Groundwater Hydrology

September 2001

This booklet is part of a series of educational brochures and slide sets that focuses on various aspects of water source protection. The series has been prepared jointly by the University of California Agricultural Extension Service and the California Department of Health Services.

For further information about this and other documents in the series, contact the project team leader (see below) or visit the following website:

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Life depends on water. Our entire living world—plants, animals, and humans—is unthinkable without abundant water. Human cultures and societies have rallied around water resources for tens of thousands of years—for drinking water, for food production, for transportation, and for recreation—as well as for inspiration.

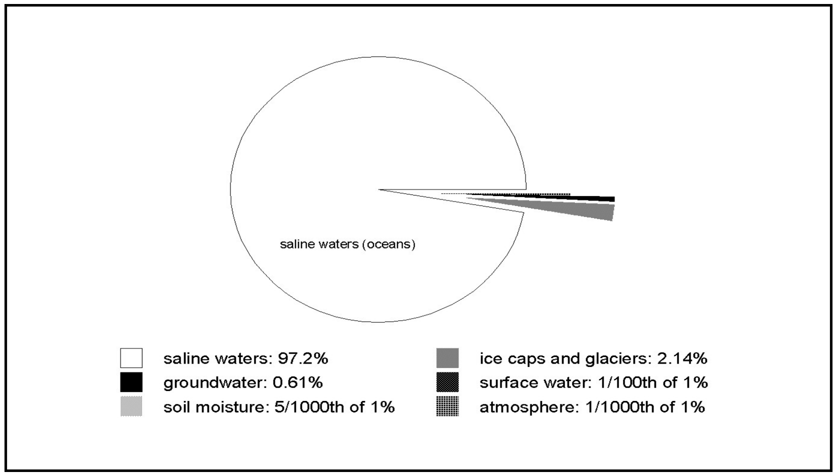

Figure 1. Groundwater comprises 0.6% of Earth's fresh water.

Ninety-seven percent of Earth's water exists in the oceans; only 3% occurs on land. Hydrologists estimate that most of the latter (about 2% of Earth's total) is bound up in polar ice-caps and alpine glaciers. Thus, less than 1% of Earth's water is available to humans. These accessible and usable waters occur as lakes, rivers, and other surface water bodies, and as *groundwater*—the water that lies beneath our feet within 13,000 ft (4000 m) of Earth's surface. Actually, water in lakes and rivers contributes only 0.009% of earth's water supply, soil moisture another 0.005%, and moisture in the atmosphere a miniscule 0.001%. Groundwater, by contrast, contributes 0.61% of Earth's total (Figure 1); it is by far the largest "reservoir" of available fresh water.

Worldwide, more than a third of all water used by humans comes from groundwater. In rural areas the percentage is even higher; according to a recent United Nations report, more than half of all drinking water world-wide is supplied from groundwater. In California, the large metropolitan areas in Southern California and the San Francisco Bay area rely primarily on surface water for their drinking water supplies. The remainder of California relies largely on groundwater. Of California's 8,700 public water-supply systems, all but 900 rely on groundwater, drawing from 15,000 individual wells or well fields. In addition, thousands of privately-owned domestic wells exist within the state.

Water is in constant flux and follows a complex hydrologic cycle (Figure 2). Evapotranspiration from land and evaporation from the ocean contribute to atmospheric moisture, which returns to the land in the form of rain and snow (precipitation). Snow accumulates in a winter snow pack, in glaciers, and on arctic ice-caps. Rainfall creates infiltration into the soil,

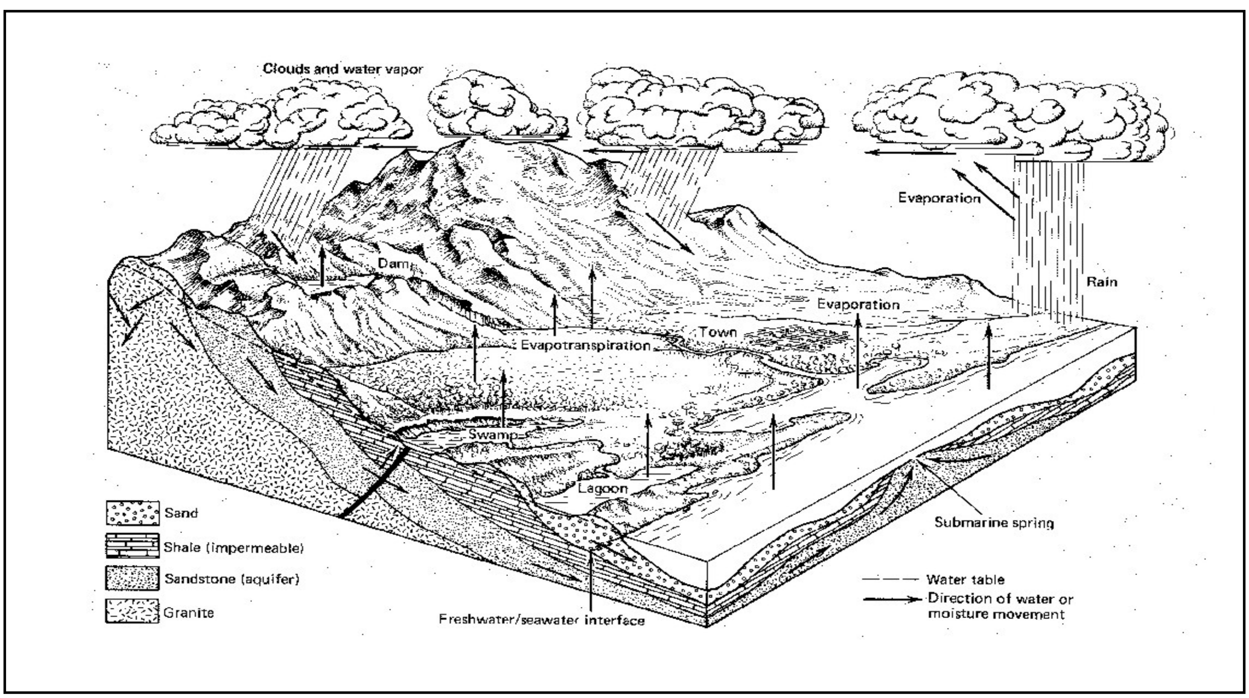

Figure 2. Earth's hydrologic cycle includes evaporation, precipitation, infiltration, and runoff.

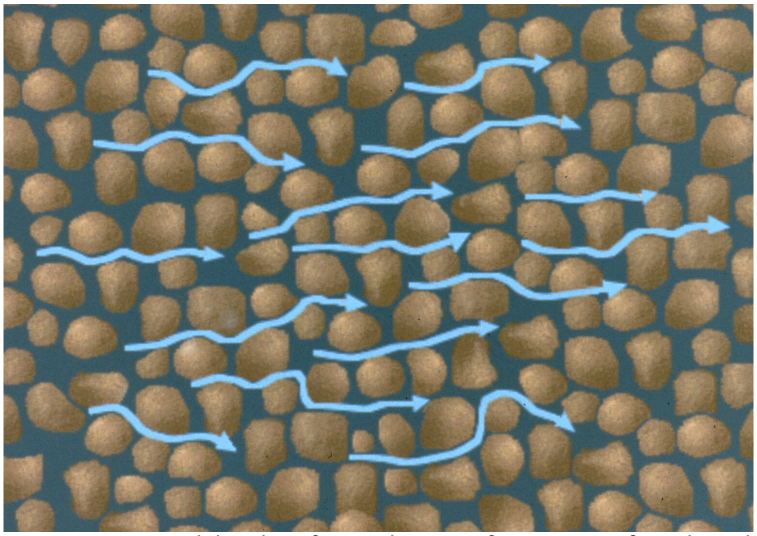

Figure 3. Unconsolidated surface sediments often consist of sands and gravels.

as well as runoff into creeks, rivers, and lakes that drain back into the ocean. Soil moisture becomes available to plants that will transpire water. Excess infiltration water percolates to groundwater, which accumulates in underground basins and eventually flows back into rivers, lakes, man-made wells, or directly into the ocean. The elements of the hydrologic cycle are all interconnected. The time a water molecule spends in one part of a cycle or another may range from seconds (in a quickly disappearing water puddle) to millions of years (in deep oceans, polar ice-caps, or deep groundwater).

What Is Groundwater?

Groundwater in California is an important and significant part of the hydrologic cycle. Yet, it remains a mystery to many people. Because we cannot see it directly, groundwater defies our experience of the land surface as a solid, rigid boundary marking the top of the earth. Many people find it hard to imagine that water can move underground at rates that allow California's largest springs to discharge almost 1 million gallons per minute (as at Fall River, in Shasta County). Likewise, it's hard for many of us to understand how California farmers and municipal utility companies manage to extract from 500 to 2,000 gallons per minute out of a pipe in the ground that is merely 1 foot in diameter. More often than not, faced with these realities, people envision that groundwater exists somehow in a mysterious, hidden system of underground rivers, reservoirs, and water "veins." Such concepts are far from the physical reality, though these terms are sometimes useful when speaking metaphorically about one purpose of groundwater or another.

In the vast majority of California's numerous valleys and intermontane basins—the areas where groundwater exists in greatest quantity—groundwater fills small, often microscopic pores between the grains of sediment (Figure 3). In the rocks that make up the hills and mountains of California's uplands, groundwater is also quite common, even plentiful. In such locales, groundwater occurs in tiny fractures and fissures of rocks hidden below the loose, or *unconsolidated*, surface sediments.

To the naked eye, soils and rocks may appear next to impenetrable. But most such materials are, in fact, a microscopic universe full of empty spaces that can be occupied by water. When such water completely fills the void space of sediment pores or rock fractures, the pores or fractures are said to be saturated. In all but truly swampy areas, the ground just beneath the land surface is aerated: water occupies only part of the pore space. The water partially filling the voids is referred to as soil moisture, or vadose zone water (Figure 4). The boundary between groundwater and vadose zone water is referred to as the water table. The zone above the water table is referred to as the "unsaturated zone" or "vadose zone."

