CHAPTER 4
GROUNDWATER

1. INTRODUCTION

1.1 I suppose I don't need to tell you that ***groundwater*** is the term used for *the liquid water that's present beneath the land surface in the pore spaces in regolith and in cracks in bedrock* in virtually all areas except in very cold climates where the subsurface water is permanently frozen in the form of what's called permafrost.

1.2 Groundwater is invisible, until it emerges at the surface from springs and wells. And there's nothing spectacular about groundwater and its movement, the way there is with, say, catastrophic landslides or floods. But much of the world's water supply is from groundwater wells, so it's a topic of great importance.

1.3 I think it's true that the general public is aware of the existence of fresh water underground. I suspect, however, that a clear understanding of the environment and movements of that water is not widespread. Where does groundwater come from? Where does it reside? How does it move? What happens to it? How deep down do we have to drill or dig to find it? How deep in the Earth does it extend? It's questions of this kind that we need to address in this chapter.

2. HOW WATER GETS TO BE GROUNDWATER

2.1 The main way that groundwater is replenished is from the Earth's surface, by infiltration of surface water down through the soil to become groundwater. As a prelude to our study of groundwater, this section deals with this first step in the process of groundwater flow: infiltration of surface water through the uppermost layer of the solid Earth.

2.2 You all know that some of the rain that falls on the soil surface runs off into streams and rivers and some sinks down into the soil. I'll define ***infiltration*** as *the downward movement of water across the upper surface of the soil layer*. The ***infiltration rate*** is *the rate at which the surface water moves downward across the upper surface of the soil layer*. The infiltration rate is measured in depth of water per unit time, the same as precipitation. The ***infiltration capacity*** is *the maximum rate of infiltration at a given point on the soil surface and under a given set of conditions*.

2.3 Once the water infiltrates into the soil, it tends to continue its downward movement into deeper layers. The term ***percolation*** is used for *the continued* *downward movement of the water that infiltrates the upper surface of the soil layer*. Not all of it keeps on moving downward, however: if the regolith layer is sufficiently dry beforehand, some of the downward-percolating water is left behind in the pore spaces between the regolith particles, mainly in two forms: little fillets at grain contacts, and absorbed by porous materials like plant residues.

3. A HOME EXPERIMENT ON INFILTRATION

3.1 Here's a home experiment that would not be difficult for you to do if you are a home gardener. It's a very realistic experiment on the controls on infiltration rate and infiltration capacity.

3.2 Set up an adjustable-rate lawn sprinkler on a very gently sloping and freshly spaded and raked plot in your garden (Figure 4-1). Measure the "precipitation" from the lawn sprinkler with a little rain gauge, and measure the surface runoff directly by collecting it or catching it at the downslope edge of the plot. You can compute the infiltration rate indirectly by the equation

$$S = P - I \tag{1}$$

where S is surface runoff, P is precipitation rate, and I is infiltration rate. You have to express S in units that are the same as I and P: depth of water per unit area per unit time. You can easily spot the point of nonzero S when you're watering your garden: water stands on the soil surface as a thin, smooth sheet instead of soaking in immediately.

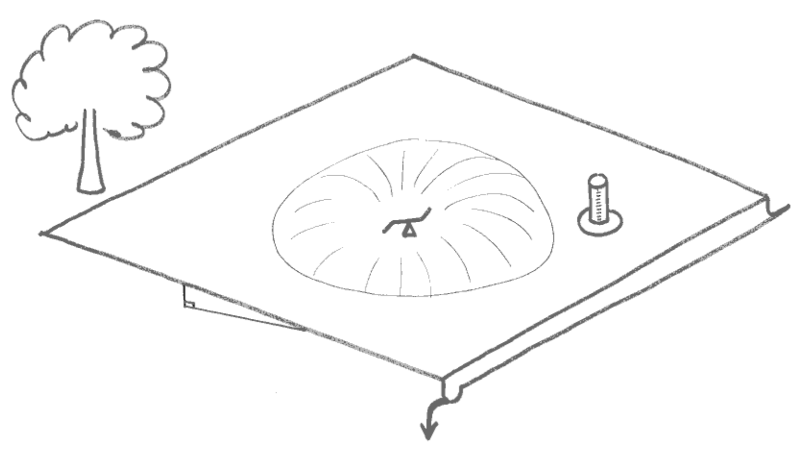

Figure 4-1. An experimental sprinkler plot for studying infiltration.

3.3 But there's a little more to it than this. Figure 4-2 is a typical plot of infiltration capacity I_C against time, for a given soil. There's a rapid decrease from very high I_C in the first few minutes, but then it levels off to a steady value. (It's hard to measure I_C at the very beginning.) Why the initial decrease?

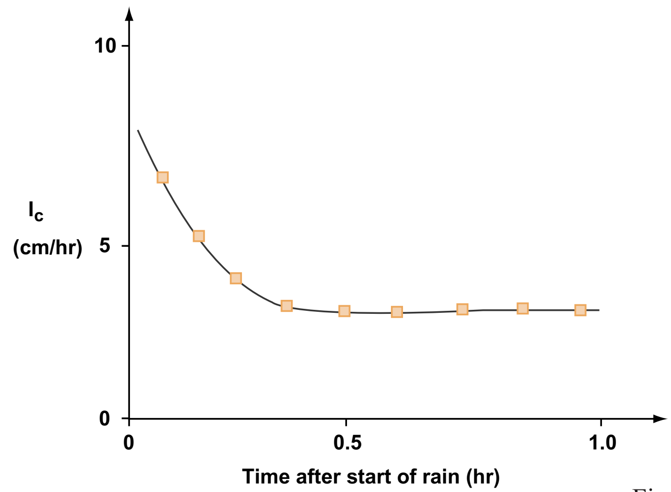

Figure 4-2. Plot of infiltration capacity against time for a given soil.

(1) Some soil particles, like clay or organics, have a tendency to swell when they come in contact with liquid water, and this tends to close off the pore spaces. And your intuition should tell you that *the smaller the passageways, the more difficult it is to push water through them.*

(2) The impact of raindrops pulverizes the soil structure at the surface and washes fine particles into the uppermost pore spaces, producing a kind of "micro-armor".

3.4 How is the infiltration capacity measured?

(1) With an infiltrometer: a tube about one foot in diameter, driven about one foot into the soil. You maintain the water level in the tube at some height above the ground surface, and you measure the rate of input of new water needed to replenish the water lost to infiltration. The advantages of this arrangement are that it's inexpensive and easy to use; the disadvantage is that it disturbs the soil.

(2) With a sprinkler on an experimental plot, as in the home experiment above. The advantage of this is that it's accurate, realistic, and non-disruptive, but the disadvantage is that it's expensive and time-consuming, and you can't do it very well on irregular ground.

3.5 What is going on, physically, during infiltration and percolation? Its simple: a raindrop falls on the granular surface and then drains down through the interconnected pores of the soil by the pull of gravity. The passageways are diverse: they might be intergranular interstices, animal burrows, shrinkage cracks, or rotted rootways. If the precipitation rate is less than the infiltration capacity ($P < I_C$), then some of the pores are occupied by downward-draining water but others are occupied by air (Figure 4-3A). If the precipitation rate is greater than the infiltration capacity ($P > I_C$), then all of the pores are occupied by downward-draining water (Figure 4-3B). In both cases the downward-moving layer of water is marked by a blurred but recognizable front or lower boundary. You can see this nicely for yourself by excavating and examining a little vertical wall of soil in your garden after a heavy brief shower or a longer rain after a dry spell (Figure 4-4). After the rain stops, the infiltrated water keeps descending as a coherent layer.

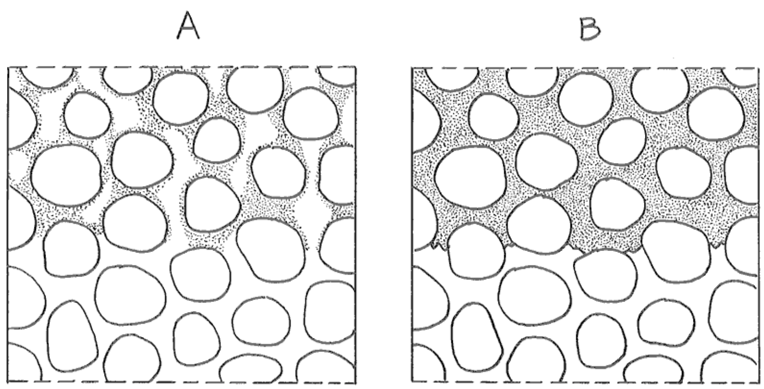

Figure 4-3. Sketch of grain-scale infiltration behavior. A) Pore spaces are occupied by both water and air. B) Pore spaces are occupied entirely by water.

3.6 You perform an experiment like this every time you water your house plants! Take a pot with established soil structure, dry it to total dryness in the oven, weigh it, and then water it very fast, to simulate the condition that P is greater than I_C. As the leading or lower boundary of the percolating layer passes the bottom of the pot, water starts pouring out onto the floor. Then, as the trailing boundary of the percolating layer passes, water stops coming out of the bottom of the pot rather abruptly. Now weigh the pot again, and convert the difference between final weight and initial weight to *volume of water per unit bulk volume of the soil*. This quantity is call the ***field capacity*** of the soil: *the maximum concentration of soil water that can be held by the soil against the pull of gravity*. The field capacity is substantially greater than zero in most soils. It's much smaller, but still nonzero, even in very coarse gravel soils.

