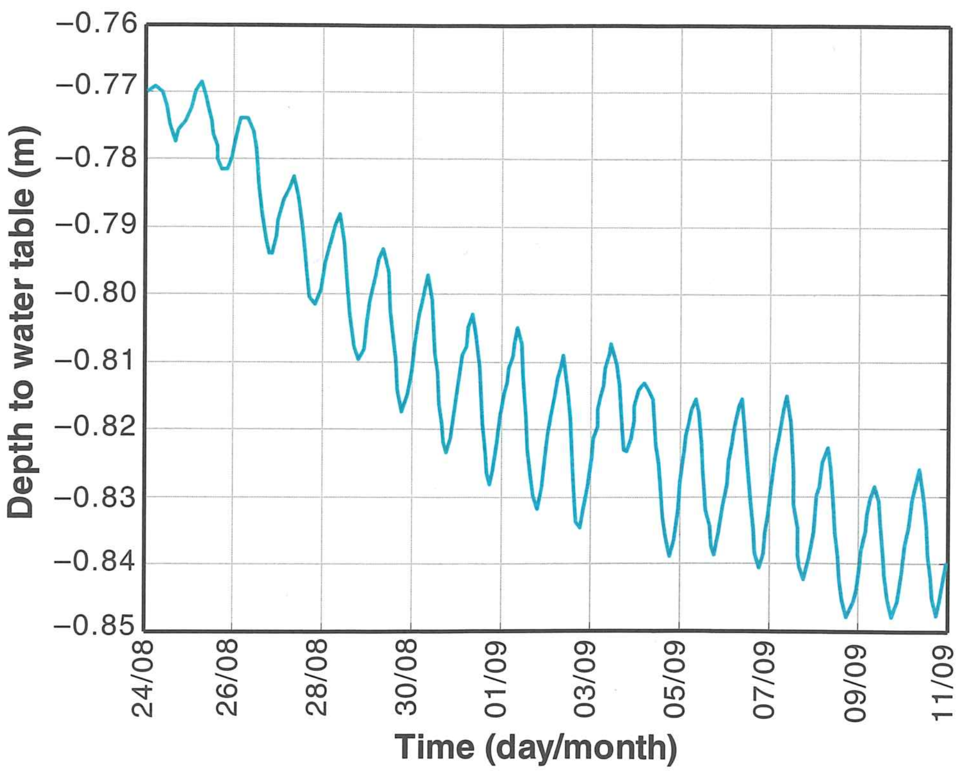

Figure 7.10 Water levels for a late summer period in 2010 in a shallow well adjacent to White Clay Creek at the Stroud Water Center near Avondale, Pennsylvania, USA.

wells) or by routing water into infiltration basins in permeable material (the idea mentioned in regard to Orange County in the introduction).

Unconfined aquifers store water; a change in the elevation of the water table indicates a change in the amount of water stored in the aquifer. Consider a portion of an unconfined aquifer of (in plan view) area 1 m^2 (Figure 7.11). A 1-m drop in the water table would result from the removal of a certain volume, V, of water. The volume of water removed depends on the porosity of the aquifer and how much water would remain behind after gravity has drained the upper cubic meter of aquifer material. (The water that remains is held by forces that will be described in the next chapter.) Hydrologists refer to this volume, V, per square meter of aquifer per meter drop in the groundwater table as the **specific yield** (S_y) of an unconfined aquifer. Because it is the volume of water produced per unit aquifer area per unit decline in hydraulic head, it is dimensionless. The specific yield is characteristic of a given aquifer and allows determinations to be made of the change in the volume of water stored over time. Values of specific yield typically are less than the porosity, and most range from 0.01 to 0.30. Not all of the water in an aquifer volume will drain under the influence of gravity. Some will be retained by forces causing adhesion of water to particle surfaces. This is why the specific yield is less than the porosity.