The term "soil" describes only the first three to six feet below the land surface that is exposed to weathering,

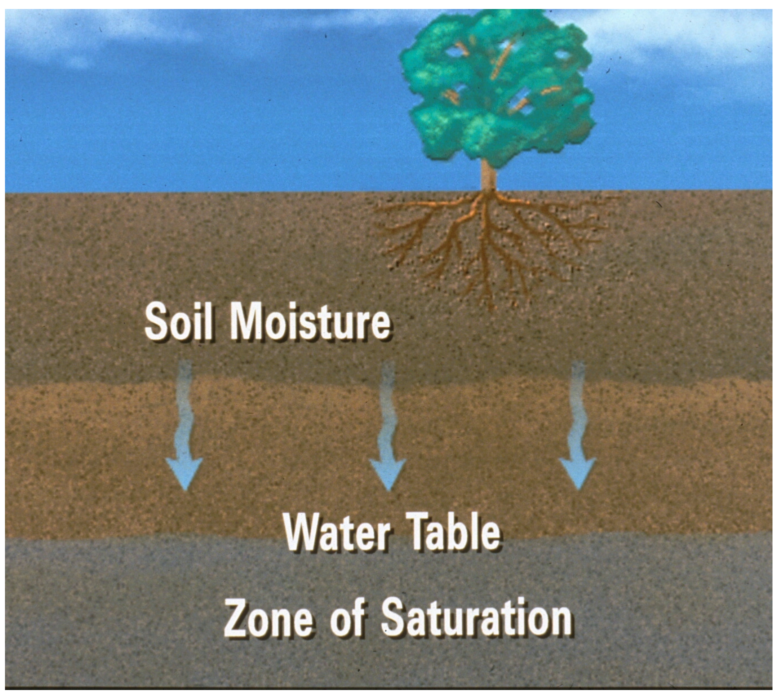

Figure 4. Soil moisure and the vadose zone.

➤ medium gravel:	0.5	inches
➤ coarse sand:	0.05	inches
➤ very fine sand:	0.005	inches
➤ silt:	0.001	inches
➤ clay:	0.0001	inches

Figure 5. Definitions of various particle sizes.

plant root growth, etc. In cold to moderate, humid climates, the vadose zone is commonly identical to the soil layer—that is, the water table is less than 6 feet from the land surface. In California's predominantly Mediterranean and semi-arid climates, it is not uncommon to have a vadose zone that is several tens of feet thick. In some areas of southern California this zone may even be several hundred feet thick. While the vadose zone is not a resource from which we can obtain water, it is an important storage area, pathway, and potential barrier for pathogens, nutrients, or contaminants traveling within the water.

Geology and Groundwater

To understand the occurrence of groundwater at a given site, one must understand the local geology. Geologists usually investigate local geology by studying the *type* and *age* of the various sediment and rock layers. That information also provides important clues about the characteristics of the groundwater system in an area.

Sediment and Rock Types

Most of California's groundwater resources are found in the *unconsolidated sediments* (in other words, loose material) that cover the floors of valleys and basins. Unconsolidated sediments consist of gravel, sand, silt, or clay, or a mixture thereof. The precise definitions for these particle size classes are given in Figure 5. A geologic unit of this type is shown in Figure 6.

Unconsolidated sediments are deposited by river floods (resulting in *alluvial* and *fluvial* deposits), in lakes (resulting in *lacustrine* deposits), in shallow oceans (*marine* deposits), by glaciers (*till* and *outwash* deposits—rare in California), or by wind (*eolian* deposits, such as sand dunes or *loess*). The often colorful particles and grains of which sediments are composed are themselves remnants of the rocks from which the sediment originated. "Well-sorted" sediments consist of grains that all have similar size (a *uniform grain size distribution*), while "poorly-sorted" sediments consist of grains with a wide range of sizes. Water moves most readily through well-sorted sediments that have relatively large pores (sand, gravel). In poorly-sorted sediments, the space between larger grains or rock pieces (gravel or coarse sand) is filled with finer material (medium or fine sand, silt). The space between the finer material is filled, in turn, with even finer material (silt, clay). That leaves only the tiniest pore spaces for water to move through.

Alluvial sediments deposited by the river floods that occasionally emanate from California's uplands provide by far the largest share of California's groundwater resources. Lakebed (lacustrine) and marine clay deposits, on the other hand, yield little or no groundwater, due to their extremely fine grain size and low permeability.

Rocks—that is, hard material—are grouped by their origin. Geologists distinguish between *lithified sedimentary rocks*, *chemical sedimentary rock*s, *igneous rocks*, and *metamorphic rocks.*

Lithified sedimentary rocks are sediments that have hardened through chemical alteration and the pressure of thousands of feet of overburden (overlying sediments). Under such pressure (accompanied by higher temperatures), sand deposits become *sandstone*, silt deposits become *siltstone*, clay beds become *shale*, and poorly sorted alluvial or fluvial sediments containing

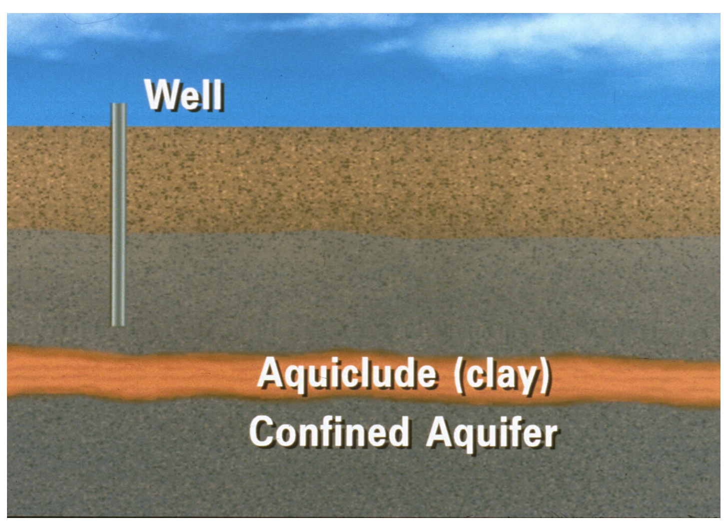

Figure 6. Geologic unit consisting of unconsolidated sediments—gravel, sand, silt, and clay.

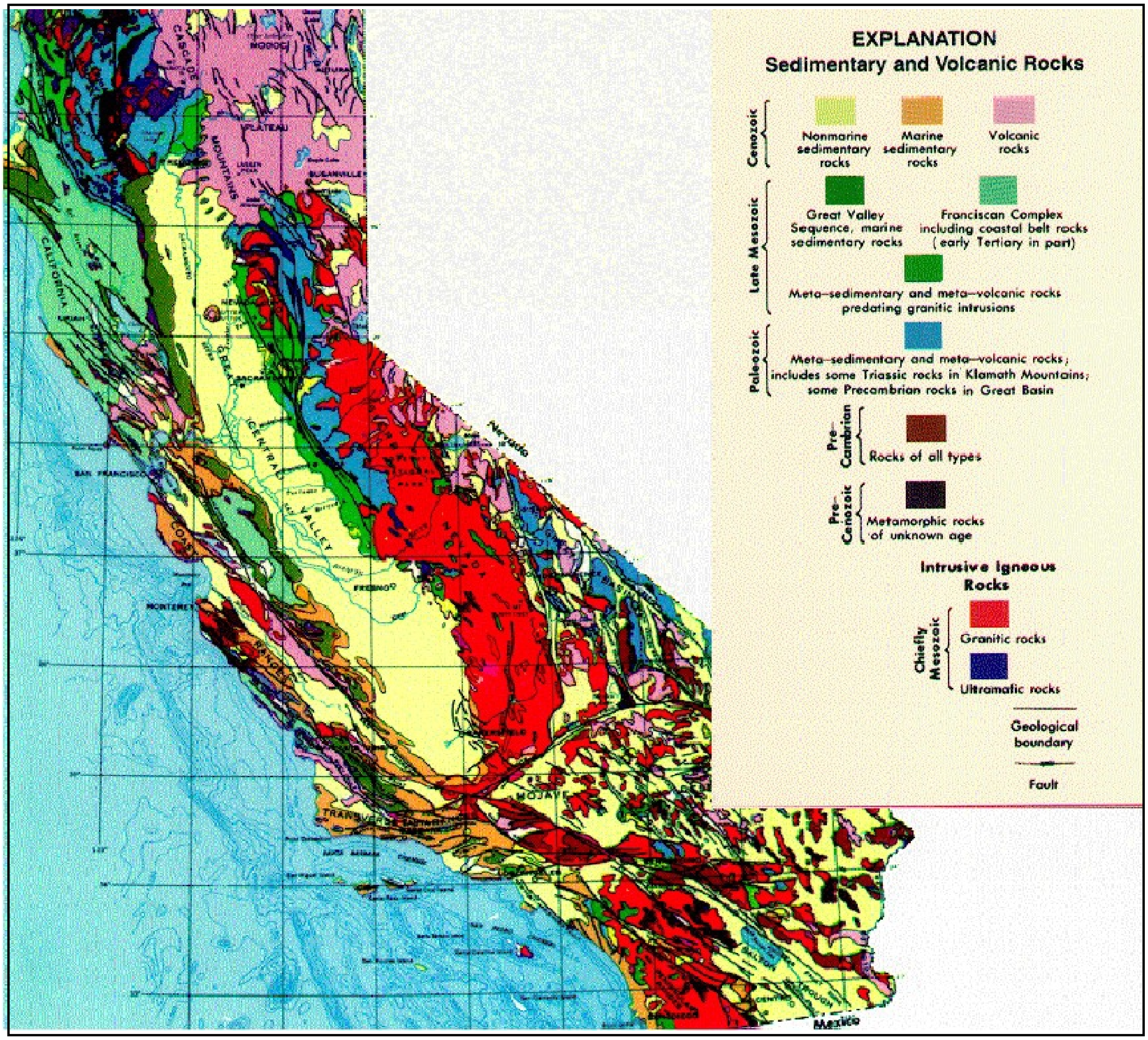

Figure 7. Chart showing ages of rocks in California (from California Division of Mines and Geology).

everything from gravels to clays become *conglomerate*.

Chemical sedimentary rocks are those deposited by chemical precipitation of carbonate and salts (*limestone*, *dolomite*, *gypsum*), typically in shallow ocean basins.

Igneous rocks are formed when magma rises from Earth's interior to near the surface of the crust. If the magma rises without penetrating the land surface and remains, it cools slowly and forms the rock known as *granite*. If magma erupts from volcanoes or other earth-surface fissures, it cools very quickly, forming *basalt*. In addition, ashes and other debris blasted out of other *pyroclastic deposits*.

Extremely high pressure from overburden or tectonic forces, combined with high temperature, can further alter rock formations buried deep below the surface. Sedimentary and igneous rocks then become metamorphic rocks. Sandstone alters to *quartzite*, shale alters to *slate*, limestone to *marble*, and granite changes to *gneiss*. Over the course of millions of years, subsequent geologic processes may bring these deep rocks back to the surface.