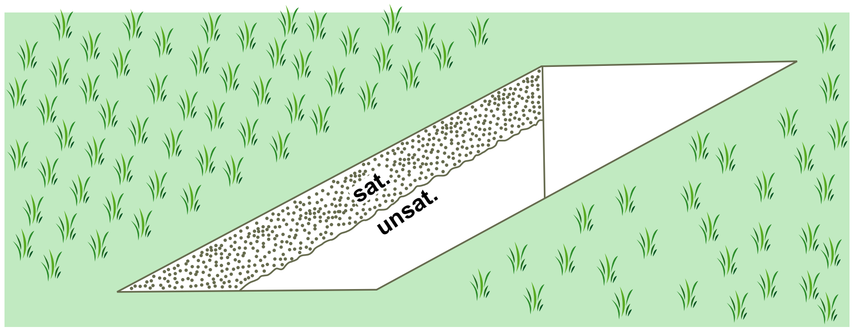

Figure by MIT OCW.

Figure 4-4. Studying infiltration after a heavy rain by means of a vertical-walled trench.

3.7 *The water that remains in the soil after initial drainage is* called ***soil moisture***. The water that constitutes this soil moisture is located in several kinds of places in the soil (Figure 4-5):

- as thin films on all particles
- as fillets at points of grain contact
- suffusing porous organic matter

3.8 Soil moisture makes up a very small percentage of fresh water, but it's of critical importance for life on the Earth's surface, because it's what plants use. Plants pull water out of the soil and transpire it, until the remaining soil moisture is held so tightly that plants can't take it up, and then they wilt. This level of moisture is called the ***wilting point***. The soil moisture is depleted in two other important ways as well:

- slow *residual drainage* downward
- evaporation at the surface, after *upward diffusion* as water vapor through the soil or by *capillary rise* as a liquid through the soil.

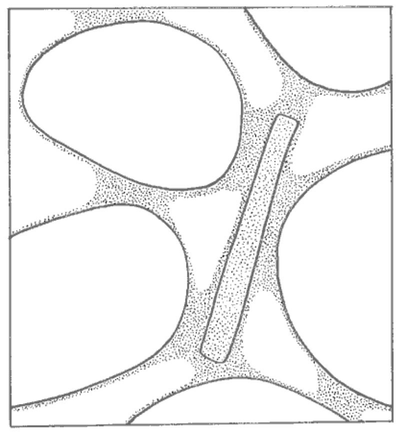

Figure 4-5. Sites of soil moisture.

BACKGROUND: SURFACE TENSION AND CAPILLARITY

1. Inside a liquid like water, each molecule is attracted by molecules all around it. A molecule at the surface of the liquid, however, is attracted by molecules below it and beside it but not above it, because none are there. The net effect of that asymmetry of forces is that the surface water molecules attract one another in the plane of the surface, and the result is analogous to the tensile force in a thin sheet of stretched rubber. That's why the effect is called ***surface tension***. It's as if the surface wants to "pull itself together". It's for that reason that globules of water, and soap bubbles as well, try to take on the shape of a sphere.

2. Now think about a small mass of water resting on a solid surface. If the attraction between the water molecules is greater than the force of attraction between the water molecules and the atoms or molecules of the solid, then the mass of water "balls up" on the surface. We say that the liquid does not wet the solid surface. If the force of attraction between the water molecules is less than the attraction between the water molecules and the atoms or molecules of the solid surface, then the liquid spreads out as a uniform thin film. We say that the liquid wets the surface.

3. If you put a hollow tube with very small diameter into a liquid that wets the surface of the solid of which the tube is composed, the liquid rises up in the tube, above the liquid level outside the tube, because of the tendency for the liquid to spread itself upward onto the solid surface. The effect is called ***capillarity*** or ***capillary rise***.

4. The smaller the diameter of the tube, the higher the capillary rise in the tube. That's because the height of capillary rise is the outcome of two competing effects: the upward force of capillarity, and the downward force of gravity. The surface area within the tube goes as the square of the diameter, whereas the perimeter of the tube goes as the first power of the diameter. Because the capillary rise depends on the perimeter but the downward gravity force depends on the cross-sectional area, the smaller the tube, the greater the upward capillary force relative to the downward gravity force.

5. You may had noticed that in such a situation the water surface inside the tube is concave upward, with the level higher at the wall of the tube than in the middle of the tube. That curved surface is called a ***meniscus***.

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4. THE PHYSICS OF GROUNDWATER MOVEMENT

4.1 Once surface water infiltrates below the surface of the soil and keeps on moving downward by percolation, it has become groundwater. At this point we have to deal with the physics of groundwater movement. This comes under a branch of fluid dynamics known as *flow through porous media*. The essential features of flow through porous media are in common with flow in pipes and channels, (Chapter 1), but certain aspects are distinctive to groundwater flow.

4.2 I'll say some things about the basic nature of groundwater movement by means of another rewarding home experiment (Figure 4-6). This one is quite simple to do and would be easy and inexpensive to set up in just a short time in your basement or in your backyard. For this experiment you will need to have a large vessel, open at the top, that will hold water without too many leaks. You could nail something together out of wood and make it leak-proof with a caulking gun, or you could borrow or buy a 55 gallon metal drum. A large plastic rubbish barrel would be good also. Mount the barrel on a stand above the floor or the ground, and attach a horizontal pipe to the wall of the vessel just above the bottom. Don't worry too much about the practical problem of how to attach the pipe to the barrel in a way that's leak-proof or nearly so. Tack a piece of fine metal screening over the entrance to the pipe as it leaves the barrel. From the downstream end of the pipe, stuff the pipe full of some natural granular material like the soil in your garden or the sand in your sandbox. When the pipe is full, attach another piece of fine metal screening to the downstream end of the pipe.

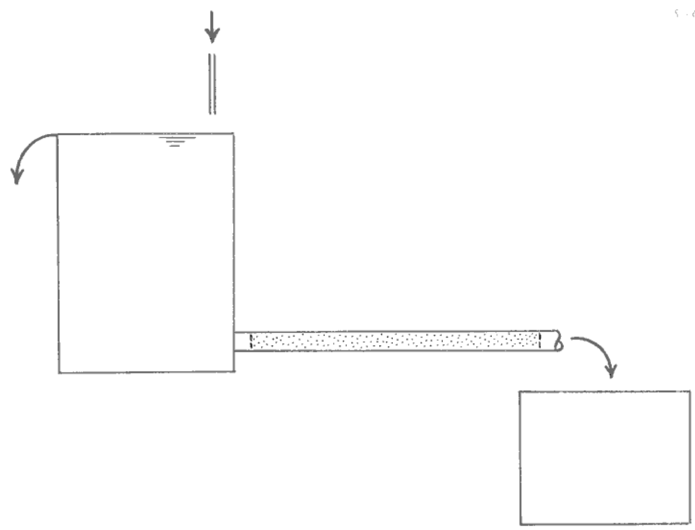

Figure 4-6. A home experiment on groundwater flow.

4.3 Run your garden hose into the vessel until water overflows, and then leave it running during the experiment to ensure a constant level of water in the vessel. What you want to do is measure the discharge (that is, the volume rate of flow) of water through the porous medium and out the downstream end of the pipe. You can easily measure this by using a catch tub at the downstream end of the pipe and measuring the volume of water caught in the catch tub with a measuring cup and measuring the time of flow with a stopwatch.

4.4 Think now about the nature of the porous granular medium in the pipe: it consists of *an enormous number of little solid particles, of more or less irregular geometry, each in contact with some small number of neighboring particles*. Look at some representative unit volume of the porous material. It's made up of two different sub-volumes (Figure 4-7): *solid granular material*, and *empty interstitial spaces*, which might be called ***void spaces*** or ***pore spaces***. Before you start the experiment, these pore spaces are filled with air. One of the important physical properties of the porous medium is the ***porosity***, defined as *the volume of pore spaces divided by the bulk volume of the material.* Loose sands or gravels typically have porosities of something like 20–30%, depending mainly on the size distribution of the material: a well sorted granular material has a larger porosity, other things being equal, than a poorly sorted granular material.

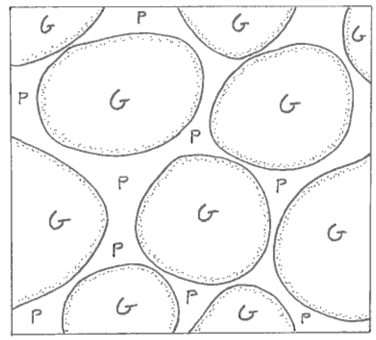

Figure 4-7. Geometry of pore spaces in a porous granular medium.

4.5 What I want you to study in your experiment is *the discharge of water through the pipe as a function of the height of water in the vessel*. You can easily vary the height of the water in the barrel by cutting holes in the side of the vessel so that the excess water runs out of the vessel at different levels above the entrance to the pipe.

4.6 Before we discuss the results of this experiment, let's think about *the nature of flow through the porous medium in the pipe*. You can't easily observe that flow, because even if you made the pipe out of a transparent material like glass or acrylic plastic you would still have a hard time observing the details of flow among the grains. But you could imagine shrinking yourself down to microscopic size and taking a submarine trip through the porous medium by drifting along with the flow.

4.7 The flow through the porous medium is like flow through a large number of tiny pipes with very irregular geometry, branching and rejoining, which narrow as they pass around grains and widen as they pass between grains. The narrow, constricted places among grains are called ***pore throats***. Although the patterns of flow are very complicated in detail, *the essential nature of the flow is not greatly different from the flow through a pipe*; it's just that the "pipe" is now narrow, tortuous (with lots of twists and turns, that is), and very intricate geometrically.