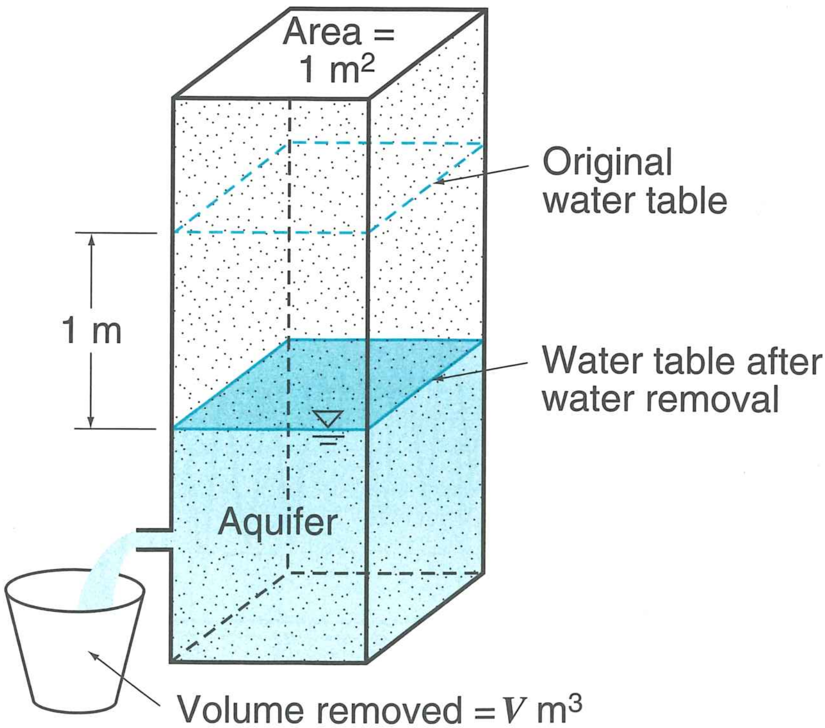

Figure 7.11 The specific yield of an unconfined aquifer. For a 1-m decline in the water table, the volume of water produced per unit aquifer area is the specific yield, S_y.

The specific yield is a hydrological parameter that determines the response of the water table to changes in inputs and outputs. In the case of an increase in evapotranspiration (Figures 7.9 and 7.10), the change in water-table level may be fairly uniform over a given area, although variations will occur due to the lateral movement of groundwater and spatial variations in evaporation rate and vegetation. Pumping a well has a different effect. Pumping produces a decrease in hydraulic head at a point, which increases the hydraulic gradient toward the well. The change in water level in the pumping well, or in observation wells nearby, is referred to as a **drawdown**. The amount of this drawdown will decrease as one moves away from the pumping well, and the pattern that is produced is called a **cone of depression** because of its characteristic shape (see Figure 7.14). The shape and extent of the cone of depression within an unconfined aquifer depend on the pumping rate, the transmissivity and specific yield of the aquifer, and time. From the definition of the specific yield, however, we know that the *volume* of the cone of depression is equal to the volume of water removed (by pumping) divided by the specific yield:

$$V_{cone} = \frac{V_{pumped}}{S_y}. \tag{7.6}$$

7.4.2 Confined aquifers

In an unconfined aquifer, the response to pumping a well is a change in the water table. Water is drained out of the aquifer as the water table declines, and the specific yield provides a measure of the volume of water released from storage. However, the upper boundary of a *confined* aquifer, the overlying aquitard, does not move substantially in response to withdrawing water from a well within the aquifer. Instead, a cone of depression is created *within the potentiometric surface.* The aquifer material is not being drained, and the aquifer remains saturated. However, water is withdrawn from storage within the aquifer (otherwise, the well would become a dry hole almost immediately). Analogous to the specific yield of an unconfined aquifer, we can define the **storativity** (S) of a confined aquifer as the volume of water produced per unit aquifer area per unit decline in the potentiometric surface (Figure 7.12). The question becomes, how is water removed from a confined aquifer without de-watering (draining) the aquifer material?

Consider the forces acting on a horizontal plane within a confined aquifer (Figure 7.13). A downward force due to the weight of the overlying material exists, which when divided by the area of the plane (A) is called the **total stress** (σ_T). Because the plane is not undergoing an acceleration, Newton's second law tells us that this downward force must be balanced by an equivalent opposing force or forces. In other words, the weight of the overlying material must be supported or held up by something. In this case the opposing forces per unit area (stresses) are the pressure (p) of the water and the upward stress exerted by the aquifer solids, called the **effective stress** (σ_e):

