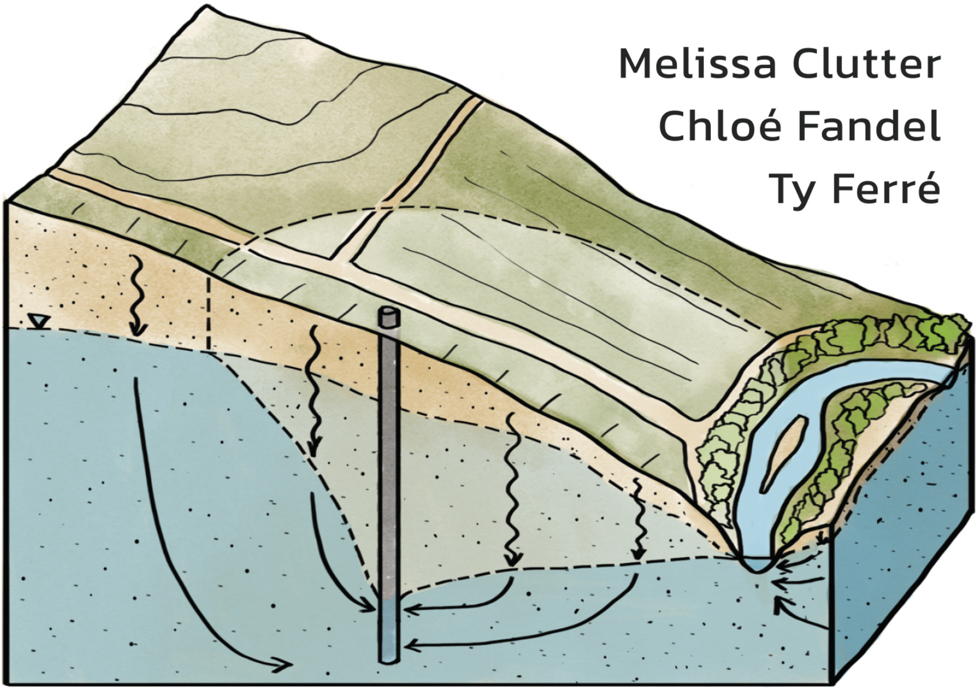
EARTH SCIENCES IN THE 21ST CENTURY

The Basics of GROUNDWATER



Melissa Clutter
Chloé Fandel
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Preface

Hydrogeology is a branch of geology that studies water, both underground and on the surface of the Earth. The goal of this book is to help you understand hydrogeologic systems on a practical level. If you end up working in hydrogeology or a related field, there is a good chance that you will be asked to predict the outcome of a water-related scenario. Will contaminated groundwater reach my water supply? Does the pumping from a well at my neighbor's house have the potential to affect my water availability? Where are the dissolved compounds in my drinking water coming from? At what rate can the city pump water to minimize impacts on local surface waters? These are all important questions that require specialized knowledge. Through this book, we want to introduce you to concepts that can help to answer these questions.

Whenever possible we will use realistic examples to explain concepts. Before diving into the big picture, there are basic building blocks to introduce. So, hold on tight through the first few chapters and keep in mind that we will eventually reach real-world scenarios. Each chapter will follow a similar path – we set the scene, introduce fundamental concepts, and then ultimately revisit the scene considering your new understanding. We hope that you enjoy reading this book as much as we enjoyed writing it.

Acknowledgments

This book was inspired by the need to improve pedagogy in introductory hydrogeology classrooms. Although we had a vision for communicating introductory-level groundwater concepts, the story was made stronger by our two editors Dr. Gigi Richard and Dr. Robert Krantz. We are grateful for their contributions.

Chapter 1

Soil Properties

Introduction

"Win a prize! How many marbles are in this jar?" This is a common game at carnivals, fundraisers, and school events. The host provides a glass container filled with marbles or M&Ms, and you must guess how many are in the container. What makes this game so difficult? And why do the hosts almost always choose small objects like M&Ms or marbles rather than filling the jar with baseballs? Do you think the game would be easier if the jar *were* filled with baseballs? Would the game be more difficult if you had M&Ms and marbles mixed together? How does the stacking of the objects make the game more or less difficult? These concepts and questions are not specific to fundraisers – they are also concepts that a hydrogeologist thinks about every day. How do solids fill a space? And what complexities arise as the shape, size, and mixture of the solids change? Throughout this chapter we will explore these topics and learn ways to quantify (and characterize) the space that is left between the solids. This chapter will not only lay a foundation for the rest of your hydrogeologic journey, but it could also help you to win a prize at the next fundraiser or school event that you attend!

Material, Soil, and Porous Media

For the "How many marbles are in this jar?" game described above, you always have some sort of material (e.g., marbles, M&Ms, baseballs, etc.) inside the jar. How the material is stacked and packed into a jar determines the amount of material that can fit in the space. Furthermore, the properties of the material determine the stacking and packing. Within hydrogeology, we are not really dealing with jars of marbles, but we are interested in materials such as soil particles and rock fragments. For example, how do the grains of a soil stack together, and what properties determine the configuration of the soil grains?

So, what do we exactly mean by *soil*? Imagine that you step outside and look down. It doesn't matter where you are – if you look beneath the surface (pavement, rocks, grass), you are almost certain to see soil. Now, some of you may object to this term ("soil") – and you would be right. To be soil, the material must have the specific requirements needed to grow plants. Therefore, the more correct term is *porous medium*. But we personally find this term rather soulless. Due to the more relaxing tone of the word "soil", we will often use it in this textbook to describe porous media.

When you look at the soil up close, you will notice that it is made up of small particles or grains. You might also see micro-organisms, plant roots, and bits of dead organic matter. All these properties will change how the particles stack together and fill a space. What's more, the characteristics of the grains and the biological components depend on where you are. The grains will be different in the city or the country, in the mountains or in the desert. Each soil has a unique history and unique properties.

Any description that you can offer for your soil is a physical property – color, size of the grains, shape of the grains, how tightly packed together the grains are. As scientists, we aim to develop terminology to describe these properties as exactly and universally as possible. The goal is that when you describe your soil, another scientist could understand everything that they need to know about the soil from your description.

Grain Size, Shape, and Sorting

The difficulty of the marble game discussed above varies based on the size of the objects in the jar. For example, we briefly mentioned how the game might be easier with baseballs. If we were to ask you why the game would be easier, you would likely say, because the baseballs are larger. You could even compare the diameter of the baseball to the diameter of a marble. In the same way that we can describe the size of these objects, we can describe soil grains. We describe and characterize the size of particles in a soil using the *grain size*. If the grain is a perfect sphere, we measure the grain size just like we measure the diameter of a baseball. But what if it isn't? The grain size would depend on which side we measure. Practically speaking, the size of a single grain is often referred to by its s*mallest side*. Sometimes we may want to know the precise size of a grain. But usually, it is good enough to know the approximate size of the grain. We do this by deciding if a grain falls in a range of sizes, say 0.25 to 0.50 mm (Figure 1). We refer to each bracket of sizes by a specific name. For example, grains between 0.25 and 0.50 mm are called medium sand (Figure 1).

With some practice, you can get good at classifying most soil grains by eye. But chances are that every grain in a handful of soil will have a different size. In this case, what do you do? As a scientist, you must describe it as well as you can. Sometimes the variation in size is small. For example, every grain might have a different size, but they may all be classified as medium sand (0.25-0.50 mm). Think of a jar of Skittles or M&Ms; the sizes of the objects are close enough to the same, that you can consider them to be in the same class. We call soils with grains of all the same size well sorted soils; the *sorting* of a soil describes the distribution of grain sizes. In other cases, the grains might fall into different, even quite different, classes; a poorly sorted soil has grains of assorted sizes (Figure 1). To use equivalent terminology for your guessing game, a jar of only marbles (all the same size) would be a well sorted sample (Figure 1). The opposite would be a poorly sorted jar, or a jar filled with different sized objects (e.g., M&Ms, golf balls, and marbles).

Thus far we have talked about the marble game with rounded objects. But what if we had a jar of square blocks rather than rounded golf balls? How would that change the stacking in the jar? Or better yet, what if we had a jar of tortilla chips? Soil particles aren't (usually) perfectly rounded. We mentioned earlier that every soil has a history. That means every grain has a history, too. If a grain was recently broken off a larger rock, it may still have sharp edges. But, if the grain is on a beach and it has been rolled in the tide for years, it is likely to be exceptionally smooth. We describe the sharpness of the edges of a grain as its *angularity*. As with everything in science, there are highly mathematical ways to define angularity. But for us, it is enough to use a few general terms that range from angular to well-rounded (Figure 1).

We have introduced soil properties such as grain size, sorting, and shape (e.g., angularity). Now, begin to think about how these properties could affect a soil sample. How does changing the object size, sorting, and shape affects your guessing game? What exactly changes within the jar when you have tortilla chips versus rounded marbles? M&Ms versus golf balls? Specifically, think about how these changes affect the stacking of the objects and the amount of empty space between the solid objects. Next, imagine that you poured water in the jar. Where would the water go? Between the solid objects, right? Not only that, but the total space available for water would be limited to the vacant space. Water within a soil works the same way. Properties of soil are important for many reasons, one of which is that they help us understand where water exists and how it moves underground.

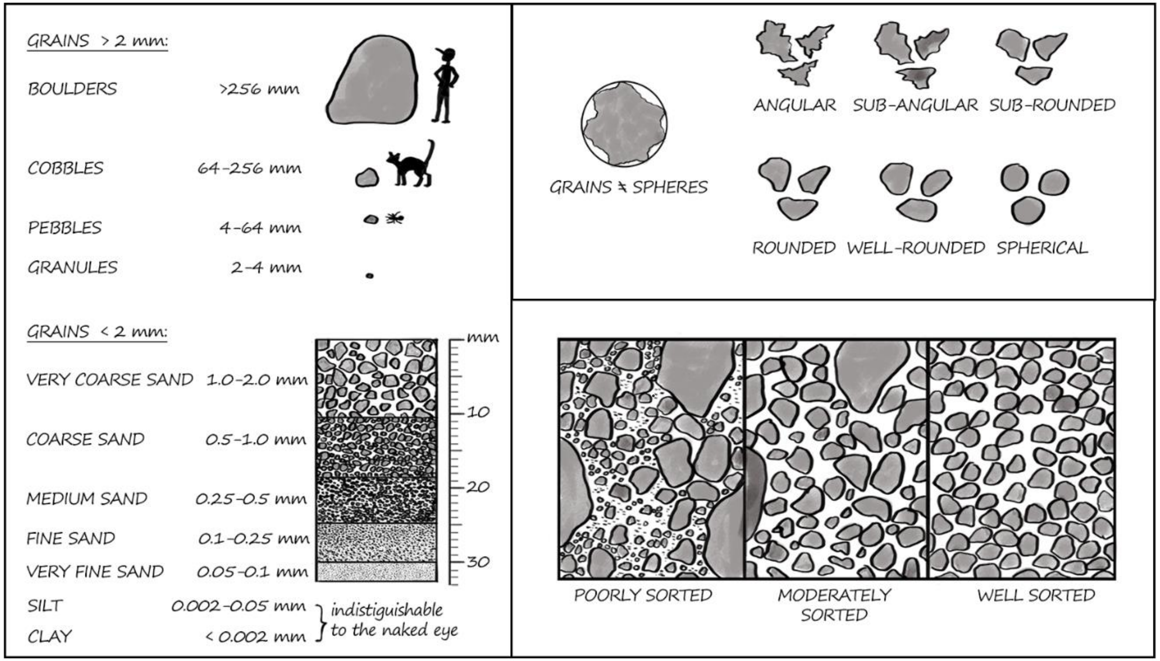

Figure 1. Examples and characterization for the various shapes, sizes, and sorting of grains in a sample material. Adapted from Forestry Suppliers, Inc.

Soil Classification

You have a few terms for describing soils (e.g., sorting, shape, grain size). However, it can be a bit of a mouthful to say that the soil beneath your feet is a moderately sorted, sub-angular to rounded mixture of medium sand, very coarse sand, and granules (for example). So, pedologists (scientists who study soil) use names for soils and the properties of the grains determine the *soil texture* (or soil name). One popular method for defining texture is the *soil texture triangle* (Figure 2). The soil texture triangle uses the relative percentages of different grain sizes – sand (>0.05 mm), silt (0.002-0.05 mm), and clay (< 0.002 mm) to determine the texture. You can plot any soil on the texture triangle based on the size percentages. More on how to read this triangle to come, but first, let's discuss how you decide the relative percentages of each grain size in a soil sample.

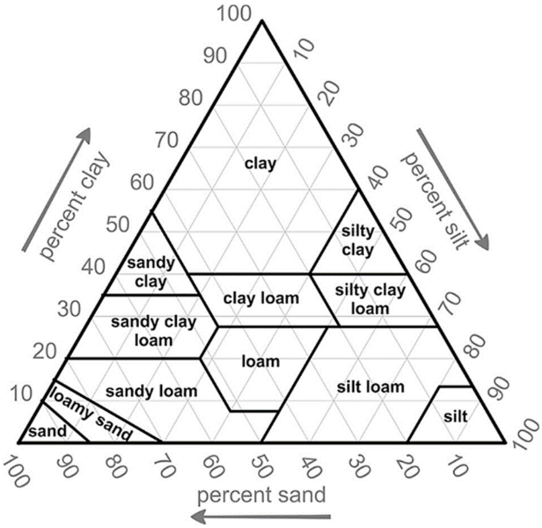

Figure 2. The soil texture triangle represents the different classes of the distribution of particle sizes (soil textures). The triangle shows how the textures are classified based on percent sand, silt, and clay. Adapted from USDA (Soil Science Division, 2017).

We use devices called *sieves* to do the job of separating grains by size. You might have used something similar at the beach when you were a kid: a pan with a mesh bottom that you put sand in and then shake. Most of the sand particles travel through the holes in the mesh; but in the pan, you are left with pieces of seaweed and shells that are larger than the holes. Similarly, sieves have assorted sizes of screened mesh that allow varied sizes of particles to pass through. The sieves are stacked in order with the largest mesh size on top and the smallest mesh size at the bottom. Particles smaller than the holes travel through to the next sieve; particles that are larger are caught in the pan. The grains that are caught in the middle are larger than that sieve's mesh size, smaller than the sieve above, and can be categorized according to the sieve size. If we wanted to use the soil texture triangle, we would use three sieves that isolate the sand, silt, and clay size particles, specifically.

Once you sieve a sample, you have three size fractions (sand, silt, and clay). To measure the volume of each size fraction there are two approaches, 1) volumetric displacement or 2) particle density.

For the volumetric displacement method, transfer each size fraction (e.g., sand, silt, or clay) from the sieve to a container of a known volume ($V_{container}$). Next, record the amount of water it takes to fill the container (V_{water}). Using these two values, subtract the volume of water (V_{water}) poured in from the total volume of the container ($V_{container}$) to get the volume of the size fraction (e.g., V_{sand}, V_{silt}, V_{clay}). Repeat for each size fraction.

As an example, the equation for V_{sand} would be,

$$V_{sand} = V_{container} - V_{water}$$

Repeat for V_{silt} and V_{clay}. Once you have the volume of each fraction (V_{sand}, V_{silt}, V_{clay}), you can add them together to get the total volume of the solids (V_{solids}).

$$V_{solids} = V_{sand} + V_{silt} + V_{clay}$$

Using the total volume of the solids, you can now figure out the percentage that each represents of the total. The sum of the %sand, %silt, and %clay of the sample will be 100%.

$$\%\ sand = \frac{V_{sand}}{V_{solids}} x\ 100$$
$$\%\ silt = \frac{V_{silt}}{V_{solids}} x\ 100$$
$$\%\ clay = \frac{V_{clay}}{V_{solids}} x\ 100$$

For example, if you have a sample that has a total volume of particles equal to 1000 cm^3, and you can use the displacement method to find that there is 300 cm^3 of sand, 400 cm^3 of silt, and 300 cm^3 of clay, then the percentages of each size fraction are:

$$\% \, sand = \frac{300 \, cm^3}{1000 \, cm^3} \times 100 = 30\%$$
$$\% \, silt = \frac{400 \, cm^3}{1000 \, cm^3} \times 100 = 40\%$$
$$\% \, clay = \frac{300 \, cm^3}{1000 \, cm^3} \times 100 = 30\%$$

The second possibility for calculating the volume of each size fraction is to use the particle density method. For this method, you must measure the mass of each fraction using a scale and divide each by the *particle density*. The *particle density* is the density of the solids that make up the sample. Mathematically, density is the mass per unit volume. Particle densities are typically expressed in grams per cubic centimeter and will vary depending on the mineral composition of the sample. Quartz is a common mineral constituent in soils and has a particle density of 2.65 g/cm^3. If you must guess the particle density of a soil, the density of quartz is often close; quartz is one of the most common minerals on Earth and it also doesn't erode easily. Therefore, a common range of particle densities is 2.55-2.70 g/cm^3 and exceptions are made when soil components are different. For example, the particle density of a soil will be lower if it has a lot of organic bits. For the particle density method, you can either assume that the density is the same for all particle sizes or use different densities if you know them.

To calculate the size fractions based on particle density, start again by sieving the soil. Then, transfer the size fractions into three containers. Using a scale, find the mass of each size fraction (m_{sand}, m_{silt}, m_{clay}) and subtract the mass of the containers. From the example above, imagine that the tared scale reads 795 g, 1,060 g, and 795 g for sand, silt, and clay, respectively; a tared scale automatically subtracts the mass of the containers. The sum of the three masses is the total mass of the solids (m_{solids}), 2,650 g.

$$m_{solids} = m_{sand} + m_{silt} + m_{clay}$$
$$m_{solids} = 795 \, g + 1060 \, g + 795 \, g = 2650 \, g$$

Assuming a constant particle density throughout (e.g., 2.65 g/cm^3 for quartz), divide each of the masses by the particle density to get volumes.

$$V_{sand} = \frac{m_{sand}}{\rho_{quartz}} = \frac{795\ g}{2.65\ g/cm^3} = 300\ cm^3$$
$$V_{silt} = \frac{m_{silt}}{\rho_{quartz}} = \frac{1060\ g}{2.65\ g/cm^3} = 400\ cm^3$$
$$V_{clay} = \frac{m_{clay}}{\rho_{quartz}} = \frac{795\ g}{2.65\ g/cm^3} = 300\ cm^3$$

Now that you know the particle size volumes, you can find the %sand, %silt, and %clay as shown above: 30% sand, 40% silt, and 30% clay.

Note: in the size fraction calculations above we used metric units (e.g., grams and centimeters). Those were simply the units chosen for the example problem. However, we also could have used imperial units (e.g., ounces and inches). Throughout this book we will use a variety of units, both imperial (e.g., feet and pounds) and metric (e.g., meters and kilograms). As a scientist, it is important to be able to use both and be able to convert between units.

After calculating the relative percentages of sand, silt, and clay, it is time to name the sample's soil type using the soil texture triangle. Using Figure 3a, first find the percentage of sand (30%) on the bottom axis. Notice how at 30% there are two lines; one line connects to 70% clay and one to 70% silt. In both cases, the lines are parallel to the adjacent sides of the triangle. Next, find the percentage of clay (30%) on the left side. Notice how the two lines from here go horizontally toward 70% silt and diagonally toward 70% sand. So, you must find where the 30% sand line and 30% clay line intersect. This might take some practice. Place one finger on the 30% sand location. From here, travel up and left along the line (toward clay). Meanwhile, using another finger find 30% clay. From here travel horizontally (toward silt). Move your fingers along the two lines until they intersect. At this intersection, visually trace the lines backwards and double check that the percent clay and percent sand are both 30%.

You now have your intersection point between sand and clay. Next, because the sand, silt, and clay percentages will always equal 100%, the intersection of sand and clay also defines your silt percentage. Double check this. Move up and right from the intersection of sand and clay and see where it intersects the silt axis. Does the point of intersection of 30% clay and 30% sand also have a percentage of 40% silt (Figure 3b)? Yes, it does. The last step is to read the name of the soil. The point of intersection lies within one of the boxes indicating the soil texture (Figure 3). In this case, the soil texture would be clay loam. As you can see in Figure 3b, there are many different combinations of sand, silt and clay that are all considered to be clay loam.

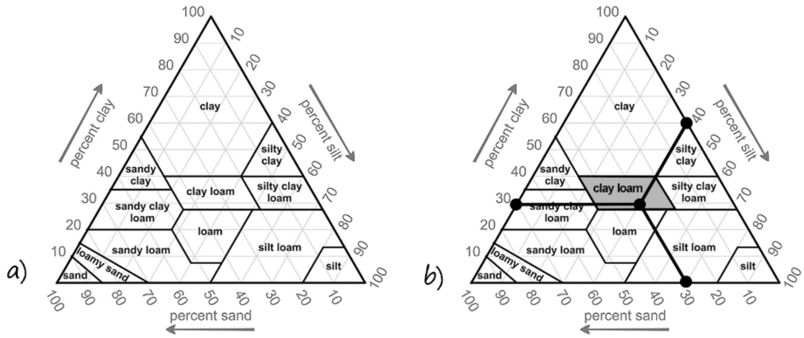

Figure 3. Two soil texture triangles: a) the soil texture triangle and b) the soil texture triangle with a clay loam highlighted (30% clay, 30% sand, and 40% silt). Adapted from USDA (Soil Science Division, 2017).

Now try it on your own! If you have a sample with 30 cm^3 sand, 5 cm^3 clay, and 15 cm^3 silt, what are the relative percentages of each grain size? What is the soil texture?

Answers:

- 60% sand, 10% clay, 30% silt.
- Soil Texture: Sandy Loam

The soil texture triangle is one of the simplest methods for classifying soils. There are scientists who have whole labs and computer programs dedicated to classifying soils, and many different disciplines (e.g., engineers, soil scientists, geologists) use different classification methods. What's more, the same soil can have a different name in different countries. In fact, some countries have entirely different-looking soil texture triangles. Although more detailed and complex methods for classifying soils exist, we will not go into them here. It is not necessary for you to know all the ins-and-outs of more complex classification methods; it is simply good to know that they are out there.

Porosity

You might be asking yourself, why does soil classification matter? Remember that by measuring the space between grains, we can better understand the quantity and location of water in a soil. We call the individual spaces between grains *pores*, and the total amount of empty space in a sample *porosity*. Porosity is the volume of the pores in a sample divided by the total sample volume.

$$n = \frac{V_{pores}}{V_{total}}$$

Equation 1: Porosity equation where n is porosity (L^3/L^3), V_{pores} (L^3) is the volume of the pores, and V_{total} is the total sample volume (L^3).

Note: look at the units in the caption for Equation 1 (e.g., L^3). You may not be familiar with this general format for units. “L” is a general representation of length (with units of centimeters, feet, inches, kilometers, etc.). Other examples of general variable abbreviations are “M” for mass (with units of pounds, kilograms, etc.) and “T” for time (with units of seconds, minutes, hours, etc.). Throughout the text we will do our best to identify key variables and provide their units in this format; we will also summarize them at the end of every chapter.

The total volume (V_{total}) in Equation 1 includes the volume of the solids (V_{solids}) and the volume of the pores (V_{pores}), which may contain air or water ($V_{total} = V_{pores} + V_{solids}$). One way that you can think of porosity is as a percentage. If the porosity is 0.34, then 34% of the sample is pores and 66% is solid material. Although “percent pore space” can be an easier way to conceptualize porosity, it is standard practice to express porosity as a fraction between 0 and 1. Additionally, even though porosity is “unitless”, it actually has units of volume per volume (L^3/L^3).

To better visualize porosity, think of your jar of marbles. If the porosity of the jar is 0.34 and the space between the marbles is the pore space, then 34% of your jar is void space and 66% of your jar is marbles. In this scenario, the size of each pore space is determined by the shape and size of the marbles (i.e., particles). The roundness and diameter of the marbles determine the way they stack on top of and next to each other.

Imagine a jar of baseballs and a jar of marbles. How do the pore spaces in the two jars compare? The marbles have a smaller diameter, creating smaller individual pores. The baseballs have a larger diameter, creating larger individual pores. This might blow your mind. But, despite the individual pores being smaller in the jar of marbles (compared to the jar of baseballs), the porosity of the baseball-filled jar and the porosity of the marble-filled jar could be the same. Wait, what? How is that possible? The jars are so different. Wouldn’t the larger pores have a higher porosity? Remember, the porosity of a sample is dependent on the total *volume* of the pore space, not the *size* of the individual pores. The diameter of the individual pores does not always correlate to the total volume of pore space. Although there are smaller pores in the marble-filled jar, there are also more of them.

To better visualize porosity, consider a pint glass of water and six small paper cups (~3 oz each) of water. If asked which of the vessels (or group of vessels) holds more water (the pint glass or the small paper cups), you might pick the pint glass. That is because it *looks like* the pint glass could hold more water. It's a bigger cup! However, if you were to pour all the small paper cups of water into an empty pint glass, it would overflow. Although the size of each individual paper cup is smaller, the *total volume* of water is larger. Many small pores (e.g., a smaller-grained soil) can have a greater pore volume than fewer larger pores (e.g., a larger grained soil). Lesson learned: don't let your eyes fool you. You cannot determine the porosity of a material by the *size* of the individual pores. Perhaps the hosts of the next fundraiser should try to get you to guess the volume of water that can be stored in jars of various materials!

Now it is time to apply this concept to soil. Let's start by considering a simple scenario: three (different) well sorted soils with subrounded grains. The largest individual pores are in the soil with the largest grains (Figure 4a) and the smallest individual pores are in the soil with the smallest grains (Figure 4c). Remember, the size of the pores (e.g., 2 mm) in a material does not necessarily tell us anything about porosity. To demonstrate this, using our well sorted samples in Figure 4, let's calculate the pore volume of each of these materials.

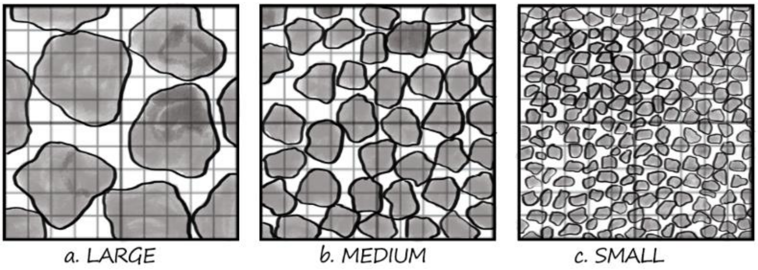

Figure 4. The pore volume for three soils with well sorted subrounded grains. The dark-colored area is the particles, and the light-colored area is the pore space. The larger grains have larger pores, and the smaller grains have smaller pores. To calculate the pore volume for each sample you can count the number of light-colored squares. The volume of each small square is = 1 mm^3. The dimensions of each sample (large square) are 18 mm x 18 mm x 1mm.

We can determine the pore volume (V_{pore}) for each soil by first counting the number of small, white squares (S) representing pore space. For example, in the large-grained sample (Figure 4a), draw a pencil dot in each white square to count them. Continue through the picture until there is a dot in each square. Do your best to guess how many full white squares there are by adding partial squares together. Note: the volume of each square is 1 mm^3.

Table 1 shows the number of white squares (S) that we counted. Remember that $V_{total} = V_{solids} + V_{pores}$. Therefore, if it is too difficult to count the white squares, you can also count the dark squares. For example, for the smallest-grained sample, it is easier to count the number of dark squares representing solid grains (each grain ≈ one small square) and simply subtract the number of grains from the total number of squares ($S_{total} - S_{particle} = S_{pore}$). Note: the dimensions of each sample are 18 mm x 18 mm x 1mm. Therefore, there are 324 total squares (18 x 18 = 324).

Table 1. Number of white squares (S) in Figure 4, for large, medium, and small grains

(a) S_{large}	99
(b) S_{medium}	90
(c) S_{small}	$S_{total} - S_{particle} = S_{pore}$ 324-180 = 144

Next, we must multiply the number of squares by the volume of each small square (V_s = 1 mm^3).

Below is the volume of pores present in each sample.

(a) $V_{pore\_large} = S_{large}\, x\, V_s = 99\, x\, 1\, mm^3 = 99\, mm^3$

(b) $V_{pore\_medium} = S_{medium}\, x\, V_s = 90\, x\, 1\, mm^3 = 90\, mm^3$

(c) $V_{pore\_small} = S_{small}\, x\, V_s = 144\, x\, 1\, mm^3 = 144\, mm^3$

The last piece of information that we need to calculate (before we can calculate the porosity) is the total volume (V_{total}) of the sample. If the dimensions of the sample are 18 mm x 18 mm x 1 mm, we can multiply these dimensions together to get the total volume.

$$V_{total} = Height\ x\ Width\ x\ Depth$$
$$V_{total} = 18\, mm\, x\, 18\, mm\, x\, 1\, mm = 324\, mm^3$$

The porosities of each sample are:

$$n_{pore\_large} = \frac{V_{pore\_large}}{V_{total}} = \frac{99\ mm^3}{324\ mm^3} = 0.31$$

$$n_{pore\_medium} = \frac{V_{pore\_medium}}{V_{total}} = \frac{90\ mm^3}{324\ mm^3} = 0.28$$

$$n_{pore\_small} = \frac{V_{pore\_small}}{V_{total}} = \frac{144\ mm^3}{324\ mm^3} = 0.44$$

Notice that the porosity is not connected to the grain size. The largest-grained material has a porosity of 0.31, the medium-grained material a porosity of 0.28, and the finest-grained material a porosity of 0.44. As mentioned above, a material with many small pores can have a greater pore *volume* than a material with only a few larger pores (think about the small paper cups and a pint glass).

Figure 5 shows common porosities for different materials. There aren't any noticeable trends between pore size and porosity. Clays and silts, although fine-grained, have higher porosities than gravel, but coarse sand typically has a higher porosity than a fine- or medium-grained sand. This gets even more complicated when geologic processes act on the subsurface. For example, karst forms when limestone is eroded by dissolution (think caves and sinkholes) which can lead to large void spaces. As a result, limestone that has been karstified can have a higher porosity than unkarstified limestone. Similarly, fractured granite can have a higher porosity than unfractured granite.

Now, Figure 5 is interesting, but it could be misleading. We cannot determine porosity simply by understanding the grain size. The porosity also depends on how closely the particles are packed, which depends on the shape and distribution of grain sizes. For example, there is a lower porosity in a jar that is filled with baseballs and marbles than in a jar of just baseballs. Why is that? Because the marbles fill in some of the pore space between the baseballs. Similarly, we would call the soils in Figure 4 well sorted; the grains are relatively similar in size. However, what would happen if we mixed the large grains (Figure 4a) and small grains (Figure 4c) together? We would get a poorly sorted sample like the one shown in Figure 6, and the small grains would fill the space between the large grains. Therefore, the porosity of the poorly sorted sample in Figure 6 is lower than the large-grained sample in Figure 4c.