4.8 The question now arises: *Is the flow through the porous medium laminar or turbulent?* In laminar flow, the fluid moves along in regular straight paths, without mixing sideways. When you pour a stiff (that is, very viscous) liquid like oil or paint, you are observing laminar flow. Flow of water or air at all but the slowest speeds, however, is turbulent. By putting little floating tracers in the flow you can observe how flow paths are sinuous and irregular. An even better way to observe turbulence is to watch smoke as it rises from a chimney or smokestack: you can see the irregular swirls, called turbulent eddies. Go back to Chapter 1 for more details on laminar and turbulent flow.

4.9 Whether a flow of fluid is laminar or turbulent depends on four factors: the speed of the flow, the depth or width of the flow, and the density and viscosity of the fluid that's flowing. The deeper and faster the flow, and the greater the fluid density and the smaller the fluid viscosity, the more likely the flow is to be turbulent. In the case of flow through the porous medium in the pipe, the passageways are very small, and the flow velocity is in almost all cases rather small, so you should expect that *flow in porous media is laminar rather than turbulent*, except in unusual situations where the passageways are very large and the velocities are very great, as for example in very coarse open gravels, or in certain kinds of basalt flows with large connected tunnels or passageways, or in solution tunnels in limestones (Figure 4-8).

4.9 What drives the flow through the sand-filled pipe is *the downstream pressure gradient* (that is, the rate of decrease of fluid pressure with downstream distance in the pipe). That's the same thing that causes the water to flow out through your home piping system! It should make good sense to you that the speed of flow through the porous medium in the pipe depends upon the pressure gradient: the greater the pressure gradient (the driving force that causes the flow), the greater the speed of flow. But you should expect two other factors to affect the speed of flow as well: the size of the solid particles of the porous medium, and the viscosity of the fluid. The particle size is important because larger particles mean wider the pore throats, and therefore less frictional resistance to flow, because the friction arises from contact of the fluid with solid surfaces. (Go back to Chapter 2 and review how the ratio of surface area to volume increases with decreasing particle size.) The viscosity of the fluid is a measure of the resistance of the fluid to a deforming force: the greater the viscosity, the more difficult it is to make the fluid flow—as all of you know from trying to make the honey flow out of the bottle, compared to water.

ADVANCED TOPIC: MATHEMATICAL ANAYSIS OF FLOW THROUGH A POROUS MEDIUM

1. It should make sense to you that there must be some mathematical relationship between the various physical quantities involved in the flow described in the preceding paragraphs: the average speed of flow V in the pipe, the pressure gradient G, which drives the flow through the pipe, the grain size D of the particles that constitute the porous medium, and the viscosity μ of the fluid flowing through the porous medium. Provided that the flow is sufficiently slow, the accelerations of the fluid as it flows through the little passageways are very small, so we can neglect the inertial properties of the flow as embodied in the density ρ. So some function of V, G, D, and μ has to be equal to a constant:

$$f(V, G, D, \mu) = \text{const}_1 \qquad (2)$$

2. Because the right side of Equation 2 is a constant number and has no physical dimensions, the variable or variables involved in the left side of the equation have to be dimensionless as well. You can easily demonstrate for yourself that the mass, length, and time dimensions of the quantity $GD^2/V\mu$ all cancel out, so it's a dimensionless quantity. So the functional relationship has to look like this:

$$\frac{GD^2}{V\mu} = \text{const}_1 \qquad (3)$$

You can easily rearrange Equation 3 to show that the velocity V of flow through the porous medium is linearly related to the quantity GD^2/μ:

$$V = \text{const}_2 \cdot \frac{GD^2}{\mu} \qquad (4)$$

(where const_2 is just the inverse of const_1). The value of const_1 (or const_2) depends on the shape and packing of the particles in the pipe.

3. If now you made a number of runs with different water levels in your supply tank, you'd get a graph of the results that looks like Figure 4-9: you'd see a *linear relationship* between the velocity V and the quantity GD^2/μ. (How do you measure G and V, you might be thinking. Well, G is just the difference between the pressure at the upstream end of the pipe, which by the hydrostatic equation is just ρgh, where h is the height of the pipe entrance below the water surface, and the pressure at the downstream end of the pipe, which is the same as the atmospheric pressure, divided by the length of the pipe. And V is equal to the discharge divided by the cross-sectional area of the pipe.) Experiments like this have been done many times, and they show that there's a function like this for all porous media. But you should expect that *the value of the constant is different for different porous media, because of the differences in particle shape and packing geometry*.

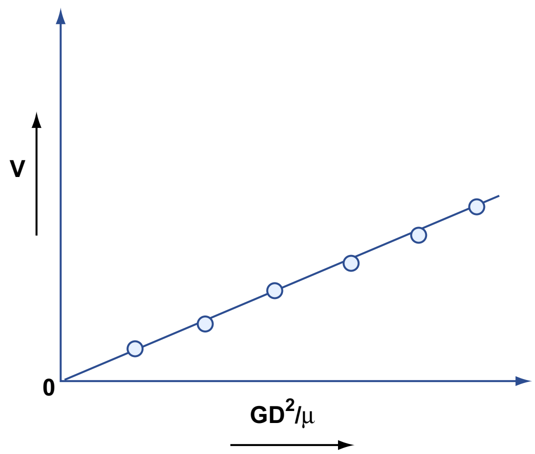

Figure by MIT OCW.

Figure 4-9. Graph of flow velocity V against GD^2/μ for flow through a porous medium.

4. Incidentally, the fact that you indeed find a linear relationship in your experiment shows you that we were correct in our assumption that the flow in the porous medium is laminar and that the density can therefore be neglected.

5. The results of the barrel experiment are a manifestation of a well-known law in the flow of porous media called ***Darcy's law***. Darcy's law states that *the flow speed through a porous medium is directly proportional to the pressure gradient through that medium, and directly proportional to the square of the characteristic size of the pore spaces of the medium, and inversely proportional to the viscosity of the medium.*

4.14 What's usually done with Equation 4 is to absorb the D^2 into the constant:

$$V = \text{const}_3 \cdot \frac{G}{\mu} \tag{5}$$

The constant const_3 depends on the size distribution of the porous medium as well as on the particle shape and packing. It's called the ***intrinsic permeability*** of the porous medium, and it's usually denoted by small k. (It's misleading to call this quantity a constant. It's constant only for the particular porous medium we used in our home experiment! Each porous medium has its own value of intrinsic permeability.)

6. One final massaging of the relationship that started out as Equation 3 leads to another measure of permeability, called the ***hydraulic conductivity***, which is more commonly used in dealing specifically with groundwater flow. You have to make use of the concept of the ***hydraulic head***, which is *the level to which a column of water would rise if a tiny test column is inserted anywhere in the flow system* (Figure 4-10). This height h is related to the pressure p in the liquid by the hydrostatic equation,

$$p = \gamma h \tag{6}$$

where γ is the weight per unit volume of the liquid. Remember that the pressure gradient G in Equation 3 should really be written $\Delta p/\Delta x$, where x is the direction down the pipe. Substitution of the expression for p in Equation 6 into $\Delta p/\Delta x$ gives $\gamma\, \Delta h/\Delta x$, and substituting this resulting quantity into Equation 5 gives

$$\mathrm{V} = \text{const}_3 \cdot \frac{\gamma \frac{\Delta h}{\Delta x}}{\mu} \tag{7}$$

where $\text{const}_3(\gamma/\mu)$ is the hydraulic conductivity. The hydraulic conductivity is usually denoted by capital K. You can easily see for yourself that the dimensions of K are *velocity*, because $\Delta h/\Delta x$ is a ratio of length variables, and therefore dimensionless; values of K are commonly cited in meters per day.

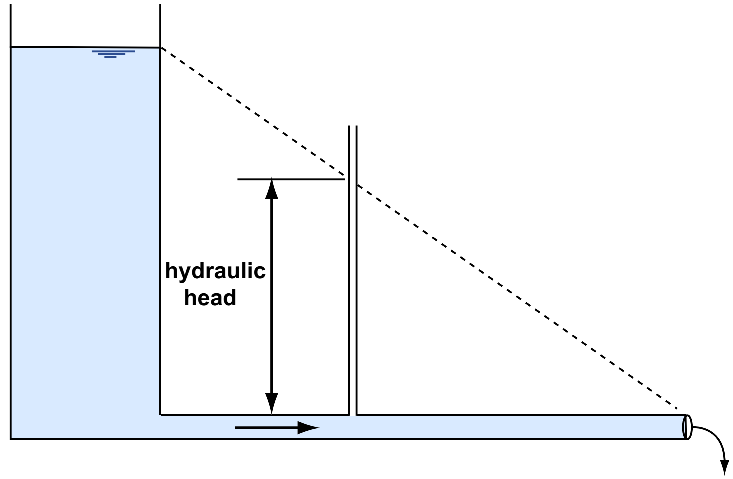

Figure by MIT OCW.

Figure 4-10. Sketch showing the hydraulic head in flow through a pipe packed with porous material.

4.10. What you would find, when you run the barrel experiment, is that the flow speed is directly proportional to the pressure gradient, and directly proportional to the size of the particles of the porous medium, and inversely proportional to the viscosity of the medium. This result is a manifestation of a well-known law in the flow of porous media called ***Darcy's law***. See the "advanced topic" above for details, if you are interested.