Groundwater may occur in any of these rocks. It moves through fractures and fissures created by weathering of rock, by the tectonic movement of rock masses, or—in the case of the chemical sedimentary rocks—by dissolution of rock minerals. Fractures in igneous and metamorphic rocks may yield sufficient water for domestic purposes. In limestone, the dissolution of calcium carbonate sometimes creates large caves, which may provide the framework for real underground rivers. In basalts, large shrinkage cracks, lava tubes, gas vesicles, and fissures often create highly permeable pore space. In northeastern California, large areas underlain by basaltic rock are so porous that no surface water drainage network exists (for example, the areas around

Mt. Lassen, Mt. Shasta, and Medicine Volcano). Groundwater may also saturate the actual rock particles. Sandstone, for example, can yield significant amounts of groundwater stored in its rock matrix.

Geologic Times

Besides classifying geologic formations by rock type and grain size, geologists classify geologic formations by age. Earth was formed approximately 4,000 million years ago. Little geologic evidence exists at the earth's surface to describe the geologic processes that occurred during the first 1500 million years of Earth's life. Most rocks found at or near the ground surface (in other words, where we can see them and inspect them) are younger than about 570 million years old. That marks the beginning of the Paleozoic Era—the period during which advanced life forms began to proliferate on Earth. Few places in California can be found with rocks of Paleozoic age, let alone older rocks (Figure 7). Located at a major boundary between two large earth-surface crustal plates, California is simply too active geologically for such old rocks to survive.

The dominant rocks of California all formed during the late part of the Mesozoic Era (240 to 65 million years ago) and during the relatively recent Cenozoic Era (65 ma to present). The latter is divided by geologists into two periods: the Tertiary (65 ma – 3 ma) and the most recent geologic time period, the Quaternary (3ma – present). The unconsolidated sediments that make up California's largest groundwater basins were all deposited in recent geologic time (late Tertiary and Quaternary).

Aquifers

A geologic formation from which significant amounts of groundwater can be pumped for domestic, municipal, or agricultural uses is known as an *aquifer*. The term is relative: it means that a geologic unit yields water relative to surrounding materials, but does not indicate that a specific amount of groundwater can be pumped. A small intermontane-valley aquifer yields significantly more water than its surrounding hard rock (bedrock) formations. Yet, wells within the intermontane valley may yield water at much lower rates than similar wells installed in a large alluvial river basin such as California's Central Valley.

Aquifers sometimes are vertically separated by geologic formations that permit little or no water to flow. The formation that acts as water barrier is called *aquitard* if it is much less permeable than a nearby aquifer but still permits flow (e.g., sandy clay). If the water barrier is almost impermeable (e.g., clay) and forms a more or less formidable flow barrier between multiple levels of aquifers, it is known as an *aquiclude*.

Aquifers can be of two major types: *unconfined* or *confined*. In an unconfined aquifer, there is no overlying aquitard or aquiclude. Where multiple levels of aquifers exist, the uppermost aquifer is typically unconfined. Vertical recharge by infiltration of rainwater or irrigation water downward to the unconfined aquifer is therefore not restricted (Figure 8). The water table at the top of the unconfined aquifer can migrate freely up and down depending on how much water is stored in the aquifer. The water level in a borehole drilled into an unconfined aquifer will be the same as the water level in the aquifer (if we ignore the effects of the capillary fringe).

A confined aquifer, on the other hand, is "sandwiched" between an aquiclude above and an aquiclude or aquitard (e.g., bedrock) beneath (see Figure 9). As a result of "backpressure", water in the confined aquifer is pressurized. Due to the pressure, the water level in a borehole drilled into a confined aquifer will rise significantly above the top of the aquifer. An *artesian well* occurs where the pressure is so large that the water level in a well drilled into the confined aquifer rises above the land surface—in other words, the well flows freely (if opened) without pumping. Note that a confined aquifer does not have a water table—it is completely filled with groundwater. The water levels in wells drilled into a confined aquifer correspond instead to the *potentiometric surface* of the aquifer, also known as *pressure head or confined head*. If the pressure head falls below the top of the aquifer, the aquifer is no longer confined; it becomes, by definition, an unconfined aquifer.

An aquifer confined by an aquitard rather than an aquiclude is referred to as a *leaky aquifer*, or a *semi-confined aquifer*. In alluvial aquifers, the aquitard rarely is a contiguous layer of low-permeability clay, loamy clay, or sandy clay. Rather, it can be thought of as a

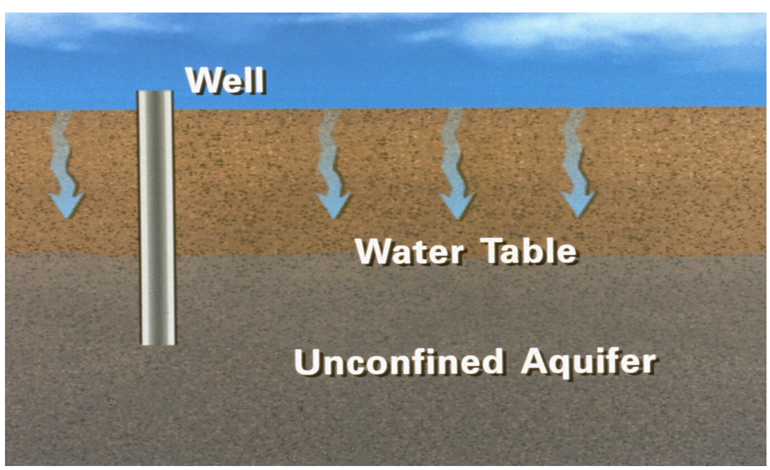

Figure 8. Unconfined aquifer.

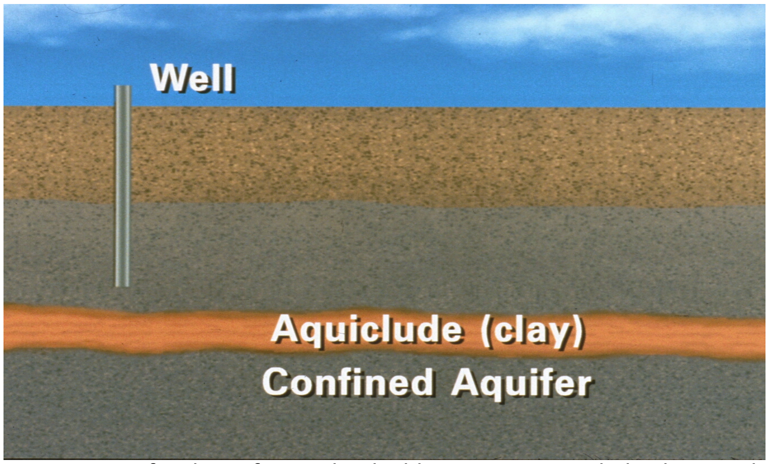

Figure 9. Confined aquifer sandwiched between an aquiclude above and an aquiclude or aquitard such as bedrock below.

local accumulation of multiple, horizontally discontinuous smaller clay "lenses" and other clay-rich or otherwise impermeable (or low permeability) layers. Though such lenses and layers are not contiguous, the overall effect on the regional aquifer below is identical to that of a solid, continuous aquitard.

Similarly, the alluvial aquifers that are so common in California typically are not contiguous, homogeneous layers of sand, clayey sand, and gravel, but, rather, a heterogeneous "sandwich package" with a significant amount of higher permeability material. Hence, the distinction between an aquifer and a nearby aquitard is sometimes based not on well-defined geologic boundaries, but on changes in the vertical frequency of the occurrence of (or lack of) less-permeable, horizontally-discontinuous layers of clay or other sediments that have a high percentage of fines (clay, silt).

Sometimes water collects above an impermeable layer or low-permeability layer within the unsaturated (aerated) zone, forming a "perched" water table. By definition, a perched water table is a saturated groundwater zone separated from the aquifer below by a zone that is unsaturated or aerated (Figure 10). This should not be confused with an unconfined shallow aquifer that is separated from a deeper confined aquifer by thick, saturated layers of clay.

Springs form where a water table intersects with the land surface. This may occur in a depression of the land surface, particularly on hillsides. It can also occur where two geologic (rock) formations outcrop at the land surface, if the lower one is less permeable than the upper one (Figure 11). A spring also may form when back pressure forces water to the surface through a sinkhole, fracture, joint, or fault zone that acts as a conduit for water movement. Depending on the geologic history and on the surrounding geologic material, fault and fracture zones may sometimes represent a barrier to flow rather than conduit for flow, in which case water may be forced to the land surface along the top of the fault or fracture zone.

Understanding Porosity, Yield & Storage

As discussed in the last section, pore space can vary tremendously depending on the rock or sediments that make up the subsurface. The size, shape, regularity, and continuity of the pore space will determine how much water can be extracted or stored in the subsurface, how fast water can move through the pore space, and how contaminants are distributed within the subsurface.

The *porosity* of a sediment or rock formation is defined as the fraction of the material's volume that is not occupied by solids:

$$\text{porosity [in \%]} = (\text{volume of pores/total volume}) \times 100$$

$$= (1 - \text{volume of solids/total volume}) \times 100$$

Groundwater fills the entire pore space, but not all of that groundwater will be available for pumping. Some pores may be entirely isolated from the rest of the pores (Figure 12). Other pores may be so fine that water molecules are held tightly to the soil, particle, or mineral surfaces by adhesion (magnet-like forces on the surface of water molecules that attract them to the surfaces of mineral grains, especially clay). Adhesion immobilizes water molecules. Water adhesion occurs in the matrix or in fine fissures of "wet" consolidated, metamorphic, or igneous rocks and in clay sediments.

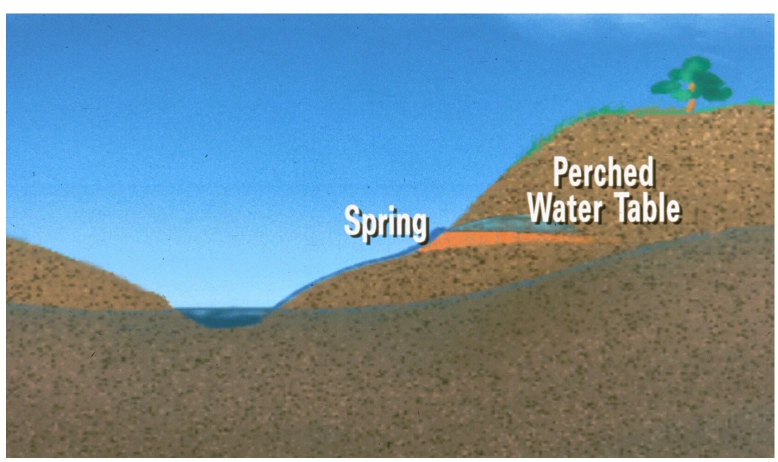

Figure 10. Perched water table.