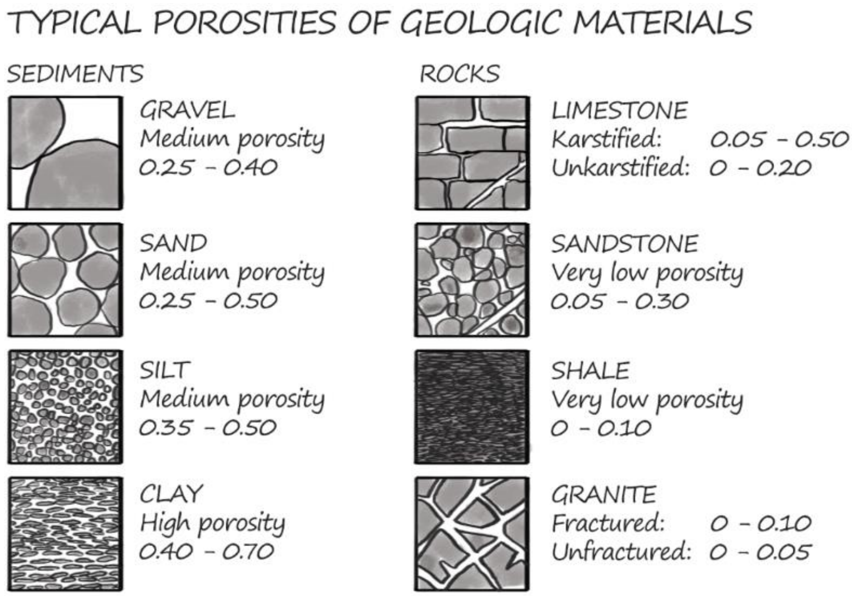

Figure 5. Typical ranges for porosity values of common materials. Adapted from Artiola & Uhlman 2009. Porosity values from Freeze & Cherry 1979.

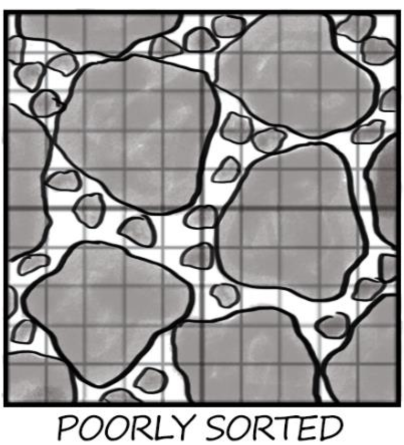

Figure 6. A poorly sorted soil with large and medium grain sizes. Notice, the pore space between large grains is filled by the medium grains.

To better understand how sorting impacts porosity, let's calculate the porosity for the poorly sorted scenario (Figure 6) and compare it to the well sorted soil with large grains above (Figure 3a). Note: the total volumes for both cases are the same (364 mm^3). The volume of each small square (V_s) is still 1 mm^3.

First, use the number of squares to calculate the volume of pore space for the poorly sorted sample:

$$S_{poorly} = 47\ squares$$
$$V_{pore\_poorly} = S_{poorly}\ x\ V_s$$
$$V_{pore\_poorly} = 47\ x\ 1\ mm^3 = 47\ mm^3$$

Next, calculate the porosity:

$$n_{poorly\_sorted} = \frac{V_{pore\_poorly}}{V_{total}}$$

$$n_{poorly\_sorted} = \frac{47\ mm^3}{364\ mm^3} = 0.15$$

Lastly, find the difference in porosity between the poorly sorted and well sorted samples:

$$Difference\ in\ porosity = n_{pore\_large} - n_{poorly\_sorted}$$

$$Difference\ in\ porosity = = 0.31 - 0.15 = 0.16$$

As we expected, the spaces between the large grains are filled with the smaller grains (Figure 6) and this reduces the pore volume significantly! The porosity of the poorly sorted soil is 0.15 and the porosity of the well sorted soil is 0.31 (Figure 4a). This is about a 50% decrease in pore volume due to differences in sorting.

By knowing the porosity, we know where the water travels right? Unfortunately, no. Typically, we are not *just* interested in the total pore volume of a sample, because pore spaces are not always *connected*. A common example of disconnected pores occurs in the rock type basalt (Figure 7a). Basalt is a rock that forms when lava cools rapidly. When the basalt cools, gas bubbles can get trapped in the material, creating holes in the rock called vesicles. Although these holes can be rather large, the holes are surrounded by rock and are not connected to other pores (Figure 7a).

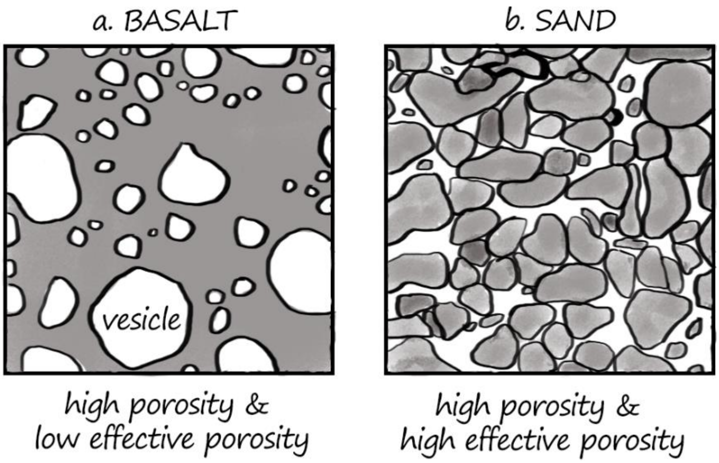

Figure 7. The effective porosity changes with the connectivity of pores. In a) basalt, the pores (small holes = vesicles) can be large but disconnected. In b) sand, the pores can be both large and connected.

When there are disconnected pores (Figure 7a), it is difficult for water to flow between the pore spaces. If you had a chunk of basalt and poured water on its surface, the water wouldn't flow through the rock and reach the pores at the bottom of the sample; the pores need to be connected for the water to flow in and fill them. The pore spaces that can contribute to fluid flow create what we call *effective porosity*. Therefore, a material like basalt could have a high porosity, but a low effective porosity. Because of the connectivity of the pores, a material like sand might have a porosity and an effective porosity that are nearly identical (Figure 7b).

Permeability and Hydraulic Conductivity

We have discussed the path that water follows through a porous medium, but we have not yet discussed how *fast* water moves. An important soil property related to the rate of groundwater movement is *permeability*. Permeability and porosity are often lumped together, but they describe two very different soil properties from a hydrologic perspective. *Porosity* is a physical description of the total pore volume in a sample and is a ratio (e.g., L^3/L^3). *Permeability* describes the ability of the porous material to transmit a fluid. It describes the ease of fluid movement through the soil and has units of length^2 (e.g., cm^2, m^2, ft^2). Fluid will move very slowly through a material with a low permeability.

An example is clay which has many small, poorly connected pores. In contrast, fluid moves easily through a highly permeable medium like a gravel with lots of large, connected pores.

Permeability is used to understand how a medium transmits *any* fluid through it. However, fluids can vary in their properties. For example, water moves through a jar of marbles differently than a more viscous fluid like honey. Therefore, in hydrogeology we use a more specific term, *hydraulic conductivity (K)*, when referring to the ability of a medium to transmit water. It is important to realize that the terms hydraulic conductivity and permeability are often used interchangeably. However, hydraulic conductivity only refers to the ability of water to move through a medium. Permeability is constant and independent of the fluid, but for the case of water (only) we combine some fluid and material properties to define hydraulic conductivity. Hydraulic conductivity has units of length/time (e.g., cm/day, ft/hr), like velocity.

We have mentioned the concept of pore size several times; pore size can be a nice clue to many properties of a soil. Although we have cautioned that the size of the individual pore spaces *does not* always relate to porosity, pore size is a reliable proxy for hydraulic conductivity. One way to visualize the impact of pore diameter on hydraulic conductivity is to imagine that the pore space is a series of interconnected tubes. When fluid flows through a tube, there is friction on the tube walls and the quickest flow is through the middle of the tube (Figure 8). The fluid velocity increases with distance from the tube walls. Therefore, narrow tubes, which have a lower maximum and average distance from a pore wall, have smaller maximum and average velocities than larger tubes.

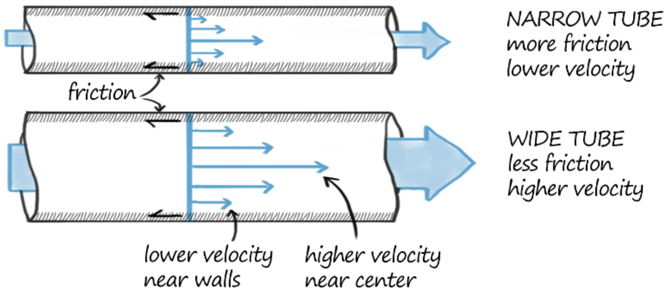

Figure 8. Represents the comparison of flow through a tube to flow through pores. Narrower tubes have more friction along the sides and slower velocities. Smaller pore spaces typically have lower hydraulic conductivities or transmission of water. Wider tubes and pores have higher hydraulic conductivities.

Imagine that you have two glasses of water. In one glass, you put a wide drinking straw. In the other, you put a wide drinking straw filled with little coffee straws. All the straws are the same length. If you wanted to suck water from the glasses, it would take more effort with the straw filled with coffee straws. This is because of the friction on the walls of the straws. Narrower tubes have more friction and typically slower flow velocities than wider tubes; wider tubes have less friction and higher flow velocities. For groundwater flow, the average diameter of the "tubes" is really the average pore diameter, which is generally proportional to the average particle diameter. Larger pores have higher hydraulic conductivities and smaller pores have lower hydraulic conductivities.

A table of hydraulic conductivities for common materials is shown below (Figure 9). In some cases, porosity (Figure 5) and hydraulic conductivity are directly proportional. Gravels have high porosity (Figure 5), and they also have high hydraulic conductivity (Figure 9); a fluid can move easily through the large, interconnected spaces. Clays can have a high porosity, but they often have a low hydraulic conductivity. Before reading further, can you figure out why clay has these properties?

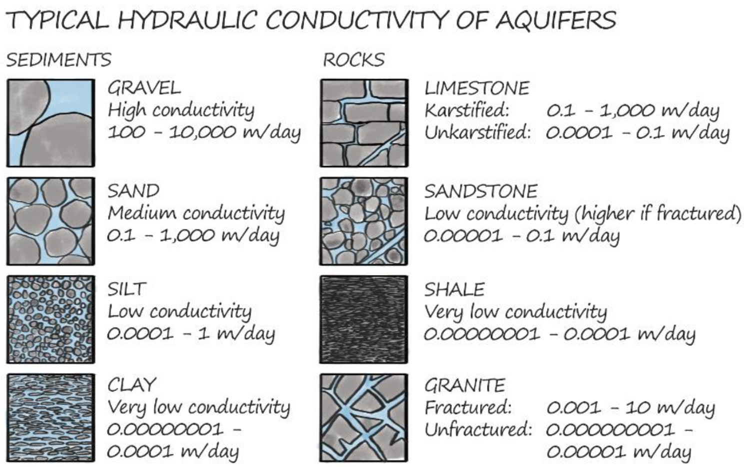

Figure 9. The hydraulic conductivity of common sediments and rocks. Adapted from Artiola & Uhlman 2009, conductivity values from Freeze & Cherry 1979.

Clay is a material that has a high porosity and a relatively low hydraulic conductivity because the grains in a clay are often platy, long, and flat like a dinner plate, and the water must make its way through the maze of grains

(Figure 10a). Although the pore *volume* of an uncompressed clay can be relatively large because of the structure of the clay (Figure 10a), the actual connections between the pores are typically small. Because the connections between pores are small, the diameters of the "tubes" (Figure 8) are small, the friction is high, and the velocities are slower. In comparison, the overall porosity of a sand is lower than that of an uncompressed clay, however, the pore size is larger, leading to a higher hydraulic conductivity (Figure 10b).

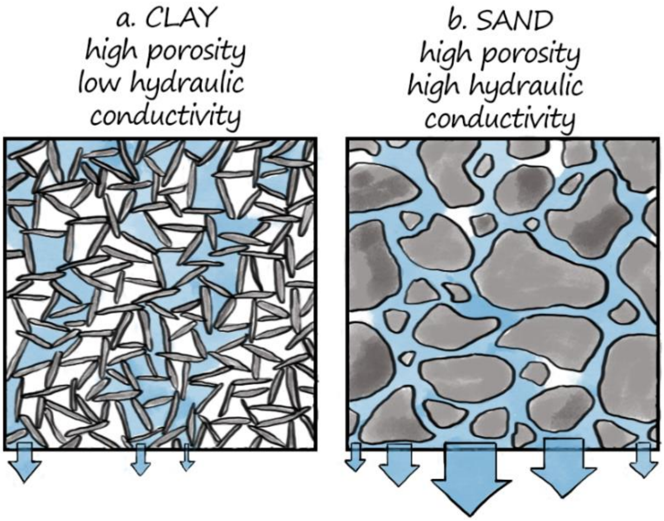

Figure 10. The porosity and hydraulic conductivity of a sample are connected, but not equivalent. For clays, you can have a high porosity (pore volume) and low hydraulic conductivity (determined by the diameter of the pore spaces). For sand, low porosity and high hydraulic conductivity.

Bulk Density

What do you think when you hear the word density? It is related to mass, but it's not the same. A truckload of marshmallows might weigh a lot, but they are not dense. A lead sinker used for fishing is dense, but it doesn't weigh much. If something is dense, it is heavy *for its size*. Previously, we discussed particle density which is the density of the solids that make up a soil sample. However, particle density only tells us about the solids in a soil – it is different from the *dry bulk density*, which includes the particles and the spaces between them. Solids and air clearly have different densities. The dry bulk density (ρ_{bulk})

combines both the particle density, the density of air, and the porosity of the material (Equation 2).

$$\rho_{bulk} = (1-n)\rho_{solid} + n\rho_{air} \approx (1-n)\rho_{solid}$$

Equation 2: Equation for dry bulk density (ρ_{bulk}, M/L^3), where n is porosity (L^3/L^3), ρ_{solid} is particle density (M/L^3) and ρ_{air} (M/L^3) is the density of the air. Note: the density of air can change with variations in atmospheric pressure, temperature, and/or humidity. But generally, you can assume that it is so small as to be nearly equal to zero. Therefore, the dry bulk density is nearly equal to the fraction of the sample that is solid multiplied by the particle density.

Calculate the dry bulk density (Equation 2) of a soil with a particle density of 2.57 g/cm^3 and a porosity of 50%. For this example (Figure 11), assume the density of the air is negligible.

$$\rho_{bulk} = (1-n)\rho_{solid} + n\rho_{air} \approx (1-n)\rho_{solid}$$

$$\rho_{bulk} \approx (1-n)\rho_{solid}$$
$$\rho_{bulk} \approx (1-0.50)\left(2.57\frac{g}{cm^3}\right) \approx 1.29\ \frac{g}{cm^3}$$

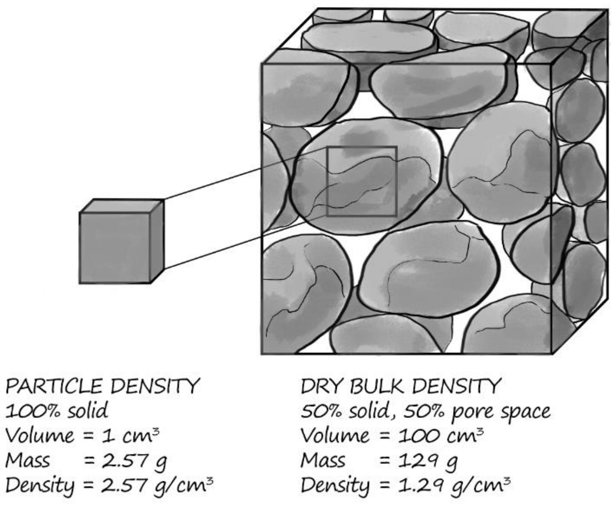

Figure 11. Particle density is the density of the granules. The dry bulk density is the density of a porous medium which includes pore space filled with air.

Not all samples are oven dried; some samples have water between the solids. Therefore, in addition to dry bulk density, you can also find the *total bulk density* of a sample. The total bulk density does not have to be a dry sample (i.e., the pore space can contain water). The total bulk density is the total mass (of the solids, air, and water) divided by the total volume.

Why might gardeners and farmers be interested in the total bulk density of a soil? There are two types of answer to this. First, think about what is likely to make a soil denser. One answer is that it has been highly compacted; in a compacted soil it is likely more difficult for water and air to flow. It can also be harder for roots to grow! In contrast, a soil might have low total bulk density if it has high porosity or if it has lots of lightweight solids (e.g., an organic rich soil – think of a piece of wood versus a piece of stone of the same size). A low total bulk density is usually a good thing for plants, and it is one of the reasons that you add mulch to soil. Generally, total bulk density increases with depth because there is less organic matter, there are fewer roots for aeration, and there is more weight from the overlying soil (which increases compaction). The second reason that total bulk density is useful is that it is relatively easy to estimate. Try it for yourself some time. Take soil from a garden (with permission). Squeeze half of it in one hand and leave the rest uncompressed in the other hand. You might not be able to estimate the total bulk density to the second decimal place, but you could learn to "feel" the soil density.

Compaction

We introduced the idea of compaction in the last section without really defining it. Compaction occurs when soil particles are pressed together, and pores space is reduced. Imagine that you put a kitchen sponge (1 cm thick) on the countertop. What would happen if you put a large book on top of it? It might get squished to half its height. Notice what changes. The mass of the sponge doesn't change. The actual sponge material doesn't change. However, the volume of the sponge does change. The weight of the book causes the solids to move closer together, decreasing the pore space and the ability of the sponge to hold water. Therefore, when you squeeze a wet sponge, water comes out! The same thing can happen to soil. The weight of overlying materials can cause soil to compact. When you compact a soil, the total bulk density changes, which affects other properties of the soil.

We use the term *compaction* to describe the degree to which a material decreases in volume when it is subjected to a load (i.e., a weight or source of pressure). Note: compression is a similar term, but it is really the verb form – you compress something, which causes compaction. In the case of the sponge, if we assume that the length and width don't change, then the change in volume is equivalent to the change in height (because volume = length x width x height). By convention, compaction is defined as the change in volume (in this case, height) divided by the initial volume (here, height).

$$Compaction = \frac{h_{initial} - h_{final}}{h_{initial}}$$

Compaction is useful if you want to *measure* how much your hydrogeology textbook causes your kitchen sponge to compact. But what if you wanted to *predict* how much compaction your chemistry textbook would cause? Physically, it makes sense to account for the different weights of the two books. Mathematically, we do this by defining an intrinsic property of the medium (one that doesn't depend on the sample size or the applied stress) known as compressibility. *Compressibility* is defined as the change in volume of a unit volume of the medium, under a unit of applied pressure. Generally, the compressibility of coarse-grained material is low because it is composed of mineral fragments that are supporting a load by grain-to-grain contact. The grains are already filling the volume quite efficiently, so it is hard for them to make room to allow further compaction. However, soil material that contains organic matter (which can compress under pression or tension) has a higher compressibility.

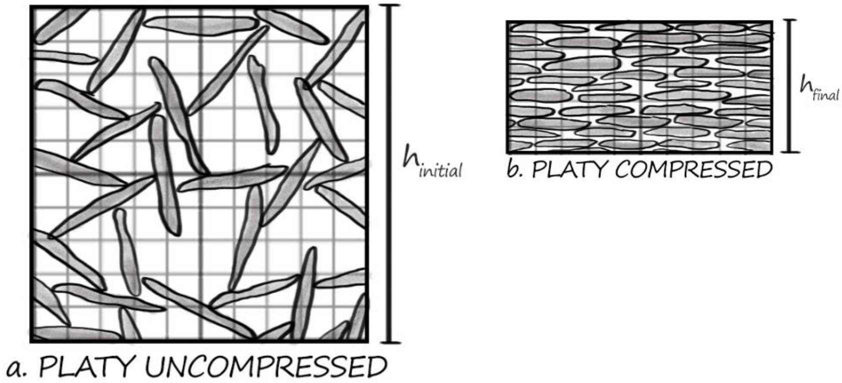

Figure 12. A clay soil with platy grains (a) uncompressed and (b) compressed.

Clay is the most compressible material. Clay particles are platy in shape (Figure 12a). So, when compressed, the plates stack on top of each other, like a house of cards being flattened, reducing the porosity of the soil (Figure 12b). This drastic rearrangement of clay particles can lead to long-lasting, even permanent compaction (i.e., the volume reduction caused by compaction will not reverse once the load is removed). Compaction can have dramatic impacts that can even be seen at the ground surface in the form of subsidence. Subsidence is when the ground caves in or sinks in a region, often caused by pumping water out of the pore spaces; the empty pore spaces are no longer supported by the water and collapse.

The compaction equation supplied above is a quick way to determine the change in height of a material due to pressure. To calculate the compaction of the material shown in Figure 12, count the number of squares in the vertical direction of each panel to get the initial height of the clay (Figure 12a, $h_{initial}$) and final height of the clay after compression (h_{final}). The height of each small square is 1 mm.

$$h_{inital} = 18\, mm$$
$$h_{final=}\, 9\, mm$$

$$Compaction = \frac{h_{inital} - h_{final}}{h_{inital}}$$
$$Compaction \; = \frac{18\, mm - 9\, mm}{18\, mm} = 0.50$$

The height of the clay material after compression is only 50% of its initial height!

Similarly, let's translate the concept of the changing porosity of a clay during compression into a math problem. Start by counting the pore space (S), or white squares, in Figure 12 for both the a) uncompressed and b) compressed clay soil.

For the uncompressed case Figure 12a:

$$S_{uncomp} = 167\; squares$$

For the compressed case (Figure 12b):

$$S_{comp} = 6\; squares$$

To calculate volume of the pores, remember that the volume of one square (V_s) is 1 mm^3 and the dimensions of a large sample square is 18 mm x 18 mm x 1mm (for the platy uncompressed clay). Notice that the dimensions of the sample square changes for the compressed case. Therefore, count the squares along the height and width to determine the total sample volume for the compressed case (V_{total_comp}).

The volume of the pores for each case:

(a) $V_{pore\_uncomp} = S_{uncomp}\ x\ V_s = 167\ x\ 1\ mm^3 = 167\ mm^3$
(b) $V_{pore\_comp} = S_{comp}\ x\ V_s = 6\ x\ 1\ mm^3 = 6\ mm^3$

The total sample volume for each case:

(a) $V_{total_{uncomp}} = h_{inital}\ x\ Width\ x\ Depth = 18\ mm\ x\ 18\ mm\ x\ 1\ mm = 324\ mm^3$
(b) $V_{total_{comp}} = h_{final}\ x\ Width\ x\ Depth = 9\ mm\ x\ 18\ mm\ x\ 1\ mm = 162\ mm^3$

Lastly, use the volumes of the pores and samples to calculate the porosity. The porosities for the two cases are:

(a) $n_{uncomp} = \frac{V_{pore\_uncomp}}{V_{total\_uncomp}} = \frac{167\ mm^3}{324\ mm^3} = 0.52$

(b) $n_{comp} = \frac{V_{pore\_comp}}{V_{total\_comp}} = \frac{6\ mm^3}{162\ mm^3} = 0.04$

$$Difference\ in\ porosity = n_{uncomp} - n_{comp}$$
$$Difference\ in\ porosity = 0.52 - 0.04 = 0.48$$

Wow! Although both materials in Figure 12 have the same grain size (based on their smallest dimension) and sorting, they have significantly different porosities before and after compression. The uncompacted soil has a porosity of 0.52 and the compacted soil has a porosity of 0.04; that is a 77% decrease in pore space during compression. To visualize this, think about a fully packed suitcase before a vacation. What do you do if you discover that you forgot to pack your toiletries? Simple, you cram the stuff in your suitcase to make more space. And where does that space come from? The “pores” between your clothes. The same thing can happen in a soil; if it gets compressed it can react by reducing its porosity. Another subtle but key point is that not all loss in volume by compaction is due to a change in porosity; the grains themselves can also be compressible. However, even soils with compressible grains lose most of their volume by rearrangement and reduction of pore space during compaction.

Not all soils are significantly affected by compression, but it is important to consider the compaction of soils for some cases. For example, subsidence can often occur in marshes, swamps, or other very wet areas underlain by soft, easily compressible soils that are rich in organic matter. Therefore, engineers must sink pillars down to more solid rock to support highways or buildings in these settings – if they didn't, the entire structure would sink over time as the material below it compacted under the stress of the overlying weight.

Water Content and Saturation

We mentioned previously that water is found in pore spaces. Imagine pouring water into your jar of marbles. It would fill the pore spaces between the marbles, right? Furthermore, if you knew the volume of the pore space (V_{pores}), you could pour that *exact volume* of water and fill the jar perfectly to the brim. If you poured less than V_{pores}, you would have some pores that would be full of water and others that are partially or completely empty. Like a jar of marbles, the pore spaces between grains can be fully or partially filled with water. We use the term water content (θ) to describe how much water is stored in the pore spaces between grains. Think of water content as the ratio between how much water you pour into your jar and the total volume of the jar (solids + pores spaces).

$$\theta = \frac{V_{water}}{V_{total}}$$

Equation 3: Equation for water content (θ, L^3/L^3) where V_{water} is the volume of water (L^3) in a sample and V_{total} is the total volume of the sample (L^3) (solids + pore spaces).

Just like porosity (Equation 1), you can think of water content as the percentage of water in a sample. If a sample has a water content of 0.34, that means the total volume is 34% water. Also note that the correct units for water content, like porosity, are volume per volume (e.g., cm^3/cm^3). But, because these units cancel each other out, they are often omitted.

The water content of a material is limited by the porosity (i.e., you can't have more water than you have space for the water!). Therefore, the water content is always less than or equal to the porosity, and the maximum water content (or the *saturated water content*, θ_s) is close to the value of porosity. Why do we say, "close to"? When water enters pores, little pockets of air can get trapped, just like when pockets of air get trapped in your swimming suit when you jump in a pool. Due to entrapped air, the saturated water content may be slightly less than the porosity.

It is rare that a soil is completely void of water. Even after a soil drains, there are some water molecules that stick onto the soil grains. If you were to drain the water from the jar of marbles, right after you drained the water, even once the void space emptied, the marbles would still be wet. After the initial draining of a soil, it might contain a fair amount of water. Some of this water drains slowly because it must flow through small, winding (tortuous) paths to get to the outlet. Eventually, all or most of this water will drain out, leaving the minimum water content that can be achieved by drainage, which is known as the *residual water content* (θ_r). In the case of soils, this residual value is water that adheres to the grains, which depends on the specific surface area of the grains. The *specific surface area* is the total surface area per unit mass. As we saw before, smaller particles have smaller pores. Another characteristic is that smaller grains have a greater particle specific surface area. These two things taken together – smaller pores and more area on which water can adsorb – lead to a higher residual water content (θ_r) and slower drainage. Here's a practical example – if you look at a soccer field after a rain, the area in front of the goal is often the last to dry out. Why? The goalie spends a lot of time walking around this area, compacting it. So, the area in front of the goal has smaller pores than the rest of the field. As a result, water cannot flow through it as quickly and more water gets stuck on the soil grains.

It is useful to have a term to describe how full or empty pore spaces are. We use the term *water saturation* (S_w) to quantify the fraction of the *pores* that are filled with water. Remember, water content is the fraction of the *total volume* that is water-filled (Equation 3). Water saturation (S_w) is simply the ratio of the water content (θ) to the porosity (n).

$$S_w = \frac{\theta}{n}$$

Equation 4: Equation for water saturation (S_w) (-) where θ is water content (L^3/L^3) and n is porosity (L^3/L^3).

Another way to think of water saturation (Equation 4) is that it is the ratio of how much water you pour into your jar of marbles (not necessarily filled to the top) to how much water you could add before it overflowed (filled to the brim). Or, in other words, water saturation is the ratio of the added water (V_{water}) to the total amount of pore space between the marbles (V_{pores}).

$$S_w = \frac{V_{water}}{V_{pores}}$$

Equation 5: Second equation for water saturation (S_w, -) where V_{water} (L^3) is the volume of water in the sample and V_{pores} (L^3) is the volume of the pore space.

If the pore spaces in a medium are all full of water, the two volumes (V_{water} and V_{pores}) are the same. That is, the water content is equal to the porosity, and the water saturation is 1 (Equation 4). If there are some partially empty pore spaces, the water saturation will be less than 1. As an example, if you had a soil with a water content of 0.28 and a porosity of 0.34 then the water saturation would be 0.82. You can interpret this saturation to mean that 82% of the pores are filled with water. If the water content is equal to the porosity ($\theta = 0.34$, $n = 0.34$) the water saturation is 1 and 100% of the pores are filled with water. The more pore spaces that are empty, the closer the water saturation is to zero. If none of the pores have any water in them, then the saturation is zero. Keep in mind that this never really happens unless you bake a soil in the oven. Even in the desert, soil will still hold some water!

The distribution of water content (θ) in the soil defines zones of saturation and partial saturation underground. Regions underground where all the pore space is filled with water ($S_w = 1$) are called *saturated zones* (Figure 13). The region where the water content is less than the porosity (the pore spaces are partially filled with water, $S_w < 1$), is called the *unsaturated zone* (Figure 13). Due to forces of adhesion, cohesion, and surface tension, water is drawn upward from the saturated zone toward the unsaturated zone by capillary action. During these processes water is held in a transition zone (between the saturated and unsaturated zone) called the *capillary zone* (Figure 13). If you really want to impress people at a party, you can pull out two additional terms. The *zone of effective saturation* is the region where the water saturation is almost complete – except for entrapped air. The *zone of residual saturation* is the region that is as dry as it can be through drainage.

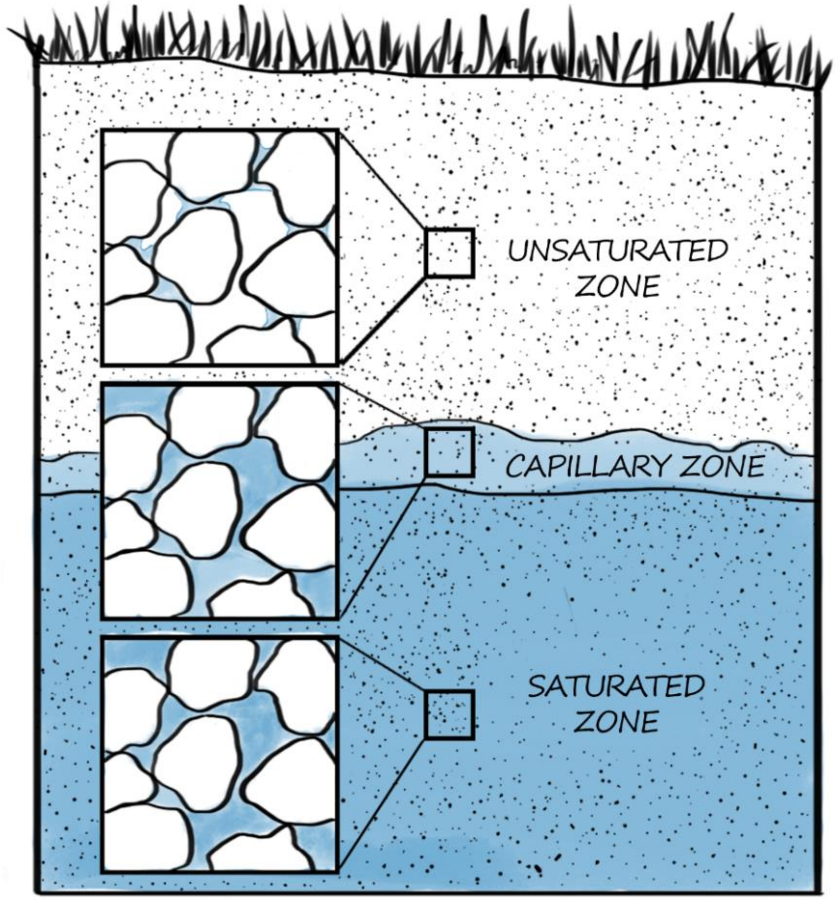

Figure 13. The water content (θ) determines underground zones of saturation. The region where all the pores are filled with water is the saturated zone. The region where $n > \theta$ is the unsaturated zone. The very top of the saturated zone is fully saturated because of capillary action and is called the capillary zone.