4.11 How the speed of flow depends on the pressure gradient, the particle size, and the fluid viscosity in this way is a reflection of a physical property of the medium that is known as the ***permeability*** of the medium. In a qualitative sense, the permeability is *a measure of how easy it is to force fluid through the porous medium by imposing a pressure gradient*. The permeability is related to the porosity (the permeability can't be very high unless there is substantial porosity), but it's not the same as the porosity. In fact, it's possible for the medium to have a high porosity but a low permeability, if the pore spaces are not sufficiently well connected; hence the concept of ***connected porosity***. The permeability of the medium is of great importance for groundwater studies, and it is also very important in the petroleum industry: you can't pump oil and gas out of deep sedimentary rock unless the both the porosity and the permeability are sufficiently great.

4.12 The home experiment on flow through a porous medium is realistic in all respects but one important one: the direction of flow is constrained to be straight down the pipe. That's not necessarily relevant to flow within a large volume of porous medium, as in the subsurface of the Earth. The critical question here is: *What determines the particular direction of groundwater flow within a large volume of a porous medium?* I can't pursue that question in any detail here, because it depends in a complex way on the dynamics of the water flow. There will be a bit more on this matter later in the chapter. If you would like to get some further insight into the patterns of flow, look into the following "advanced topic".

ADVANCED TOPIC: WHAT CONTROLS THE PATTERN OF FLOW OF A FLUID

1. Think about the pressure of the water in some large tank like the supply tank for your home experiment when the water isn't moving. Think about a little unit area at the bottom of the tank. The pressure of the water at the bottom of the tank is equal to the weight per unit area of the water in the column overlying that small unit area, times the height of the column of water above that unit area. If the weight of the water per unit volume is γ and the depth to the bottom of the tank is h, then the pressure p at bottom of the tank is just γh. And by extension of that argument, the pressure of the water at any depth *h within* the tank is also equal to γ times that depth h. This water pressure in still water is called the ***hydrostatic pressure***.

2. The hydrostatic pressure within the still water in the tank is a manifestation of a balance between the weight of the water overlying a given point, which acts vertically downward, and the pressure gradient at that point, which acts vertically upward. So although there's a gradient of pressure in the tank, the water doesn't move, because that pressure gradient is offset by an equal and opposite force, namely the weight of the water.

3. Now suppose you took your ice pick (does anybody still have an ice pick?) and poked a hole in the side of the supply tank. Water would squirt out of the hole. In the interior of the tank in the vicinity of the hole the water is now in motion toward the hole. What you've done is impose a low pressure (namely, the atmospheric pressure) on the water at the hole, just the same as at the upper surface of the water in the tank. In doing so, you have disrupted the previously hydrostatic distribution of pressure near the hole-to-be and caused non-hydrostatic pressure gradients in the water near the hole—that is, pressure gradients that are no longer balanced by the weight of the water. That's the mechanical reason why the water flows from the tank!

4. The exact distribution of pressure in the vicinity of the hole and the resulting patterns of water motion are much too complicated for us to deal with here, but this example suggests that *the direction of water movement at any point is in the direction in which the difference between total pressure and hydrostatic pressure (a quantity called the **dynamic pressure**) decreases most rapidly*. It's differences in this dynamic pressure that cause fluid in any situation to move. So you can be confident that, whenever you are dealing with groundwater flow, the flow will always be in the direction of most rapid decrease in dynamic pressure. Figure 4-11 shows qualitatively the distributions of total pressure, hydrostatic pressure, and dynamic pressure in the tank once you've punched the hole and the water is flowing out. Note how *the flow lines are everywhere normal to the contours of dynamic pressure*.

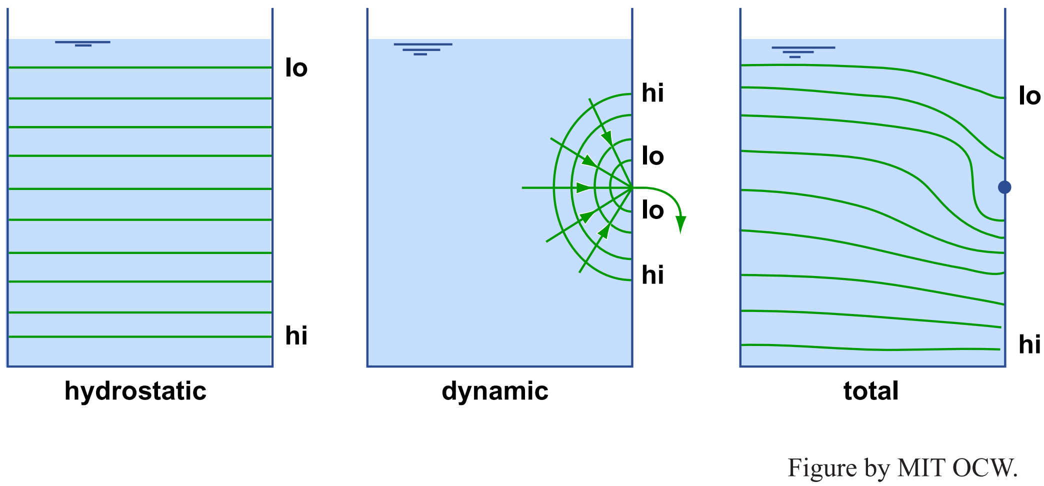

Figure 4-11. Distributions of hydrostatic pressure, dynamic pressure, and total pressure in outflow through a small hole in the wall of a barrel (qualitative).

5. Of course, the reasons for the distribution of dynamic pressure are always very complicated, and really beyond the scope of this course. Prediction of the spatial distribution of the dynamic pressure, and therefore the spatial pattern and speeds of water movement in the porous medium, are one of the major topics in the study of groundwater hydraulics. I'm not doing anything more than giving you the barest flavor of this endeavor.

4.13 I don't know what your intuition tells you about representative speeds of flow in your home experiment as a function of the nature of the porous medium, but Table 4-1 gives some representative values for various common kinds of porous media within the Earth. The values range enormously from coarse gravel, in which speeds are of the order of centimeters per second, to solid rock (which in reality is porous because of tiny spaces at grain boundaries and other miscellaneous rock fractures), in which speeds are of the order of a thousandth of a millimeter per second. Table 1 also gives corresponding values of hydraulic conductivity K, discussed in Paragraph 4.14.

Material	Typical values of K (m/day)
gravel	5000
coarse sand	50
fine sand	5
silt	0.1
clay	0.0002
sandstone	1
"hard rock"	0.1

Figure by MIT OCW.

Table 4-1. Representative values of hydraulic conductivity K for various kinds of porous media.

4.14 Figure 4-12 shows a practical application of the principles we dealt with on the basis of the home experiment. There's a sloping land surface in which the deeper bedrock is covered by a fairly uniform but perhaps rather thick layer of loose and much more permeable material. High up on the slope is a source of pollutants, and farther down the slope, ten meters let's say, is your home or summer place, where you might have a water well. The groundwater flow is directly down the slope through the porous surficial layer at some speed that depends on the intrinsic permeability of the porous medium. If you know the slope of the ground and the intrinsic permeability of the material, you can compute the travel time of a pollutant tracer from the input point to the water well beneath your house. Assume for the sake of discussion that the slope of the ground is one in ten. The gradient in hydraulic head, $\Delta h/\Delta x$, within the porous medium is then 0.1. Using Equation 7 we find that V, the characteristic velocity of the groundwater, is $0.1K$. Using the representative values for K given in Table 4-1 you can obtain travel times for various kinds of porous medium (Table 4-2). You can see that depending on the permeability of the medium the grace period between the time of introduction of the pollutant and the time it pollutes your water well varies enormously.

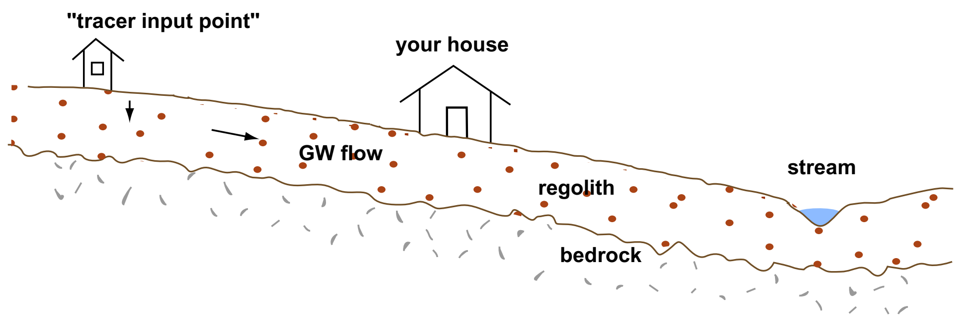

Figure by MIT OCW.

Figure 4-12. A tracer experiment in groundwater flow.

Material	Travel Times, hr, 10m down 0.1 slope
gravel	0.5
coarse sand	50
fine sand	500
silt	240×10^4
clay	$120{,}000 \times 10^4$
sandstone	24×10^4
"hard rock"	240×10^4

Figure by MIT OCW.

Table 4-2. Ten-meter travel times for the same kinds of porous media as in Table 4-1.

5. A HOME EXPERIMENT ON GROUNDWATER FLOW

5.1 Now it's time to do a more realistic and ambitious experiment on the flow of groundwater (Figure 4-13). For this purpose you will need to build a very large shallow square tank in your backyard or use an entire spare room in your home. If you decide to use the spare room, you had better shore up the floor with some big timbers, because otherwise it's likely to collapse.