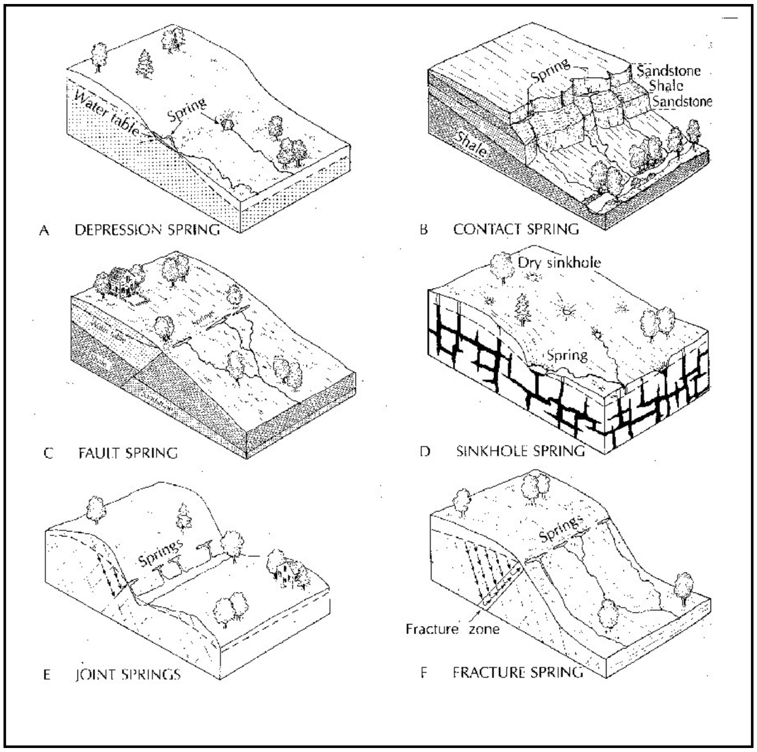

Figure 11. Types of springs.

Even if pores are not fine enough for adhesion, capillary forces can hold water back against gravity drainage. (That's why moisture stays in a flower pot and doesn't immediately drain into the saucer.)

The total porosity (total pore space) of a rock or sediment formation is therefore divided into *effective porosity*, consisting of the fraction of pore space that is interconnected, and *isolated pore space*. Effective porosity is further subdivided into *specific retention* and *specific yield*. Specific retention is that volume fraction of water that is held back by adhesion and capillary forces when an aquifer is drained. Specific yield is the percentage of water that is actually available for pumping when sediments or rocks are drained out due to the lowering of the water table near a well.

Different sediments and rocks not only have different effective porosities, but also different specific yields (Figure 13). Peat has by far the highest effective porosity: 90% of its total volume consists of pores, half of which is drainable (specific yield: ~45%). Clay, silt and fine sand also have relatively high effective porosities. But their specific yields vary widely: The specific yield of clay is only 2% or less, while that of fine sand is over 20%. The highest specific yield is obtained in coarse sands and gravels (20% - 30%), where little water is held back by retention (large pores have only small capillary forces). This makes sand and gravel aquifers good storage reservoirs for groundwater. Most production aquifers in California contain a mixture of sands, gravels, and finer sediments. Typical specific yields in these aquifers range from under 10% to over 20%.

In an unconfined aquifer (sometimes also called a *water table aquifer*), the water pumped from a well is obtained by lowering the water table and draining the volume of rock or sediment just above the water table. Pumping water from the well creates a lower-than-normal water level surface in the vicinity of the well; this is known as the *cone of depression* (Figure 14). The volume of water pumped from the well is equal to the specific yield of the sediments (or rocks) times the volume of the cone of depression (the total volume of sediments that have been drained). As explained above, most of the water contained in an aquifer becomes part of the the specific yield, when the water level is lowered and sediments (or rock fractures) are drained.

Example 1: To illustrate the meaning of specific yield, let's assume we have an "aquifer box" that measures 1 acre in area by 100 feet deep. Let's further assume that the box is entirely enclosed except at the surface. If the specific yield is 15% (0.15), how much water do we have to pump to lower the water table in the entire aquifer box by 1 foot?

The answer is:

Water pumped = Volume to be drained x Specific yield

= (1 acre area x 1 foot drawdown) x 0.15

= 0.15 acre-feet (~ 50,000 gallons)

In a confined aquifer, water and the sediments or rocks containing the water are under pressure and therefore slightly compressed. When pumping from a well in a confined aquifer, the water table inside the well is lowered, resulting in lower water pressure in the deep

Porosity is the percentage of a soil or rock volume that is **NOT** occupied by solids.

- SIZE
  - –Very fine pores: water is held to soil/rock particles by adhesion
  - –fine pores: water is held back against gravity by capillary forces
  - –large pores: water moves only by gravity
- SHAPE AND IRREGULARITY
- DISTRIBUTION
  - –connected pores ⟷ isolated pores
  - –even distribution ⟷ skewed distribution

Figure 12. Definition of porosity.

part of the well (Figure 15). This, in turn, depressurizes the water in the (confined) aquifer around the well. Water that was compressed by the pressure on the aquifer expands and replaces the water pumped in the well. Also, the sediment structure expands elastically due to the depressurization, thus expulsing water. The water available for pumping as a result of lowering the pressure head and thus decompressing water and sediment structure is, of course, only a tiny portion of the total water contained in the confined aquifer. In fact, only about one-millionth of the amount of water contained in a cubic foot of aquifer is released from the aquifer per 1-foot water pressure drop. That tiny amount of water that is packed into the confined aquifer by pressure is called the specific storage (or storativity) of the aquifer. It is measured in cubic feet of water released per cubic foot of aquifer for each foot of pressure drop. It typically ranges from 10^{-6} to 10^{-5} (1/ft).

In a confined aquifer, the pressure drop induced by a pumping well occurs simultaneously over the entire thickness of the aquifer. To compute the amount of water available for pumping from a confined aquifer, hydrogeologists multiply the specific storage coefficient by the thickness of the aquifer. This number is called the storage coefficient of the aquifer.

storage coefficient = specific storage x aquifer thickness

Like the specific yield, storage coefficient usually is reported in percent. Typically, it ranges from 0.00005 (0.005%) to 0.005 (0.5%).

Example 2: Let's assume a confined aquifer is 100 feet thick and that the aquifer material has a specific storage of 0.000005 (in engineering notation: $5\text{x}10^{-6}$) [1/ft]. The storage coefficient of the aquifer, then, is 100 [ft] * 0.000005 [1/ft] = 0.0005 ($5\text{x}10^{-4}$), which is the same as 0.05%. How much water do we have to pump from each acre area of that aquifer to lower the confined pressure head by 1 foot? The answer is obtained from the following equation:

pumpage = aquifer area x pressure drop (drawdown) x storage coefficient

Using the numbers for our sample aquifer, we get:

P = 1 acre x 1 foot x 0.0005

= 0.0005 acre-feet, or 162.5 gallons

This amount, obtained be depressuring the confined aquifer, is much, much less than the amount that would have been available if the aquifer were unconfined and we had drained it by 1 foot (48,500 gallons; see Example 1).

This illustrates how important it is to distinguish between the water level changes associated with an unconfined aquifer (where water level change means actually draining the aquifer) and those associated with a confined aquifer (where a water level change is actually just a pressure change).

We can now understand why the drop in pressure head and the spreading of the cone of depression in the pressure head surface of a confined aquifer occurs much more rapidly than the lowering of the water table in an unconfined aquifer. To sustain a given pumping rate (e.g., 500 gpm), the water pressure in the confined aquifer has to drop much faster and over a much larger area than the water table in the unconfined aquifer, where water is actually draining from pores.

Note that the water level in a confined aquifer well corresponds to the pressure head of the confined

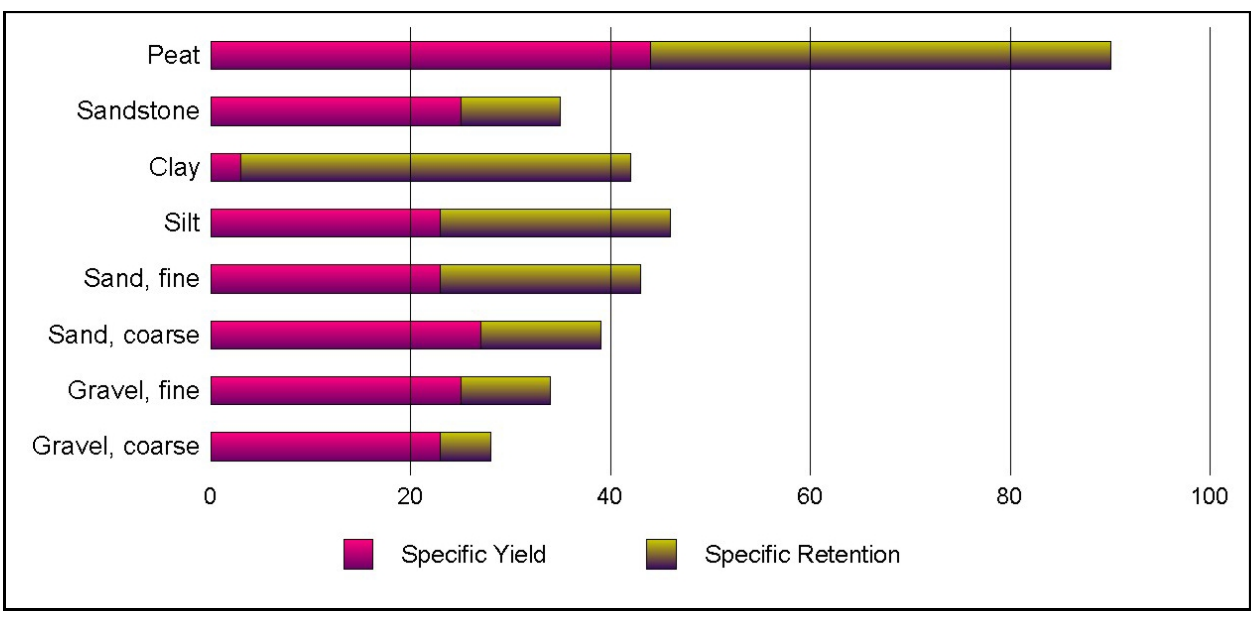

Figure 13. Effective porosity and specific yield for various sediments and rocks.