You don't have to understand all the ins and outs of capillary action, but you have likely seen it. Where? A candle. A candle burns the melted wax that is right at the tip of the wick; the heat from the flame melts some wax, which moves up to the tip of the wick by capillarity. Similarly, if you take a piece of string and dip just the tip of it in a cup of water, surprisingly, it isn't just the end of the string that gets wet. Some of the water travels up the string by capillary action. In the same way, water travels upward, against gravity, through the pores in a soil.

The capillary zone sits on top of the saturated zone (Figure 13). In the capillary zone, the pores are completely filled with water by capillarity. Above the capillary zone, capillarity can only fill some of the pores. Which ones get filled? The smallest ones. Why? Because they have a higher specific surface area. See the pattern here? This is the very start of an introduction to water flow in unsaturated soils (which is an especially important field) – without unsaturated flow we wouldn't have plants or dry ground to walk on. But unsaturated flow also gets complicated quickly. Therefore, in this book, we will focus on the *saturated zone*. However, we strongly encourage you to take a course in unsaturated flow (sometimes called vadose zone hydrogeology) if you want to continue in hydrogeology or soil-related professions.

Water Storage

As discussed above, if you start with a dry soil and pour water on it, the water will fill the pore space between particles; the water in the material is referred to as being in storage or as *water storage*. Water storage can change with time. The ability of a soil to store water (and to release water from storage) will become important later when we discuss pumping wells. But for now, it is enough to understand that the materials underground can store water.

So, water can be stored between grains in the pore spaces. If this isn't clear, take some time to re-read the sections above. Next, we wish to introduce two ways that the amount of water stored in a fully saturated soil can change. Stop here for a minute and think about how strange that statement was. How can the water content of a fully saturated medium change if the saturated water content equals porosity? In other words, if water is removed from the subsurface, how is it possible for a soil to remain saturated? The trick here is that porosity can change. Remember compaction? With compaction, we can take water out of a soil and have the soil remain fully saturated. However, this can only happen if the pore volume changes by *exactly the same amount* as the volume of water removed.

So far, so good? Now it gets a bit counterintuitive. To start, let's ignore changes in porosity and consider the water itself. Water is compressible. That is, its volume will decrease if you increase the pressure applied. Imagine a fully saturated sample and think of all the water molecules as little "springs". If you force more water into the pore spaces (by increasing the water pressure), you will compress these springs to make room. Of course, water molecules aren't actually springs, but this is close to what happens. What if you were to remove some water from the pores? That would decrease the water pressure and the water expands to fill the pores, meaning that the soil could still be fully saturated even though you took water out.

The compressibility of water is relatively tiny. Therefore, the compressibility of water doesn't usually have a large effect on storage. What has a larger impact on storage is how the soil particles respond to the removal or addition of water. From the viewpoint of the soil particles, when the springs are compressed, they actually push on the soil particles, keeping them apart. However, when water is removed and the water springs relax a bit, if the soil isn't already highly compacted, then reducing the water pressure will allow the soil to compact. This is because when you remove water, you reduce the pressure. When you reduce the pressure, the soil compacts. When the soil compacts, the soil can be fully saturated even though you removed water.

It is important to understand that the amount of water stored in a soil can change by 1) the filling and draining pores, 2) compressing the water (a little bit), and 3) changing the porosity. Take a guess, what do you think might happen if the subsurface has a lot of clay present and you removed water from underground? That's right, you could get a lot of compaction. In fact, you can get so much compaction that the ground surface can drop by several meters. Look for pictures of the effects of subsidence in Mexico City as an example. What's more, even if you stopped removing water from the subsurface, you would not recover the surface deformation. Compaction is irreversible!

Conclusion

Before you can build a house, you must lay a foundation. Before you can study groundwater, you must understand soil properties. Soil properties such as: grain size, sorting, shape, bulk density, and compaction all characterize the void space (i.e., pore space) between soil grains. Why is pore space important? Remember, pore space is where water resides! Therefore, the size and connectivity of pores control factors such as storage (how much water is underground) and hydraulic conductivity (how easily water can move through the pores). Can you begin to see how this chapter is foundational to your groundwater journey? By understanding the properties of soils, you can start to understand influences on groundwater quality, quantity, and movement. This is only the beginning!

What to Remember

Important Terms	
angularity	saturated zone
capillary zone	sieves
compaction	soil
dry bulk density	soil texture
effective porosity	soil texture triangle
grain size	sorting
hydraulic conductivity	storage capacity
particle density	total bulk density
permeability	unsaturated zone
pores	water content
porosity	water saturation
porous medium	water storage
residual water content	zone of effective saturation
saturated water content	zone of residual saturation

Important Equations
$V_{total} = V_{pores} + V_{solids}$
$n = \frac{V_{pores}}{V_{total}}$
$\rho_{bulk} = (1-n)\rho_{solid} + n\rho_{air} \approx (1-n)\rho_{solid}$
$\theta = \frac{V_{water}}{V_{total}}$
$S_w = \frac{\theta}{n}$

Important Variables		
Symbol	**Definition**	**Units**
V	volume	L^3
n	porosity	L/L
ρ	density	M/L^3
θ	water content	L/L
S_w	water saturation	-

Chapter 2

Systems

Introduction

So far, we have talked about hydrologic properties that you could see in a small soil sample. But the real complexity in hydrogeology comes when you think about real hydrologic systems (e.g., lakes, rivers, underground water storage, etc.), each with different local properties. Systems that you are familiar with are all around us: the solar system, the immune system, computer systems, political systems, ecosystems – just to name a few. A *system* is a group of interdependent items that form a unified whole; changes in one part of the system impact other components in the system. For this section we will introduce the general concept of systems before diving into hydrologic systems, specifically. Our hope is that by starting with familiar systems, you will form a strong basis to understand hydrologic systems.

Boundaries and Inputs/Outputs

All systems have boundaries or limits. Your body stops at your skin, which is the boundary of the system that is your body. A computer stops at the plastic and glass casing. A bathtub stops at the porcelain walls of the tub (Figure 14). These physical boundaries (skin, casing, and porcelain walls) might be considered obvious physical limits of each of these systems. However, boundaries are just *conceptual lines* that divide a system from everything else. What this means is that we (as observers) are the ones defining boundaries; they are not necessarily objectively determined by specific physical features. Boundaries are a useful mental construct to divide the system from the rest of the world. We will start with a few simple examples that have physical boundaries and explain conceptual boundaries as we move forward.

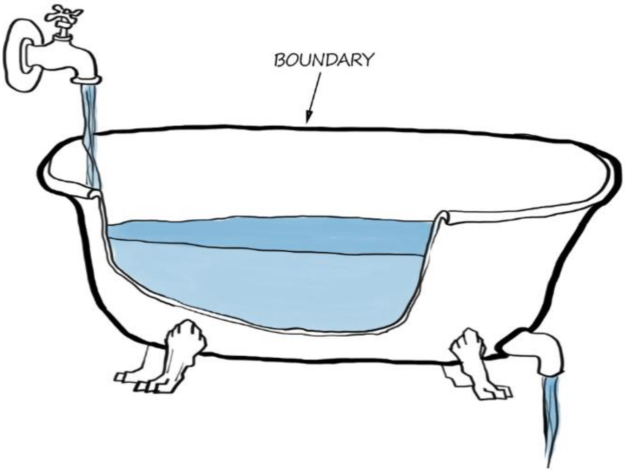

Figure 14. Bathtub system with the tub's boundary highlighted. The porcelain walls of the tub are the boundary dividing it from the rest of the world.

There are three categories of systems: closed, isolated, and open systems. A closed system allows transfer of energy, but not transfer of mass. Consider an uninsulated cup of coffee with a tight lid. The boundaries are the walls of the cup and lid. Heat (energy) can be lost through the sides of the cup. However, with the lid sealed, coffee (mass) is not lost. In other words, if you put this cup of coffee on a scale, the coffee would get cooler with time, but the scale would read the same weight throughout the process. For isolated systems, neither energy nor mass is transferred. For an isolated system, consider a cup of coffee in a perfectly insulated thermos; neither energy (heat) nor coffee (mass) is lost. Lastly, an open system allows mass and/or energy to be transferred to the surroundings. Consider a hot cup of coffee without a lid being jostled, or a barista giving you a refill; liquid is transferred in/out of the system and the addition of hot coffee can bring heat (energy) with it.

Anything that crosses the system boundaries from the outside to the inside is an *input*. Anything that crosses from the inside to the outside is an *output*. All open systems can have inputs and outputs (Figure 15). For the open coffee system, an input would be a refill, and an output would be evaporation or a person sipping from the cup. For a bathtub system, all the water coming into the tub from the faucet is an input and all the water leaving the tub through the drain is an output (we will assume that you are not splashing). For a digestive system, inputs include food and water. The outputs… we won't list them in polite company.

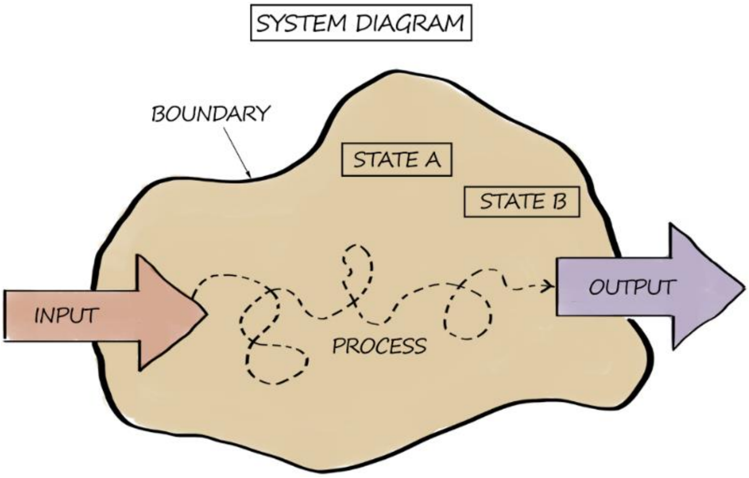

Figure 15. Generic system diagram with the major components of all systems: inputs, outputs, boundaries, internal processes, and states.

System Conditions

Once inputs cross a system boundary, a series of internal processes can occur within the system (Figure 15). These internal processes can be flow (e.g., the path that milk takes once it is poured into coffee), mixing (e.g., stirring the coffee and the milk becoming evenly distributed), or chemical reactions (e.g., sugar dissolving into the coffee).

- If (absolutely) nothing is changing over time, we call a system *static*; for this to occur, there can be no inputs, outputs, or internal processes.
- When internal processes are occurring and their rate changes anywhere in the system over time, then the system is *transient*.
- If internal processes are occurring at the same rate at every point in the system over time, then we call the system *steady state*; this also requires that the inputs and outputs are constant in time.

The terms static, steady state, and transient refer to the *system state*. The system state is the condition of the entire system. The inputs, outputs, and internal processes together control the system state. To determine the system state, think about if you were to step back and look at the entire system. How would you describe it over a prescribed time period? Is it changing? Is it staying the same? Is it gaining mass? Is it losing mass? For gaining and losing mass, we apply two additional terms to describe the system state that are more specific: gaining and losing. This can get a bit complicated, so let's give some examples.

Generally, begin by thinking of system states as:

- If there are no inputs or outputs for the time period of interest, the system state is *static*.
  - Example: a cup of room temperature black coffee sitting on a table – the coffee level is not changing over time.

- If the inputs are greater than the outputs for the time period of interest, the system state is *transient* and *gaining*.
  - Example: you are pouring coffee into a cup, so the coffee level in the cup increases over time.

- If the outputs are greater than the inputs for the time period of interest, the system state is *transient* and *losing*.
  - Example: you are drinking your coffee, so the coffee level in the cup decreases over time.

- If the inputs equal the outputs for the time period of interest, the system may be in *steady state.*
  - Example: a broken coffee machine continuously pours black coffee into a pot, but you drink it exactly as fast as the machine pours it, keeping the coffee level in the pot constant. (We do not recommend this.)

- If the inputs equal the outputs for the time period of interest (but there are internal changes in the system) the system may be *transient.*
  - Example: you are walking to class with a coffee in hand, you are neither drinking nor pouring coffee into the cup, but the coffee is sloshing around (i.e., the coffee level is not constant).

There are more complicated factors involved in determining system states. But we will begin here. As we learn terminology throughout the section, we will dive deeper into fine-tuning our definitions.

State variables define the system state. When you step back to look at the behavior of your system what are you observing? For example, in the coffee cup example above, to determine how the system was changing you observed the *level of coffee* in the cup. Was it higher? Lower? The same? The level of coffee is an example of a state variable. Generally, state variables are qualities of a system that you can measure and that can change in time and space within the system.

There can be more than one state variable for a system, but you typically choose one to observe. In the example above, the state variable was the level of the coffee. Another example of a state variable for the coffee cup system is the temperature of the coffee. What is the initial temperature of the coffee? Is it decreasing over time? Does it remain the same? The state variable depends on your question. If you are interested in how quickly coffee is refilled in your cup, you would choose the coffee level as your state variable. If you are interested in the effectiveness of your thermos in keeping your coffee warm, you would choose temperature as your state variable. Common examples of state variables are pressure, temperature, volume, mass, and density.

To properly understand if a system is static, gaining, losing, or steady state, you must look at the state variable at every location in the system. A system is only at steady state when the state variable is constant at every location (over time). If the state variable is changing, even at one location in the system, then the system is not at steady state. Rather, the system is in a transient state. The word transient implies that the state variable is not constant or is changing in time. Transient is the general term for systems that are changing – as mentioned previously, gaining and losing systems are specific cases of transient systems.

Let's look more closely at the bathtub system. In a bathtub, water pours into the tub through a faucet (input) and water leaves the tub through a drain (output). Take a moment to define a state variable and the system states for the bathtub. What are you observing in the bathtub (state variable) to determine the overall condition of the system (system state)? While answering this, also think about what series of events would cause the bathtub to be transient (gaining or losing), static, or steady state. During your brainstorm, notice that there are many different state variables you can choose from to define this system. Your choice depends on your interest: the volume of water in the tub, the temperature of the water in the tub, or the cleanliness of the water in the tub (just to name a few). However, for this example, we suggest you observe the water level in the bathtub.

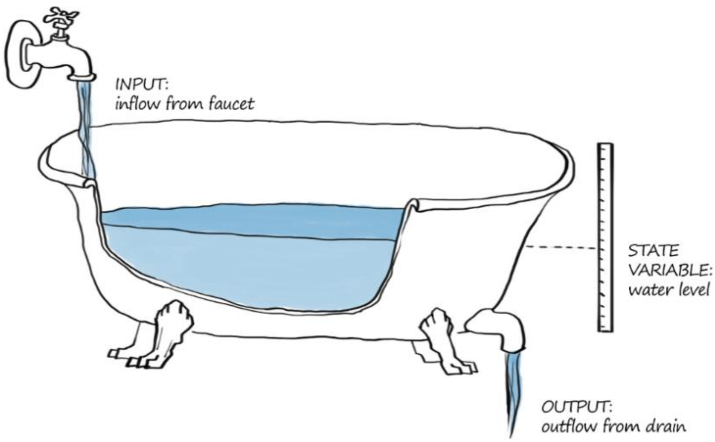

Figure 16. System diagram for a bathtub.

Answers:

- State variable: water level in the bathtub
- Input: water coming into the tub from the faucet
- Output: water leaving the tub through the drain
- System states:
    - Steady state (Figure 17a): the amount of water coming into the tub from the faucet is equal to the amount of water leaving through the drain (e.g., the faucet input = the drain output). The water level (state variable) is not changing anywhere in the system.
    - Transient (Figure 17b): the state variable is changing somewhere in the system
        - Gaining: more water is coming into the tub than leaving it (e.g., the drain is plugged completely or full of hair; there is a high-flow faucet and a small drain hole). The water level (state variable) is increasing at one or more places in the system.
        - Losing: more water is leaving the tub than coming in (e.g., pulling the plug after a long bath while not adding any water). The water level (state variable) is decreasing somewhere in the system.
    - Static (Figure 17c): there is no water entering or leaving (e.g., faucet input = drain output = 0).

As mentioned previously, there are many different state variables (e.g., mass, volume, density, etc.) that you can observe for a single system. However, you must choose only one to observe, because for each state variable the system states will be different. For the bathtub system above (Figure 17), if you had chosen temperature as your state variable, the conditions for each system state (e.g., steady state, transient, static) would change. Let's look at this more closely. Figure 17a shows a steady state system with respect to the water level. Steady state means the water level is not changing anywhere in the system. However, what if the water coming into the bathtub is warmer than the water leaving the bathtub, and the state variable is temperature? The system would be gaining with respect to temperature (the state variable), but steady state with respect to water level. Similarly, an empty bathtub could be considered static with respect to water level (Figure 17c). However, over time if the bathtub remains empty and collects dust, the bathtub could be considered losing with respect to cleanliness (a potential state variable). This means that the same system, with the same series of outside forces acting on it, could have different system states depending on which state variable you choose to observe.

To summarize, to define the system state it is best to understand both the state variable at every location in the system and what is happening at the boundaries of the system (e.g., inputs/outputs). Keep in mind that static state is a special category of steady state where the system has zero inputs/outputs and no processes are occurring anywhere in the system. Really think about how steady state is different than static for the bathtub (even though the water level stays the same for both). For static, no new water is coming into the tub from the faucet or leaving the tub through the drain. The state variable (water level) is constant everywhere; but also, there is no movement of water at any location in the tub (Figure 17c). However, if the volume of water coming in from the faucet is the same as the volume leaving through the drain, the water level doesn't change, but the molecules of water in the tub are constantly leaving and being replaced by new ones; there is flow in the tub and water crosses system boundaries. The system is steady state (Figure 17a). These are subtle but crucial differences when you apply them to hydrologic systems.

One last note about system states. System states are relevant to another important property of the system: the storage. We discussed water storage in the previous chapter; most systems have some sort of storage (not just

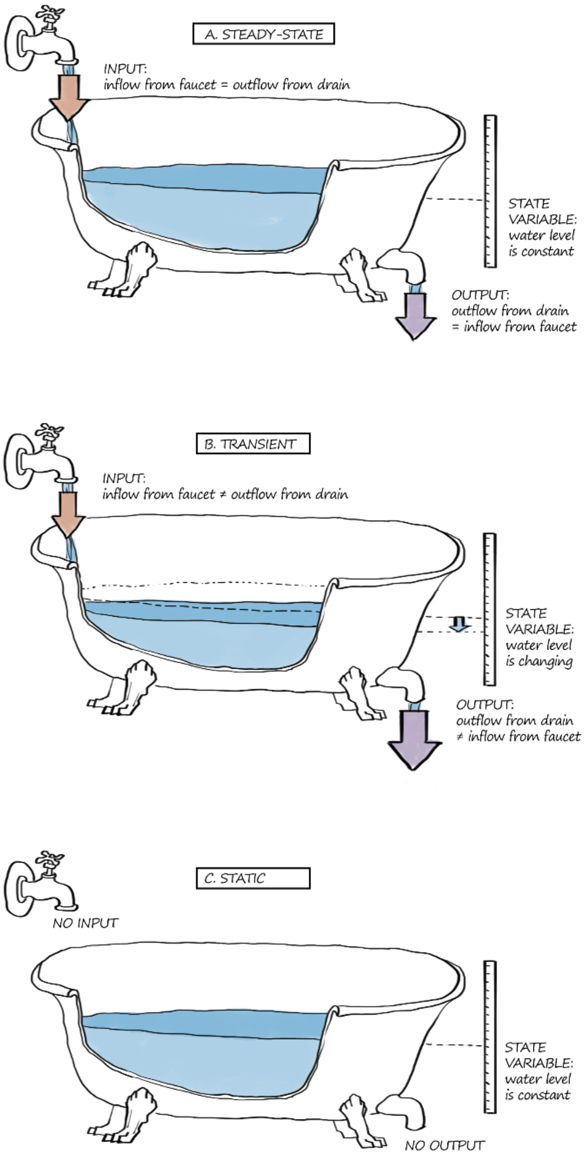

Figure 17. Bathtub example with (a) steady state, (b) transient state and (c) static state.

groundwater systems). One way that a system can be transient is by having a change in storage (anywhere in the system) over time. The system is steady state if (and only if) there are no changes in storage anywhere in the system over the time interval of interest. For example, imagine that you have two pockets with a storage of marbles. You have five marbles in one pocket and zero in the other pocket. Next, imagine that you move two marbles to the empty pocket. You always have marbles in your system, so there is no change in storage (i.e., the two-pocket system is in steady state). However, part of your system is gaining marbles into storage, and another is losing marbles from storage.

Below are a few simple steps and a summary table for determining the system state using input-output relationships, boundary flow, and the local storage within the system.

- If the input and output rates are different from one another, then the system is transient.
  - If more water is entering than leaving (inputs > outputs), then the system is gaining.
  - If more water is leaving than entering (outputs > inputs), then the system is losing.
- If there is no change in storage with time anywhere in the column, then the system is static.
- If the flow into and out of the system is the same (but not zero) and there is no local change in storage, then the system is in a steady state.

Table 2. Input-output relationships, local storage, and flow status for different system states

System State		**Input-Output Relationship**	**Boundary Flow**	**Local Storage**
Transient	Gaining	Inputs > Outputs	Yes	Change
	Losing	Outputs > Inputs		
Steady State	Static	Inputs = Outputs = 0	No	No Change
	Steady State	Inputs = Outputs	Yes	

Observation Time

When examining a system, you must start at some point in time. Observing a system isn't like observing a movie. There isn't usually an "official" start and stop time to a system; systems most often function continuously over time. As the observer, you must choose a time period of interest or an *observation time* (i.e., when you want to start and for how long you want to observe the system). Of course, because there isn't any one moment when time stops, your start (and end time) must be something you the observer decide. Your decision will depend on the question(s) you want to answer about your system.

The moment right before your observation time starts is time zero. At time zero, your system is at an *initial condition*. The initial condition is simply a description of the system immediately before your start time. In other words, it is a list of all the state variables at time zero. Of course, remember that most systems function continuously. Therefore, the system state depends on all the inputs and outputs and internal processes that have occurred up to that time. At time zero, the initial condition describes and is characterized by the history of the system up to time zero.

Similar to how we can only describe what happens within the boundaries of our system, we can only describe what happens during our observation time. You define the initial conditions everywhere in the system. Then you define what happens at every point on the boundaries during the observation time (inputs and outputs). Finally, your understanding of the system's internal processes allows you to understand, even predict, how the system will evolve through your observation period. For example, if I said that Dominique drove in a straight line in the direction that he was facing at 55 mph for one hour, could you tell me where he would end up? No. You would need to know where he started and which way he was facing (i.e., his initial conditions). However, if you knew the initial conditions and the processes that occur – you could predict the system states at every time during the observation period, including the *final condition* (e.g., where Dominique would end up) at the end of the observation period.

To further explore system concepts, let's look at a bathtub system over the course of five minutes. The state variable of the system is the water level (Figure 18):

- *Minute 0 (initial condition)*: The bathtub is empty. The faucet is off, and the drain is unplugged. The system is static.

- *Minute 1:* You turn the faucet on (the drain is still unplugged). The flow rate is low enough that the inflow from the faucet is exactly matched by the flow out the drain. The water level is constant. You have a steady state system, and the change in water storage is zero.
- *Minute 2:* You then plug the drain but leave the faucet running. Now, you have a transient (gaining) state as the bathtub fills. The water level and the water storage are increasing.
- *Minute 3:* You turn the water off (with the drain still plugged). Now, there is no flow coming in or out of the tub for a minute. At first, the water may continue to swirl around in the tub causing the water level to fluctuate (transient). But, if you wait a few seconds, you will have a static system. There is no change in water storage or water level.
- *Minute 4:* You pull the plug. Now, the flow out is greater than the flow in (which happens to be zero). You have a transient (losing) state. The water level and the water storage are decreasing.
- *Minute 5 (final condition):* All the water has flowed out of the tub, so the inflow and outflow are zero. The system is static, and the water storage is zero.

For this system, just over the course of five minutes, notice how much the system state changed. It was static, steady state, gaining, and losing. However, if you only look at the system for the entire observation time, from initial condition (minute 0, empty) to final condition (minute 5, empty), the system is steady state (specifically, static). In other words, if you looked at the tub at minute 0, left the room and came back five minutes later, you might even think the tub was empty (static) the whole time. But, in fact, the system experienced an entire range of system states (e.g., static, gaining, losing) during the observation time.

Figure 18 highlights an important point regarding observation times. You must carefully choose your observation time based on your research question. For example, if you are curious how quickly you can fill the bathtub in Figure 18, you wouldn't want to observe the system from minutes 3 to minute 5 (when the system is losing). Rather, you would want to match your observation time to the expected dynamics of the system. For example, from minute 1 to minute 2, the system is gaining— this could help you infer how quickly you could fill the tub. Or perhaps you are curious if the drain plug is fully watertight. To answer this question, you should observe the tub from minute 3 to 4 to see if the tub remains static or if your plug is leaking water. Keep in mind that to understand if a system is transient, you need to make multiple (at least two) measurements in time. If you only looked at the system during minute 2 to see if the tub is filling, you might reasonably assume that the system is in a steady state. Therefore, you must observe the system *between* minute 1 and minute 2 to determine if it is gaining. Always choose your observation time based on your research question and make multiple measurements for transient systems.

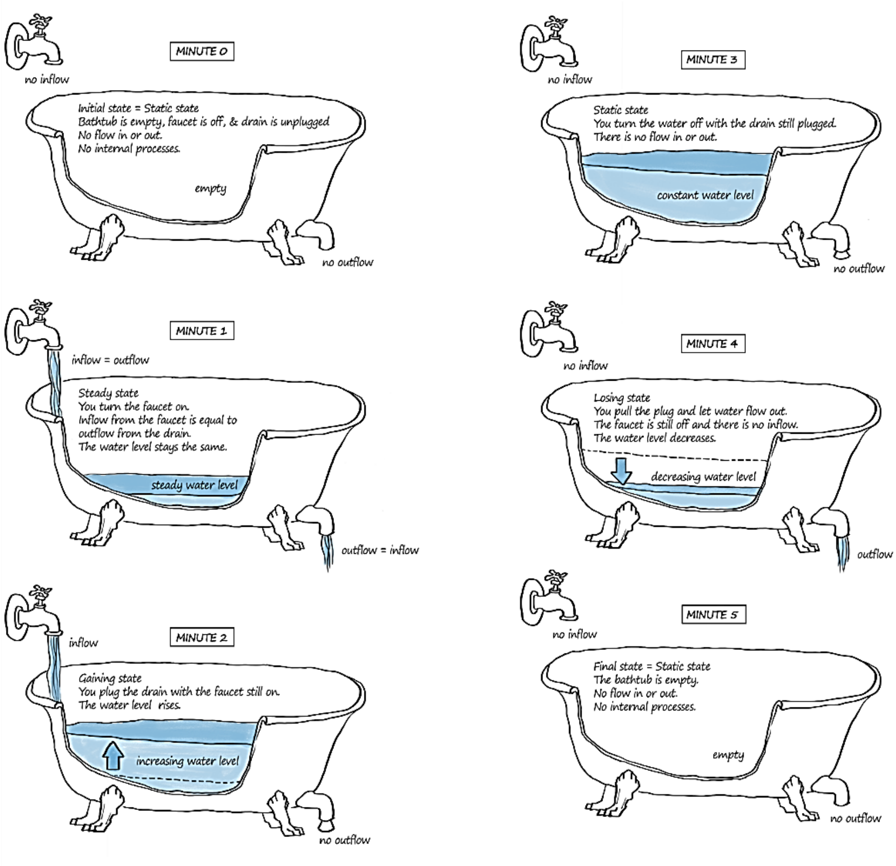

Figure 18. A bathtub system over the course of five minutes from initial static state (empty) to final static state (empty). The state variable is the water level in the tub.

System Examples

To further understand system states, let's look at a few examples. Earlier in this chapter, we briefly mentioned the inputs and outputs of a digestive tract.

Let's take a closer look at the digestive system, specifically for a bear. The state variable that we will observe is the bear's body mass. When more food is going into the bear's digestive system than is being converted to waste, the bear's mass is gaining. In the summer and fall prior to hibernation, a bear does this to gain mass as muscle or fat (Figure 19). In other words, prior to hibernation the bear is experiencing an increase in storage. As a bear hibernates in the winter, it burns some of its body mass as heat (e.g., a bear's metabolism converts mass to energy) (Figure 19). The bear loses its stored mass and the system state is losing. Note: this may be more than you want to know about bears, but they generally do not urinate or defecate during hibernation.

It may seem relatively simple to figure out the states of a bear leading up to and during hibernation (gaining and losing, respectively). But, what about during the rest of the year? If the bear is not gaining or losing mass, is it at steady state? Well, that depends on how often you measure. If you measure every minute, then no; a bear is never ingesting the exact same amount as it is digesting. After a bear eats, the molecules are digested and potentially stored; they are not fully released across the system boundaries (skin of the bear) until hours (or even months) later as CO_2 or scat. But keep in mind — steady state does not have to happen instantaneously. All system states are dependent on your observation time. Therefore, if you only measured the bear's weight once a year, then (just like viewing the water level in your bathtub from minute 0 to minute 5 in Figure 18), if the bear maintained the same mass from year-to-year, the bear would appear to be in a steady state over the course of a year. To visualize this concept, view the illustrations in Figure 19. *Note:* in contrast, to an adult bear that maintains its same mass from year-to-year, a bear cub would be in a gaining state (even on an annual basis) until it is grown.

Let's look at one last example. On Manhattan Island, New York, there is a complex network of tunnels, railroads, subways, ferry systems, walkways, and bike lanes that make up the transportation system that allows people to move into and out of Manhattan. This is necessary because there are far more people working in Manhattan than living there. The population of the island nearly doubles during the workday! The weekday population is approximately 4 million during the day and is approximately 2.05 million at night, all within the same day.