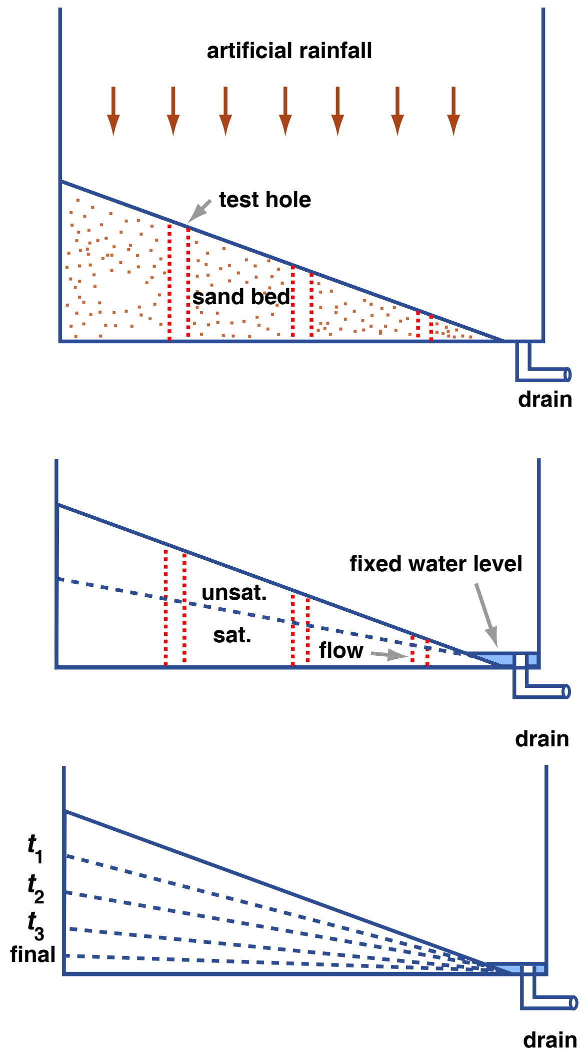

Figure by MIT OCW.

Figure 4-13. Another home experiment on groundwater flow.

5.2 Line the tank or the room with a large new polyethylene tarpaulin tucked in neatly at the corners, and fill the space with something like a meter of sand. Taper the layer of sand so that it has maximum depth at one side and zero depth at the other side, and along that latter side provide a drain or a sump pump at one corner. Put in several test holes or wells across the space, and line the holes with window-screen cylinders so that the sand doesn't fall into the holes. Now connect the hose to your kitchen sink and spray the sand surface to simulate a brief and heavy rainstorm.

5.3 Here's a summary of the results you would obtain:

- There will be water in the drain, and it will flow long after the rain stops.
- There will be water in all the test holes.
- The profile connecting the water level in the test holes will show the same sense of slope as the surface of the sand; this defines the groundwater table. The ***water table***, also called the ***groundwater table***, is *the locus of points where the water pressure is equal to the atmospheric pressure*. It's the top of the permanently saturated zone.
- The slope of the groundwater table is less than the slope of the sand surface, and it decreases with time (see the lowest part of Figure 4-13).
- The movement of groundwater is in the downslope direction, toward the drain. You can tell this by injecting food coloring in the uppermost test hole and seeing it appear in successive holes and finally at the surface, at the drain.
- If the rainfall is too heavy there will be some surface runoff down to the drain, but even in this case most of the water will infiltrate and become part of the groundwater flow.

5.4 This is a very realistic experiment in groundwater flow. The only problems with it are these:

- The scale is too small, and things happen too fast.
- The material is uniform and has a fixed and permeable floor. In real life there's usually a gradual downward decrease in both porosity and permeability—although in glaciated areas the structure of the subsurface is often just the same as in this experiment.

6. QUALITATIVE ASPECTS OF GROUNDWATER AND GROUNDWATER FLOW

6.1 Figure 4-14 shows a vertical cross section through a representative area of the Earth's surface, showing some of the important features of distribution of groundwater. The part of the subsurface lying *above* the groundwater table is called the ***vadose zone*** or ***aerated zone*** or ***unsaturated zone***. Most of the time, the pore spaces in the vadose zone are occupied mostly by air (plus locally generated gases); only right after a heavy rain are the pore spaces filled with downward-percolating water. The part of the subsurface zone lying *below* the groundwater table is called the ***phreatic zone*** or the ***saturated zone***. In the phreatic zone the pore spaces are always filled with water.

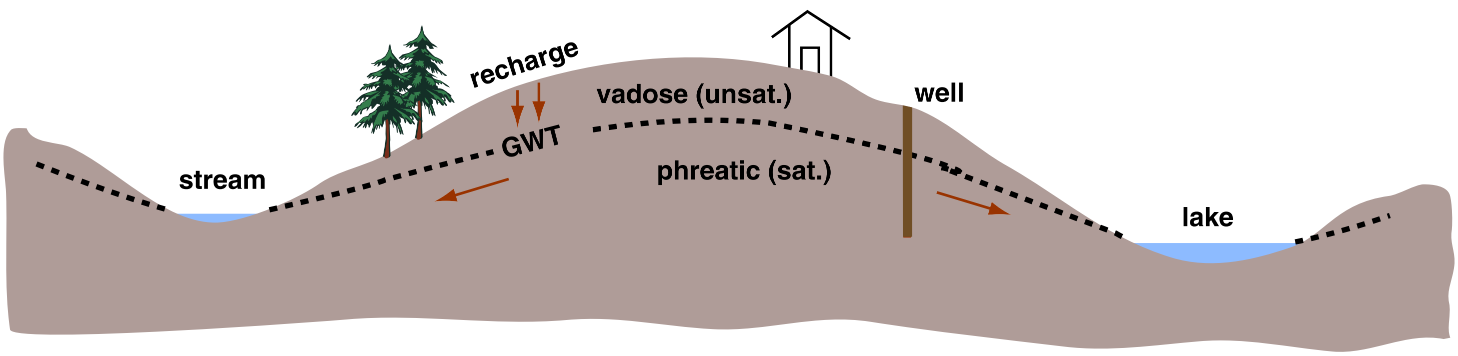

Figure by MIT OCW.

Figure 4-14. Vertical cross section through a representative hill-and-valley area of the land surface, showing features of distribution of groundwater.

6.2 In reality the interface between the vadose zone and the phreatic zone (that is, the groundwater table) is not a sharp and well-defined surface: it's a fuzzy zone of transition. That's because of capillary rise of water up into the pore spaces of the lowermost part of the phreatic zone. (See the background section on surface tension and capillarity earlier in the chapter.) The zone of partial saturation is called the ***capillary fringe***. Its thickness ranges from just a centimeter or two, in coarse gravels, to as much as a couple of meters, in fine silt-rich sediments. The reason that there is a gradation in the degree of saturation is that, owing to the spread of particle sizes, adjacent pore spaces vary widely in their effective size.

6.2 The topography of the groundwater table mimics the topography of the land surface itself: it's high under hills and low under valleys. But the topography of the groundwater table is more subdued than that of the land surface, because the depth to the groundwater table is greatest under hills and least under valleys. In fact, *the groundwater table intersects the land surface at rivers and lakes and springs*. This should not be surprising to you if you think of rivers and lakes as just the places where the permanently saturated zone emerges from beneath the mantle of porous medium!

6.3 There's no necessary relationship between the groundwater table (the surface of contact between the vadose zone and the phreatic zone), on the one hand, and the contact between bedrock and regolith, on the other hand. In areas with a relatively thin mantle of regolith and a relatively deep water table, the water table lies mostly within bedrock (Figure 4-15A). In areas with a relatively thick mantle of regolith and a relatively shallow water table, however, the water table lies mostly within the regolith (Figure 4-15B). In New England, both extremes are common, basically because of the highly variable depth of regolith resulting from glacial erosion, transportation, and deposition of regolith during the last Ice Age.

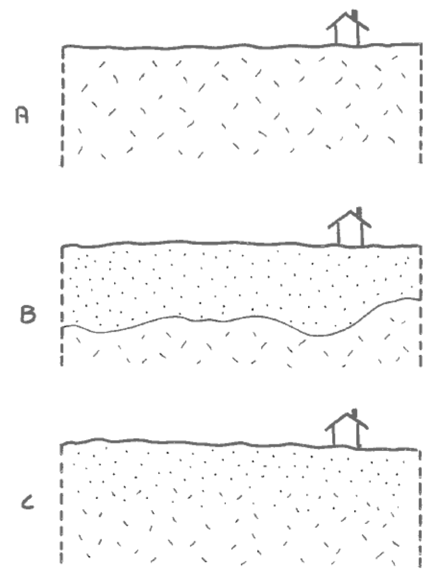

Figure 4-15. Groundwater table vs. position of regolith–bedrock contact.

6.4 Figure 4-16 shows a record of the level of the water table in a well located as shown in Figure 4-14 during a representative year. In this area, the water table fluctuates vertically by about ten feet in the course of the year. Fluctuations might be considerably smaller or larger than that, depending partly on the *variability of rainfall* but also very importantly on the *permeability of the subsurface material*. Obviously, the water table tends to be high during spring and fall rainy periods and low during summer droughts. But the water table tends to be *low during the winter* also, because the ground at the surface freezes to a depth of a meter or even more, preventing, or at least impeding, recharge, while the groundwater at greater depths continues to flow and thus lower the water table. (The term ***recharge*** is used to describe *replenishment of groundwater in a* *subsurface region from which groundwater was previously withdrawn*, either naturally or by human activities.) Another noteworthy thing about the record in Figure 4-16 is that *the water table rises faster than it falls*. This is because recharge involves percolation vertically downward through a relatively thin layer of porous material, whereas drainage involves slower groundwater movement through long distances down low gradients of hydraulic head.