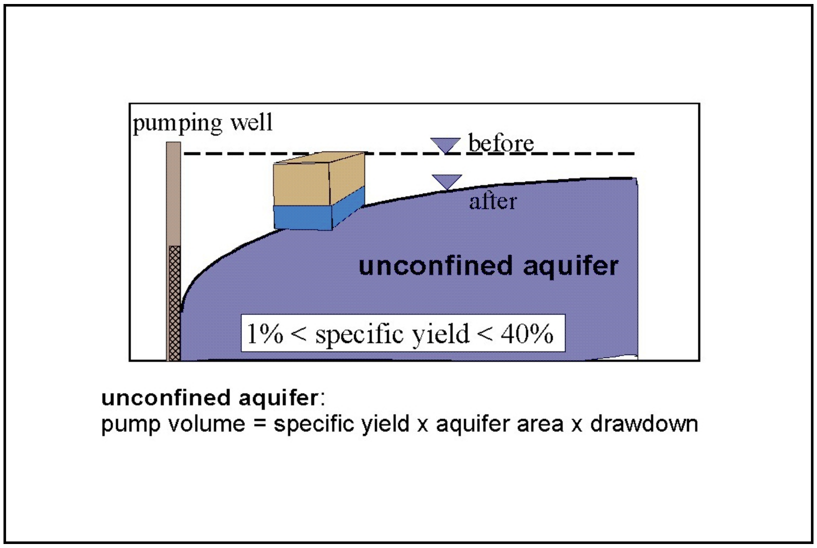

Figure 14. Unconfined aquifer and cone of depression.

aquifer. Therefore, for the same pumping rate, the water level in a well penetrating a confined aquifer will drop much faster and by a much larger amount than the water level in a well penetrating an unconfined aquifer. For the same reason, the water level (pressure head) in a confined aquifer well is much more sensitive to nearby groundwater extraction than a well in an unconfined aquifer.

As a result of continuous pumping, the pressure head in the confined aquifer will drop until the confined aquifer either becomes partially unconfined (at which point actual drainage of the aquifer begins), or until the lowering of the pressure head has created a steep pressure gradient between the aquifer being pumped and an overlying or underlying source aquifer. The steep pressure gradient across the aquitard that separates the overlying (or underlying) source aquifer from the pumped aquifer accelerates the rate of water flow through the aquitard. The pressure gradient increases until water flow through the aquitard to the pumping well is sufficient to match the depletion created by the quantity of water removed from the well.

Direction & Speed of Groundwater Movement

Groundwater moves from higher elevations to lower elevations and from locations of higher pressure to locations of lower pressure. Typically, this movement is quite slow, on the order of less than one foot per day to a few tens of feet per day. In groundwater hydraulics (the science of groundwater movement), water pressure surface and water table elevation are referred to as the *hydraulic head*. Hydraulic head is the driving force behind groundwater movement. Groundwater movement is always in the downward direction of the hydraulic head gradient (i.e., it moves down-slope, or "down-gradient"). If there is no hydraulic head gradient, there is no flow.

In the mid-19th century, the French engineer Henry Darcy demonstrated through a series of experiments that the rate of water flow through sediments is directly proportional to the hydraulic head gradient (i.e., to the slope of the water table). He found that when the hydraulic gradient flattened by 25%, the flow rate through the sediments dropped by 25%. Likewise, when he increased the hydraulic gradient by 90%, the flow rate also increased by 90%:

Flow rate is proportional to the hydraulic head gradient.

Darcy also found that the proportionality factor relating flow rate to hydraulic head gradient differs for different sediments. For a given hydraulic head gradient, coarser sediments produced higher flow rates than fine sediments. Darcy called the proportionality factor *hydraulic conductivity*. The fundamental equation relating these quantities came to be known as "Darcy's

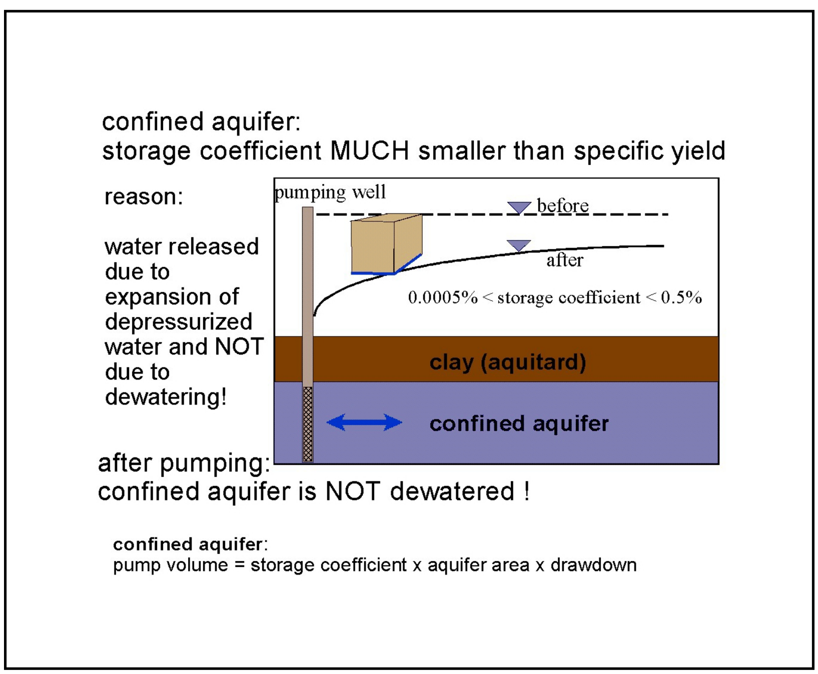

Figure 15. Confined aquifer.

Law":

"The flow rate through a one-square-foot cross-section of saturated porous medium equals the hydraulic conductivity times the change in hydraulic head per unit distance."

Hydraulic conductivity has the same units as the flow rate per unit cross-sectional area: length per time. Some common ways of expressing hydraulic conductivity are: gallons per day per square foot (gpd per sq. ft), feet per day, meters per day, or meters per second.

The hydraulic conductivities of different sediments and rocks can be vastly different; the range of values spans ten orders of magnitude (Figure 16). Depending on the percentage of fines in a sediment, the uniformity of the sediment's grain sizes, and other factors, even the hydraulic conductivity within a single sediment category (sand, gravel, etc.) can vary by several orders of magnitude.

The flow rate, or velocity, of groundwater is controlled primarily by the type of sediment through which the water flows. The hydraulic conductivity of sandy or gravelly aquifers, which have only small amounts of fine sediments, typically ranges from 100 to 100,000 gpd per sq. ft. (10 to 10,000 ft per day). On the other hand, the hydraulic conductivity of clays, which consist of tiny particles that stick together and block water movement, is a tiny fraction of that in sandy aquifers: one millionth to one thousandth of a gallon per day per sq. ft. Many aquitards consist of a high percentage of clay with also some coarser material. These typically have hydraulic conductivities only a little higher than that of clay. The hydraulic conductivity of fractured rock depends greatly on the degree of fracturing. It may be as high as 10 to 100 gpd per sq. ft. (1 to 10 ft/day).

Recall that the hydraulic conductivity of an aquifer is not the same as the actual flow rate. To obtain the flow rate, we must multiply the hydraulic conductivity by the hydraulic gradient (the slope of the water table or pressure head). The hydraulic gradient is often similar to that of the land surface. In most areas of California's valleys and basins that means that the hydraulic gradient is in the range of 1 to 10 feet per thousand feet (0.1% - 1%). A typical flow rate in a sandy aquifer therefore ranges from 0.1 to 100 gpd per sq.ft (0.01–10 ft/day).

Note also that the flow rate is not the same as the actual velocity of water, although both have units of distance per time. The flow rate gives the volume of water passing through an entire unit cross-section of aquifer per unit time. Groundwater, however, flows only through the porous part of the aquifer. It has therefore only about 10%–20% of that cross-sectional area to flow through. This means it must travel from 5 to 10 times faster than the (bulk) flow rate. To determine the velocity with which groundwater travels from point A to point B (referred to as "linear groundwater velocity" or "pore velocity"), we must divide the flow rate by the effective porosity of the aquifer (approximately 0.1 to 0.2, or 10% to 20%):

Groundwater velocity = flow rate divided by effective porosity

At the regional scale, typical linear groundwater velocities in sandy aquifers range from a few feet per year to several hundred or even a few thousand feet per year if the hydraulic gradient is high.

Figure 17 illustrates how the orientation of the sediment grains can influence the hydraulic conductivity. In the

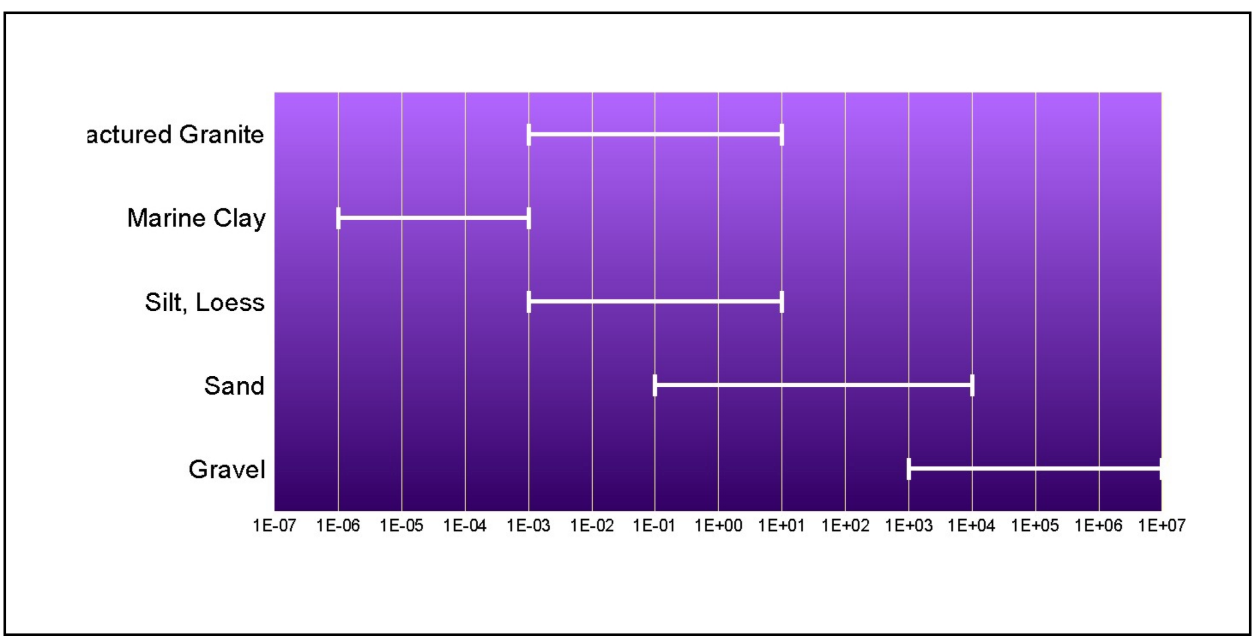

Figure 16. Hydraulic conductivity of sediments and rocks.