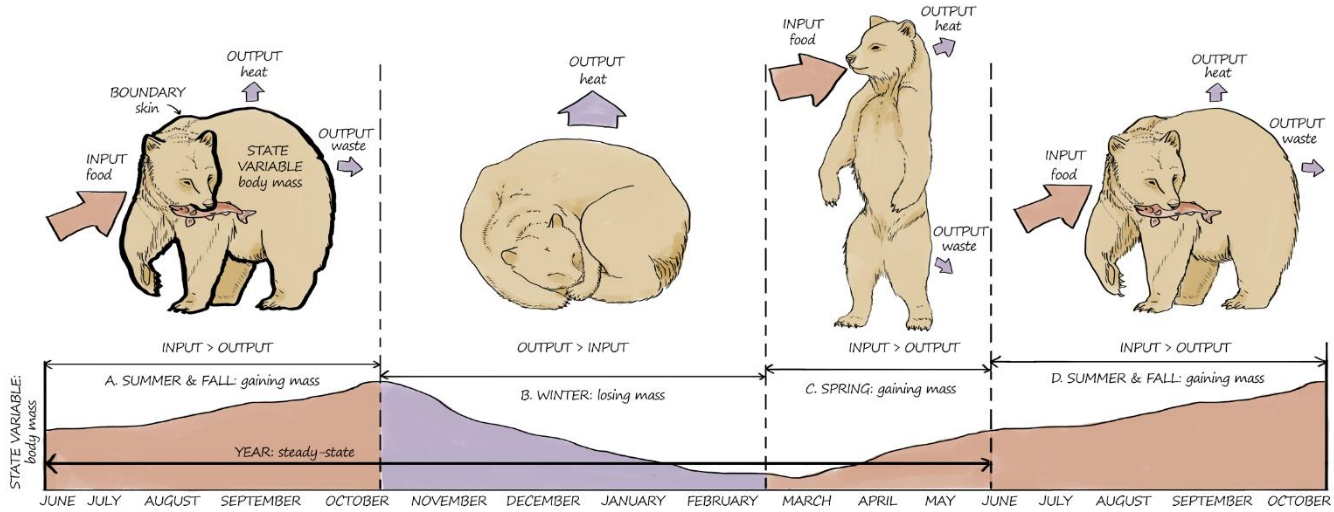

Figure 19. System diagram for a bear. Its state variable is its body mass. During the spring, summer and fall the bear is gaining mass (inputs > outputs). During hibernation in the winter the bear is losing mass (outputs > inputs). Over the course of the entire year, the bear returns to its original mass and is in a steady state.

If you were to look at the number of people (i.e., the state variable) in Manhattan between 9:00 am and 9:15 am, the system would be gaining; more people are coming onto the island than are leaving. Remember, you need multiple observation times (e.g., 9:00 am and 9:15) to observe a change in storage. The opposite would be true between 5:00 pm and 5:15 pm, more people would be leaving than are coming onto the island. However, if you looked over the course of 24 hours (10:00 pm on Tuesday and 10:00 pm on Wednesday); generally, the same number of people who come onto the island also leave the island. This is because all the commuters go home (off the island) to sleep. Therefore, the Manhattan population is generally steady state from 10 pm one day to 10 pm the next.

Let's analyze steady state for Manhattan Island a little more closely. Is the Manhattan population *always* in a steady state during a 24-hour time period? Compare the interval from Tuesday at 10:00 pm to Wednesday at 10:00 pm, to the interval from Friday at 12:00 pm to Saturday at 12:00 pm. Both are 24-hour observation periods, but is the system steady state for both? Who is coming into the city and who is leaving during the two intervals? Well, during the week (Monday-Friday) commuters are coming onto the island and leaving the island — this could be considered a consistent population fluctuation over a 24-hour period. However, on the weekends, tourists and visitors (non-residents) come onto the island for Broadway shows, shopping in Times Square, and picnics in Central Park. Therefore, the number of people on the island at noon on a Friday (during the week), is likely to be different than the number of people who are on the island at noon on Saturday. The system is not in steady state from Friday at 12:00 pm to Saturday at 12:00 pm. We won't say this enough. Observation times matter!

For the Manhattan population system to be gaining or losing during a defined observation time, the residential population (the number of people that live and sleep in Manhattan) must increase or decrease. For example, between 2000 and 2010 the residential population of Manhattan increased by approximately 3%. Therefore, the Manhattan population was gaining for the observation period of 10 years (between 2000 and 2010). This shows how a system can be gaining over one observation period (10 years) but can be approximately steady state for a different observation period (24 hours).

Remember how observation times are related to your research question? For the Manhattan example, if a city planner is curious about the population growth in Manhattan to ensure future water supply — they could choose an observation time of multiple years. If they want to balance sales tax versus property tax, they could choose an observation time that helps better quantify tourist and resident populations. There are many reasons to observe a system over different time intervals, but the researcher must choose what to measure, how often to measure, and for how long to measure. For your own research, you (the researcher) must make these decisions carefully to answer the specific questions that you want to address.

Making It Real

In all the previous examples, you defined the most obvious (physical) system boundaries. You imagined boundaries and put physical limits on the internal processes. The processes of a bathtub draining/filling, the bear's digestion, and citizens moving/working in Manhattan – they all functioned within the boundaries of the system: the porcelain walls of the bathtub (Figure 14), the skin of the bear (Figure 19), and the shoreline of Manhattan Island. But remember that boundaries are simply *conceptual*. They do not have to be as physically obvious as porcelain, skin, or a shoreline. They can be defined elsewhere. For example, although you chose to look at the entire digestive system of a bear, you could choose different boundaries to look at certain organs of the bear. Previously, the boundaries were the skin of the bear, but you could change the boundaries to a limited set of organs or one specific organ and this would change which inputs, outputs, and internal processes are relevant.

There are several reasons that you might change the boundaries of your system. Perhaps you are interested in understanding the function of one part of a system in greater detail. If there is a malfunction happening in a bear's stomach, you might want to limit the boundaries to the stomach. If the liver is not working properly, perhaps you would look at inputs, outputs, and processes specifically within the liver. Similarly, if you want to understand the population fluctuations in Greenwich Village (a historic neighborhood on the west side of Manhattan), rather than more broadly studying the changes in the entire population of Manhattan, you could change the boundaries to 14th Street, Broadway, Houston Street, and the Hudson River (the streets bounding Greenwich Village). Once you change your defined boundaries, the inputs, outputs, and internal processes of the system also change to fit the new limits of your system.

Now things are getting a bit circular. As we discussed above, the choice of boundaries is entirely up to you. But there is no point in defining boundaries unless you know the conditions on them. For example, if you do not understand the conditions at the Manhattan shoreline (the number of people coming to and from the island), you can't have a correct count of the population and how it is changing with time. The fluctuations at the defined boundaries are where it all happens. Similarly, if you were interested in changes on Manhattan Island but measured fluctuations along the boundaries of Greenwich Village (rather than the Manhattan shoreline), you would get a different answer... to a different question. For this reason, we often choose boundaries where we can define the *boundary conditions*.

Lastly, where you define system boundaries can also determine if a system is open or closed. The population fluctuation on Manhattan Island is a typical open system – there are flows of people across the boundary. What would make this a closed system? If all the roads, tunnels, bridges, boats, and planes into and out of the city were stopped for one day. Chaos! People would still move around in the city, but the system would be closed as we have defined our boundaries. What about making the system isolated? All the routes of transportation would need to be closed and all forms of communication outside of the city blocked – phones, email, letters, all of it. People could only move around inside the city and talk to each other face to face (gasp!).

Mass Balance

Hopefully, you are starting to understand that systems are all around us. Even your checking account is a system (state variable = account balance). If you are spending the same amount of money that you deposit, your checking account is in a steady state. If you are saving money, your checking account is gaining. If you are spending more than you are earning, your checking account is losing. This of course all depends on the observation time. Your account might be in steady state from the beginning of the month to the beginning of the next month, but it may vary between gaining, losing and steady state from day to day.

To keep track of changes in your checking account, you can review your payment history online. What is coming in and what is going out during the observation time? However, your payment history is just another name for something we more generally call the mass balance of a system. Money does not magically appear in bank accounts without coming from somewhere, and it doesn't disappear from bank accounts without going somewhere (although it may sometimes feel like it). Changes in other systems are similar and we can keep track of those changes using a *mass balance*. For example, if a pollutant in a lake increases, the pollutant comes from somewhere. It is either carried into the lake from someplace else or produced via a chemical reaction of existing compounds that were already in the lake. This is the idea behind a mass balance – mass is conserved and it can be tracked!

We can quantify changes in a system using a *mass balance equation*. The state variable is the total mass of a substance. For a mass balance equation, all inputs are positive (mass added to the system), and all outputs are negative (mass removed from the system).

$$Final\ Condition = Initial\ Condition + (Inputs) - (Outputs)$$

This can appear complicated, so let's take a closer look. For the mass balance equation of a checking account, all deposits are positive (mass added to the system), and all withdrawals are negative (mass removed from the system).

At the beginning of August, the initial balance (initial condition) in your checking account is $350. You get two paychecks (inputs) throughout the month of $800 each. Your birthday is August 5th, and your grandma gives you $50 (input). You spend $400 on food, your rent is $700, electric bill is $100, phone bill is $75, car insurance is $70, and health insurance is $400 (all outputs). You want to know the state of your account from the beginning of the month (initial condition) to the end of the month (final condition).

$$Final\ Balance = Initial\ Balance + (Paycheck + Birthday\ Money) - (Food + Rent + Electric\ Bill + Water\ Bill + Phone\ Bill + Car\ Insurance + Health\ Insurance)$$

Using the mass balance equation, we can quantify whether our system is gaining, losing, or steady state. For a steady state system, the state variable is constant over the defined observation period. Therefore, if you want to quantify that a system is in steady state, the state variable must have a constant value. Stated another way, the input exactly balances the output and there is no change in storage (Table 2). Note that this does not necessarily mean that you are earning zero and spending zero, although that is one kind of steady state (i.e., static condition, Table 2). For transient systems (e.g., gaining and losing), the input is different than the output and the total mass (state variable) changes with time (Table 2).

Let's further construct a mass balance equation for your checking account to determine if the account is gaining, losing, or steady state.

$$Final\ Balance = \$350 + (\$800 + \$800 + \$50) - (\$400 + \$700 + \$100 + \$75 + \$70 + \$400) = \$255$$

Your checking account balance is a positive value – that must mean that your checking account is gaining, correct? Well, not necessarily. If you consider the initial time as the beginning of the month and the final time as the end of the month, you must look at the state variables at those two times to determine the system state. The initial balance is $350, and the final balance is $255. Both are positive values, but the change between the two is negative (-$45). This means there was a *loss of storage* in the system and the account is losing (Table 2).

Here, again, the boundaries matter. Imagine that your observation time is a month. If you earn $2,000 throughout the month and spend exactly $2,000 that month, your checking account is steady state. However, let's say you also transfer $500 (of the $2,000) from checking to savings. If the boundaries are still around your checking account, then the $500 transfer is an output from the checking system and the system is transient (losing). However, if your boundaries are around your entire bank holdings (checking and savings) then the total mass in your bank system is steady state (no change in storage, you still have $2,000 dollars in the bank).

Next, let's use mass balance to understand one of our previous systems – a bathtub. What is the mass balance equation for a bathtub? Think about the deposits (inputs) and withdrawals (outputs) for this system. Previously, the state variable that we used for this system was water level (Figure 16). For the mass balance equation, we are going to assume that the water level in the tub is an indicator of the amount of water stored in the tub, which is the volume.

Answers:

- Boundaries: porcelain walls of the tub
- Initial condition: volume of water in the tub prior to inputs/outputs
- Inputs: volume of water from faucet
- Outputs: volume of water down the drain
- Mass Balance Equation:

$$Final\ tub\ volume = initial\ tub\ volume + volume\ from\ faucet - volume\ down\ the\ drain$$

Note: for mass balance equations, all variables must have the same units. For example, for the bathtub example above all the variables are volumes of water (L^3).

Hydrologic Systems

Introduction

The *hydrologic cycle* is the continuous movement of water in the Earth-atmosphere system. If you wanted to create a hydrologic mass balance for the valley in which you live, that could sound overwhelming. How do you quantify the hydrologic components and processes for the valley? What are the relevant processes, boundaries, and observation times? Before you get too overwhelmed, remember how we changed the boundaries of a bear's digestive system to study the stomach, specifically? Or how we changed the boundaries within Manhattan Island to Greenwich Village to understand the neighborhood's population fluctuation? Well, we can do the same thing with the water cycle. We can define the boundaries within the Earth-atmosphere system to isolate a smaller hydrologic system.

Hydrologic systems can range from an entire continent to the area around a specific lake or river, or even to a small sample of soil within an agricultural field. The size of a hydrologic system should match the question that you are asking and the observations that you can make. You can't draw conclusions regarding the hydrologic function of the Nile River with a system that is limited to one neighborhood in Cairo, Egypt. Similarly, you can't use five observations in a cubic centimeter of soil to analyze the hydrogeology of the Colorado River Basin. You can begin to see why defined boundaries are important – and so difficult to define! Throughout this section we will gradually and systematically introduce concepts that you need to understand hydrologic systems. You already have basic terminology for systems. Now, it is time to apply the terminology to water and to learn some new hydrologic terms and concepts.

Clay-Lined Lake

Our first example of a hydrologic system is a lake with an impermeable clay liner, one inlet stream, and one outlet stream (Figure 20). *Impermeable* is a novel word, but it uses terminology that you have already learned. Remember that *permeability* is the ability of a material to transmit a fluid. Therefore, a soil that is impermeable does not transmit fluid. In this case, we will assume the water cannot flow through the clay liner; the boundaries of the system are the surface of the water and the top of the liner. What does this hydrologic system remind you of? It behaves much like the bathtub system introduced above (Figure 16). The clay layer (with an exceptionally low permeability) behaves like the porcelain walls of the bathtub and the inlet and outlet streams like the faucet and drain, respectively.

The mass balance equation for a bathtub is

$$\begin{aligned} &Final\ tub\ volume \\ &\qquad = initial\ tub\ volume \\ &\qquad + volume\ from\ faucet - volume\ down\ the\ drain \end{aligned}$$

Therefore, the mass balance for the impermeable clay-lined lake is

$$\begin{aligned} &Final\ lake\ volume \\ &\qquad = initial\ lake\ volume \\ &\qquad + volume\ from\ inlet\ stream - volume\ from\ outlet\ stream \end{aligned}$$

Note: as with the bathtub systems, all the variables in the mass balance for the clay-lined lake are volumes of water.

Keep in mind that for the bathtub example (Figure 16), we only considered water leaving or coming into the tub via the faucet and the drain. However, a lake system is exposed to more external processes; water isn't only entering and leaving the lake via the inlet and outlet streams ("the faucet and the drain"). Perhaps water is falling onto the lake as precipitation or leaving via evaporation. *Precipitation* is the water that condenses and falls from clouds due to gravity. You are likely familiar with many forms of precipitation such as drizzle, rain, sleet, snow, and hail. *Evaporation* is the process of converting liquid water into vapor due to an increase in temperature and/or a decrease in pressure. Over the course of 24 hours, the sun rises and sets and the amount of solar energy going into a hydrologic system rises and falls, driving weather patterns – this causes water to leave a system via evaporation (when the temperature rises), or enter a system via precipitation (when there is rain or snow) (Figure 20).

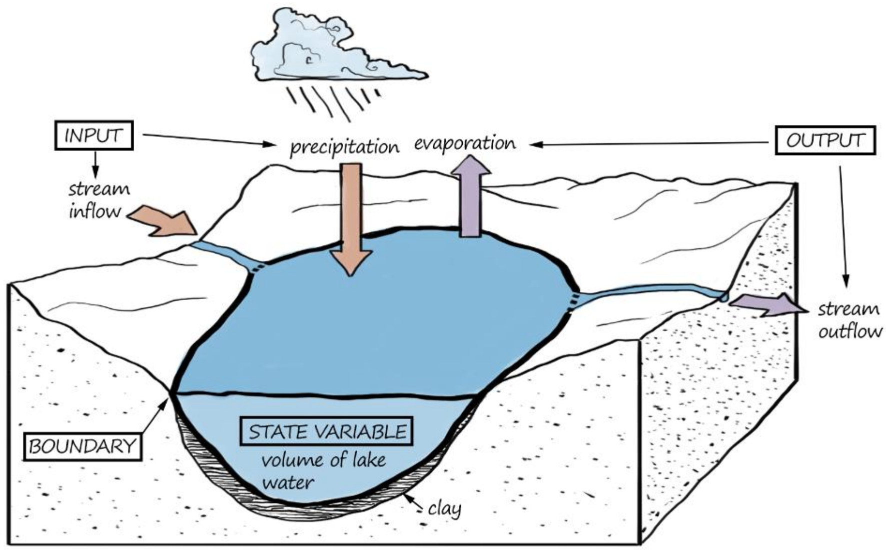

Figure 20. System diagram of a lake with inputs and outputs.

If we consider precipitation and evaporation for our clay-lined lake, these additional processes lead to a slightly more complex mass balance equation for the lake system:

$$\begin{aligned} Final\ lake\ volume \\ &= initial\ lake\ volume \\ &+ (volume\ from\ precipitation \\ &+ volume\ from\ inlet\ stream) \\ &- (volume\ from\ evaporation \\ &+ volume\ from\ outlet\ stream) \end{aligned}$$

Remember, for a mass balance equation we need all the variables (inputs and outputs) to be in the same units. In this case, we want volumes of water. The stream flow is usually reported as volume per time (e.g., cfs). Therefore, we must multiply the stream flow values by the duration of the observation period to convert them to volumes. For precipitation and evaporation, we often receive rates (e.g., mm/day). To include these inputs/output rates, we need to first multiply the rate by the area of the lake and then multiply by the duration of the observation period.

There are two inputs into the lake (inlet stream and precipitation) and two outputs (outlet stream and evaporation) (Figure 20). Inputs and outputs control the state of the system and collectively determine the mass balance. Because the inputs and outputs are independent of each other, there are many different possible scenarios for the same mass balance equation; some in which the lake is losing, others gaining, and others steady state. The only way to know… is to do the calculation. Of course, there is another way to figure out whether the lake system is gaining or losing. Simply look at the volume of water at one time, then again later (as we did for the bathtub water level in Figure 18). If the volume increases, the lake is gaining – if the volume decreases, the lake is losing – if the volume stays the same, the lake is steady state.

Let's take a closer look at a clay-lined lake system (e.g., Figure 21). Table 3 provides hourly values for the inputs (inlet stream and precipitation) and outputs (outlet stream and evaporation) of the system. Hourly values are the volume of water that goes into/out of the lake during each hour (ft^3/hr). The net flow (ft^3/hr) is the sum of all the inputs and outputs for each hour; positive values represent inputs > outputs, negative represent inputs < outputs. The lake volume (ft^3) is the total volume of the lake at the end of each hour. Notice, all the input and output rates have the same units (ft^3/hr) in Table 3, we did all the calculations and conversions for you! Lastly, the summation of the columns is included at the very bottom of the table (Observation Time Total (ft^3)). This shows how the total inputs and outputs provide a net flow of zero for the observation period. Figure 22 shows the net flow and lake volumes through time. The initial and final volumes are noted near the top and bottom of Table 3 (initial, final) and as large dots in Figure 22.

The inlet stream has the highest flow from 12:00 pm to 5:00 pm and the outlet stream is highest from 2:00 pm to 4:00 pm (Table 3). There is a monsoon storm that occurs from 2:00-3:00 pm which provides the area with a high level of precipitation. Additionally, the highest evaporation values are during the hottest time of the day from 12:00 pm-7:00 pm. Notice how the volume of the lake initially is 50,000 ft^3 (see initial time, 7:00 am).

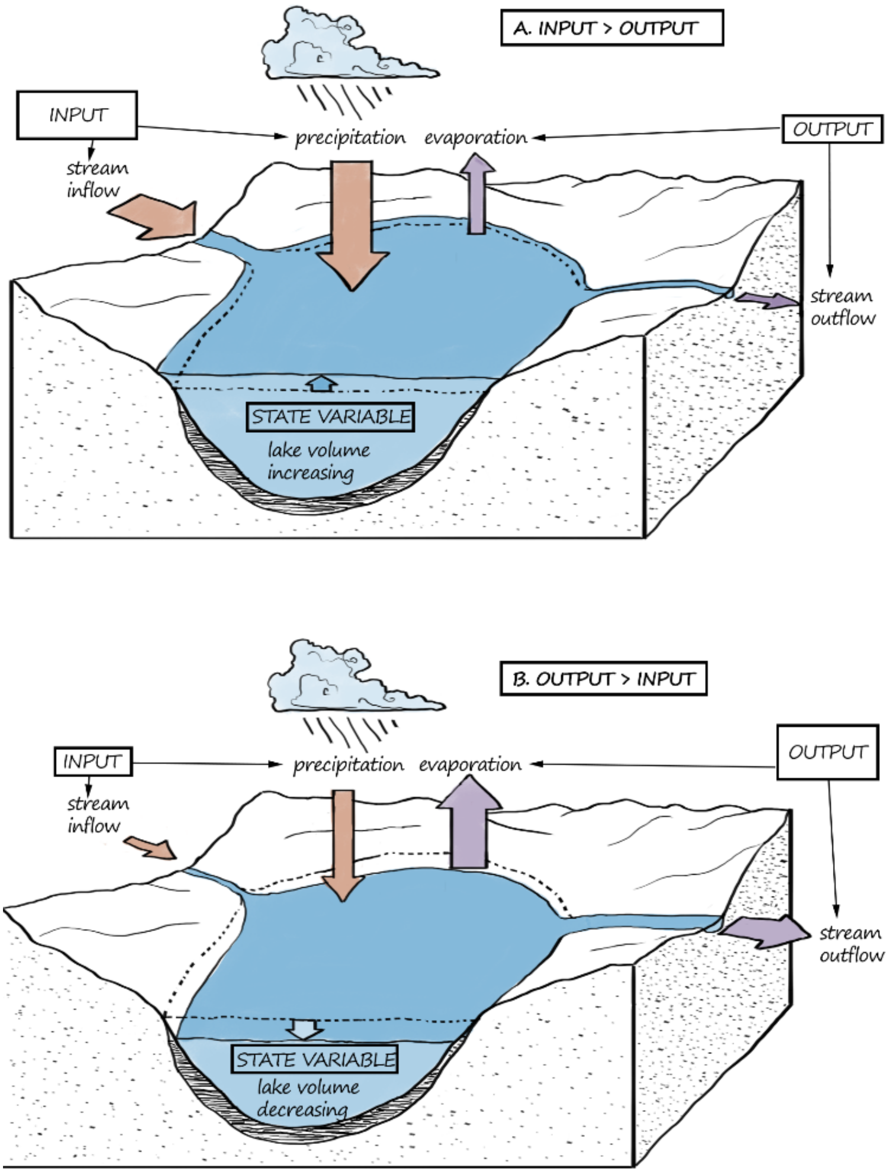

Figure 21. A lake system under two different transient states: (a) gaining and (b) losing.

Table 3. Hourly values for a clay-lined lake system. Inputs include an inlet stream and precipitation. Outputs are an outlet stream and evaporation. All input/output values are given in ft^3/hr. Inputs are positive. Outputs are negative

Time	Inlet Stream (ft^3/hr)	Outlet Stream (ft^3/hr)	Evaporation (ft^3/hr)	Precipitation (ft^3/hr)	Net Flow (ft^3/hr)	Lake Volume (ft^3)
7:00 AM (Initial)					0	50,000
8:00 AM	435,600	-399,600	-36.18	0.00	35,964	85,964
9:00 AM	435,600	-403,200	-37.83	0.00	32,362	118,326
10:00 AM	439,200	-406,800	-42.76	0.00	32,357	150,683
11:00 AM	439,200	-399,600	-46.05	0.00	39,554	190,237
12:00 PM	576,000	-414,000	-52.63	0.00	161,947	352,185
1:00 PM	720,000	-424,800	-60.86	0.00	295,139	647,324
2:00 PM	918,000	-900,000	-70.72	639.80	18,569	665,893
3:00 PM	756,000	-756,000	-74.01	261.51	188	666,080
4:00 PM	540,000	-536,400	-75.66	0.00	3,524	669,605
5:00 PM	507,600	-435,600	-67.43	0.00	71,933	741,537
6:00 PM	493,200	-439,200	-64.14	0.00	53,936	795,473
7:00 PM	432,000	-478,800	-54.28	0.00	-46,854	748,619
8:00 PM	439,200	-504,000	-46.05	0.00	-64,846	683,773
9:00 PM	435,600	-558,000	-32.89	0.00	-122,433	561,340
10:00 PM	439,200	-576,000	-16.45	0.00	-136,816	424,523
11:00 PM	435,600	-558,000	-14.80	0.00	-122,415	302,109
12:00 AM	439,200	-547,200	-13.16	0.00	-108,013	194,095
1:00 AM	435,600	-504,000	-4.93	0.00	-68,405	125,690
2:00 AM	439,200	-468,000	-4.93	0.00	-28,805	96,886
3:00 AM	439,200	-468,000	-3.29	0.00	-28,803	68,082
4:00 AM	432,000	-435,600	-1.64	0.00	-3,602	64,481
5:00 AM	428,400	-432,000	-1.64	0.00	-3,602	60,879
6:00 AM	428,400	-432,000	-16.45	0.00	-3,616	57,263
7:00 AM	430,200	-432,000	-29.61	0.00	-1,830	55,433
8:00 AM (Final)	428,400	-433,800	-32.89	0.00	-5,433	50,000
Observation Time Total (ft^3)	12,342,600	-12,342,600	-901.32	901.32	0	

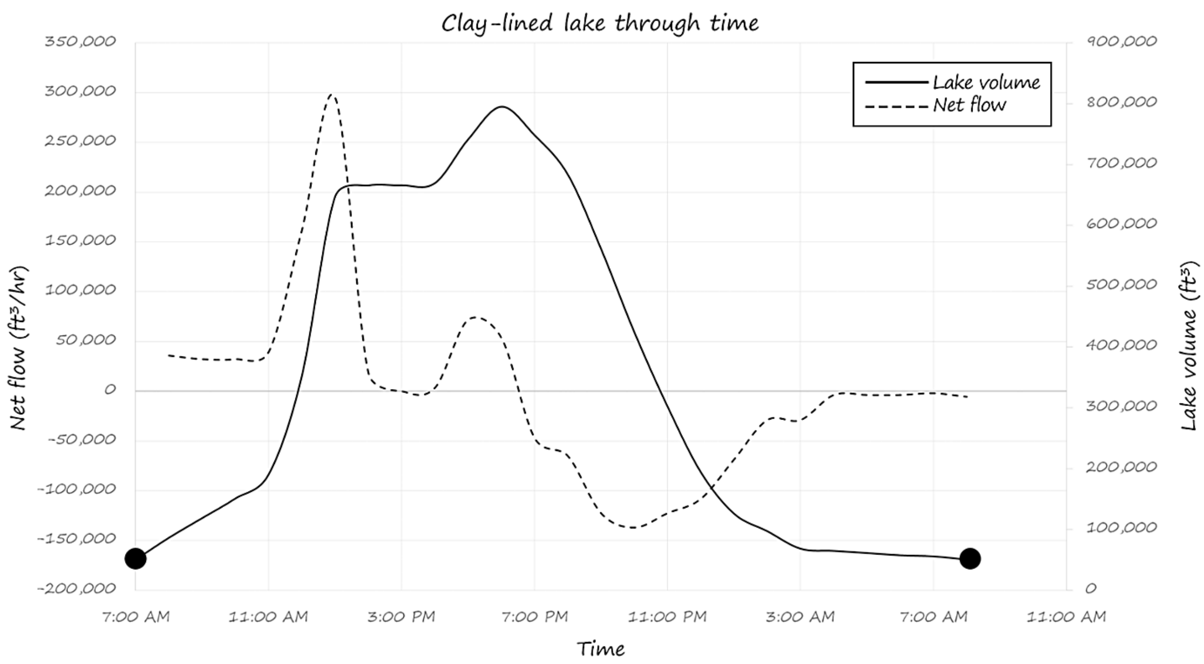

Figure 22. Clay-lined lake through time. The dashed line is the net flow, and the solid line is the lake volume (Table 3). The large dots represent the initial and final volume in the lake (equal to 50,000 ft^3).

Use the inputs, outputs, and state variable (i.e., the lake volume) in Table 3, to define the system state during the following time intervals:

1. From 7:00 am (the first day) to 8:00 am (the next day)
2. From 8:00 am to 6:00 pm?
3. From 6:00 pm to 7:00 am?

Answers:

1. Steady state
2. Gaining
3. Losing

The inputs and outputs vary throughout the day, and so does the state variable (the lake volume). At 7:00 am on the first day (initial) the volume of the lake is 50,000 ft^3. At 8:00 am the next day (final), the volume of the lake is once again 50,000 ft^3. What does this mean? It means that the lake is in a steady state during this time period. However, if you only observe the lake from 8:00 am to 6:00 pm the input is greater than the output, and the system is gaining as the total water volume in the lake increases. From 6:00 pm to 7:00 am the system is losing, as the total water volume in the lake decreases. This shows how the system isn't steady state during all time intervals, much like the bathtub example above (Figure 18). This is quite common. The state of hydrologic systems often changes over time and is dependent on the observation period.

Permeable Lined Lake

A lake lined with an impermeable clay liner (Figure 20) is one of the simplest hydrologic systems we could imagine. So, let's make the system a bit more complex. What would happen if we added another output to the system? Each time you add inputs or outputs to a hydrologic system, you are complicating the mass balance equation. Just as when you add more sources of income (inputs) or authorized users (outputs) to a checking account (e.g., a family versus an individual checking account) – it complicates the task of tracking the balance.

What happens if the liner of our lake is *permeable*? Water can leave the lake via infiltration through the sides and the bottom boundary. *Infiltration* is when water seeps into subsurface soils from the ground surface through cracks and pore spaces. Imagine a bathtub made of a cloth that allows water to seep out of it, slowly. If we remove the porcelain and replace it with cloth, we keep the same boundaries – the walls, sides, and the top of the bathtub. However, the boundary condition changes. For the porcelain walls, there was no flow, anywhere along the boundary, for all of time. For the cloth walls, the flow is low, but constant, over time. In other words, the flow out of the walls and bottom of the tub is no longer zero. Our mass balance has an extra output term! This is the same for a lake with a permeable liner.

What is the mass balance equation for the permeable lake system? Remember that outputs are negative values and inputs are positive values.

$$\begin{aligned} \textit{Final lake volume} \\ &= \textit{initial lake volume} \\ &+ (\textit{volume from precipitation} \\ &+ \textit{volume from inlet stream}) \\ &- (\textit{volume from outlet stream} \\ &+ \textit{volume from evaporation} \\ &+ \boldsymbol{volume\ from\ infiltration}) \end{aligned}$$

Notice the equation is almost exactly the same as for our clay-lined lake, with one added term (infiltration). This seems relatively simple. From here, we just use the volume from infiltration, plug it into the equation, and calculate the mass balance, right? Well, what if I told you that it is difficult (maybe even impossible) to measure infiltration directly? Is that it? Game over? Not necessarily. Let's brainstorm. How could you use the mass balance equation (in a different way) to infer the infiltration? Here's a hint – what if you could *measure* the change in the volume of the lake over a time period (i.e., measured lake volume, Table 4), as well as everything else except the infiltration?