$$\sigma_T = p + \sigma_e. \tag{7.7}$$

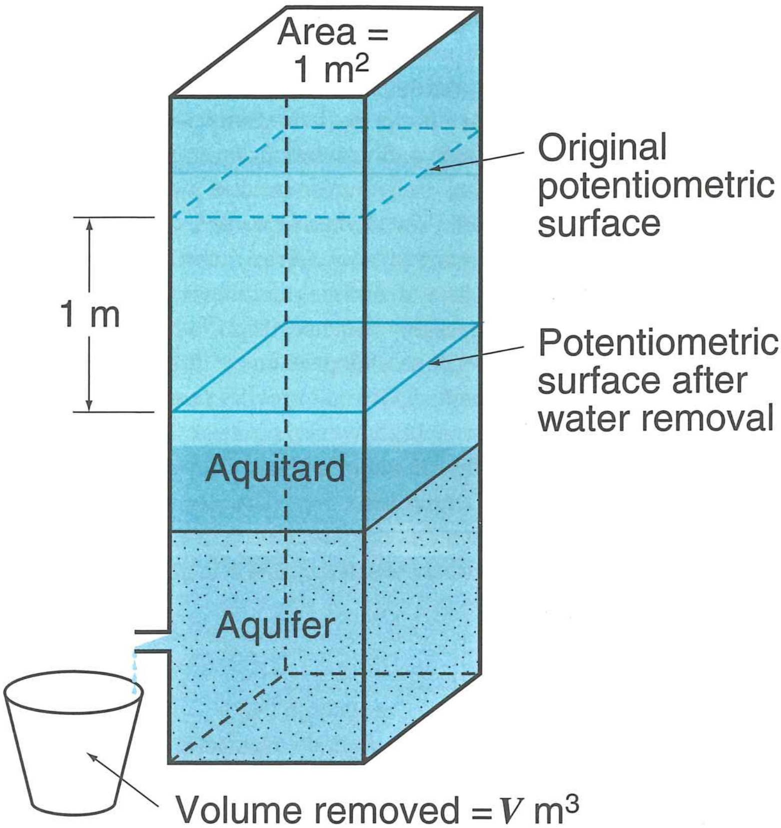

Figure 7.12 The storativity of a confined aquifer. For a 1-m decline in the potentiometric surface, the volume of water produced per unit aquifer area is the storativity, S. The aquifer material is not drained and remains saturated.

This idea may be stated in another way: the weight of the overlying material on a horizontal plane is supported in part by the fluid and in part by the solid. The total stress won't vary over time:

$$d\sigma_T = dp + d\sigma_e = 0.$$

Therefore, any change in pressure must be offset by a change in effective stress:

$$dp = -d\sigma_e. \tag{7.8}$$

If water is being withdrawn from a well within a confined aquifer, the hydraulic head, and, hence, the fluid pressure, is being reduced at that point (dp is negative). As a result, the fluid will expand slightly. This is because water is (slightly) compressible. This is one mechanism by which water is released from storage in a confined aquifer. The additional *volume* of water produced may flow to the well and be withdrawn by the pump.

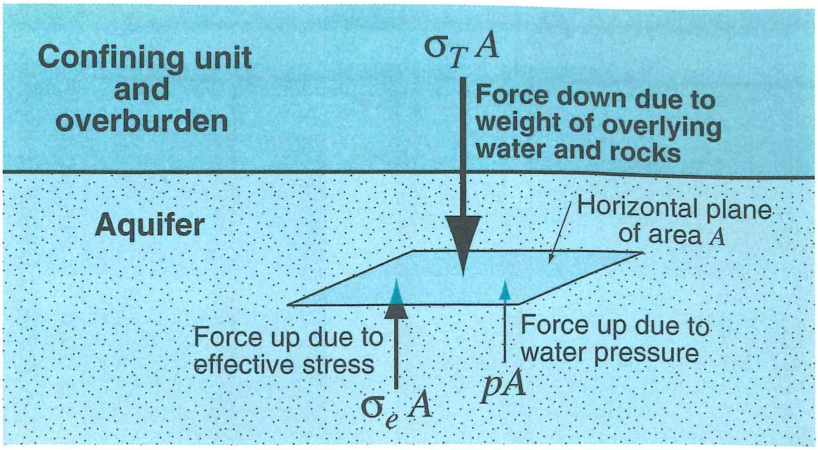

Figure 7.13 Force due to water pressure and effective stress balance the downward force due to the weight of the overburden.

Equation 7.8 indicates that the decrease in fluid pressure must be accompanied by an *increase* in the effective stress (positive $d\sigma_e$). A part of the weight of the overlying material is being transferred from the fluid to the solid. This results in the compression of the aquifer material, just as decreasing the fluid pressure resulted in expansion of the fluid. Compressing the aquifer material is similar to squeezing a sponge, and produces water that may be pumped from the well. This is the second mechanism by which water is removed from storage in a confined aquifer.