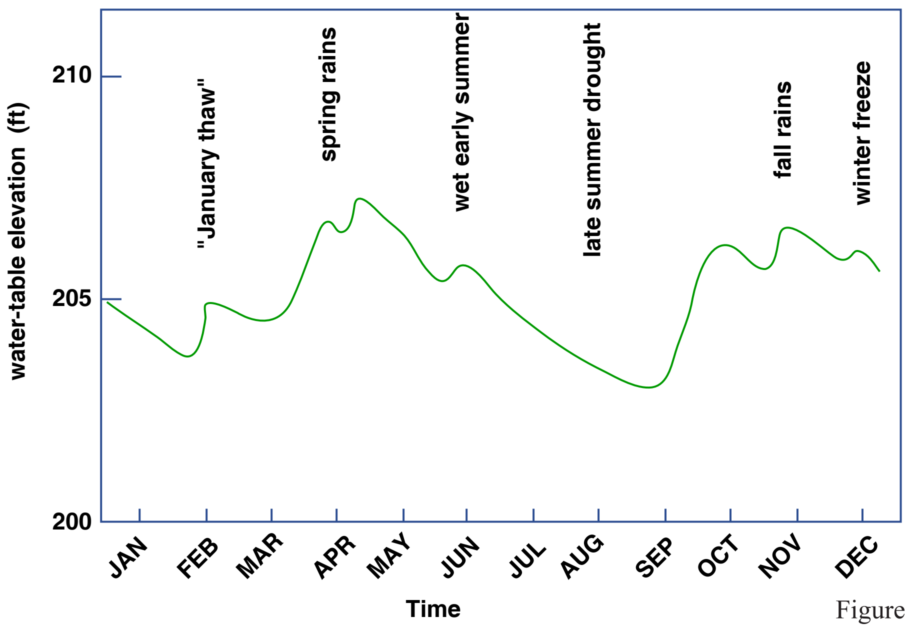

Figure by MIT OCW.

Figure 4-16. Representative record of the level of the groundwater table in a well.

6.5 Here's an incidental note on the depth of water wells. Usually a well is drilled far below the level of the local water table, for two reasons:

- to ensure that the bottom of the well stays below the water table even during severe droughts;
- to get below the level of polluted near-surface waters (but the problem is that even the deeper waters don't stay unpolluted forever!).

6.6 The prices one pays for deeper wells (aside from the dollar cost per foot of well) are:

- generally, the deeper the well the smaller the porosity and permeability of the medium, and so the lower the rate of flow into the well;

- The slower the flow into the well, the longer the residence time of the groundwater in the porous medium, so the longer the time available for uptake of ions from the medium, so the harder the water.

6.7 An ***aquifer*** is *any body of porous material in the Earth (rock or regolith) with sufficient volume and connected porosity to yield appreciable water to wells or springs*. Aquifers contain relatively large drainable porosity, relatively large volume, and relatively high permeability. The concept of an aquifer is a loose one, partly because of those sneaky words *appreciable* and *relatively*: The term is used for any volume of subsurface material that's a good producer, or a potentially good producer, of water, relative to surrounding volumes of subsurface material. The minimum lateral dimensions of what are called aquifers might be as little as tens of meters, and the minimum vertical dimensions might be as little as some meters. Large aquifers, however, might have lateral dimensions of hundreds of kilometers and vertical dimensions of many tens of meters.

6.8 The opposite concept is that of an ***aquiclude***: *any body of subsurface material through which water can move at only negligible rates*, or at least at rates much smaller than through adjacent aquifers. Also, the term ***aquitard*** is used for any body of subsurface material through which groundwater travels slowly, relative to some adjacent aquifer, but not so slowly as to be negligible, as in an aquiclude.

6.9 Aquifers can be classified into four kinds (Figure 4-17):

unconfined aquifer: an aquifer in which the groundwater is in direct contact with the overlying atmosphere through connected pore space.

confined aquifer: an aquifer overlain by an aquiclude so that it is not in contact with the atmosphere except in some upstream area of recharge.

leaky aquifer: a confined aquifer whose overlying aquiclude allows some non-negligible passage of groundwater into or out of the aquifer.

perched aquifer: an unconfined aquifer present above a shallow and laterally restricted aquiclude.

6.10 Aquifers and aquicludes can exist on regional as well as local scales. Figure 4-18 shows a cross section through such a system. A highly porous and permeable sedimentary formation, like a well sorted and poorly cemented sandstone, is overlain by a highly impermeable shale in a large region with very low angles of dip of the formations. Wells may tap the confined aquifer at distances of hundreds of kilometers from the recharge area, where the aquifer formation is exposed at the surface. Note that there's also a much shallower unconfined aquifer that derives its water much more locally.

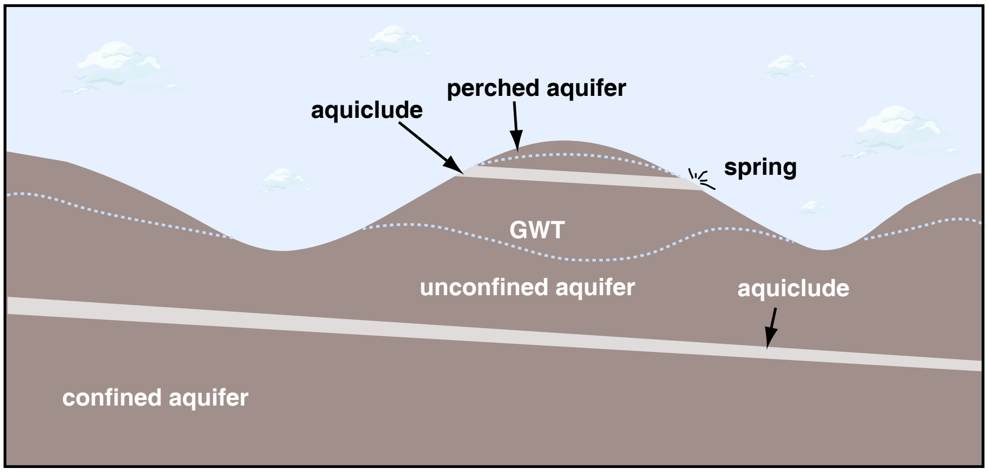

Figure by MIT OCW.

Figure 4-17. Schematic vertical cross section showing various kinds of aquifers.

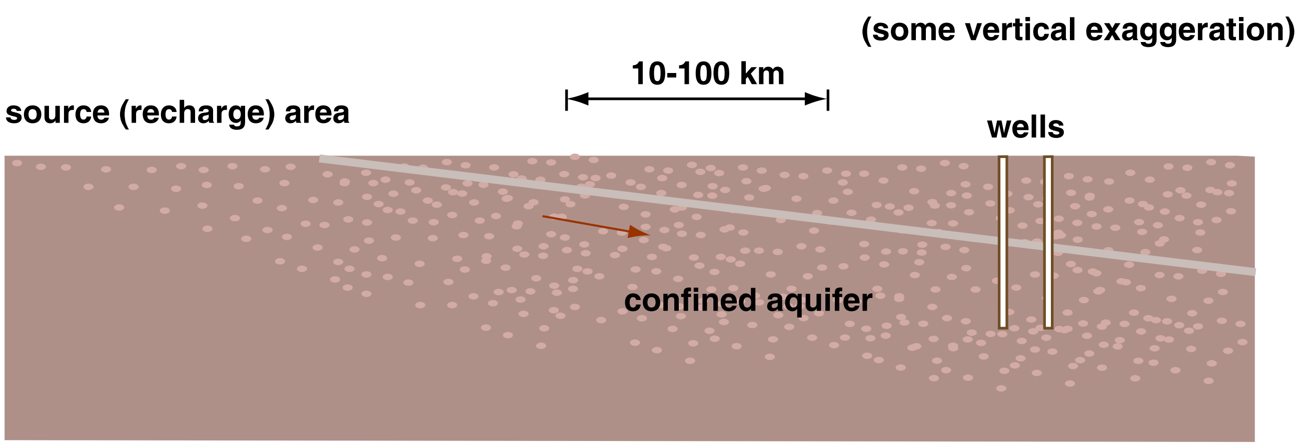

Figure by MIT OCW.

Figure 4-18. Schematic vertical cross section showing regional extent of aquifers and aquicludes.

6.11 If the surface slope of the region is greater than the slope of the surface of hydraulic head associated with the regional confined aquifer, a well drilled to the confined aquifer will produce *a water flow at the ground surface*, with no need for pumping (Figure 4-19). Such a well is called an ***artesian well***. Figure 4-20 shows a homey analogy in which you water your garden with a hose leading from an elevated tank of water. In this situation, the nozzle of the hose is a kind of artesian well. And, in a broad sense, if you live in an area with a central municipal water supply all of the faucets in your home or apartment are artesian wells!