The predicted lake volume in Table 4 is the volume from the mass balance equation using all the known inputs/outputs (this is the same as the lake volume in Table 3).

$$\begin{aligned} \textit{Predicted lake volume} \\ &= \textit{initial lake volume} \\ &+ (\textit{volume from precipitation} \\ &+ \textit{volume from inlet stream} \\ &- (\textit{volume from outlet stream} \\ &+ \textit{volume from evaporation}) \end{aligned}$$

Table 4. Input and output values for the lake system shown in Table 3. This table shows the predicted lake volume (same as the lake volume in Table 3) and the measured lake volume. The measured lake volume is lower than the predicted lake volume due to infiltration. Infiltration values can be inferred by subtracting the measured volume from the predicted volume

Time	**Inlet Stream (ft^3/hr)**	**Outlet Stream (ft^3/hr)**	**Evaporation (ft^3/hr)**	**Precipitation (ft^3/hr)**	**Predicted Lake Volume (ft^3)**	**Measured Lake Volume (ft^3)**	**Infiltration (ft^3/hr)**	**State**
7:00 AM (Initial)					50,000	50,000		**Initial**
8:00 AM	435,600	-399,600	-36.18	0.00	85,964	85,884	-80	Gaining
9:00 AM	435,600	-403,200	-37.83	0.00	118,326	118,134	-112	
10:00 AM	439,200	-406,800	-42.76	0.00	150,683	150,347	-144	
11:00 AM	439,200	-399,600	-46.05	0.00	190,237	189,717	-184	
12:00 PM	576,000	-414,000	-52.63	0.00	352,185	351,318	-346	
1:00 PM	720,000	-424,800	-60.86	0.00	647,324	645,816	-641	
2:00 PM	918,000	-900,000	-70.72	639.80	665,893	663,726	-660	
3:00 PM	756,000	-756,000	-74.01	261.51	666,080	663,254	-660	
4:00 PM	540,000	-536,400	-75.66	0.00	669,605	666,115	-663	
5:00 PM	507,600	-435,600	-67.43	0.00	741,537	737,312	-735	
6:00 PM	493,200	-439,200	-64.14	0.00	795,473	790,458	-789	

Table 4. (Continued)

Time	Inlet Stream (ft^3/hr)	Outlet Stream (ft^3/hr)	Evaporation (ft^3/hr)	Precipitation (ft^3/hr)	Predicted Lake Volume (ft^3)	Measured Lake Volume (ft^3)	Infiltration (ft^3/hr)	State
7:00 PM	432,000	-478,800	-54.28	0.00	748,619	742,862	-742	Losing
8:00 PM	439,200	-504,000	-46.05	0.00	683,773	677,338	-678	
9:00 PM	435,600	-558,000	-32.89	0.00	561,340	554,350	-555	
10:00 PM	439,200	-576,000	-16.45	0.00	424,523	417,116	-418	
11:00 PM	435,600	-558,000	-14.80	0.00	302,109	294,405	-296	
12:00 AM	439,200	-547,200	-13.16	0.00	194,095	186,204	-188	
1:00 AM	435,600	-504,000	-4.93	0.00	125,690	117,680	-119	
2:00 AM	439,200	-468,000	-4.93	0.00	96,886	88,784	-91	
3:00 AM	439,200	-468,000	-3.29	0.00	68,082	59,919	-62	
4:00 AM	432,000	-435,600	-1.64	0.00	64,481	56,259	-58	
5:00 AM	428,400	-432,000	-1.64	0.00	60,879	52,603	-55	
6:00 AM	428,400	-432,000	-16.45	0.00	57,263	48,935	-51	
7:00 AM	430,200	-432,000	-29.61	0.00	55,433	47,056	-49	
8:00 AM (Final)	428,400	-433,800	-32.89	0.00	50,000	41,580	-44	**Final**
Observation Time Total (ft^3)	12,342,600	-12,342,600	-901.32	901.32			-8,420	

Compare the two volumes (measured volume and predicted volume) in Table 4 and notice how they are different. Here is where the real value of mass balance calculations lies. For the permeable lake, the predicted lake volume is always greater than the measured lake volume. In this case, we know that this is because there is an output that is not measured, infiltration. If we can be certain that infiltration is the only thing that we missed measuring, we can rely on the mass balance equation to infer this rate. We just rearrange the equation and solve for infiltration.

Or another way to write this is:

$$\boldsymbol{Volume\ from\ infiltration} = predicted\ lake\ volume - measured\ lake\ volume$$

To further explore the mass balance, let's explore potential materials for the permeable lake system. What type of material would need to be at the bottom of the lake for infiltration rates to be high? A material with a high hydraulic conductivity or a material with a low hydraulic conductivity?

Answer:

- For infiltration rates to be high, the hydraulic conductivity of the material at the bottom of the lake would need to be high. For example, sand would have a higher infiltration rate than clay because water flows more easily through sand. See Figure 9 for a comparison of hydraulic conductivity values of different materials.

Once the water infiltrates from the lake into the subsurface, where does it go? It doesn't just stay directly around the lake, does it? Depending on the amount of water in the soil, the water can continue to percolate through the subsurface. *Percolation* is when water moves through cracks and pores underground in the unsaturated zone (Figure 13). The difference between infiltration and percolation is subtle. Infiltration is like percolation. However, it occurs at the surface and percolation happens beneath the surface. Water will continue to percolate through the subsurface and can even travel all the way down to the saturated zone (Figure 13). The saturated zone is located at different depths in different environments; in this case the saturated zone is located at some depth below the lake (Figure 23). Regardless of where it is located, water that reaches the saturated zone is called *recharge*. This is a wonderful place to re-iterate the importance of system boundaries. Infiltration is when water crosses the air-soil boundary. Percolation is when water moves through unsaturated soil and doesn't cross any boundaries. Recharge is when water crosses the unsaturated/saturated boundary. Generally, the top of the saturated zone is represented by a small upside-down triangle (∇) (Figure 23). We will continue to use this symbol throughout the text. The surface or top of the saturated zone is referred to as the *water table*. Many non-experts think of the water table as the boundary between the saturated and unsaturated zones. In truth, it is often close to the top of the saturated zone. We will explain a more precise definition of the water table in a later chapter.

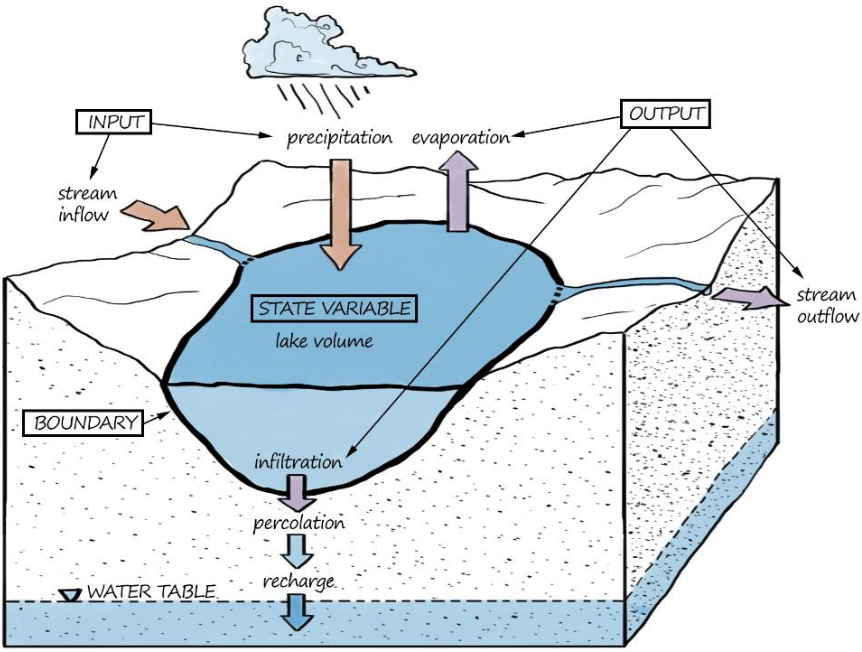

Figure 23. A lake system with multiple inputs and outputs, including infiltration as an output. The water table is deep underground and is represented by a small upside-down triangle (∇). Processes are the same as Figure 21. However, water also infiltrates through the lake bottom. Some of the water travels down and recharges the saturated zone.

Watersheds

Thus far, we have limited our hydrologic system to the boundaries of a lake. The lake system(s) above helped introduce basic hydrologic concepts such as infiltration and recharge, and further clarified the importance of system boundaries. But what if we wanted to study the processes occurring on the land surrounding the lake, and not just within the lake itself? Remember how we changed the boundary locations in the bear system? We first limited the boundaries to the skin, then the stomach or the liver. Similarly, for a hydrologic system, we can limit the boundaries to the lake, to the perimeter of the watershed, or even to a larger hydrologic system like an entire continent (Figure 24). Remember that you get to choose the boundaries of your system, but some boundaries are easier to deal with than others.

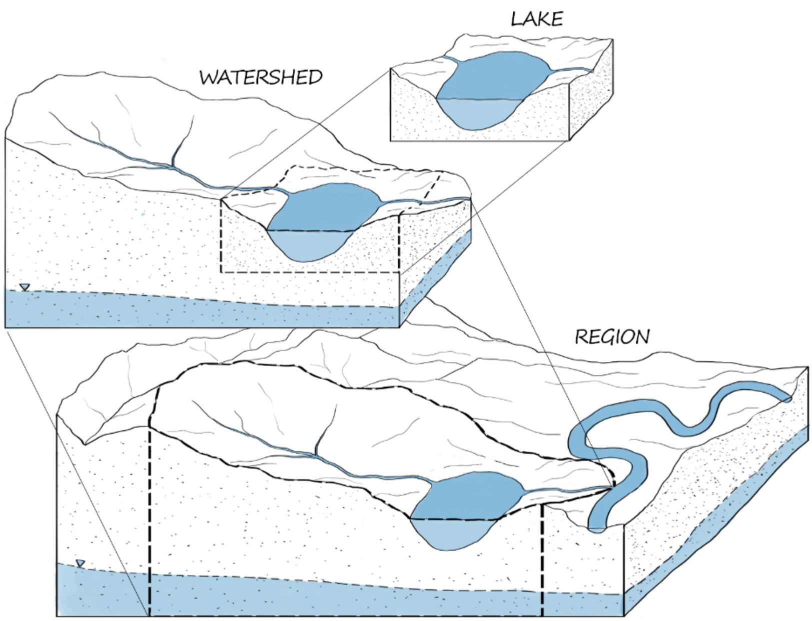

Figure 24. Hydrologic system with several nested sub-systems: a small watershed within the larger region and a lake within the small watershed.

A watershed or drainage basin is a geographical area where all the streams and rainfall drain to a common outlet point; everything upstream drains downstream (Figure 24). Watersheds can encompass a small, inland lake with a couple of streams or can cover thousands of square miles. The largest watershed in the United States is the Mississippi River Watershed, which extends across 31 U.S. states and two Canadian Provinces. This means that water that falls anywhere in this watershed could ultimately end up in the Mississippi River. We say, “could end up”, because not all water that falls as precipitation runs off directly into rivers. Can you guess some of the reasons why? What natural processes and human interventions prevent precipitation from running off into rivers? There are some natural processes such as evaporation that happen prior to the water reaching the rivers. Furthermore, human impacts such as dams, diversions, and groundwater pumping can also interfere with water travel. Perhaps human interference could move the water from the Mississippi watershed to another watershed through diversion projects. Or groundwater could be pumped from the ground, bottled, and shipped to another location.

Our lake system (Figure 23) is part of a small watershed (Figure 24). If we wish to expand the boundaries to include the perimeter of the watershed, we must look at the hydrologic processes that occur within those boundaries (Figure 25). For example, precipitation that falls on the land outside the watershed is not an input to the watershed system. Only the precipitation that falls *on or within* the watershed boundaries contributes to flow in the streams that feed the lake. Same goes for evaporation, infiltration/recharge, and evapotranspiration. *Evapotranspiration*, that might be an unfamiliar word. You might have heard (or rather experienced) *transpiration* before – this is when you lose water vapor through your sweat glands. Plants have a similar process and lose water vapor through pores in their leaves (i.e., evaporation of water from plant leaves). Have you ever hiked in a rain poncho? If you get too warm and begin to sweat, the poncho collects water vapor on the inside. Similarly, if we tied a plastic bag around the leaves of a house plant, we would get water vapor inside the bag from plant transpiration. However, if you tied a plastic bag around a wetted pot of soil (without the plant), the water vapor inside the bag would be from evaporation. *Evapotranspiration* is simply a summation of transpiration and surface evaporation: imagine tying a bag around the plant and the potted soil; the water vapor in the bag would be the evapotranspiration. You might imagine how on the surface of the earth it is difficult to separate these subtle processes (evaporation and transpiration). Because we don't always wish to separate these two processes (particularly on the land surface), we often lump them together in mass balances as evapotranspiration. In this case, we can still measure evaporation from surface water bodies separately (Figure 25) and add it to evapotranspiration from the land surface.

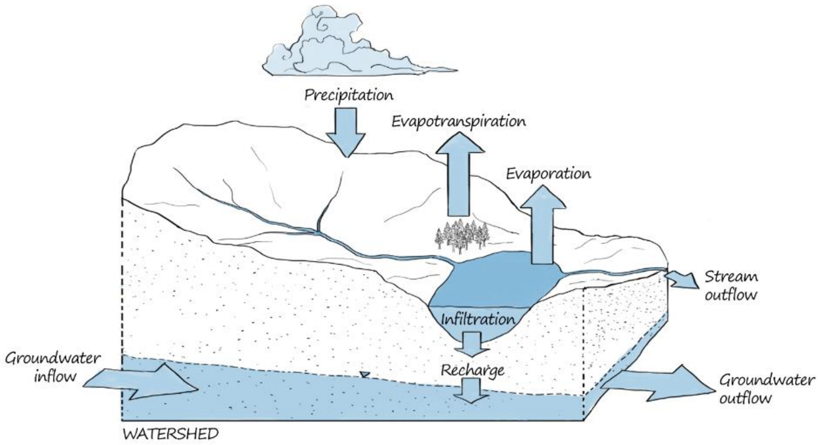

Figure 25. Mass balance of a watershed.

What is the mass balance of our watershed system in Figure 25? What are the inputs and outputs that occur within the defined boundaries?

Mass balance:

$$\begin{aligned} &Final\ volume\ of\ water\ in\ the\ watershed \\ &= initial\ stored\ volume \\ &+ (volume\ from\ precipitation \\ &+ volume\ from\ groundwater\ inflow) \\ &- (volume\ from\ evapotranspiration \\ &+ volume\ from\ evaporation \\ &+ volume\ from\ stream\ outflow \\ &+ volume\ from\ infiltration \\ &+ volume\ from\ groundwater\ outflow) \end{aligned}$$

Once again, system boundaries matter. The boundaries determine the components of the mass balance. For example, if we wanted to study the larger hydrologic system, which has multiple watersheds, the boundaries and inputs/outputs of the hydrologic system would need to be different than for a single watershed. In the same vein, for all the systems discussed, we chose the bottom boundary to be the water table. However, that meant we assumed that plants could not access water directly beneath the water table or draw water up from the water table. If the plants could access this water, we would need to include transpiration in the mass balance. Lastly, look at the left side of Figure 25. Sometimes, even when we carefully chose our system, we can have groundwater flowing across the watershed boundary (an input that is tough to account for). It is difficult, sometimes impossible, to define boundaries such that you know all the inputs and outputs exactly, all the time. You just must do your best to choose boundaries that will help you to answer your research questions.

Conclusion

The interconnected components and internal processes of systems might go unnoticed, that is until you decide to study them. Once you begin studying systems, you'll realize both their complexity and abundance (they are all around us!). We know that studying systems can feel overwhelming at first, which is why we suggest breaking them down using concepts such as: boundaries, state variable(s), observation times, inputs, outputs, and system states. If you continue studying hydrogeology, "systems thinking" or "systems theory" will be prevalent in your studies. What is the big deal with systems? Systems thinking is part of the approach that scientists use to create conceptual and computer models. Unfortunately, groundwater is not easily visible (it is underground after all), which makes it difficult to study. Therefore, groundwater models are useful for understanding concepts, predicting outcomes, and communicating ideas. Systems thinking will continue to come up throughout the book and throughout your hydrogeologic journey. We suggest coming back to this chapter regularly to aid in your understanding of groundwater.

What to Remember

Important Terms		
boundaries	inputs	recharge
closed system	isolated system	state variables
evaporation	mass balance	static
evapotranspiration	observation time	steady state
final condition	open	system
hydrologic cycle	outputs	system state
impermeable	percolation	transient
infiltration	permeable	transpiration
initial condition	precipitation	watershed

Chapter 3

Mechanisms of Groundwater Movement

Introduction

Water flows! Water slides send water from the top of the slide to the bottom. If you wash your car in the driveway, water travels down the pavement and into the street. Water moves from higher elevations to lower elevations in a streambed. And when it rains, water moves down the surface of a hillside. These are all similar examples, right? Well, there is a distinct difference between a water slide and a hillside. Or a driveway and a streambed. Can you think about what it might be? Think about the types of materials in the four examples. The plastic of the slide and the pavement of the driveway are impermeable. Water moves on top of the surface and downward because it doesn't have anywhere else to go. However, a streambed or a hillside can be permeable, allowing some of the water to infiltrate into the ground and some of the water to flow along the ground surface. Once the water infiltrates into the soil, where does it go?

Previously, we discussed how water percolates downward. But there is a complicating factor – it doesn't always move straight downward. For our lake example, the water underground can also move laterally, or side-to-side. One consequence of this is that groundwater is not always an output from a lake, as it was represented in Figure 23. Under some conditions, it can also be an input across the bottom of a lake. Groundwater inflow and outflow are shown in Figure 26. Generally, the water underground in Figure 26 is moving from left-to-right, but on the left it flows into the bottom of the lake, and on the right, it flows out of the bottom of the lake. The surface water and groundwater are intimately connected. Remember this! The interaction between groundwater and surface water is an important concept in hydrology. It is common for people to think of groundwater and surface water as separate – in some states, they are even different by law. However, in reality, groundwater and surface water are a single, connected resource.

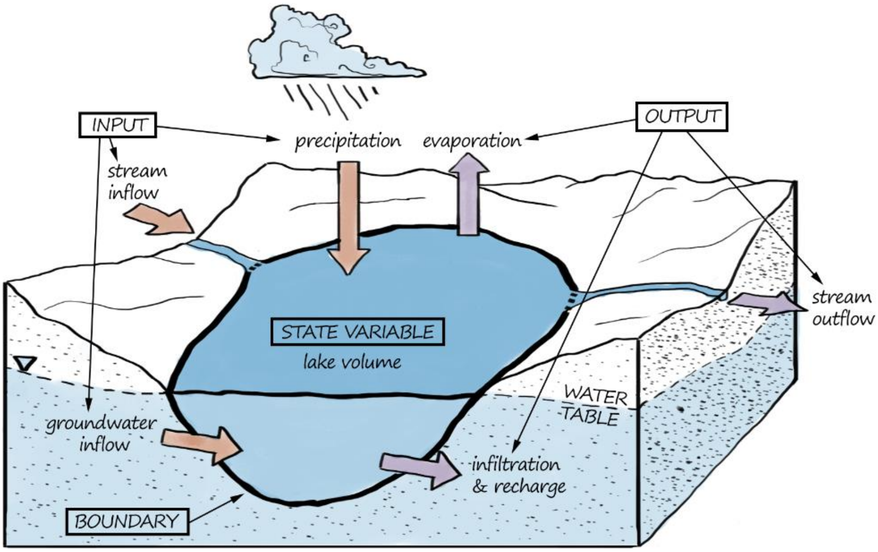

Figure 26. Lake system with groundwater as both an input (left) and an output (right) to the lake system.

So, what exactly causes the lateral movement of groundwater as shown in Figure 26? We often think of water as flowing “downhill”. But it is more correct to say that water flows from higher energy locations to lower energy locations. For groundwater, the energy that we are considering is potential energy – the ability of the water to perform work. Water flows from locations with high potential energy to locations with low potential energy. Similarly, heat flows through solids from areas at higher temperatures to areas at lower temperatures. Electrical currents flow from areas with higher voltage to areas with lower voltage.

If you know the potential energy distribution, you can figure out the direction of flow. But how can we figure out the *rate* of flow? The flow rate between two points is proportional to the difference in potential energy over the distance between those two points. This is known as the *potential gradient.* The greater the difference in potential, all other factors being equal, the faster the flow. For example, if you have two water slides of the same length – water will travel more quickly down the steeper water slide! The slope of the slide is an example of a gradient (change in potential per unit distance). You can define a gradient for any form of potential energy – elevation, temperature, voltage, etc.

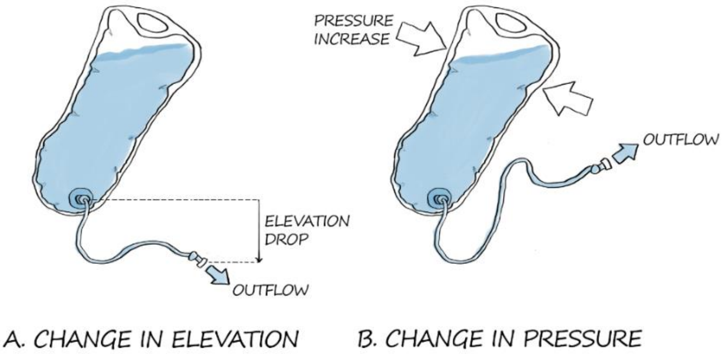

Figure 27. Water bladder hydration system used to demonstrate the two drivers of water flow (a) changes in elevation and (b) changes in pressure.

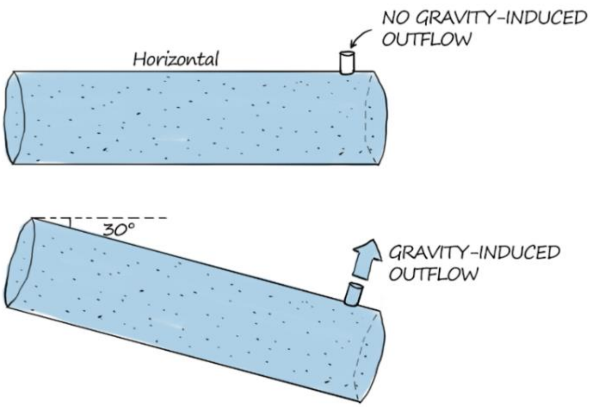

Figure 28. Flow of water through two saturated tubes of soil. When the tube is oriented horizontally, there is no flow through the tube due to gravity. Only when the tube is oriented at an angle is there flow.

There are two forms of potential energy that are important for groundwater flow. We have already introduced the first, elevation; water will flow from higher elevations to lower elevations. The second might seem less obvious; it is fluid pressure (p). Fluid pressure is the pressure you feel when you dive into the bottom of the swimming pool. The deeper you dive, the greater the weight of the water above you. But you aren't the only one feeling the weight – all the water molecules around you (at that depth) also "feel" the weight of the water above. You can imagine each drop of water as a small spring, compressed by all the water above it, storing energy. Therefore, if you poke a hole in the bottom of the pool, the springs expand into the air outside the hole, causing water to flow. This is a similar process to draining a bathtub!

To further explain the individual impacts of elevation and fluid pressure on fluid movement, let's think about a water bladder hydration system with a reservoir and a drinking tube (Figure 27). If you want water to come out of the tube, you have two options: 1) you can hold the outlet tube below the bag, which causes water to drain due to gravity, just like the draining of a tub (Figure 27a); or 2) you can suck on the straw (to reduce fluid pressure at the outlet) or squeeze the bag (to increase pressure in the bag compared to the outlet) (Figure 27b).

Now imagine a column of soil: a cylinder of saturated sand with an outlet tube (Figure 28). When you hold the cylinder at an angle where the outlet tube is pointing somewhere between 0 and 90°, water will eventually release from the pores and flow out of the tube due to the forces of gravity (Figure 28). However, if you eliminate the forces of gravity, say your column of soil is at exactly 90° (parallel to the ground), the water will not flow out the outlet tube due to gravity (or changes in elevation). To get water to flow through the outlet tube you must either increase the pressure at the inlet (e.g., push air or water through) or decrease it at the outlet tube (e.g., sucking/using a vacuum).

Hydraulic Head and Darcy's Law

We can begin to think about how water moves due to changes in elevation and/or fluid pressure, but let's dig a bit deeper into the mechanics of groundwater flow. How and why does the groundwater move? And how can you quantify groundwater movement?

The fundamental experiment used to explain groundwater flow is the Darcy experiment. In 1856 Henry Darcy published a report describing a laboratory experiment that is now generalized into the law that bears his name: *Darcy's law.* Darcy's law describes flow through a porous medium. It was derived from a series of rather simple experiments – the cool thing about Henry Darcy is that he formulated his law through observation, before fully understanding all the mechanics of why it worked. It is possible to discover new things in science *by observing them* before you can fully explain them! Since its original formulation, Darcy's law has been independently derived from other well-known equations and has been found to be analogous to other basic laws in physics.

The Darcy experiment (Figure 30) is like the cylinder analogy used above (Figure 28). The movement of water is caused by a combination of pressure and elevation. Darcy's experiment considers a cylinder filled with saturated sand (remember this means that all the pores are filled with water) of length (L) and cross-sectional area (A). The cylinder has caps at each end with an inflow tube and an outflow tube sticking through them. The flow of water through the inlet tube (Q_{in}) is equal to the flow of water through the outlet tube (Q_{out}). The flow rate is controlled by the overall potential energy difference between the two ends of the tube, which is controlled by the heights of water in two reservoirs (H_1 and H_2) above a common datum. See the callout box labeled "Datums" for further explanation on datum elevations.

Question:

- In the Darcy experiment, if $Q_{in} = Q_{out}$ (Figure 30), what is the system state? What is happening to the storage over time? Note: if you need a reminder about system states go back to the "Systems" chapter.

Answer:

- The Darcy experiment is in a *steady state* and there is no change in the volume of water stored in the column throughout the experiment.

What Henry Darcy recognized through a series of experiments was that:

1. The flow rate (Q) was directly proportional to the difference in the height of the reservoirs (H_2-H_1) divided by the length of the flow path (L),
2. The flow rate (Q) was directly proportional to the cross-sectional area of the column (A),
3. The flow rate (Q) was independent of the column angle (θ), and
4. The flow rate (Q) was different for different soils, which could be described by a soil-specific multiplier: the hydraulic conductivity (K).

Darcy's Law:

$$Q = -K\frac{\mathrm{H}_2 - \mathrm{H}_1}{\mathrm{L}}A$$

Equation 6: Darcy's law, where Q is flow (L^3/T), K is hydraulic conductivity (L/T), H_2 is the final hydraulic head (L), H_1 is the initial hydraulic head (L), L is the length of the flow path (L), and A is the cross-sectional area (L^2).

Datums

Datum elevations can be confusing, but if we want to measure an elevation (how high up something is)- we *must have* a datum elevation as a reference point (how high above *what* is the thing we're measuring). Mt. Everest is 29,029 feet. What does that really mean? It is 29,029 feet above the agreed-upon datum elevation, which is the average global sea level. If the bottom of the ocean were the datum elevation, Mt. Everest would have a much different elevation. Datum elevations are arbitrary, but for exact, relative measurements, they *must* be clearly defined and constant for your experiment.

To illustrate the arbitrary nature of a datum elevation, let's think about the difference in elevation of the two tallest mountains in the world. We will use the variable "z" to represent elevation. If you chose the datum elevation (z = 0) to be the elevation of the nearest city to Mt. Everest (Kathmandu, Nepal: elevation 4,593 feet) rather than sea level, the height of Everest would only be 24,436 feet rather than 29,029 feet. Because your datum elevation (or reference state) must be consistent for all measurements, the second tallest mountain in the world (K2) would be 23,658 feet (datum elevation: Kathmandu) rather than 28,251 feet (datum elevation: sea level). In both cases, the height of Mt. Everest is 778 feet taller than K2 and the relative heights are documented appropriately (Figure 29).

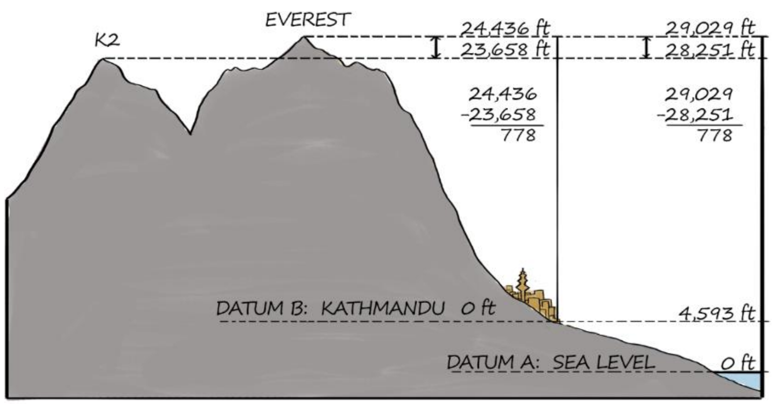

Figure 29. Elevation of Mt. Everest and K2 using two different datum elevations. Datum A: sea level and Datum B: Kathmandu, Nepal. For both datum elevations Mt. Everest is 778 feet taller than K2. This is true for all datum elevations.

Datum elevation sea level:
Mt. Everest: 29,029 feet
K2: 28,251 feet
Difference in elevation: 29,029 feet—28,251 feet = 778 feet

Datum Elevation Kathmandu, Nepal (4,593 feet):
Mt. Everest: 24,436 feet
K2: 23,658 feet
Difference in elevation: 24,436 feet — 23,658 feet = 778 feet

If all mountains in the world were measured based on the elevation of Kathmandu – their heights compared to one another would be the same as when measured based on sea level. Therefore, the elevation gradient between mountains of different heights is consistent for all possible datum elevations.

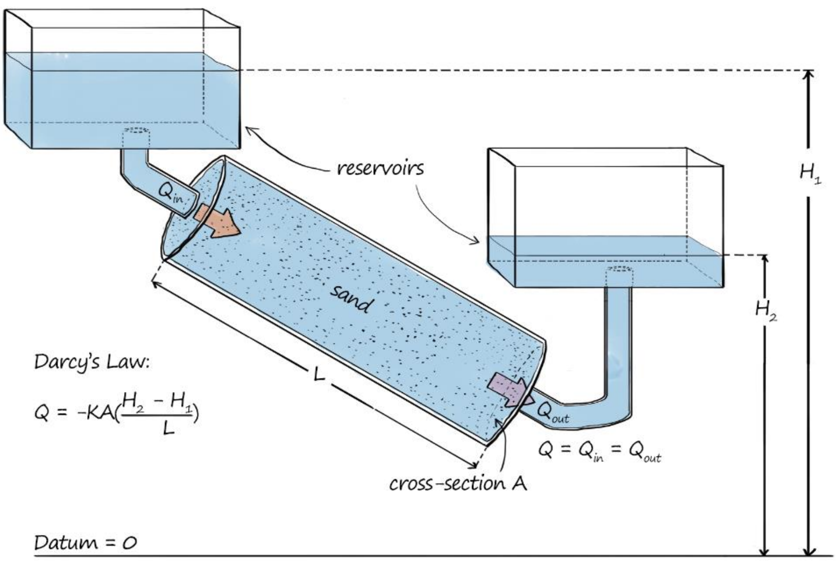

Figure 30. Darcy's experiment with reservoirs.