Water and most aquifer materials are not very compressible. As a result, storativity values tend to be lower than values of specific yield. Storativity generally ranges from 0.005 to 0.00005, as compared with 0.01 to 0.30 for specific yield. In this discussion of storage in confined aquifers, we have assumed that the aquifer is perfectly confined, or that all the flow occurs within the aquifer and none within the confining layers. If an aquifer is only semi-confined, with some water coming from surrounding aquitards, then it is referred to as a **leaky aquifer**. Because of this additional source of water, the storativity will often be higher in leaky aquifers.

The extent of the cone of depression depends, among other things, on the value of the storage parameter (either specific yield or storativity). Consider two aquifers, one unconfined and one confined, that are being pumped at the same rate. Because the specific yield of the unconfined aquifer will be greater than the storativity of the confined aquifer, the drawdown in the unconfined aquifer will be less than that in the confined aquifer (Figure 7.14). That is, the volume of the cone of depression in a confined aquifer is equal to the volume of water pumped divided by the storativity [change S_y to S in Equation 7.6], a very small number. Thus removal of water from confined aquifers produces substantial drawdown of the potentiometric surface. In confined aquifers that are heavily pumped over long time periods, the cone of depression can be quite extensive, approaching tens of kilometers in lateral extent, with tens of meters of drawdown in the vicinity