example shown, the grains are mostly elongated and oriented along the horizontal. The horizontal flow path of a water molecule is therefore much less tortuous (wiggly) than if the water molecule had to travel vertically through the same material. Hence, the hydraulic conductivity in the horizontal direction is larger than that in the vertical direction. This situation is often observed in alluvial and fluvial aquifers. There, in addition to preferential orientation of sediment grains, small, thin horizontal lenses of lower permeable material interbedded with the coarser sediments of the aquifer have a similar effect on the overall hydraulic conductivity of the aquifer: horizontal conductivity is greater than vertical conductivity. A similar phenomenon is observed in fractured rocks, where fracturing may be more dominant in one direction than another. If hydraulic conductivity of a sediment or fractured-rock aquifer is dependent on the flow direction, the aquifer hydraulic conductivity is called *anisotropic*, meaning unequal in different directions. The vertical hydraulic conductivity in most of California's layered alluvial aquifers is typically one to three orders of magnitude lower than the horizontal hydraulic conductivity.

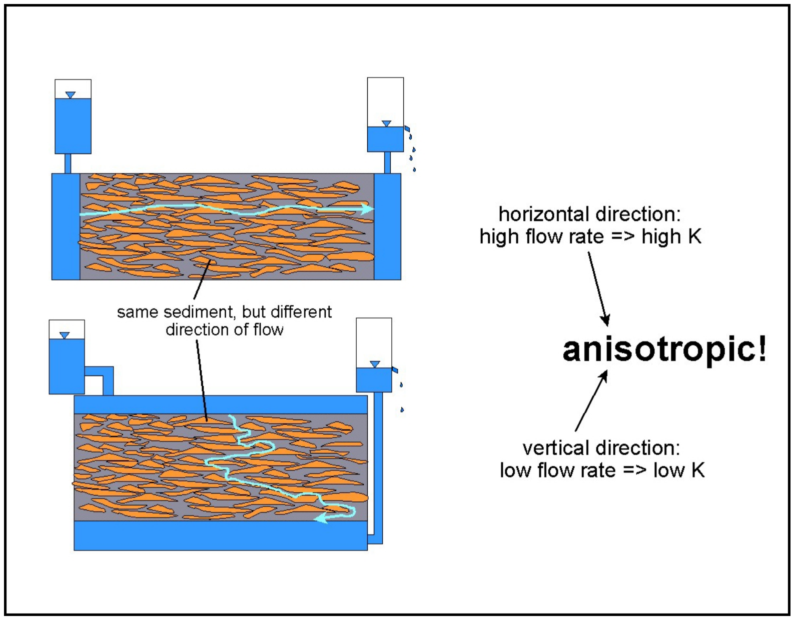

Figure 17. Orientation of sediment grains can influence hydraulic conductivity.

Transmissivity

The transmissivity of an aquifer is defined as the hydraulic conductivity times the thickness of the aquifer:

transmissivity = hydraulic conductivity x aquifer thickness

Transmissivity has units of area (length squared) per time, for example: gpd per ft., or sq. ft. per day. Transmissivity is often a better indicator of the water-production capacity of an aquifer than hydraulic conductivity. To see why, consider a thin aquifer, e.g., a sand bed interbedded (sandwiched) between thick clay layers. The bed has a very high hydraulic conductivity because it consists of clean sand; however, it is only 1 foot thick and will not sustain a large production well (its transmissivity is low).

To sustain municipal or agricultural production wells designed for pumping rates ranging from 50 gpm to 2,000 gpm, the hydrogeologist generally looks for an aquifer with a transmissivity ranging from 1,000 gpd per ft. (considered low) to over 100,000 gpd per ft. (considered high). Transmissivity (or hydraulic conductivity) has an important effect on the cone of depression created by a pumping well. In an aquifer with low transmissivity, the cone of depression is deep but of limited extent. In an aquifer with high transmissivity, the cone of depression is much shallower, but its boundaries extend much farther out from the well.

Specific Capacity

For purposes of designing a production well, drillers and hydrogeologists often determine the *specific capacity* of a well. Specific capacity is defined as the well pumping rate divided by the total drop in water level from pre-pumping conditions (drawdown):

specific capacity = pumping rate / drawdown

For example, if a well pumps at 1,000 gpm and the water level stabilizes at 50 feet below the pre-pumping water level (50 feet of drawdown), the specific capacity of the well is 1000 divided by 50, or 20 gpm/ft. A specific capacity of 1 gpm/ft or less is generally considered low (useful for domestic wells at best). On the other hand, a specific capacity of 100 gpm/ft is considered very high.

Specific capacity of a well is not the same as transmissivity of the aquifer, although the two are related. Specific capacity is approximately proportional to the hydraulic conductivity (or transmissivity), but it also depends on the construction of the well, the hydraulic efficiency of the well (a properly designed well has a higher specific capacity than a poorly designed well), the pumping rate (higher pumping rates yield lower specific capacity), and the location of the well relative to other wells and to rivers, lakes, and groundwater recharge or discharge zones. It is also a function of the duration of pumping: after the first few hours or days of rapid drawdown in a pumping well, specific capacity continues to decrease slightly over time.

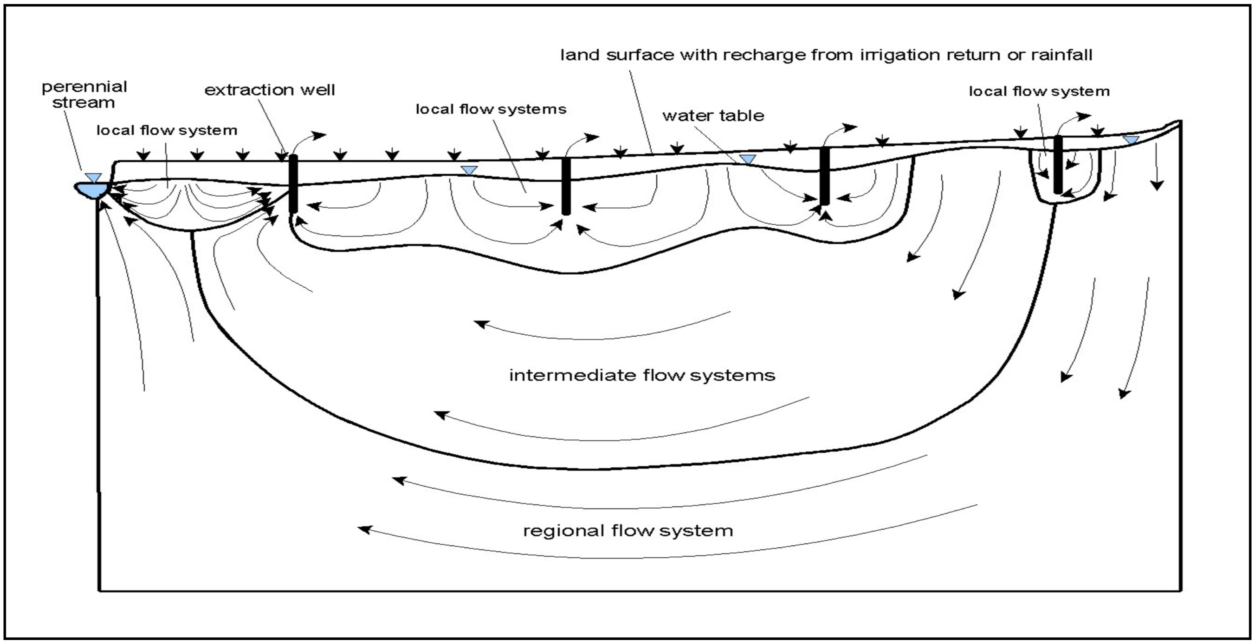

Figure 18. Local recharge-discharge systems.

Understanding Regional Groundwater Flow

A thorough understanding of regional groundwater flow patterns is essential for persons engaged in managing local and regional groundwater resources. It's equally important for those who seek to protect groundwater quality and to delineate wellhead protection zones or drinking water supply source areas. The technical procedures necessary for delineating recharge areas and wellhead protection areas are described in a separate booklet. The following is intended to serve as an introduction to general principles of regional groundwater movement.

In California, practically all groundwater consumed by municipalities, industry, and agriculture originates from historic and recent rainfall, from rivers, from unlined irrigation canals and lakes, from intentional recharge basins, or from deep percolation occurring beneath irrigated fields and orchards. From those areas of the aquifer where recharge is the predominant process, groundwater moves toward the areas where groundwater is discharged. Since most groundwater basins in California are well developed, groundwater discharge occurs predominantly via pumping. However, groundwater discharge also can occur via springs, direct discharge into streams or lakes, and subsurface outflow into a neighboring groundwater basin.

Sometimes the recharge and discharge areas overlap (recharge occurs in the area where wells pump groundwater). In other cases, the recharge and discharge areas may be as far apart as several tens of miles.

How does groundwater find its way from recharge areas to discharge areas? Simply by following the hydraulic gradient downward! In recharge areas, water is added to the aquifer. This increases the hydraulic head (water level or pressure head). In discharge areas, water is removed from the aquifer, thus lowering the hydraulic head around the discharge areas. Figure 18 shows several examples of local recharge-discharge systems.