In other words, with the same change in reservoir water height (H), wider and/or shorter columns have faster flow. With everything else held constant, higher hydraulic conductivity soils (e.g., sand) have faster flow; lower hydraulic conductivity soils (e.g., clay) have slower flow. Lastly, the greater the difference in heights of water (H_2-H_1) in the two reservoirs, the faster the flow through the same soil (think about a water slide with more elevation drop over the same distance).

But why is there a negative sign in the Darcy equation (Equation 6)? This is a difficult concept. It has to do with the common definition of a potential energy gradient. Think of the most common experience that we have with a gradient – walking up a hill. Up implies that the potential energy gradient of your path is positive (your final elevation is greater than your initial elevation). But water would flow in the opposite direction, downhill. In other words, given our common definition of a gradient (the final condition minus the initial condition divided by the distance), positive water flow occurs in the direction *opposite* to a positive gradient. So, we need the negative sign.

If you were to research the Darcy experiment, you might find a figure like Figure 30, but with a pair of *manometers* included (Figure 31). A manometer is an open tube that tells you about the pressure in a system at the specific point where it is installed (P). The water in a manometer is static (not flowing). At the surface of the water in the manometer, the water pressure is zero. The pressure at the bottom of the manometer is equal to the height that the water rises in the manometer. Therefore, the height of water in the manometer *relative to the point of insertion* is a direct measure of the water's pressure at the point of insertion. The height of water in the manometer *above the datum* is the total potential energy (elevation and pressure combined) at the point of insertion.

In hydrogeology, we refer to the potential energy driving groundwater flow as head. *Head* is a metric of how far something can move against the force of gravity, in units of length (cm, m, ft, etc.). It is expressed in terms of energy divided by weight. It is simplest to think of head as a height. The *pressure head* (ψ) in a column (like a manometer) is the height to which water rises in the column due to the pressure exerted at a certain point (P) (i.e., the potential energy at that point). In a swimming pool, this would be equivalent to the depth you are underwater, which corresponds to the weight of water above you that you feel. The height in the manometer above a specific point in the column (P) is the pressure head (ψ) at that point (Figure 32). In other words, the weight of the water in the manometer tube is pushing down as hard as the weight of the water above point P is pushing up, so the height within the manometer is how far from point P the energy can move water upwards against the force of gravity (remember, this is the formal definition of head).

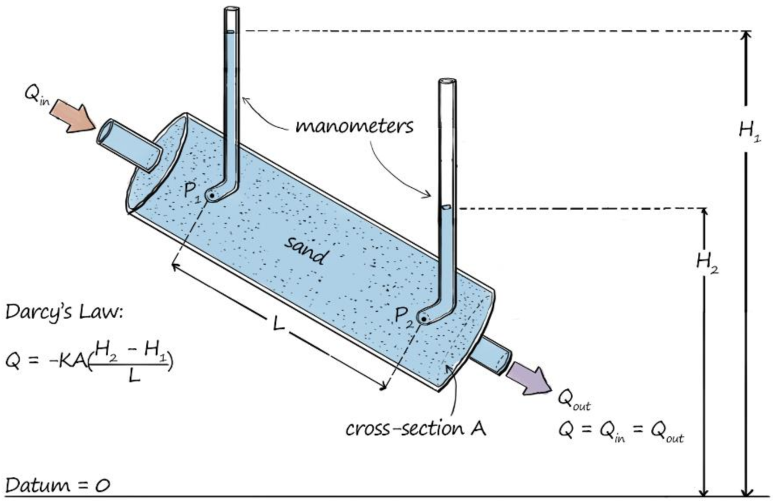

Figure 31. Darcy experiment with manometers.

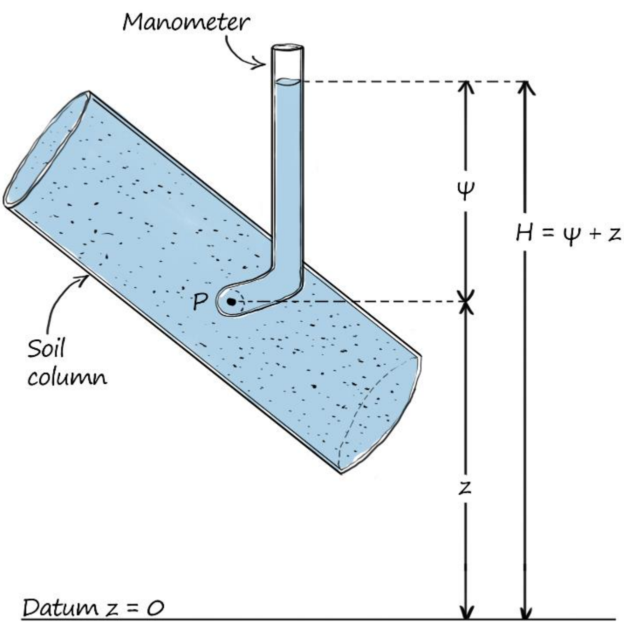

Figure 32. Soil column with a manometer that measures pressure head. The combination of pressure head (ψ) and elevation head (z) is the hydraulic head (H).

Pressure head relates to fluid pressure (p) using the following equation:

$$p = \rho g \psi + p_0$$

Equation 7: Equation for fluid pressure p (ML/T^2), where ρ fluid density (M/L^3), g is gravity (L/T^2), ψ is the pressure head (L), and p_0 is the atmospheric pressure or the pressure at some reference state (ML/T^2).

The *elevation head* (z) is the height of location "P" above the datum elevation. It represents the potential energy arising from elevation per unit weight. The *hydraulic head* (H) is the combined height of pressure head and elevation head (Figure 32).

$$H = \psi + z$$

Equation 8: Hydraulic head (H, units L) where ψ is the pressure head (L) and z is the elevation head (L).

Compare Figure 31 and Figure 32. Henry Darcy was simply observing differences in *hydraulic head* (H) in his experiment, although he had not yet defined it as such.

Hydraulic Head Gradient and Flux

Darcy noticed that the flow (Q) was proportional to the change in water heights in the manometers (H_2-H_1) divided by the distance between the manometers along the column (L) (Equation 6) (Figure 31). Reminder: the change in the potential quantity (e.g., hydraulic head) divided by the length is called the potential gradient. In this case, because the type of potential energy we are considering is hydraulic head, it is the *hydraulic head gradient* (ΔH/L). You may not be familiar with the delta symbol (Δ). This symbol is used to denote the difference between two quantities. In this case, the difference between the hydraulic head at point 2 (H_2) and point 1 (H_1), or ($H_2 - H_1$), is ΔH. But be careful: an upside-down triangle (∇) identifies a gradient (a change in something over a length); therefore, the hydraulic head gradient (ΔH/L) can also be represented simply as ∇H. Just as slides at the park with the greatest change in elevation over the shortest length are the fastest, the soil columns with the greatest change in hydraulic head over the shortest distance have the fastest flow (but only if they are filled with the same medium).

Darcy also noted that the flow rate (Q) was proportional to the cross-sectional area of the tube (A) (Equation 6). Think about two tubes of different diameters where one tube has twice the cross-sectional area of the other (i.e., A and 2A). In all other ways, the tubes are identical. They have the same material (K) and the same hydraulic head gradient (ΔH/L or ∇H).

How would you expect the flow rate to differ between the two tubes?

$$Q_{smaller\ tube} = -K\frac{\Delta \mathrm{H}}{\mathrm{L}}(A) = -K(\nabla \mathrm{H})(\mathrm{A})$$
$$Q_{larger\ tube} = -K\frac{\Delta H}{\mathrm{L}}(2A) = -K(\nabla \mathrm{H})(2\mathrm{A})$$
$$Q_{larger\ tube} = 2Q_{smaller\ tube}$$

Because everything else in the two equations is the same, the flow through the larger tube is exactly twice the flow through the smaller tube. Think about this in volumes. If you have two tubes of different diameters (with the same hydraulic head), and you place buckets at the end of each tube (for the same amount of time), you will collect *twice* as much water at the outlet of the large tube. If you wanted to have the *same* volumetric flow rate for the two different tube sizes, the water would need to move through the smaller tube much faster than it moves through the larger tube to fill the bucket with the same volume of water. This introduces another important part of Darcy's experiment: the difference between the volumetric flow rate (Q, in units of volume over time), and the flux (q, in units of length over time).

In groundwater, the rate at which water is moving through a cross section is called the *Darcy flux (q).*

$$q = \frac{Q}{A}$$

Equation 9: Flux (q, units L/T) where Q is the flow (L^3/T), and A is the area (L^2).

Imagine a garden hose with a constant flow. When you put your finger over the hose opening, you decrease the area, but the volume of water coming out of the hose each second doesn't change (Q is constant). Therefore, the water must come out of the hose more quickly to emit the same total volume. For two tubes with different areas (A and 2A), the flux would have to be twice as large to get the same flow through the smaller tube.

The difference between q and Q can be confusing at first. To start, imagine that you are standing on the side of a river and can draw a cross-section line perpendicular to the flow path (the cross section is a slice of the river across the channel width). The flow (Q) for this system is the volume of water passing through your cross section at any given time (L^3/T) or *how much water* is flowing through your slice of the river. The flux (q) is the flow divided by the cross-sectional area (of the channel) at that given location (L/T). For a river, the flux is the same as the velocity or *how quickly* water flows through your slice of the river.

What happens when we consider a groundwater system? If you consider a slice through a groundwater system, you can also describe the flow or how much water flows through a cross section per time. And if you divide Q by the area of the cross section, you get the flux. However, q is NOT the velocity of the water. Why? Because most of the cross section is blocked by soil particles! As a result, the water must move through the pore space much faster than it flows through the cross-sectional area; the actual water velocity is higher than the flux, even though they have the same units. We will cover this in more detail when we discuss solute transport.

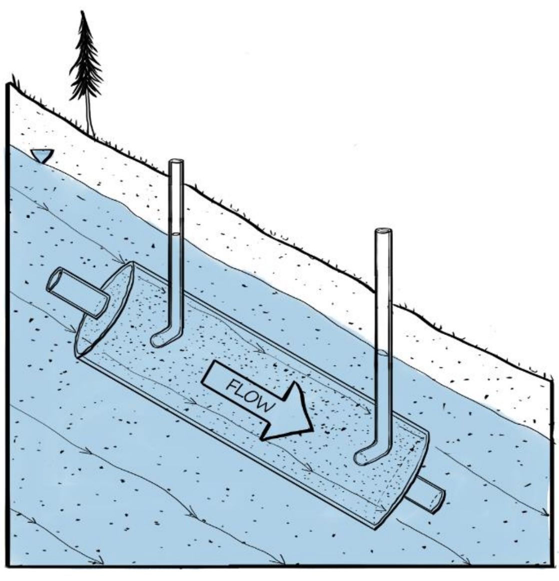

Figure 33. Darcy's experiment shown as a small representation of a hillside.

Darcy's experiment may seem inapplicable to real groundwater systems. However, imagine a hillside with a consistent material and a constant slope. Now, imagine holding a soil tube with the same material at the same angle of the hillslope (Figure 33). Can you see how the soil tube is a small representation of the hillside, just with different boundaries? These are the types of strategies hydrologists use to study complex groundwater systems. All the soil tube experiments we present throughout the text contribute to a greater conceptual understanding of real groundwater systems.

System States, Hydraulic Head, and Calculations

Previously, we discussed groundwater flow into and out of a lake (Figure 26), but what if you wanted to know *exactly* how quickly the groundwater was moving? Groundwater calculations may seem a bit intimidating at first, but we will walk you through a series of them starting with the simplest system and building up to more complex ones. Beware, the systems we first introduce might appear so simple that you are not even sure how they apply to real life. Bear with us. In the end, our hope is that you can use both your conceptual understanding and a host of equations to better describe the groundwater movement in a real system.

Static State and Hydraulic Head

Before we dive into quantifying groundwater movement, let's connect the meaning of hydraulic head (H) to system states. A static state is when there is no input or output to the system, no boundary flow, and no change in storage (Table 2). To connect hydraulic head and static state, imagine a small, square bucket filled with water (Figure 34). The boundaries of the system are the walls of the bucket, the bottom of the bucket, and the top of the water. The state variable is the hydraulic head.

In our bucket example (Figure 34), the condition is static, and the hydraulic head is the same everywhere in the bucket (no change in hydraulic head = no gradient = no flow = static). However, if the water in the bucket was flowing, there would need to be a hydraulic head gradient (a change in the state variable). We do not yet know the value of the hydraulic head for the static system (Figure 34), but the actual value doesn't really matter. All we need to know is that the hydraulic head is *constant* at every location over time. The hydraulic head is a combination of the pressure head (ψ) and elevation head (z) (Equation 8).

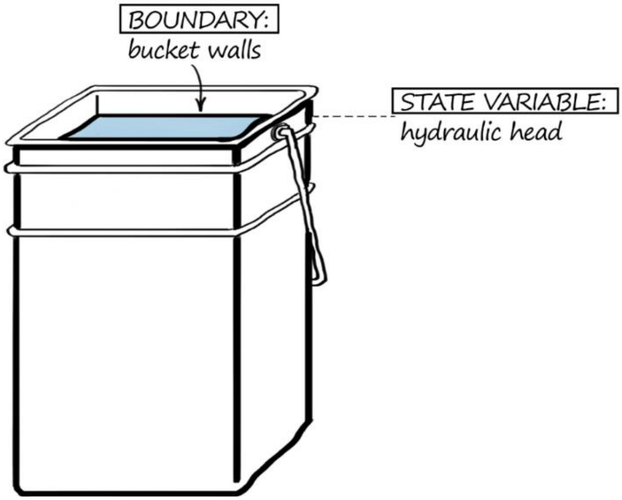

Figure 34. Bucket system with hydraulic head as the state variable.

For the elevation head of our static bucket system, we must choose a datum elevation. At the datum, the elevation head = 0. In this case, we choose the bottom of the bucket as the datum elevation (Figure 35). Therefore, all locations *above* the bottom of the bucket have a positive elevation head; just as all mountain tops above sea level have a positive elevation. If the bucket is filled to 10 cm, the elevation head at the top of the water column is 10 cm (green line, Figure 35).

For the pressure head, start by looking at the surface of the water in the bucket. At the surface (i.e., the air-water interface), the only pressure that the water in the bucket "feels" is that of the atmosphere (there is no pressure from water). Therefore, the pressure head at the air-water interface in the bucket is zero (red line, Figure 35). The pressure head is independent of the datum elevation.

Continue to analyze the bucket system to make sense of the hydraulic head distribution (specifically, look at the graph on the right side of Figure 35). At the surface, the pressure head (red line) is 0 cm, and the elevation head (green line) is 10 cm. Therefore, the hydraulic head (blue line) is 10 cm (H = 0 cm + 10 cm = 10 cm). At the bottom of the bucket the elevation head is 0 cm, the pressure head is 10 cm, and the hydraulic head is 10 cm. Because the system is static the hydraulic head is constant (10 cm) at all depths and the hydraulic head gradient is zero.

To demonstrate the effect of the datum elevation on hydraulic head values, let's change the datum elevation of our static bucket system (Figure 35). Imagine that you change the datum elevation to the top of the bucket (Figure 36) rather than having the datum at the bottom of the bucket (Figure 35). How would your pressure, elevation, and hydraulic heads differ from the example above? The pressure head distribution is not dependent on the datum elevation; the water always "feels" the same weight regardless of the datum. Therefore, the pressure head would still be 0 cm at the top of the bucket and 10 cm at the bottom (red line, Figure 36). Note: the pressure head also doesn't depend on the shape or diameter of the container. If you compare the pressure between the bucket and other containers of various diameters, if each container is filled with the same liquid (e.g., water) to the same height, and each is static – the pressure at point P below the surface will be equal in all the containers.

For the new datum elevation (at the top of the bucket) (Figure 36), the elevation head at the top of the bucket is now 0 cm (green line, Figure 36). Elevation head is always zero at your datum. Death Valley in California has a negative elevation of roughly -282 feet below sea level, or 282 feet below the datum. The same convention holds true for other systems; after you define your datum elevation, any locations *below the datum* will have a negative elevation head. In this case, if the datum is at the top of the bucket, the elevation head at the bottom of the bucket is -10 cm (green line, Figure 36) or 10 cm below the datum.

With the datum elevation at the top of the bucket, what is the hydraulic head at the top (H_{top}) and bottom (H_{bottom}) of the bucket?

$$H = \psi + z$$
$$H_{top} = 0\,cm + 0\,cm = 0\,cm$$
$$H_{bottom} = 10\,cm - 10\,cm = 0\,cm$$

The hydraulic head is 0 cm everywhere (blue line, Figure 36). This is the same bucket system as Figure 35, but we have a different *value* for hydraulic head (0 cm vs. 10 cm). Is that OK? Remember that the potential head gradient is what determines flow, not the absolute value of the hydraulic head. Although the hydraulic head numbers are different between scenarios (Figure 35, Figure 36), the *relative value* of hydraulic head (the difference between H_{bottom} and H_{top}) is the same. Therefore, the hydraulic head gradient (of zero) is the same!

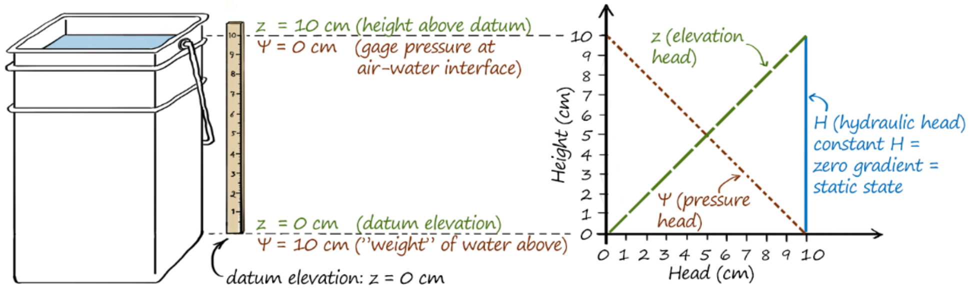

Figure 35. Elevation head, pressure head, and hydraulic head for a static bucket with the datum elevation at the bottom of the bucket.

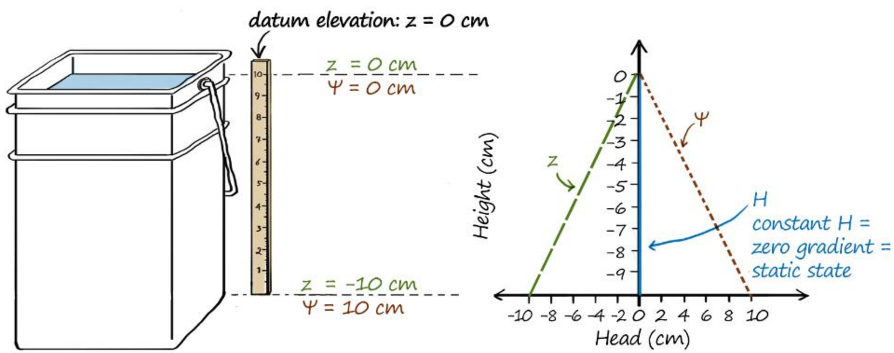

Figure 36. Elevation head, pressure head, and hydraulic head for a static bucket with the datum elevation at the top of the bucket.

To illustrate this, let's calculate the hydraulic head gradient for the two scenarios.

- The hydraulic head gradient when the datum elevation is at the bottom of the bucket (Figure 35):

$$\frac{\Delta H}{L} = \frac{H_{bottom} - H_{top}}{L} = \frac{0\ cm - 0\ cm}{10\ cm} = \frac{0\ cm}{10\ cm} = 0$$

- The hydraulic head gradient when the datum elevation is at the top of the bucket (Figure 36):

$$\frac{\Delta H}{L} = \frac{H_{bottom} - H_{top}}{L} = \frac{10\ cm - 10\ cm}{10\ cm} = \frac{0\ cm}{10\ cm} = 0$$

For static conditions, the pressure head perfectly compensates for the elevation head, and the hydraulic head is constant everywhere. The hydraulic head gradient is zero and there is no potential gradient that drives flow.

Let's scale this up to a real-world system. Using the information above, what is the *hydraulic head gradient* for a static, clay-lined lake with no input or outputs? This is not a trick question. You have all the information you need. You don't need to know the depth of the lake or even the pressure head distribution in the lake. If you know that the system is *static*, you know that the hydraulic head is *constant*; therefore, the hydraulic head gradient is zero and there is no flow.

As a final thought experiment for the static bucket (Figure 35), imagine that the bucket is filled with a saturated sand (Figure 37). How does this change the hydraulic head gradient of the system? Or does it? Even with sand present, under static conditions, the weight of the water is perfectly compensated for by the pressure of the water. This is why sediments at the ocean floor aren't compressed to solid rock even if they are under a mile of water! As a visual, think about a vertical metal spring. What happens if you put a weight on top of the spring? It compresses. When does it stop compressing? When the force applied by the spring exactly matches the force applied by the mass. The same thing happens in water, but we measure the equivalent of the compression of the spring as water pressure. Therefore, in our static bucket filled with sand, the pressure head at the bottom (green line, Figure 37) perfectly compensates for the elevation head (red line, Figure 37) and the hydraulic head gradient is 0.

We intuitively know that for a static bucket full of saturated sand, there isn't flow. That's one of the very definitions of a static state! However, if we wanted to quantitively show that the flow through the porous medium in Figure 37 is zero, we can use Darcy's law (Equation 6).

The hydraulic conductivity of the sand in the bucket is 0.5 cm/s. The cross-sectional area of the bucket is a square (which is equal to the length of the sides squared). The sides of the bucket are 20 cm long. The hydraulic head gradient for the static bucket is zero.

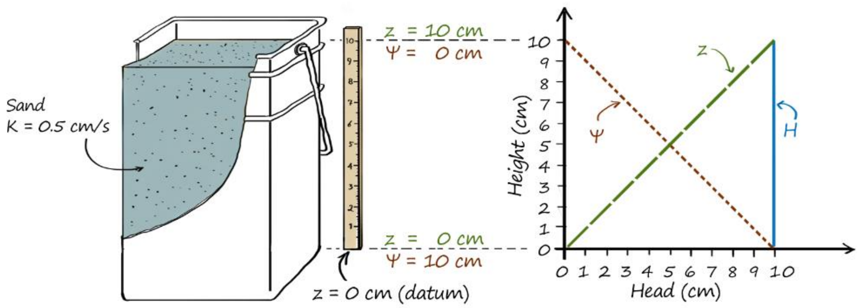

Figure 37. Elevation head, pressure head, and hydraulic head for a static bucket filled with sand (K = 0.5 cm/s) with the datum elevation at the bottom of the bucket.

Therefore, using Darcy's law (Equation 6) the flow through the bucket is:

$$Q = -K\frac{\Delta H}{\mathrm{L}}A$$

$$Q = -\left(0.5\ \frac{cm}{s}\right)\left(\frac{0\ cm - 0\ cm}{10\ cm}\right)(20^2\ cm^2) = 0\frac{cm^3}{s}$$

Steady State and Hydraulic Head

To connect *steady state* to hydraulic head, let's imagine a saturated soil tube much like the Darcy experiment introduced above (Figure 31). Take a moment to reflect on the meaning of steady state conditions. For steady state, the flow going into the tube is equal to the flow going out of the tube, there is no change in local storage within the pores (the pores are saturated for the entire observation time), and the state variable (e.g., hydraulic head) does not change with time (Table 2). It is important to notice that for steady state there is flow through the column, and the flow is constant throughout the system.

Before we do any calculations, compare a steady state soil tube (Figure 38b) to a static one (Figure 38a). For both steady state and static conditions, the value of the hydraulic head at every location within the system doesn't change over time, and the soil is saturated. This means, if you were to take a snapshot of the hydraulic head at the very middle of the column at the beginning of the experiment and at the end of the experiment (or any time in between), the hydraulic head would not change *in that location*. The difference is, for static conditions, the hydraulic head is constant (the exact same value) throughout the entire system (Figure 38a). But, for steady state conditions (Figure 38b), the hydraulic head value *does change with location*, but remains constant at each location over time.

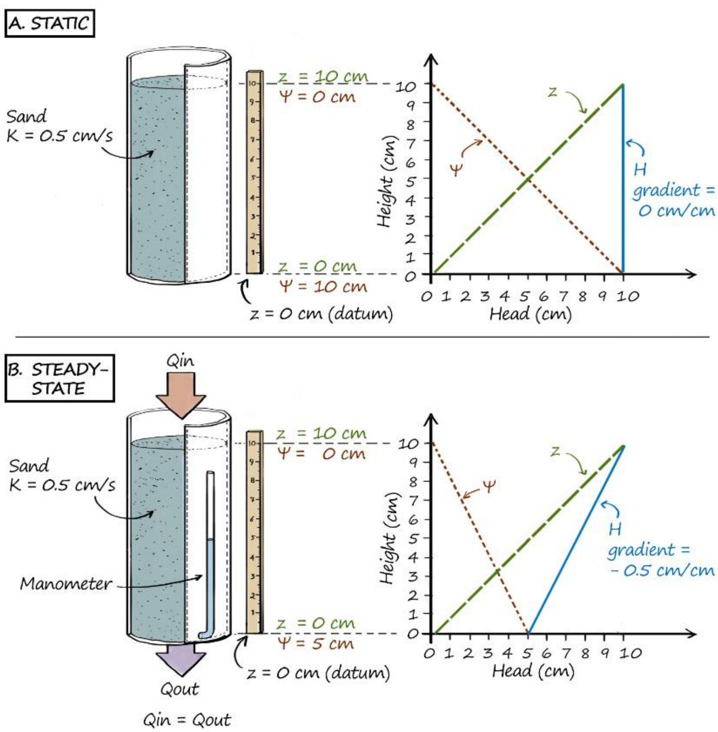

Figure 38. A 10 cm soil tube filled with sand under (a) static and (b) steady state conditions.

As an example, let's look at the hydraulic head at the top and bottom of the column in Figure 38. For static conditions (Figure 38a), the hydraulic head at the top and bottom of the column are the exact same value at both locations (H_{top} is 10 cm and H_{bottom} is 10) for all of time (Figure 38a). For steady state (Figure 38b), the hydraulic head at the top of the column and bottom of the column are different values (e.g., H_{top} is 10 cm and H_{bottom} is 0 cm), but they are still these same values for all of time (Figure 38b). Another way to think about this is that for static conditions the hydraulic head gradient is *always zero* (blue line, Figure 38a) and for steady state, the hydraulic head gradient is *always a constant nonzero value* for all of time (blue line, Figure 38b). For flow to occur, you must have a spatial change in hydraulic head (i.e., you must have a hydraulic head gradient).

Like in the static bucket example above, we can use the elevation and pressure heads to determine the hydraulic head for a steady state soil tube (Figure 38b). The datum elevation is the bottom of the soil tube. If the soil tube is 10 cm long, the elevation head at the top of the tube is 10 cm (because the top of the tube is 10 cm *above* the datum elevation). For pressure head, the top of the tube is 0 cm. We can use a manometer to determine the pressure head at the bottom of the tube. The height of water in the manometer is 5 cm. Therefore, the pressure head at the bottom of the tube is 5 cm.

H at the top of the tube (H_1):

$$H_1 = \psi + z$$
$$H_1 = 0\,cm + 10\,cm = 10\,cm$$

H at the bottom of the tube (H_2):

$$H_2 = \psi + z$$
$$H_2 = 5\,cm + 0\,cm = 5\,cm$$

From our calculations we can see that the hydraulic head at the top of the tube is greater than the hydraulic head at the bottom of the tube (Figure 38b). What does this mean? Remember water flows from high hydraulic head to low hydraulic head. Therefore, we know the water is moving from the top of the tube (H_2 = 10 cm) to the bottom of the tube (H_1 = 5 cm).

Hydraulic head gradient:

$$\frac{\Delta H}{L} = \frac{5\ cm - 10\ cm}{10\ cm} = -\frac{5}{10} = -\frac{1}{2} = -0.5$$

Notice the hydraulic head gradient is negative (-0.5). This is because the hydraulic head H_2 is less than H_1. This is what we would expect! The negative sign in Darcy's equation will correct for this when we calculate flow.

Now let's use the hydraulic head gradient calculated above (-0.5) and Darcy's law (Equation 6) to calculate the flow through the soil tube. The soil tube has a radius of 1 cm, and the hydraulic conductivity of the soil is 0.5 cm/s. We can use the formula for the area of a circle ($A = \pi r^2$) to determine the cross-sectional area of the tube.

$$A = \pi r^2 = \pi(1\ cm)^2 = 3.14\ cm^2$$

Next, we use Darcy's law to calculate flow:

$$Q = -KA\frac{\Delta H}{L}$$

$$Q = = -\left(0.5\frac{cm}{s}\right)(3.14\ cm^2)\left(-\frac{1}{2}\right) = 0.79\frac{cm^3}{s}\ downwards$$

The flow in, out, and through the soil column is 0.79 cm^3/s downwards. In fact, the flow rate is always 0.79 cm^3/s, because the system is steady state. Notice that Q is a positive number. In this case, positive means that the water is flowing from H_1 to H_2 (top to bottom). If Q were negative, the water would be flowing in the opposite direction. It is easy to make mistakes related to the sign (+/-). However, the absolute value of the hydraulic head gradient is the same regardless of how you order the hydraulic head values in your equation. Therefore, after you finish your calculations, check your work by making sure that the direction of flow is from high hydraulic head to low hydraulic head.

Practice Problems

Problem 1

You have a 20 cm soil tube (radius = 1 cm) filled with soil. The hydraulic conductivity of the medium is 1 cm/s and the height of water in the manometer is 15 cm. In a previous example, the manometer was located at the bottom of the tube (Figure 38b). However, now the tip of manometer is in the *middle* of the tube (at 10 cm) (Figure 39); the pressure head in the middle is 15 cm and the pressure head at the top is 0 cm (red line, Figure 39). Because the system is steady state, the pressure head increases 1:1 with elevation and is 30 cm at the bottom of the tube (red line, Figure 39). The elevation head at the bottom is zero and the elevation head at the top is 20 cm (green line, Figure 39).