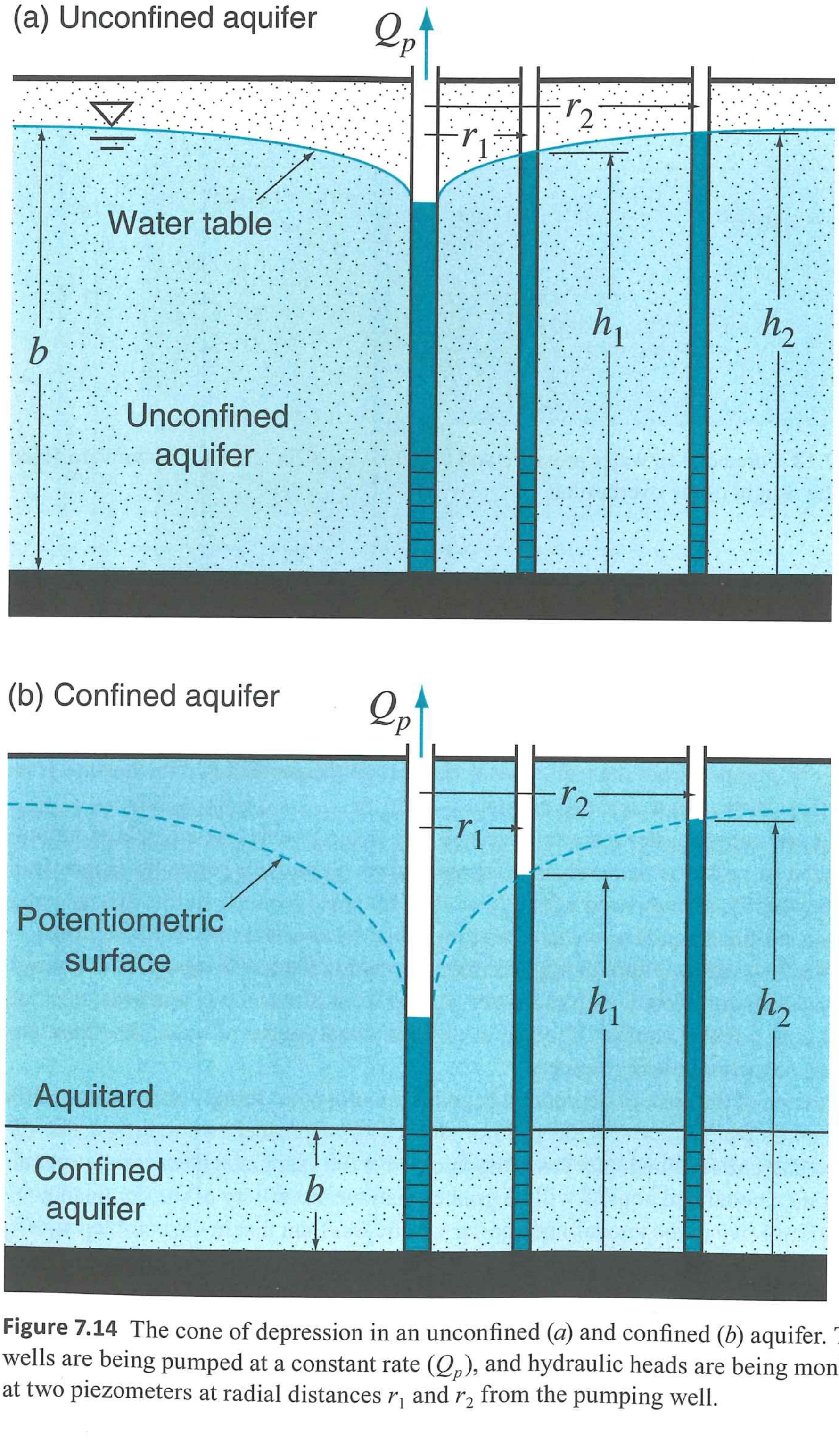

Figure 7.14 The cone of depression in an unconfined (*a*) and confined (*b*) aquifer. The wells are being pumped at a constant rate (Q_p), and hydraulic heads are being monitored at two piezometers at radial distances r_1 and r_2 from the pumping well.

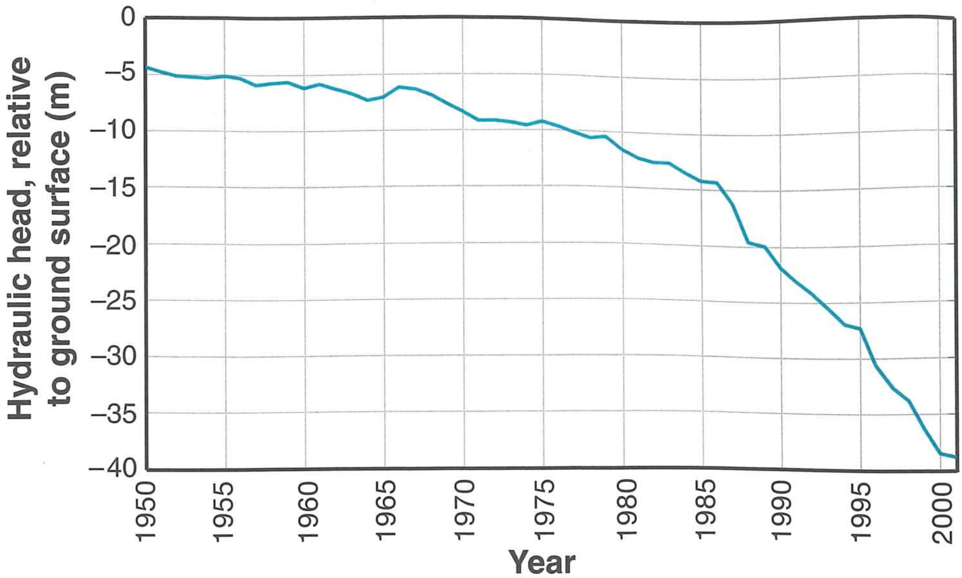

Figure 7.15 Water levels in a well in a confined aquifer in Calvert County, Maryland, USA. Data from DePaul et al. (2008).

of the well (Figure 7.15) and a cone of depression that extends a long distance from the major pumping center (Figure 7.16).

Other things can affect the balance between water pressure and effective stress in aquifers and so affect the water levels in wells. Atmospheric pressure varies as low-pressure and high-pressure systems move across the landscape. An increase in atmospheric pressure creates an additional force on the top of a confined aquifer that is distributed between water pressure and effective stress. That is, only a portion of the force from the increased atmospheric pressure will be transmitted to the water in the aquifer. The increased atmospheric pressure also acts on the water standing in the well. In the well, the full increase in the atmospheric pressure is transmitted to the water. Thus, for a confined aquifer, an increase in atmospheric pressure causes a decrease in the water level in a well and, conversely, a decrease in atmospheric pressure results in an increase in water level. For an unconfined aquifer, changes in atmospheric pressure are transmitted equally to the water table and the water in a well, and so do not produce changes in water levels.

7.5 Well Tests to Estimate Aquifer Properties

The rate at which water levels change in wells depends in part on the aquifer properties. For example, if a quantity of recharge water is suddenly added to an unconfined aquifer, the water levels will decline as the water drains out to streams. The rate of decline will depend on the hydraulic conductivity and the storativity of the aquifer. Thus, if the water table or potentiometric surface in an aquifer is purposefully disturbed, by removing or adding water for example, measurements of the time variation of water levels in wells