The regional pattern in many of California's larger groundwater basins is dictated by topography and by the availability of surface water for groundwater recharge. Most of California has a semi-arid to arid climate; therefore, the largest amount of surface water available for recharge is near the mountain fronts, where perennial, intermittent, or ephemeral streams draining California's uplands (Coast Range, Sierra Nevada, mountain ranges of the desert southeast, etc.) flow onto the highly permeable, unconsolidated sediments filling California's valleys and basins and provide groundwater recharge. (This process is not unlike the milk flowing onto the surface of a deep bowl of dry granola—the milk disappears into the pores of the granola.)

Winter rains also supply recharge water. However, in all of California's large groundwater basins, direct recharge from precipitation accounts for only a small fraction of the total recharge. In the agricultural valleys, deep percolation from summer irrigations is a much more important source of recharge than winter precipitation. Recharge from natural precipitation is only significant in some of the coastal valleys and inland basins in northern California; and in the small groundwater basins nestled in California's mountain ranges.

At the regional scale, groundwater is recharged primarily near the margins of the basins or valleys. These are also the highest parts of the valley floor topographically (Figure 19). From there, groundwater moves toward the central axis of the valley or basin, often following more or less the topographic gradient. Near the edges of the groundwater basins, groundwater typically is unconfined. These unconfined zones also often feed deeper confined aquifers that occur beneath the central portion of a basin (Figure 20).

Water tables decline or rise as a result of an imbalance between the amount of water recharged into an aquifer and the amount of water withdrawn or discharged. It is a simple mass-balance problem and works just like a banking account. As part of any groundwater investigation or assessment, it is therefore an important practice to develop a solid understanding of the amount of water stored, transferred into, and taken out of an aquifer at the site, region, or basin of interest. To prepare a mass balance, the first step is to define accurately the boundaries of the site, region, or basin of interest. These include the land surface boundary, lateral subsurface (groundwater flow) boundaries along the edges of the site, region, or basin, and boundaries at the bottom of the aquifer region of interest. The next step is to estimate the water flux across each of these boundaries. Physical boundaries of an aquifer basin—for instance, the boundaries between permeable aquifer material and relatively impermeable bedrock—act most often as barriers to flow, since water fluxes across these boundaries are often known to be negligibly small. Other fluxes that must be estimated include: the recharge and discharge rates from and to rivers and lakes, the net aquifer recharge from precipitation and from irrigation percolation, the pumping discharge, the discharge rate of groundwater to or from overlying or underlying aquifers, and the discharge to or from laterally adjacent aquifer basins

Boundaries of California's Regional Aquifer Systems

An important part of understanding regional groundwater flow is knowledge about the boundaries of the regional groundwater flow system. What is the extent of a groundwater basin? How does it connect to other groundwater basins? The physical boundaries of California's groundwater basins, as depicted, for example, on the Water Education Foundation's "California Groundwater Map," are determined mainly by the rather abrupt rise of hills and mountains around the edges of California's large valleys and basins. These mountain ranges consist of consolidated rocks (granites, volcanics, and metamorphic or sedimentary rocks), which generally have much lower hydraulic conductivity than the alluvial, unconsolidated sediments of

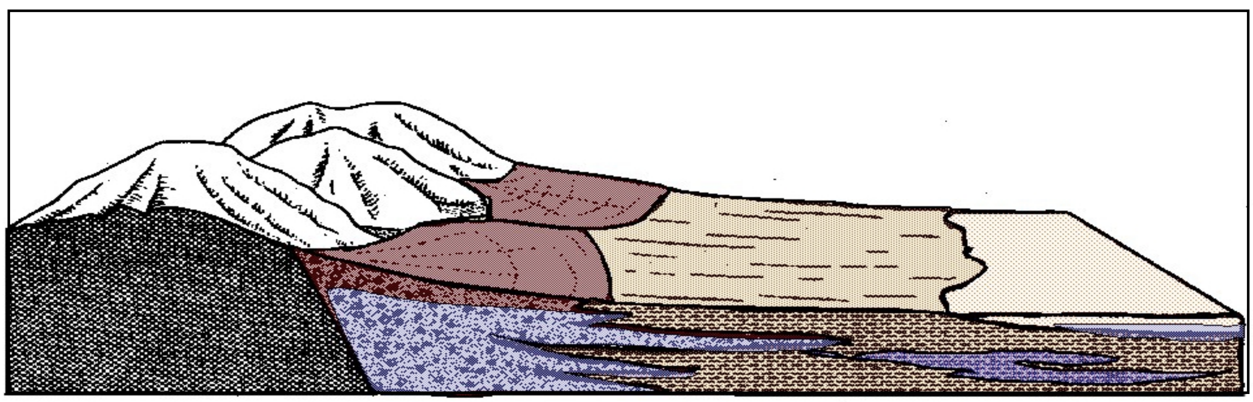

Figure 19. Typical recharge and discontinuous aquifers at margins of one of California's groundwater basins.

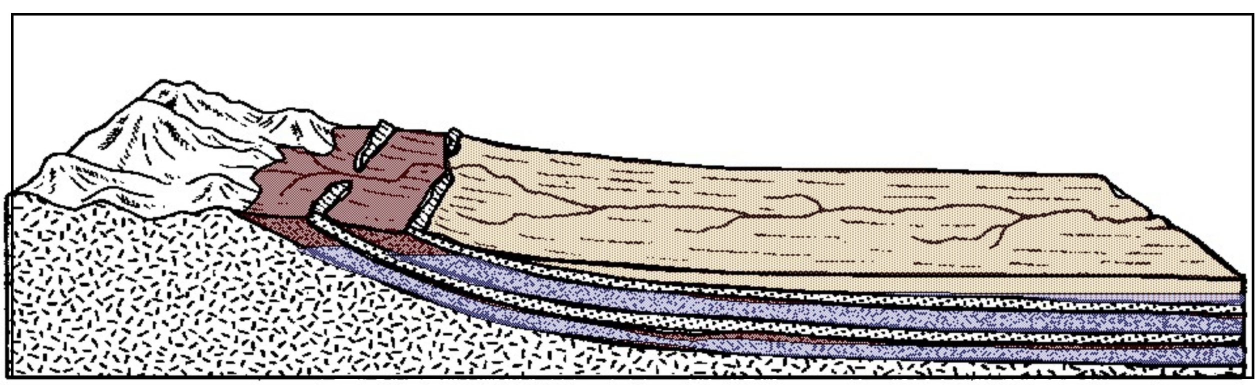

Figure 20. Typical multiple confined aquifers in the central portion of one of California's groundwater basins.

California's groundwater basins. Because even these mountain rocks can transmit water, they are—strictly speaking—not true boundaries. But, for most practical purposes, the foot of a mountain range can be designated as the boundary of an alluvial groundwater basin. Where adjacent rock formations or fracture systems are known to contribute significant recharge to a groundwater basin, that recharge should be taken into account when computing a water mass balance.

The boundaries of aquifers in fractured rocks, semi-consolidated porous sedimentary rocks, and unconsolidated sediments underneath an undulating (hilly) topography are more difficult to define, because they may not coincide with topographic features. The boundaries of these aquifers are defined by the extent of the geologic strata and by natural *groundwater divides.* Geologic mapping and interpretation are the primary tools for defining such aquifer boundaries. Natural groundwater divides can be found from maps showing the regional water table or pressure head surface. Aquifer boundaries rarely coincide with watershed boundaries (defined by the actual topographic ridges around a watershed).

Two types of fractured rock aquifers often are particularly productive: volcanic (primarily basaltic) rocks, and limestone or dolomite rocks. The gas bubbles initially present in lava flows can leave a highly porous structure or even large cavities in the basalt that results from the cooling lava. Limestone and dolomite rocks form their porosity in a very different way: just below the water table, such rocks are subject to *karst corrosion*—that is, the slow dissolution into water of the carbonate minerals that make up the rock. This process results in ever-widening fissures that eventually may develop into entire cave systems and sometimes actual underground rivers.

With respect to groundwater and wellhead protection, the above types of aquifers provide particular challenges, because the travel time of contaminants can be exceedingly fast—up to several hundred yards per day—yet the capability of the aquifer to attenuate and neutralize contaminants by physical filtering and microbial biodegradation is very small. The source areas for water wells in these types of aquifers are likely to be large and subject to significant uncertainty. Boundaries of these aquifers are best defined by geologic interpretation and with the aid of other hydrogeologic tools, such as geochemical analyses and tracer tests. If these aquifers are in direct communication with surface water systems, they are often considered to be part of the surface water resource ("under the direct influence") and treated accordingly with respect to source water delineation and protection.

Understanding Wells & Well Construction

A water well is a hole, shaft, or excavation for the purpose of extracting groundwater from the subsurface. Water may flow to the surface naturally after excavation of the hole or shaft; such a well is known as an *artesian well.* More commonly, water must be pumped out of the well, manually or mechanically.

Most wells are vertical; however, they may also be horizontal or at an inclined angle, where necessary. (The oldest known wells, known as "Qanat," were horizontal shafts built into the mountains of the Persian Empire several thousand years ago. Many are still in use.) Some wells are used for purposes other than obtaining drinking water. Pumping of oil or gas is one example. Monitoring of groundwater levels (hydraulic head) and groundwater quality is another. Still others purposes are: investigation of subsurface conditions, shallow drainage, artificial recharge (water injection), and waste disposal.

The first step in designing and constructing a well is to find a suitable location to meet the specified purpose of the well. Next, a hole or shaft is drilled, the well casing and surrounding packing are designed and installed, and the well is "developed" and tested. Finally, a pump is installed and a sanitary seal (the well head) is completed. Proper design and completion of a well helps to ensure that the well has a long life and that it operates efficiently. This section provides an overview of the most important design criteria and construction techniques employed by hydrogeologists and water well drillers in building a well.

The location of a well is determined primarily by the purpose of the well. For drinking water production wells, groundwater quality and long-term groundwater supply are important considerations. In most cases, the location is limited by property ownership, the need for minimal surface transportation of the pumped groundwater, and access restrictions for the drilling equipment. When locating a well, one must also consider the proximity of potentially contaminating activities—such as fuel or chemical storage areas, nearby streams, and leach fields or septic tanks. Also important: the presence of a significant barrier between such potential contamination sources and the well itself. Hydrogeologically, a well's location may be constrained to certain areas by the availability of groundwater and by the quality of the groundwater. For drinking water wells, the quality criteria to use are the federal and state drinking quality water standards. For irrigation wells, the primary chemical parameters of concern are salinity and trace metals concentrations.