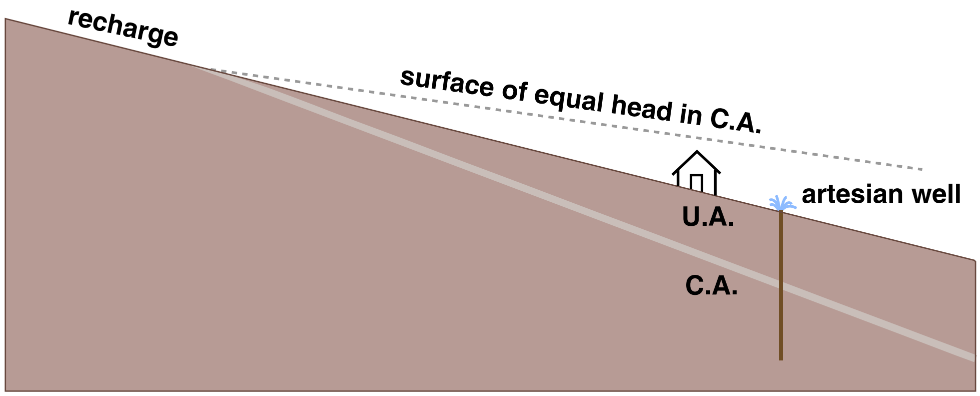

Figure by MIT OCW.

Figure 4-19. Schematic vertical cross section showing the nature of artesian wells.

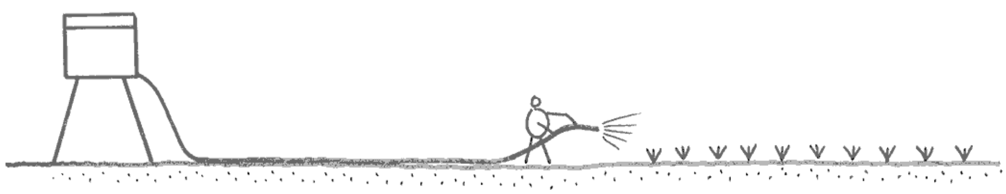

Figure 4-20. A home analogy for artesian wells.

6.12 Here's one more concept that's useful in dealing with groundwater supplies: The ***specific yield*** of an aquifer is *the ratio of the volume of water that drains out of the aquifer (when the groundwater table is lowered) to the total volume of the aquifer subjected to drainage*. The specific yield obviously depends on the *porosity* of the aquifer, but it also depends on the *size of the pore spaces*, because some water always adheres to the surfaces of the solid materials of the aquifer, both as thin films on surfaces and as fillets in reentrants, and the smaller the size of the pore spaces, the greater the percentage of the total porosity that remains occupied by this adhering water. This concept of specific yield is analogous to that of the field capacity of a soil.

6.13 I haven't said anything yet about the patterns of flow within a typical aquifer. The complexities lie in flow through deep unconfined aquifers, because shallow unconfined aquifers (as in Figure 4-12) or confined aquifers (as in Figure 4-18) can be treated approximately as flow in rivers or closed ducts. The patterns of flow in a porous medium are a response to the spatial distribution of gradients of dynamic pressure. No simple statement about the patterns of motion can be made without setting up the geometry of the problem and solving the equation (essentially Newton's second law written in the appropriate form) that governs the motion.

6.14 Figure 4-21 shows qualitatively the flow patterns in a typical deep unconfined aquifer in an area of irregular surface topography. Just below the water table in areas where the water table is sloping, the flow is as you might expect: downward and parallel to the water table. But near the crest of the water table the flow is nearly vertically downward, and directly beneath where the water table merges with rivers and lakes the flow is nearly vertically upward. At points deep within the aquifer, the directions of flow bear no obvious resemblance to the surface topography of the water table.

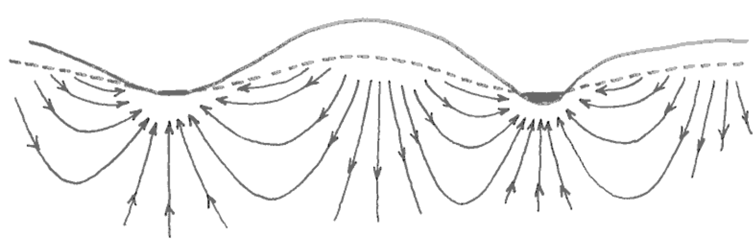

Figure 4-21. Qualitative flow patterns in deep groundwater.

7. SOME PRACTICAL THINGS ABOUT WATER WELLS

7.1 A great many households in rural or semi-rural areas depend upon wells for their water supply. In earlier times, shallow wells were dug by hand to depths of a few tens of feet. In more recent times, wells have been driven by various techniques to much greater depths of many hundreds of feet to tap deeper groundwater. Most such driven wells are lined with a jacket or casing that's porous in the lower part of the well. This casing keeps the walls from collapsing or crumbling inward, without inhibiting the water flow. This section presents some aspects of the flow of water in the immediate vicinity of wells.

7.2 The first thing to keep in mind is that *the removal of water through a well disrupts the normal or background flow of groundwater in the vicinity by producing a locally lower water pressure in the lower part of the well.* The relatively strong pressure gradients near the well cause a flow of water radially inward toward the well. The difference in elevation between the depressed water level in the operating well and the elevation the water table would have if water were not being drawn from the well is called the ***drawdown*** of the well (Figure 4-22). The locally cone-shaped surface of the water table in the vicinity of the well is called the ***cone of depression***. In areas with closely spaced wells, the cones of depression of the wells may interfere with one another to produce a general lowering of the water table in the area (Figure 4-23).

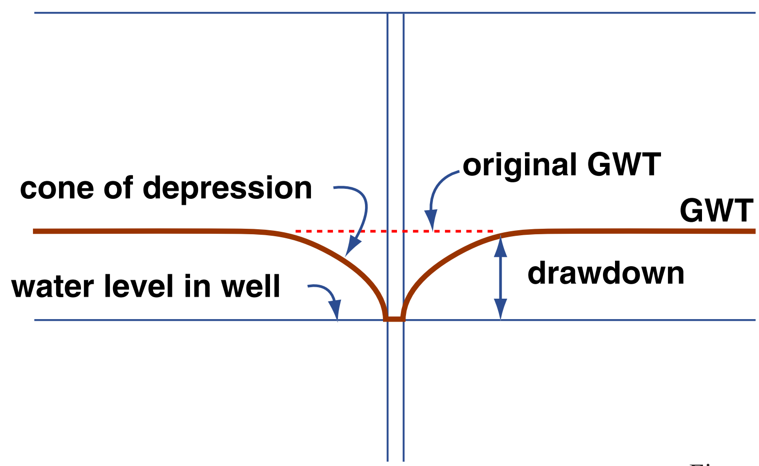

Figure by MIT OCW.

Figure 4-22. Geometry of the groundwater table around an active water well.

7.3 Your well becomes polluted from an upstream pollution source only if its intake lies in the path taken by the pollutant. A pollutant introduced at a point source reaches the groundwater table by downward percolation and then travels with the groundwater flow as a plume, narrow at first but gradually widening because of slow lateral diffusion (Figure 4-24). (A ***plume*** is a mass of fluid that moves through ambient fluid, as a consequence of a difference in density or just because that fluid is injected, over some period of time, into the flowing ambient fluid at a point, as is the case here. A good example of a plume of the former kind is smoke that rises out of a chimney or a smokestack.)

7.4 A well upslope of the point of pollutant injection can tap a pollutant plume, if the distortion of the natural flow pattern is so great as to produce *locally upslope flow toward the well.* The situation shown in Figure 4-25, in which all of the pollutant plume is drawn up by the well, is an extreme example.

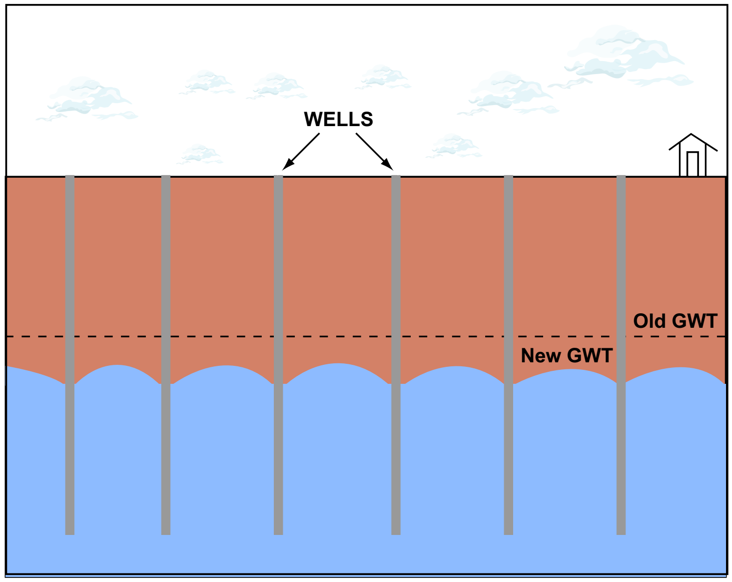

Figure by MIT OCW.

Figure 4-23. Geometry of the groundwater table in the vicinity of a number of closely spaced active water wells.

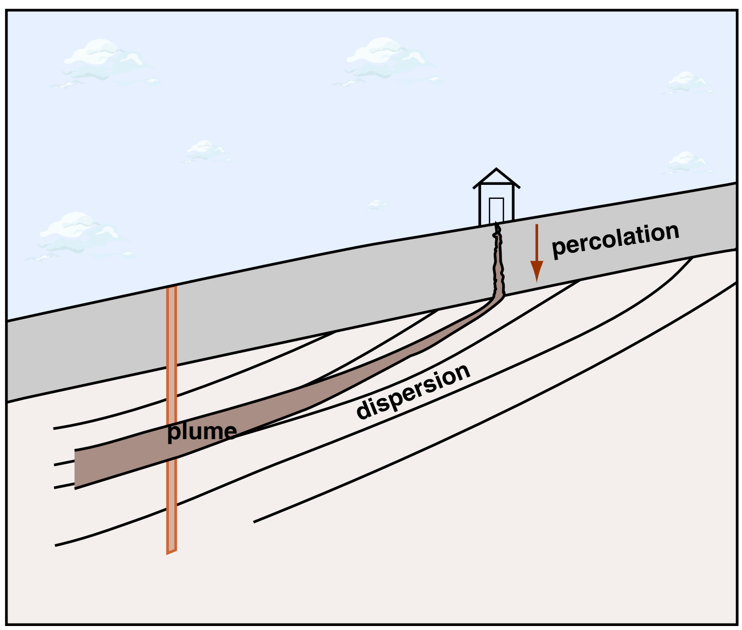

Figure by MIT OCW.