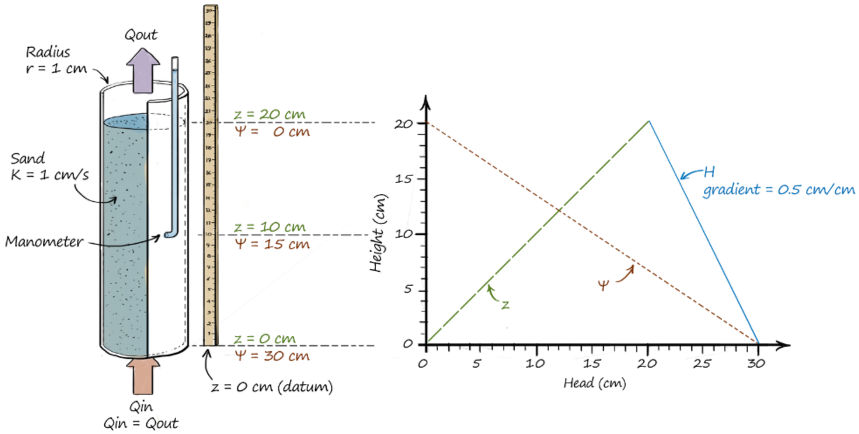

Figure 39. Steady state conditions in a 20 cm tube. The manometer is measuring the pressure head in the middle of the tube at 10 cm.

For demonstration purposes, let's assume that *we think* that the water is flowing downward (even if we know it is moving upward) and choose the H_1 to be the top of the tube and H_2 as the bottom of the tube. First, calculate the hydraulic head gradient and then the flow through the tube using Darcy's law (Equation 6).

H at the top of the tube (H_1):

$$H_1 = \psi + z$$
$$H_1 = 0\ cm + 20\ cm = 20\ cm$$

H at the bottom of the tube (H_2):

$$H_2 = \psi + z$$
$$H_2 = 30\text{ cm} + 0\text{ cm} = 30\text{ cm}$$

Hydraulic head gradient:

$$\frac{\Delta H}{L} = \frac{30\ cm - 20\ cm}{20\ cm} = \frac{10}{20} = \frac{1}{2}$$

Flow:

$$Q = -KA\frac{\Delta H}{L}$$
$$Q = -\left(1\frac{cm}{s}\right)(\pi\ 1^2\ cm^2)\left(\frac{1}{2}\right) = -1.6\frac{cm^3}{s}\ upwards$$

Notice, the hydraulic head gradient is positive. Therefore, when we calculate flow, we get a negative Q value (-1.6 cm^3/s). What does this mean? It means that the water is flowing opposite of our originally defined gradient (H_1 to H_2). In all cases, water flows from higher hydraulic head to lower hydraulic head. In this case, the water is flowing vertically *upwards* because the hydraulic head at the bottom of the tube is 30 cm and the top is 20 cm. After finishing your calculations, always double check the direction of flow and state it with your answer (e.g., upwards, downwards).

Problem 2

For the examples above, you calculated the hydraulic head and flow. But remember you can also find other variables. For example, if you know the flow rate through a steady state system, the hydraulic head values, and the dimensions of the soil column – you can calculate K. Try it! You have a 10 cm soil tube with a steady state downward flow of 3 cm^3/s (Figure 40). The soil tube has a radius of 1 cm. The hydraulic head at the top (H_1) is 12 cm and at the bottom is 5 cm (H_2). What is the hydraulic conductivity of the medium? To calculate hydraulic conductivity (K), you must rearrange Darcy's law (Equation 6):

$$Q = -KA\frac{\Delta H}{L}$$

$$3\frac{cm^3}{s} = -K(\pi 1^2\ cm^2)\frac{5\ cm - 12\ cm}{10\ cm}$$
$$\frac{3\ \frac{cm^3}{s}}{(\pi 1^2\ cm^2)\left(\frac{5\ cm - 12\ cm}{10\ cm}\right)} = -K$$
$$K = 1.36\frac{cm}{s}$$

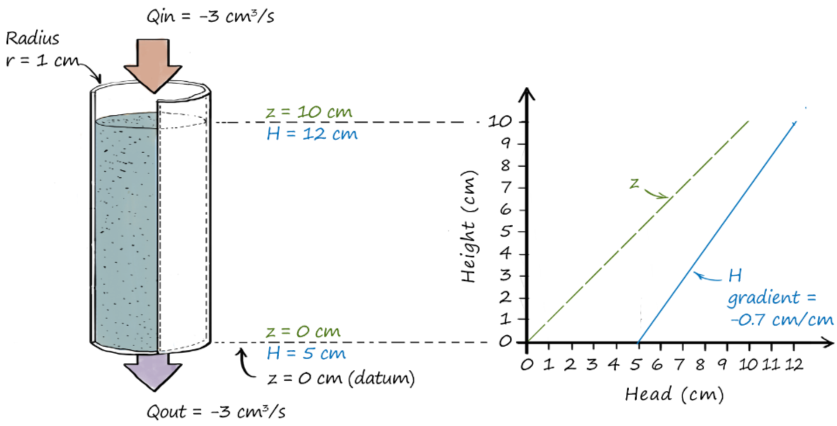

Figure 40. Soil tube with defined hydraulic head values and a hydraulic head gradient of -0.7.

Soil Type and Flow Calculations

For all the examples above (e.g., Figure 31, Figure 37, Figure 38, Figure 39, Figure 40), we used homogenous media. A *homogeneous medium* is the same throughout; at all locations, the medium has the same properties. For example, a homogeneous medium has a consistent hydraulic conductivity (K) throughout. Heterogeneous media are not the same throughout (e.g., a layered system with alternating layers of sand and clay).

For now, we will stick to homogeneous media. However, let's explore the variability of *different* homogeneous media. Imagine that you have three different 15 cm tubes filled with 1) sand (K = 0.5 cm/s), 2) silt (K = 0.002 cm/s), and 3) clay (K = 5×10^{-8} cm/s) (Figure 41). The radius of each of the soil tubes is 2 cm, and the datum elevation is defined at the bottom of the tubes. As before, use the elevation head and pressure head to define the hydraulic head in each of these systems.

The pressure head at the top of all tubes is 0 cm and the pressure head at the bottom is 10 cm (as measured by the manometers) (red line, Figure 41). The elevation head at the top of all tubes is 15 cm and the elevation at the bottom is 0 cm (green line, Figure 41). Lastly, for all three tubes, the hydraulic head at the top is 15 cm (H_1) and at the bottom is 10 cm (H_2) (blue line, Figure 41). Therefore, the hydraulic head gradient is the same for all three systems

($\Delta H/L = -1/3$). Does this mean the flow is the same? Let's calculate the flow (Q) for each of the tubes.

Sand (K = 0.5 cm/s)

$$Q = -KA\frac{\Delta H}{L}$$

$$Q = -0.5\frac{cm}{s}\,(\pi 2^2\ cm^2)\frac{(10\ cm - 15\ cm)}{15\ cm} = 2.09\frac{cm^3}{s}\ downwards$$

Silt (K = 0.002 cm/s)

$$Q = -KA\frac{\Delta H}{L}$$

$$Q = -0.002\frac{cm}{s}\,(\pi 2^2\ cm^2)\frac{(10\ cm - 15\ cm)}{15\ cm} = 0.0084\frac{cm^3}{s}\ downwards$$

Clay (K = $5x10^{-8}$ cm/s)

$$Q = -KA\frac{\Delta H}{L}$$

$$Q = -5x10^{-8}\frac{cm}{s}\,(\pi 2^2\ cm^2)\frac{(10\ cm - 15\ cm)}{15\ cm} = 2.09x10^{-7}\frac{cm^3}{s}\ downwards$$

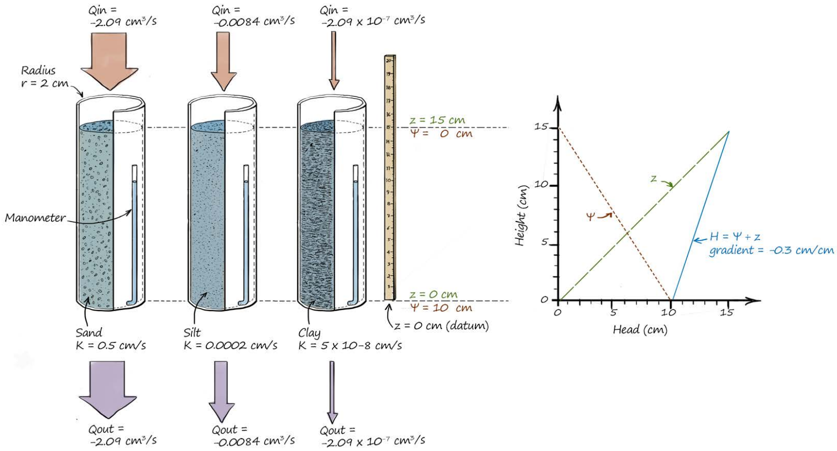

Figure 41. Hydraulic head gradient of -1/3 through three columns of the same size filled with different media (sand, silt, and clay).

What do you notice about the flow through each of the tubes? The hydraulic head gradient is the same for each soil tube (-1/3). The area is the same for each (4π cm^2). However, the hydraulic conductivity is different. Therefore, the flow is different; the flow is significantly slower through the clay (lower hydraulic conductivity) than through the sand (higher hydraulic conductivity). Also, notice that because the flow is positive for all three cases, the flow direction is downward through the tube.

From the previous example (Figure 41), we see that a more conductive medium (e.g., sand) has a higher flow with the same hydraulic head gradient. Wait, why is that? Think about a block sliding down a ramp (Figure 42). The block slides down the ramp because of changes in elevation head. If you have two ramps of different textures, one rough and one smooth, but the same change in elevation head, the block will slide at different speeds down the two ramps. For the rough ramp, the block will not slide very fast because of friction. For the smooth ramp, the block will slide more quickly. Therefore, the same potential gradient can have different block speeds depending on the material of the ramp. The same is true for water. When water flows it loses energy to friction against the sediment grains and a portion of the initial energy (hydraulic head) is lost to overcome the resistance to flow. In fact, for steady state the head loss exactly equals the loss to friction. Therefore, the same hydraulic head gradient can have significantly different flow rates through different materials (Figure 41).

For the flow rate to be the same for the three tubes above (Figure 41), you would need a significantly higher hydraulic head gradient through the tubes with the lower K values (e.g., silt and clay). In the case of the blocks, if you wanted the blocks on the rough ramp to slide down at the same speed as the smooth ramp, you would need to increase the elevation gradient for the rougher ramp (Figure 42). In other words, if you have a rougher ramp you'll need a higher the potential gradient to overcome the friction. The same is true for water. If you wanted the soil tube with silt to have the same flow rate as the soil tube with sand, you would need a higher potential gradient (hydraulic head gradient) in the silt to overcome the resistance to flow.

To better understand this concept, let's imagine we have a steady state flow rate of 3 cm^3/s (downward) in all three of our soil tubes (Figure 41). What would the hydraulic head gradient need to be for the three different media to all have a flow rate of 3 cm^3/s? Use Darcy's law (Equation 6).

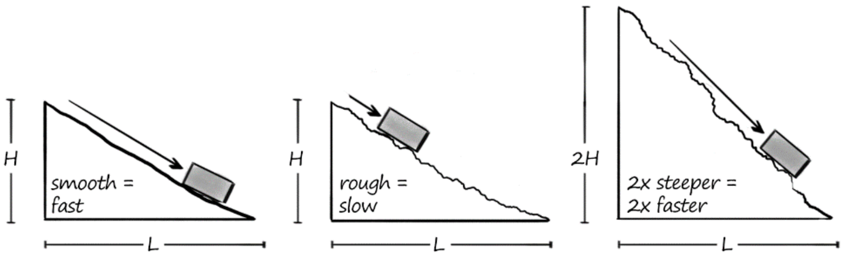

Figure 42. Block sliding down ramps of different textures and gradients.

Sand (K = 0.5 cm/s)

$$Q = -KA\frac{\Delta H}{L}$$
$$3\frac{cm^3}{s} = -0.5\frac{cm}{s}\ (\pi 2^2\ cm^2)\left(\frac{\Delta H}{L}\right)$$
$$\frac{\Delta H}{L} = -0.48$$

Silt (K = 0.002 cm/s)

$$Q = -KA\frac{\Delta H}{L}$$
$$3\frac{cm^3}{s} = -0.002\frac{cm}{s}\ (\pi 2^2\ cm^2)\left(\frac{\Delta H}{L}\right)$$
$$\frac{\Delta H}{L} = -119.37$$

Clay (K = 5x10^{-8} cm/s)

$$Q = -KA\frac{\Delta H}{L}$$
$$3\frac{cm^3}{s} = -5x10^{-8}\frac{cm}{s}(\pi 2^2\ cm^2)\left(\frac{\Delta H}{L}\right)$$
$$\frac{\Delta H}{L} = -4.77x10^6$$

Notice, the hydraulic head gradient for the clay must be 7x greater than the hydraulic head gradient in the sand to have the same flow rate! This is because more energy is lost as it travels through the clay.

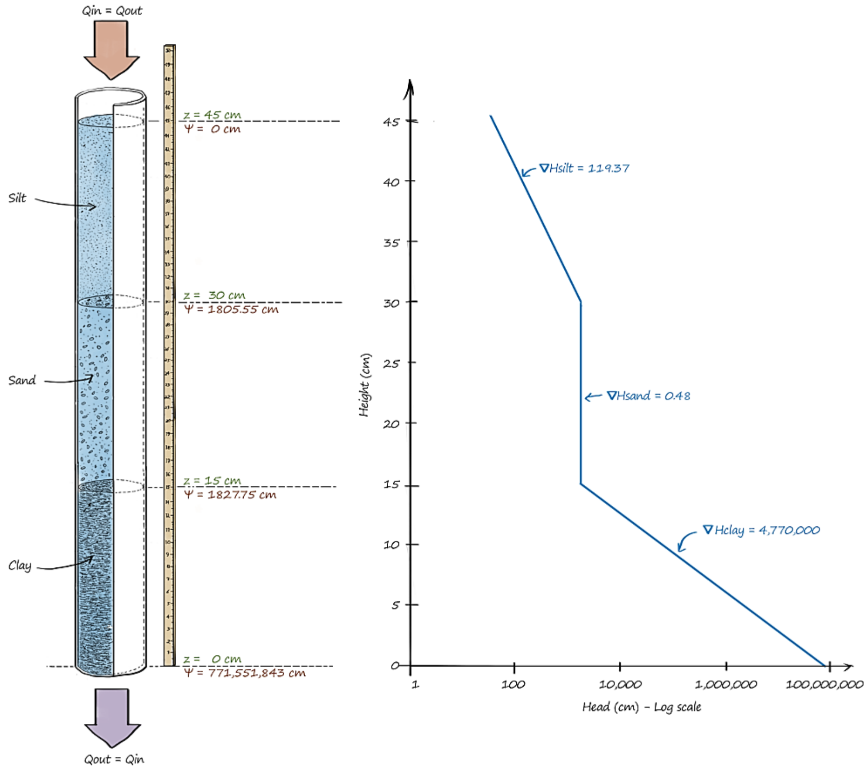

Figure 43. Column of soil with three different soils stacked on top of each other.

For flow calculations through a layered (i.e., heterogeneous) soil under steady state conditions (Figure 43), you can use a similar approach as above (Figure 41). Let's consider a case where the top layer in your soil is the silt introduced above, the middle layer is sand, and the bottom layer is clay; each layer is 15 cm thick. Essentially all three of the sediment tubes from Figure 41 are stacked on top of each other. Remember, for steady state conditions the flow into the column must equal the flow out of the column. Therefore, the flow into the top of the silt layer is equal to the flow out of the silt layer and into the sand. The flow into the sand is equal to the flow out of the sand and into the clay. And the flow into the clay is equal to the flow out of the clay. None of the water traveling through the column is lost to storage and the same volume of water that is going into the silt is leaving the tube through the clay. The medium with the lowest hydraulic conductivity controls the overall flow rate just like the slowest person in the grocery line controls how fast everyone gets through! What does this mean for the hydraulic head gradient? There must be a higher gradient through the lower K layers to have the same Q (think about the blocks and ramps in Figure 42), and the total energy must be distributed among the layers to establish steady state flow. The three layers stacked on top of each other (Figure 43) behave much like the three soil tubes behave separately (Figure 41).

Piezometers

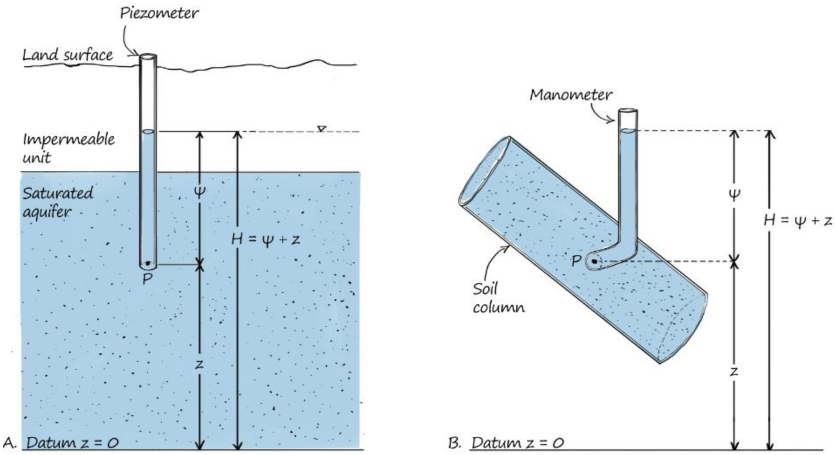

Figure 44. (a) A piezometer used to measure hydraulic head (H) and (b) a manometer for comparison. Both tools are used to measure hydraulic head (H), one in the field (a) and one in the lab (b).

Hydraulic head is a convenient variable to measure for understanding groundwater flow, not just in the lab, but also in field experiments. In the field, scientists use a tool called a *piezometer* to measure hydraulic head. Piezometers (Figure 44a) are remarkably like manometers (Figure 44b) but are placed in boreholes in the ground rather than in smaller reservoirs in the lab. A piezometer is open to the atmosphere at the top, sealed along its length, and open to groundwater flow at the bottom. The point of measurement (P) is at the bottom of the piezometer where it is slotted and allows water to flow but prevents the medium (e.g., sand or clay particles) from entering. The elevation head is the height from the datum elevation to P. The pressure head is the height of water in the tube above P. The hydraulic head is the combination of these two heights or the height of the water in the piezometer above the datum (Figure 44a).

Notice the upside-down marker (∇) in Figure 44a. Previously, we discussed that the water table is roughly the boundary between the saturated and unsaturated zone in the subsurface. However, a more precise definition is that the water table is the elevation at which the water pressure equals zero in the subsurface. In Figure 44a, the elevation of the water in the piezometer is different from the elevation of the upper surface of the saturated zone; this occurs when the saturated zone is confined.

Aquifers are underground layers where water is stored within rock fractures or unconsolidated materials (e.g., sand, gravel). For *confined aquifers*, the height to which water rises in the piezometer is the height of the potentiometric surface or the height to which the water would rise if it could get through the overlying confining layer. If the height of the potentiometric surface is above the ground surface, you could technically form a fountain of water (by puncturing through the confining layer and allowing the pressure to naturally push water upwards). For confined aquifers, the upside-down triangle in Figure 44a more specifically represents the potentiometric surface (rather than the water table or the upper surface of the saturated zone).

Unconfined aquifers (e.g., Figure 13, Figure 23, Figure 26) are not bounded by impermeable units and the upper surface of the saturated zone is at the same elevation as the water table; the height of water in the piezometer represents the height of the water table. Unconfined aquifers are connected to the surface and thus the water table can rise and fall due to surface processes (e.g., infiltration, evapotranspiration). Because groundwater is invisible from the surface, piezometers are useful for mapping the water table (unconfined aquifers) or the potentiometric surface (confined aquifers) in an area.

Multiple piezometers are often installed in an area to determine the direction of groundwater flow (Figure 45). For example, the hydraulic head measurements for three piezometers installed in a confined aquifer are shown in Figure 45. For this case, all three piezometers are installed at the *same depth* (i.e., the elevation head is the same for each piezometer). If the pressure head were the same in each piezometer, the heights of the water columns would all be the same and the water would be static (not flowing) in the *horizontal* direction. However, notice how the height of the water in each piezometer is different. Therefore, the pressure heads (and hydraulic heads) vary across the piezometers and there is horizontal flow! Remember that water flows from a high hydraulic head to low hydraulic head. So, in Figure 45, the water is moving from right to left.

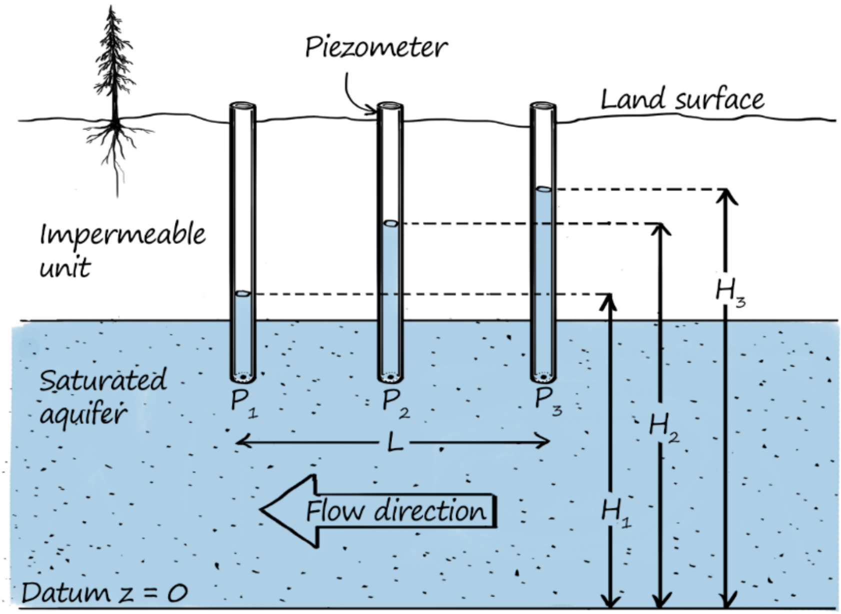

Figure 45. Several piezometers installed in a confined aquifer to understand groundwater flow direction. Groundwater flows from high hydraulic head (H_3) to low hydraulic head (H_1).

Let's apply this concept (Figure 45) to a real-life scenario. Imagine that you have two piezometers installed near a lake (Figure 46a). Both piezometers are installed in an unconfined aquifer at the same depth (the elevation head is the same for both piezometers). The pressure heads are different between the two piezometers and there is groundwater flow. The hydraulic head measured at point 1 (P_1) is 100 m and the hydraulic head at point 2 (P_2) is 75 m. We know that the groundwater flow is moving from high hydraulic head to low hydraulic head (in this case left to right). Notice how the groundwater is flowing into the lake as an input. This is due not only to the changes in hydraulic head within the soil, but also because of the hydraulic head of the lake; the lake is a part of the hydraulic head gradient.

The hydraulic head gradient of a groundwater system can help us understand the flow direction and rate of groundwater flow near the lake (Figure 46a). Because we know both the hydraulic head values and that the two piezometers are 100 m apart (L) – we can calculate the hydraulic head gradient (ΔH/L) between the piezometers in Figure 46a.

$$\frac{\Delta H}{L} = \frac{75 - 100}{100} = -\frac{1}{4}$$

Why is this information useful? Let's think back to our mass balance of a lake system from Figure 26. If we determine the flow direction and rate of groundwater flow, we can determine if the lake is gaining or losing. These shifts in water levels can have important implications for aquatic species!

Unfortunately, it is uncommon to have the luxury of (directly) measuring the height of water in piezometers above a datum (as shown in Figure 46a); if only it was that easy to obtain hydraulic head information! It is more common to measure the depth to water (DTW) as measured from the land surface at each piezometer location (Figure 46b). Then, with this information, you calculate the hydraulic head by subtracting the DTW from the elevation of the land surface (E) at each piezometer location (Figure 46b). Why is DTW a more common measurement? Well, it is often easier to measure from the top of a pipe downward than to measure upward from the datum elevation (which is often underground).

Notice the hydraulic head values are the same for both Figure 46a and Figure 46b – the methods are just slightly different. In real-world scenarios, it is common to have a variety of information available to you. With this in mind, you must always know how data is collected (if you didn't collect it yourself) and how to convert data from one form to another. A common error with DTW measurements is to forget about the height of the piezometer casing at the surface (the part of the piezometer that sticks up out of the ground). Therefore, if the DTW is measured from the top of casing (rather than from the land surface elevation) always make sure to relate the top of the casing to the datum.

In the previous example (Figure 46), we focused on lateral groundwater flow. But sometimes, groundwater flows vertically (up and down) (Figure 47) rather than laterally (left to right) (Figure 46). If this is the case, how can you use the water level in piezometers to determine flow direction? Well, if the water were *only* moving vertically, then the hydraulic head gradient would also only be in the vertical direction, meaning that the water level in the piezometers would change from one elevation to another but not laterally (at the same elevation). To better understand this, look at Figure 47a. All the piezometers have the same hydraulic head – it would appear there is no flow. However, the hydraulic head gradient could be changing in the vertical direction, and indeed this is the case.

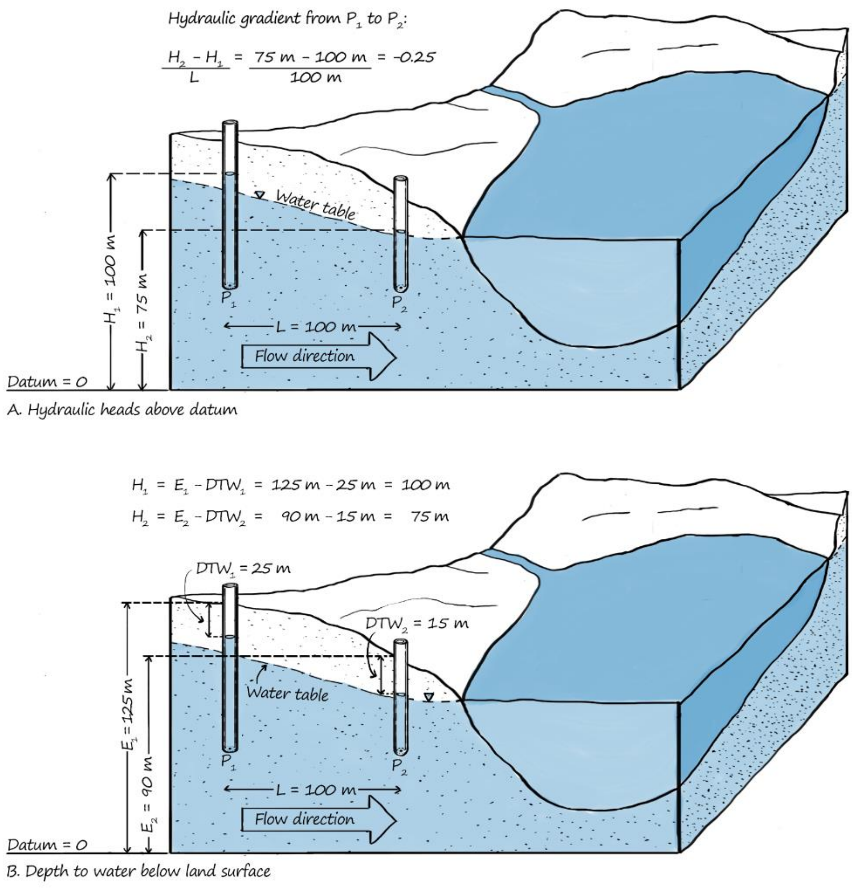

Figure 46. Two piezometers installed in an aquifer. The piezometers are used to determine the flow direction (left to right) using the change in hydraulic head (H). The hydraulic head is a) measured directly as the height of the water above a datum and b) calculated using the depth to water from the land surface (DTW) and the elevation of the land surface above the datum (E).

A piezometer tells you information about the elevation and pressure head at point P (the open screened bottom of the piezometer). Therefore, if you want to know if the water in your system has a *vertical* hydraulic head gradient, you need to install the piezometers at different elevations (Figure 47b). Furthermore, if you want to measure the vertical hydraulic head gradient at one location (point P), you need to install the piezometers right next to each other, or perhaps even in the same hole; this is called a piezometer nest or a multilevel piezometer.

For vertical flow, nested piezometers installed at different depths (Figure 48) will have both varying pressure heads and varying elevation heads (height above the datum elevation). To calculate the vertical hydraulic head gradient, you must know the elevation of the bottom opening (elevation head) and the hydraulic head in each of the nested piezometers. The hydraulic head gradient is the difference in the hydraulic heads divided by the difference in the elevation heads. For example, in Figure 48, the vertical hydraulic head gradient between piezometer 1 and piezometer 2 is -1; the change in hydraulic head is –5 m (85 m – 90 m) and the change in the elevation of the piezometers is 5 m (65 m – 60 m). To find the vertical hydraulic head gradient between the other piezometers, you can repeat this calculation. The vertical hydraulic head gradient between piezometer 2 and piezometer 3 is also -1. More generally, because the change in hydraulic head is constant between the nested piezometers, and the piezometers are installed at equal depth intervals (e.g., 55 m, 60 m, and 65 m) – the hydraulic head gradient is *constant* vertically.

It is uncommon to install piezometers all at the same elevation head at a field site (e.g., Figure 46). This is because logistically, it is difficult to install piezometers at the same depth. Additionally, it is not essential to drill piezometers at the same depth if the groundwater is shallower in one zone and deeper in another. Installing piezometers deeper than necessary could cost you more time (hand drilling) or money (if using a mechanical drill). But this complicates things. If your piezometers are installed at different elevations, you can't just use the depth of water in the piezometer to determine flow direction (like in Figure 46) – you must also know the elevation head of each piezometer.

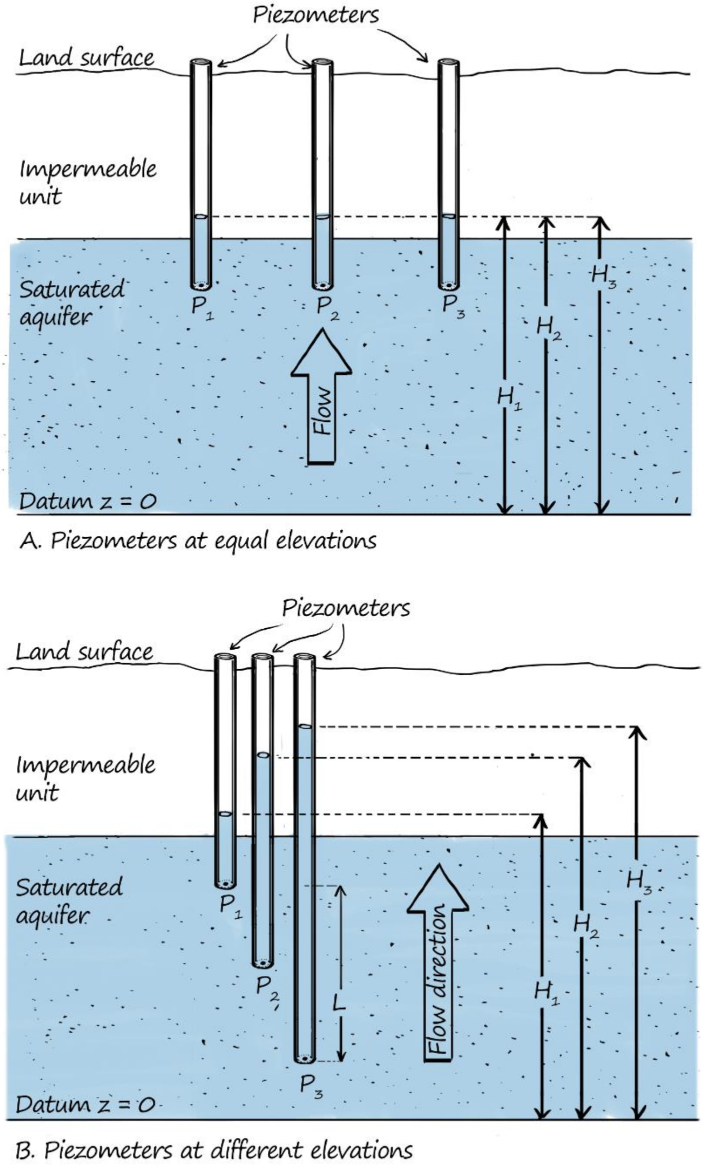

Figure 47. Vertical flow in a system where (a) piezometers are all installed at the same elevation head but do not show differences in hydraulic head and (b) nested piezometers at different elevation heads do show the differences in hydraulic head.