Of course, groundwater availability must be sufficient to meet the required pumping rates of the wells. For large municipal and agricultural production wells, pumping rate requirements range from 100 gpm to 4000 gpm. Small- and medium-sized community water systems may depend on water wells producing from 10 gpm to 100 gpm. Domestic wells may meet their purposes with as little as 0.5 to 5 gpm. To determine whether the desired amount of groundwater is available at a particular location and whether it is of appropriate quality, drillers and groundwater consultants rely on their prior knowledge of the local groundwater system, plus a diverse array of information such as land surface topography, local vegetation, rock fracturing (where applicable), local geology, groundwater chemistry, information on thickness, depth, and permeability of local aquifers from existing wells, groundwater levels, satellite or aerial photographs, and geophysical measurements. For many large production wells, a smaller test hole will be completed prior to drilling the production hole, to obtain more detailed information about the depth of the aquifer, the hydraulic conductivity and storage coefficient of the aquifer, water levels, and groundwater quality.

Once a well location is determined, a preliminary well design is completed. The final design is subject to site-specific observations made during the borehole drilling (drilling log). The primary objective of the well design is to allow water to move freely and sediment-free into the well at the desired amount and to ensure that the well remains open (does not cave in) and provides enough space to house pumps (or other extraction devices).

The principal components of a production or monitoring well in California are shown in Figure 21. A well consists, at a minimum, of well casing (pipe) surrounded by a well pack, a well pump, and appropriate surface and borehole seals. Water enters the well through perforations or screens in the well casing. Wells can be screened continuously (i.e., without regard to the depths at which highly permeable sediments or rocks occur). Or they can be screened at specific (often multiple) intervals that are matched to the aquifer layers from which production is to occur. Note that in highly permeable sand and gravel aquifers, inexpensive wells have sometimes been constructed as "open-bottom" wells consisting of blank casing: a solid pipe with no side perforations and an open bottom.

The purpose of the screen is to keep sediments and fines in the aquifer and out of the well. The purpose of the well casing is two-fold: (1) to provide an open pathway from the aquifer to the surface, and (2) to protect the pumped groundwater from interaction with shallower groundwater that is potentially of worse groundwater quality. Even in hard rock wells where no casing is deemed necessary at depth (for example, when a borehole intersects large, debris-free fissures in a fractured-rock aquifer), a shallow surface casing is necessary to facilitate construction of a well seal. The well seal is an integral part of the protective surface installation (well cap) that prevents contamination from entering the borehole.

The annular space between the well casing and the borehole wall is filled with a natural or artifical well pack (often referred to as *gravel pack*). The uppermost section of the annulus is sealed with a grout to ensure that no water or contamination can enter the annulus from the surface. (The depth to which grout must be injected varies by county.) Minimum requirements for well design in California are defined in the California Well Standards (Bulletin 74, California Department of Water Resources). Local county ordinances may require more stringent well designs. A simple decision-tree chart for designing a well is shown in Figure 22.

Wells can be constructed in a number of ways. The most common drilling techniques in California are: direct driving, auger drilling, cable tool drilling, rotary drilling, reverse rotary drilling, and air drilling. Auger drilling is employed mostly for shallow wells or small diameter wells. Cable tool, rotary, and reverse rotary drilling (Figure 23) are the most common drilling methods employed in unconsolidated and semi-consolidated material (although also used in hard rock). Many hard rock wells are drilled with the air drilling method.

While drilling the well, drillers keep a detailed log of the material encountered in the subsurface at any given depth. After completing the drilling, it is often desirable to obtain more detailed data of the subsurface geology

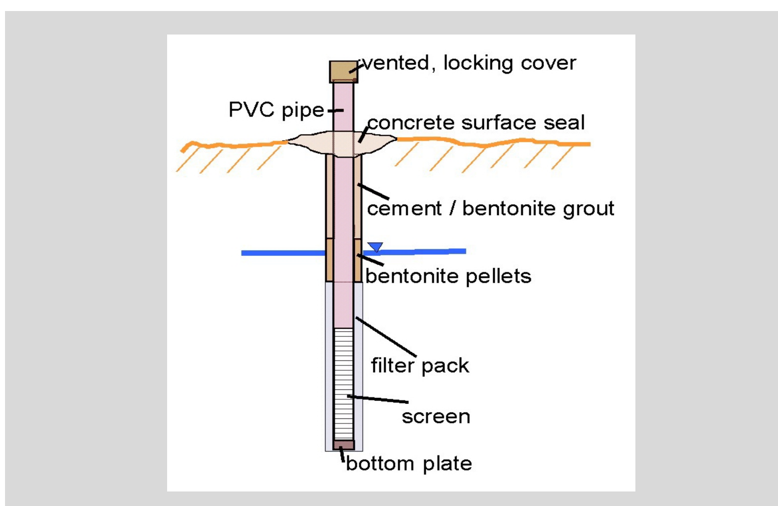

Figure 21. Principal components of a production or monitoring well in California.

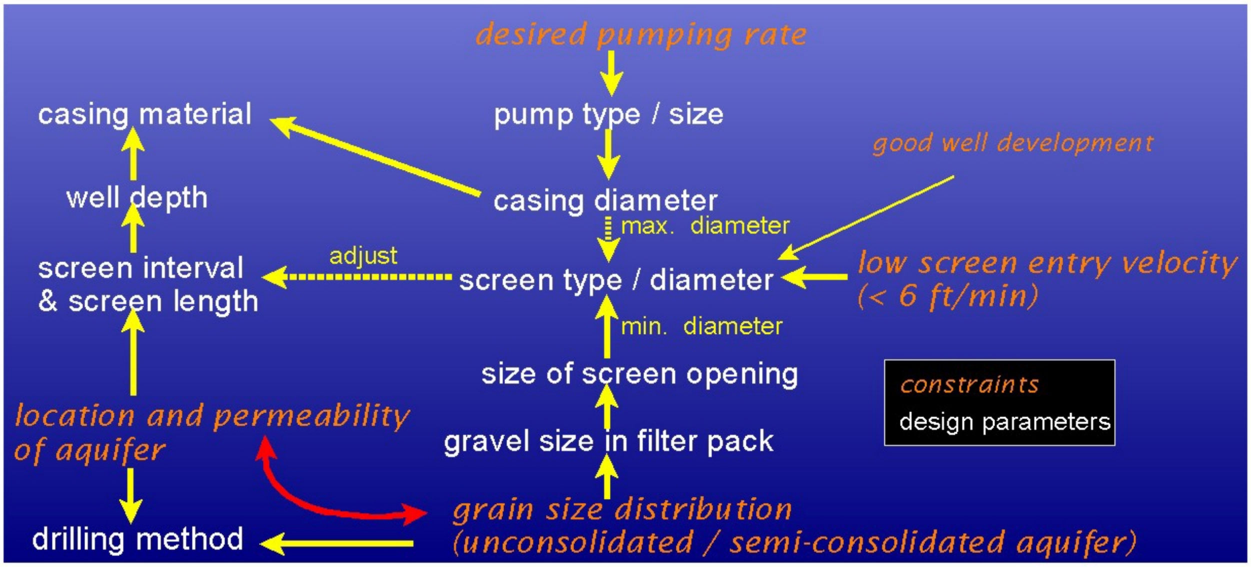

Figure 22. Decision-tree chart for designing a well.

by taking geophysical measurements along the borehole. Most commonly, drillers measure the *electric resistivity* of the geologic material along the borehole wall (producing an *e-log*). (Sand has a higher resistivity than clay; high salinity reduces the electric resistivity of the geologic formation.) They also measure the *self-potential* or *spontaneous potential* of the geologic formations along the borehole wall. Careful interpretation of the e-log and the spontaneous potential provide important information about water salinity and about the location and thickness of sand bodies and clay layers or clay lenses.

With this information in hand, the driller or water well consultant prepares the final design of the well screen, including its thickness and location, the type of screen, and the size of the screen's openings. He or she will also determine the makeup of the gravel or sand pack (filter pack) that will fill the annulus around the well screen, where necessary. A "natural filterpack" is obtained where clean sand and gravel formations collapse against the well screen (and casing) after the borehole equipment has been removed.

After the well screen, well casing, and filter pack have been installed, the well is "developed" to clean out the casing from any drilling muds and to properly settle the filter pack around the well screen. Well development typically is implemented by surging or jetting water in and out of the well. It ensures that the gravel is tightly packed around the well screen. A well developed filter pack provides an efficient filter to keep fine sediments out of the well while not impeding the flow of water to the well screen and into the well. Once completed, all drilling equipment is removed, and the final surface seal and well pump are installed. Typically, a pumping test lasting from 12 hours to 7 days is then carried out, to determine the efficiency and capacity of the well and to provide information about the permeability of the aquifer.

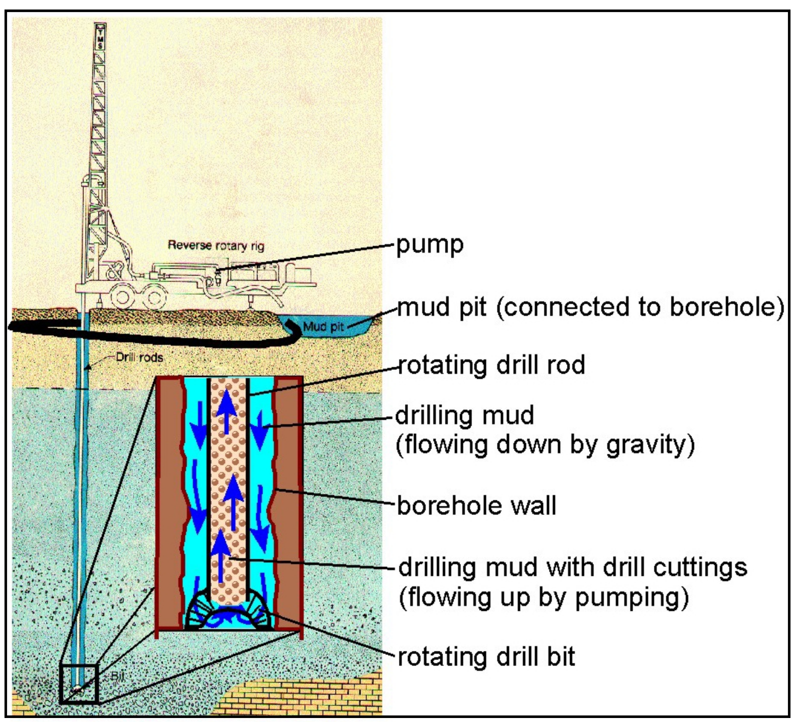

Figure 23. Reverse-rotary drilling method.