Figure 4-24. A well lying in the path of a pollutant plume.

7.5 If in the long term one or more wells in a given area remove water faster than it can be supplied by recharge by percolation from surface rainfall, the water table is lowered. In areas with shallow wells and abundant recharge, where natural flow rates are high, the water table responds rapidly, in hours or days. But in areas with deep wells and slow recharge, the depressed water table might take many years or even decades to readjust, even if no water at all is drawn from the wells. In a very real sense, groundwater is being *mined* by these wells, because recharge is on a time scale much longer than the lifetime of the wells. This is the case today in many areas of the arid and semiarid central and western parts of the United States.

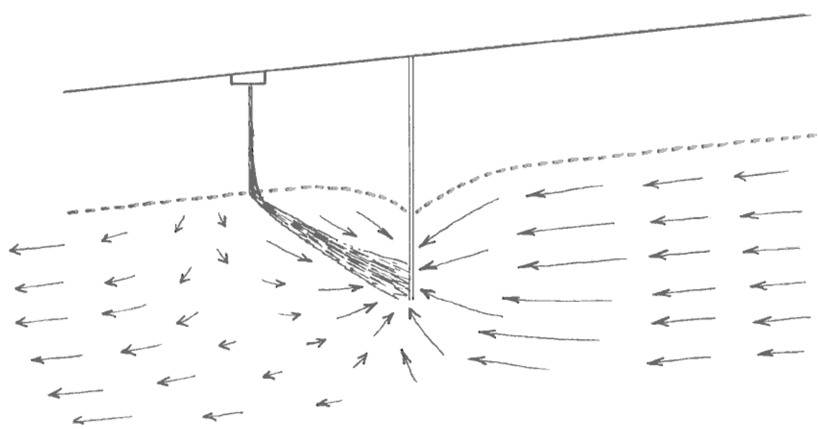

Figure 4-25. Locally upslope movement of a pollutant plume toward an active well.

8. GROUNDWATER IN COASTAL REGIONS

8.1 So far we have been concerned with groundwater in inland areas well away from the seacoast. In coastal areas, *what's the interaction of groundwater with the sea?* You know that fresh water is less dense than salt water, by a small but important few percent. So where groundwater is in contact with seawater, the groundwater should tend to float upon, or rise through, the seawater.

8.2 Perhaps you can get the clearest picture by considering a small, porous, rainy island in the ocean (Figure 4-26). (Nantucket and Martha's Vineyard are good examples—although Chamber of Commerce members probably would not appreciate that description.) Assume that the seabed all around the island is saturated with seawater. Provided that the seabed consists of porous material, this is an excellent assumption. Then the body of fresh water that constitutes the groundwater under the island is itself floored by an even lower layer of denser seawater. The sharpness of the boundary between the groundwater and the seawater is determined by the relative importance of groundwater flow, which maintains the sharpness of the boundary, and diffusion of the salt from the seawater into the groundwater, which tends to blur the boundary. The boundary emerges at the seabed some distance from shore, and landward from that line groundwater flows out to the seabed and then rises and mixes with the water of the ocean.

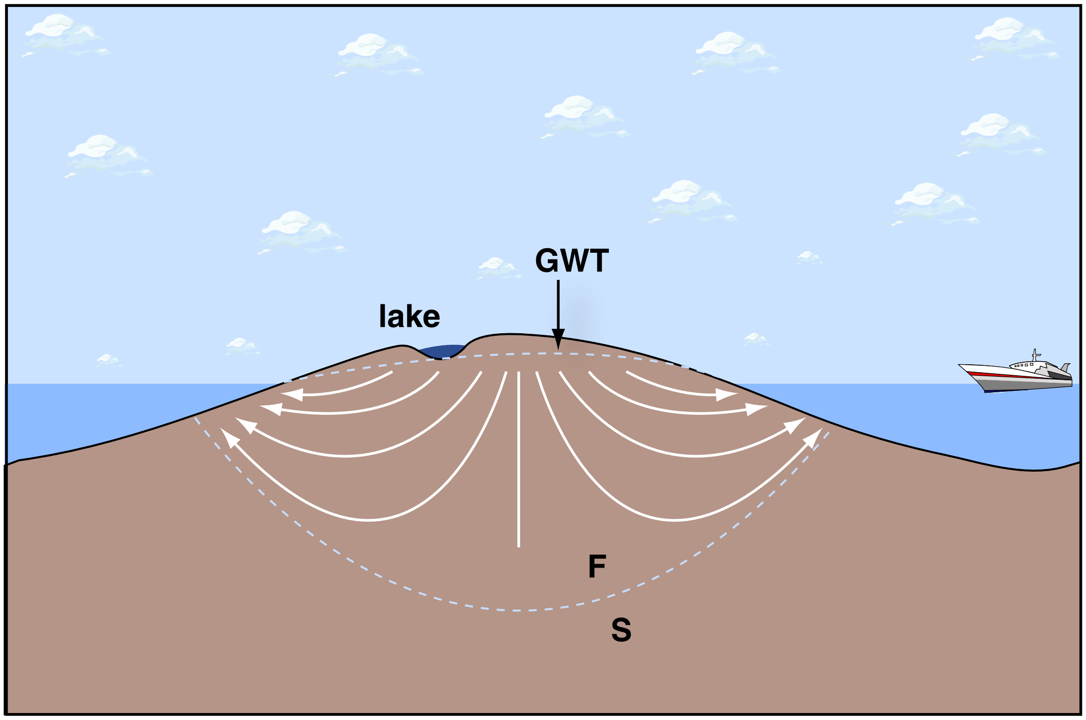

Figure by MIT OCW.

Figure 4-26. Pattern of groundwater flow in a coastal region.

8.3 How deep is the boundary between groundwater and seawater under the island? That depends on the elevation of the groundwater table above sea level within the island. *The depth of the boundary below sea level is just about 40 times the height of the groundwater table above sea level*—because the ratio of sea-water density to fresh-water density is about 41/40. It's a simple problem of buoyancy. If you are confused, think about floating a balloon full of fresh water in a tub of seawater. The ratio of submergent depth to emergent height of the balloon is governed by exactly the same principle.

8.4 Now suppose you drilled a well somewhere on the island. At first you are tapping only fresh water. But now you have to worry about two cones: not just the cone of depression that affects the groundwater table, but also an inverted cone that represents what might be called the "drawup" of the bounding surface between the groundwater and the seawater (Figure 4-27). If you pump fresh water out of the well so vigorously that the boundary is drawn all the way up to the base of the well, you start to pump seawater as well as fresh water from the well. This effect is called ***salt-water intrusion***. Salt-water intrusions of this kind are common in coastal areas underlain by porous materials, as on Cape Cod and Long Island. The problem is reversible, but there is only one remedy: *reduce the rate of pumping*.

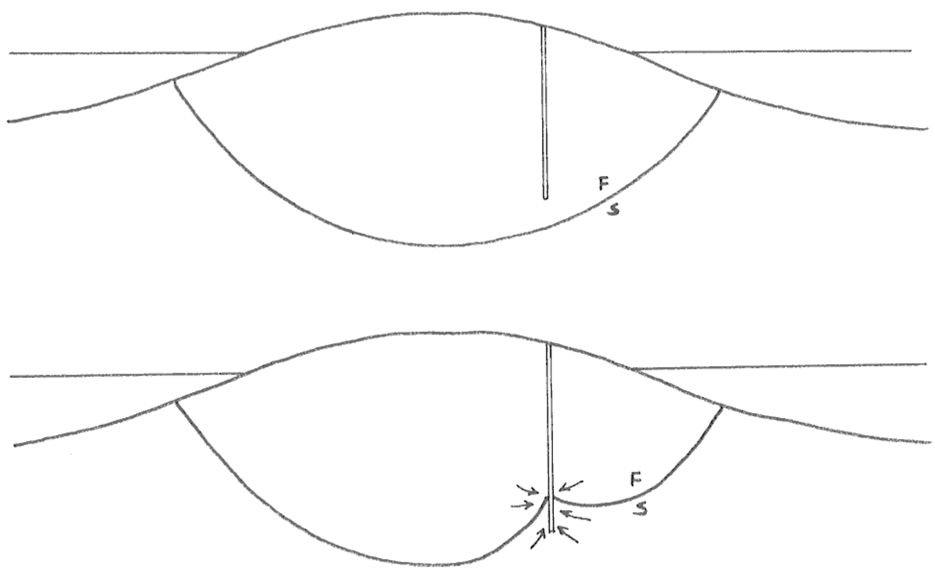

Figure 4-27. Effect of pumping on the position of the fresh–salt boundary beneath a coastal region.

READINGS

Easterbrook, D.J., 1999, Surface Processes and Landforms, Second Edition. Prentice Hall, 546 p. (Chapter 7)