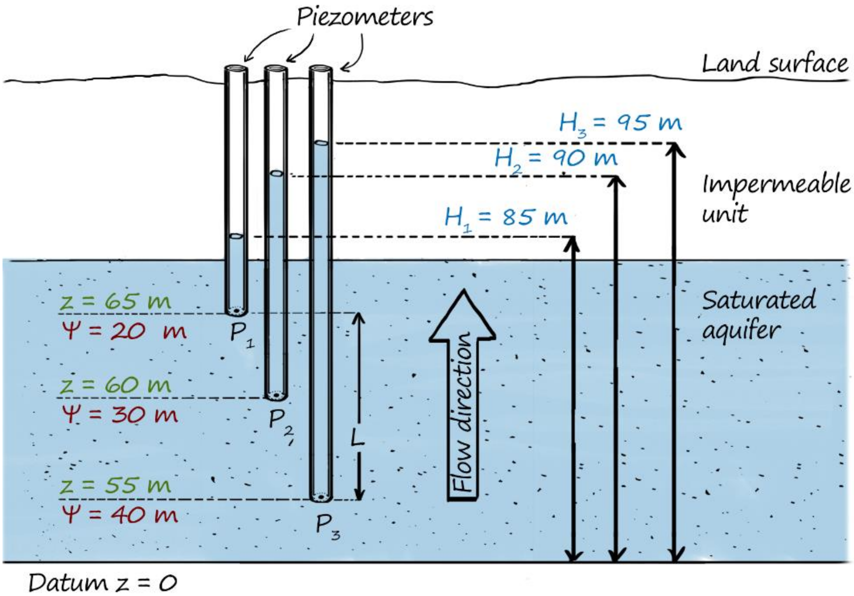

Figure 48. Nested piezometers with a constant hydraulic head gradient in the vertical direction.

Let's look at our lake system again, but with two piezometers installed at *varying depths* (Figure 49); piezometer 2 is closest to the lake and piezometer 1 is furthest from the lake. The height of the water column above P_2 in piezometer 2 is 100 m, and the height of the water column above P_1 in piezometer 1 is 65 m. Using only the height of the water columns in the piezometers (piezometer 2 = 100 m and piezometer 1 = 65 m), we might think the flow is going from right to left (piezometer 2 to piezometer 1). However, the height of the water columns in the piezometers is only the height above the bottom of the piezometer - which is the pressure head (Figure 44). Remember that groundwater flow is determined not by pressure head (the height of the water column *above the bottom of the piezometer*) but by hydraulic head (the height of the water column above the *datum elevation*) (Figure 44).

To understand the hydraulic head values surrounding our lake system (Figure 49), we must use both the pressure head and elevation head at each piezometer (Figure 44a). We already discussed the pressure heads (piezometer 2 = 100 m and piezometer 1 = 65 m). Now, let's move on to the elevation heads. To determine the elevation heads, we must first choose a datum elevation. Remember, we can choose any datum elevation – but we should make it reasonable for our system. For example, if the bottom of piezometer 2 (P_2) is installed 110 m beneath the ground surface and the bottom of piezometer 1 (P_1) is installed 80 m beneath the ground surface, we probably wouldn't want to choose 100 m beneath the ground surface (at piezometer 2) as the datum elevation. If this were the datum, piezometer 2 would have a negative elevation (e.g., an elevation of –10 m below the datum). This datum elevation isn't wrong, it might just be unnecessarily confusing.

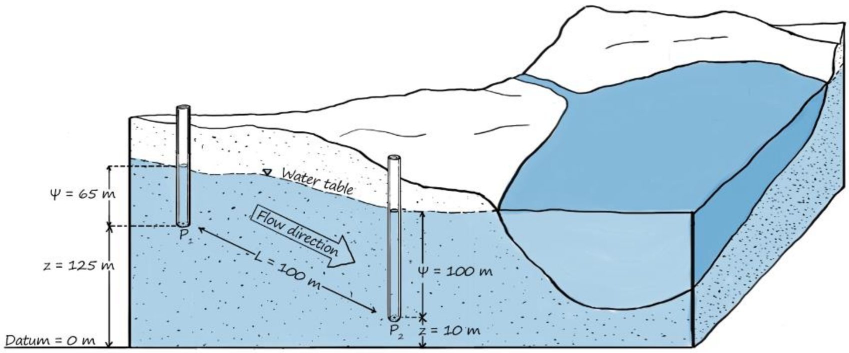

Figure 49. Two piezometers installed in an aquifer at varying depths. The piezometers are used to determine the flow direction (left to right) using the change in hydraulic head (H).

For simplicity, we will choose a datum elevation of 120 m *beneath the surface at piezometer* 2 (so that all the elevation heads are positive) (Figure 49). For piezometer 2, the bottom of the piezometer is installed 110 m below the land surface. Therefore, the elevation head is 10 m above the datum elevation, and the elevation head is 10 m. The bottom of piezometer 1 is 125 m above the datum, and the elevation head is 125 m (Figure 49). Now that we know the pressure head and elevation head for each piezometer, we can calculate the hydraulic head and determine the direction of groundwater flow.

Piezometer 1:

$$H = \psi + z$$
$$H_{piezometer1} = 65\ m + 125\ m = 190\ m$$

Piezometer 2:

$$H = \psi + z$$
$$H_{piezometer2} = 100\,m + 10\,m = 110\,m$$

Notice that the hydraulic head is greater in piezometer 1 ($H_{piezometer1}$ = 190 m) than in piezometer 2 ($H_{piezometer2}$ = 110 m), even though the height of the water column in piezometer 1 is smaller ($\psi_{piezometer1}$ = 65 m, $\psi_{piezometer2}$ = 100 m). Therefore, the water is moving from left to right (piezometer 1 to piezometer 2). This example (Figure 49) is important to keep in mind for field measurements. Piezometers are (mostly likely) not all the same length and their bottom openings will all be at different elevations. Therefore, you must always know the elevations of the bottom of the piezometers. Unfortunately, in the field you can't simply cut into the Earth to see how deep the piezometers are relative to one another (like in Figure 49). Therefore, you must look up the drilling logs (from when the piezometers were installed) to find the depth below the land surface to which they were drilled. Once you collect these data, you will have all the information you need to draw a diagram like Figure 49 and begin to visualize the groundwater.

Conclusion

This chapter was a big leap in your hydrogeologic journey. Not only did you learn the fundamental equation for flow through a porous medium (Darcy's law), but you also began visualizing hydraulic head distributions and gradients in the subsurface (using piezometers). With these two fundamental concepts in hand, you are starting to understand the basics of groundwater movement. However, this is only the tip of the iceberg. For all the examples above, we presented flow in one dimension. Flow from one piezometer (or one manometer) to the next. But what if the piezometers in Figure 49 were not all installed in a straight line extending out from the lake? How would you visualize the groundwater movement in multiple dimensions? Or how do we know what the hydraulic head gradient looks like between piezometers? We will give you more tools for visualizing groundwater movement in the next section.

What to Remember

Important Terms	
aquifers	hydraulic head
confined aquifer	hydraulic head gradient
elevation	manometer
elevation head	piezometer
fluid pressure	potential energy
flux	potential gradient
head	potentiometric surface
heterogeneous medium	pressure head
homogeneous medium	unconfined aquifer

Important Equations
$Q = -KA\frac{H_2 - H_1}{L}$
$p = \rho g \psi + p_0$
$H = \psi + z$
$q = \frac{Q}{A}$

Important Variables		
Symbol	**Definition**	**Units**
Q	Flow	L^3/T
K	Hydraulic conductivity	L/T
H	Hydraulic head	L
L	Length	L
A	Cross-sectional area	L^2
p	Fluid pressure	ML/T^2
ρ	Fluid density	M/L^3
g	Gravity	L/T^2
ψ	Pressure head	L
ρ_0	Atmospheric pressure	ML/T^2
z	Elevation head	L
q	Darcy flux	L/T

Chapter 4

Visualizing Groundwater Flow

Introduction

Groundwater is one of our most valuable water resources. All living things require water to grow and reproduce. Humans use groundwater for irrigation, industrial, and domestic uses. Although we can't *see* groundwater, it is important to *visualize* it; in other words, if we could see underground, what would the water be doing? If we remove water from the ground – how fast would the ground refill? Slowly? Quickly? We must visualize groundwater to understand how much is available and how it might respond to the stresses that we impose.

Imagine what you could learn if you could put a tiny GPS device on a water molecule and track its path. Where does it go? How fast does it move? In what direction? Answering these questions about an individual molecule (or a group of molecules) could supply valuable information about the groundwater system. Furthermore, what if that water molecule was contaminated (i.e., it was mixed with some nasty chemical). Wouldn't you want to know where the contaminated molecule is moving? I mean – what if it was moving toward your house!?

Now that you understand how important it is to see groundwater, you must also understand that hydrologists most often do not want to just *see* groundwater – they want to *watch* it! To watch groundwater, you must collect data over a period of time. This is called monitoring. This chapter introduces visualizing and monitoring groundwater for real-world scenarios. We have already learned that the hydraulic head gradient controls groundwater flow: water flows from areas of high hydraulic head to areas of low hydraulic head. We have also learned how piezometers are one common method for measuring hydraulic head in the field. You may not realize it, but you already have most of the tools you need to visualize groundwater in real-world scenarios. All that remains is some imagination.

Visualizing Hydraulic Head Gradients

Before we dive into applications, let's take a moment to visualize head gradients. Imagine that you are holding a plastic rectangle (perhaps it's the lid of a plastic storage tote). Hold the rectangle by two opposing corners (corners that are diagonally across from each other). One corner is in your left hand, and one is in your right hand (Figure 50a). Now lower the corner in your left hand so the rectangle is at an angle (Figure 50b); you now have a plane with a constant gradient. If you lower the left-hand corner more, it will be a steeper gradient; if you lower it less it will be a shallower gradient. If you stand outside in the rain with this rectangle, where does a raindrop go? If a raindrop lands on the rectangle near your right hand, it flows to where your left hand is – in the direction of the highest gradient (Figure 50b). If you switch and lower your right hand, the raindrop moves from your left hand towards your right hand.

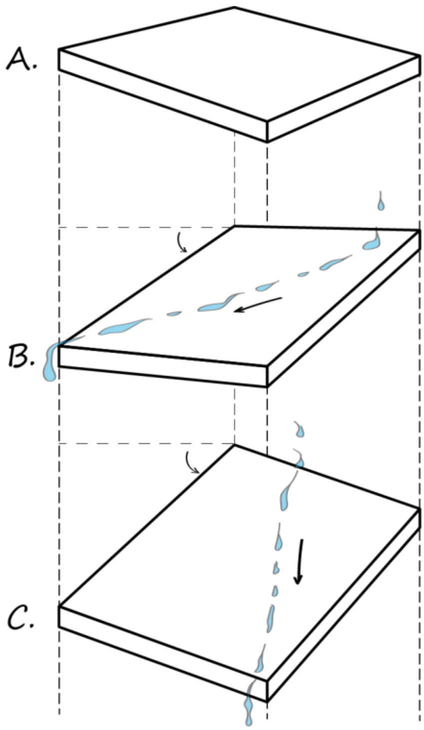

Figure 50. A rectangular plane tilted in multiple directions. Water that falls from the sky will flow in the direction of the greatest gradient.

OK, next imagine that you were to not only lower the corner of the rectangle in your left hand, but also tilt the corner of the rectangle (that's near your belly button) towards you. The rectangle now has a gradient in multiple directions. So, if you stand out in the rain, where does a raindrop go? Well, the answer is – it depends. The exact path of the raindrop is dependent on both where the raindrop lands and how steeply you are tilting the rectangle in each direction. If you tilt the rectangle steeply towards your belly button and only slightly toward your left hand, the water flows more towards your belly button (Figure 50c). If you tilt the rectangle evenly in both directions, the raindrop will choose a path somewhere in the middle (e.g., flow off the side toward your left hip). However, no matter where a raindrop falls on the rectangle, it will flow from its landing point in the direction of the *steepest gradient.*

To further visualize gradients, imagine that you are building a house. Some modern homes have intricately designed roofs with a complex configuration of ridges and valleys. When designing the roof, your architect considers the steepness and orientation of the roofing angles. A poorly designed roof directs all the water (or snow) from the front of the house towards your porch; come winter or monsoon season, you will not be happy with this design. Understanding the head distribution in a groundwater system is no less important. What if the hydraulic head gradient causes water in your yard to move toward your house – pooling water and saturating the soil at the foundation? Or what if the gradient is causing your septic tank (your wastewater system) to drain toward your drinking well? In both cases, you would surely want to know the direction of the steepest hydraulic head gradient to try to mitigate risk.

Unfortunately, predicting groundwater flow is more complicated than observing water flowing off a roof. We can't directly see hydraulic head gradients underground. Therefore, predicting groundwater movement is more like predicting where water is flowing off a roof without getting to look at the roof. This is where measurements and your imagination come in. As hydrogeologists, we use clues from piezometers to visualize gradients. Remember, piezometers provide a snapshot of the hydraulic head in one specific location (Figure 44a). We can piece together these snapshots to make a best guess as to what the hydraulic head gradient looks like underground.

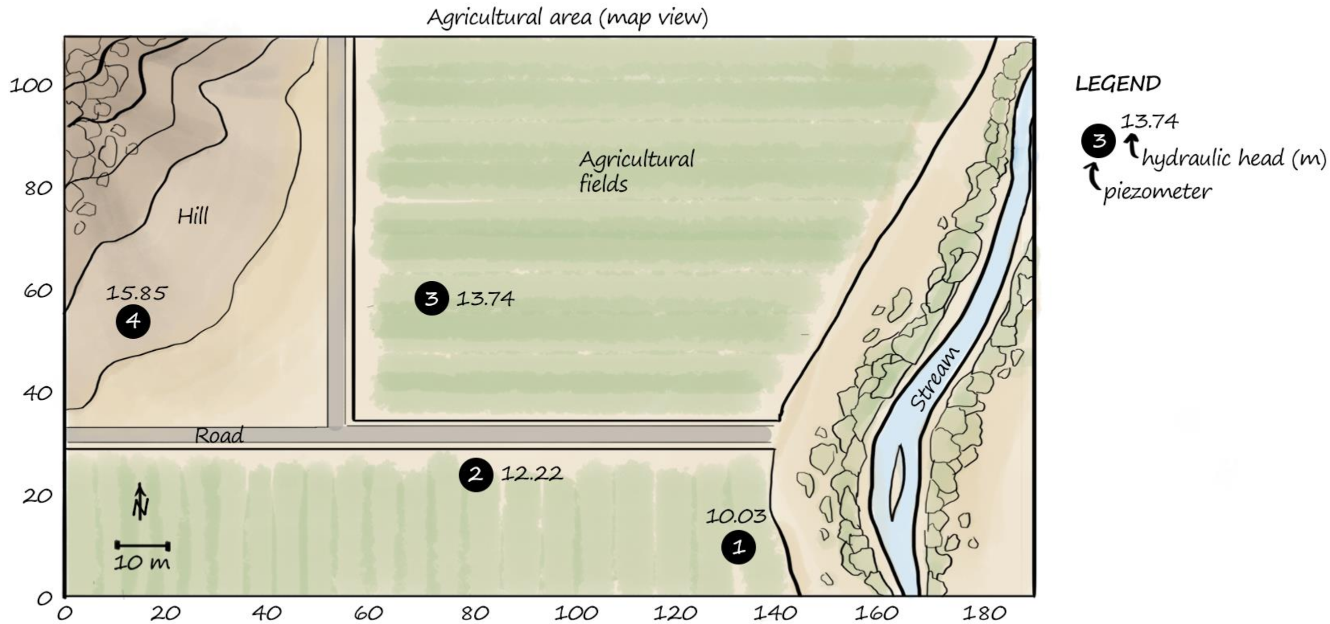

Figure 51. Aerial view of four piezometers on an agricultural field adjacent to a stream. The numbers along the bottom are distances west to east in meters. The numbers along the left side are distances south to north in meters. A scale is also provided in the bottom left corner. The circled numbers represent the piezometer labels, and the numbers next to the circles represent the hydraulic head in meters. Example: piezometer 1 has a hydraulic head of 10.03 m.

Agricultural Field with a Few Piezometers

Imagine that you have an agricultural field adjacent to a river, with four piezometers installed. Figure 51 shows an aerial map of your agricultural field (i.e., what the landscape looks like from an airplane). The piezometers allow you to measure the hydraulic head at each location. Note: if you need a review of piezometers, take a moment, and refer to the previous chapter. Notice how each circle in Figure 51 has a number inside of it; this is the piezometer label. We will use these labels as a reference, so that you know which location we are referring to in our examples. Note: labeling groundwater monitoring locations is a widespread practice. However, identification numbers are often much more complicated on real sites (e.g., NW3-456b) and many locations must be registered using these labels.

Let's discuss how we can use piezometer observations to interpret something about groundwater. Begin by looking at piezometers 1 and 2 in Figure 51. The hydraulic head at piezometer 2 is 12.22 m and the hydraulic head at piezometer 1 is 10.03 m. For now, assume piezometers 1 and 2 are the only hydraulic head values that you know. What is the hydraulic head gradient between the two piezometers? If there are only two measurements, we must assume that the gradient is uniform and aligned in the direction of the piezometers. To visualize this, go back to the plastic rectangle example discussed above (Figure 50). Imagine holding an invisible rectangle underground so that the plane goes through the two piezometers and is tilted such that it slopes downhill from 1 to 2 (Figure 52). If water were to fall on the rectangle, which direction would it travel? Toward piezometer 2. This is a simple, initial approximation of the hydraulic head gradient between piezometers 1 and 2.

Using a similar approach, how can you visualize the hydraulic head gradient between piezometers 1, 2, and 3 (Figure 51)? A rigid triangle (rather than a rectangle) might be helpful here (Figure 53). Place one corner of the triangle at piezometer 3 (13.74 m), another at piezometer 2 (12.22 m) and the last at piezometer 1 (10.03 m). How is the triangle tilted in space? The corner at piezometer 3 is the highest and the corner at piezometer 1 is the lowest. But don't forget about piezometer 2 – the triangle should also be slightly tilted downward toward piezometer 2. The water moves towards the lowest side of the triangle between piezometers 1 and 2, rather than directly towards piezometer 1 as we initially assumed. With only two piezometers, we had to assume a uniform two-dimensional gradient between piezometers 1 and 2 (Figure 52), but we couldn't account for the gradient perpendicular to that direction. The total gradient, which defines flow, is three-dimensional. Using a rigid triangle, we can better visualize the water table in the area surrounded by the piezometers (Figure 53). However, the hydraulic head gradient outside of this area is still not defined.

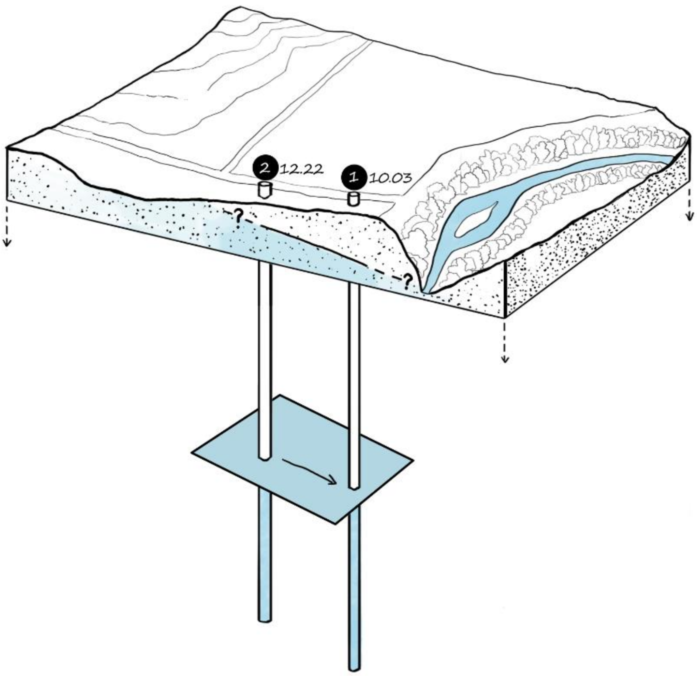

Figure 52. Agricultural field with a constant gradient between piezometers 1 and 2. The rectangle underground represents the estimated hydraulic head gradient of the water table (from piezometer 1 to piezometer 2 only). Note: the rectangle *is not* the actual water table, but rather a visual representation of the predicted gradient based on piezometer information, shown projected downward on this figure so that it is visible. The actual water table is closer to the surface (shown as a black line along the face of the block diagram).

To visualize the hydraulic head gradient using the information from all four piezometers (1, 2, 3, and 4) (Figure 51), you can use two rigid triangles (much like a complex roof system); one triangle flows to the next, as one roof tile flows onto the next. Place one triangle between piezometers 4, 3, and 2 (Figure 54) and another between 3, 2, and 1 (same as in Figure 53). For the first triangle, the highest corner is at piezometer 4, the lowest corner is at piezometer 2, and piezometer 3 is in between. The second triangle's orientation is the same as in Figure 53. With this configuration, water will continuously flow from triangle 1 to triangle 2 (i.e., the molecule doesn't change direction abruptly from one level to the next). Specifically, the water will follow a continuous path determined by the hydraulic head gradient of the triangles (Figure 54). Using all four piezometers, our estimate of what the water table looks like is gradually improving!

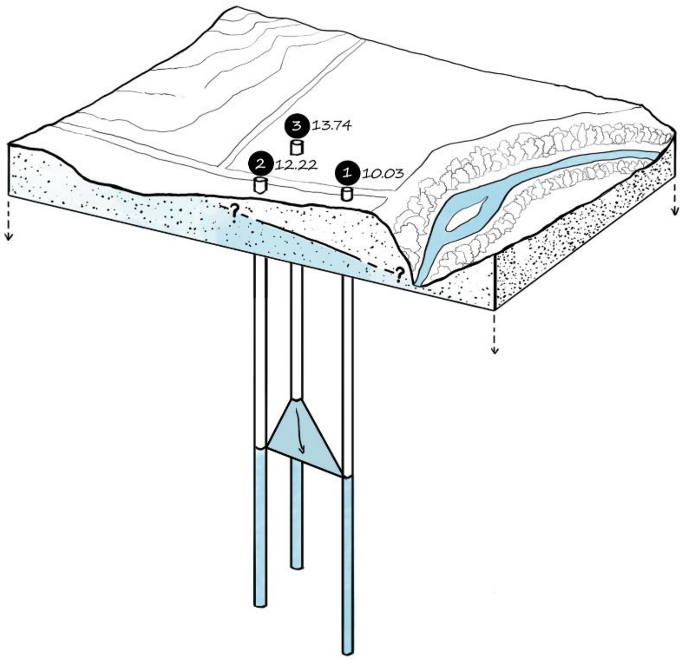

Figure 53. Agricultural field with piezometers 1, 2, and 3. The triangle underground is constructed to show the hydraulic head values at each location on a plane. Groundwater will move in the direction of the total hydraulic head gradient (black arrow). The gradient of the water table outside of the piezometers is unknown. Note: the plane representing the water table is projected downwards so that it is visible. The actual water table is closer to the surface (shown as a black line along the face of the block diagram).

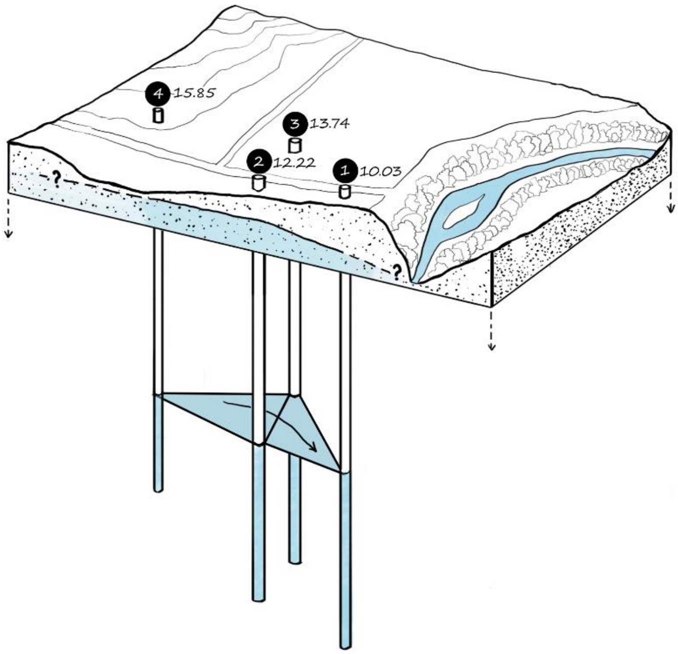

Figure 54. Agricultural field with piezometers 1-4 installed. Two rigid triangles are considered, much like a complex roof system. The hydraulic head gradient is constant along the plane of the triangle. Direction of groundwater flow is shown by a black arrow. Note: the planes representing the water table are projected downwards so that they are visible.

Now, consider the same agricultural field, but with 13 piezometers (Figure 55). You have even more information! Take a closer look at Figure 55. You will notice a new symbol: black triangles representing stream gages. A *stream gage* records *the stage* or the elevation of the water surface in a stream. Have you ever been to the beach and dug a hole in the sand near the ocean? Even if the sand at the surface is relatively dry, after some digging, your hole will eventually fill with water. Why is this? It is because the water level in the ground (i.e., the water table) is nearly the same height as the surface of the ocean. In fact, the water underground is connected to the ocean. As the level of ocean changes, the water level under the ground changes. Our stream system in Figure 55 is similar. The stream flowing across the east side of our field is connected to the water table, so the stream level provides additional information about the water table elevation.

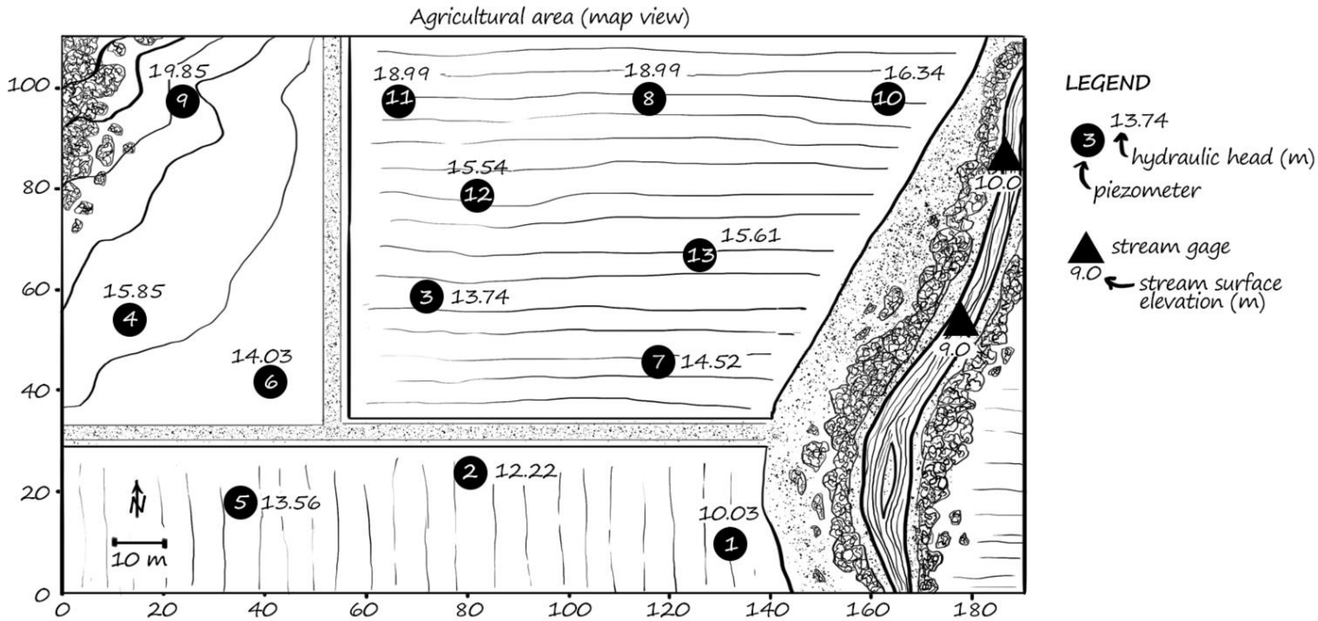

Figure 55. Aerial map of agricultural field adjacent to a river (same field as Figure 51) with 13 piezometers installed, and stream gages (black triangles) showing the water level in the stream. The datum elevation is the same for both the stream gages and piezometers. Therefore, for this example, the stage is equal to the water table elevation.

The stage of a stream is measured from a datum (just as piezometer levels are measured relative to a datum). The elevation of the water *above* the datum elevation is the stage. However, typically, stream gages are not measured using the same datum elevation as nearby piezometers. For example, when measuring the stage, you might choose a local elevation (like the stream bed) as your datum, so that when a stream is "dry" the stage is zero. For piezometers, you might choose a datum that best relates water levels in multiple piezometers to a common hydraulic head definition (e.g., Figure 49). Therefore, in the real world, if you wish to use stage values to infer water table elevations, you will often need to account for differences in datum elevations. For simplicity, we are going to assume that the datum elevation is the same for both the stream gages and the piezometers throughout this book. Therefore, the value next to the triangles in Figure 55 is specifically the stream surface elevation above a common datum; this means you can use the stream gage values to infer the water table elevation, directly.

To visualize how the stream connects to the groundwater – first, use the stream surface elevations (from the two gages) to infer the gradient of the stream. The stream surface elevation at one gage is 10 m and the stream surface elevation at the other gage is 9 m (Figure 55). Knowing this, which direction is the stream flowing? Begin by visualizing the gradient of the stream using the same rigid plane method from above (e.g., Figure 52). Because you only have two measurements, you will use a rectangular plane. The plane will tilt so that it is lower at the gage with a value of 9 m and higher at the gage with a value of 10 m. Therefore, the stream is flowing southward. Cool!

To visualize the water table in Figure 55 begin by using the same rigid triangle method that you used for four piezometers (Figure 54) but for all 13 piezometers. Place triangles between adjacent piezometers to create a mosaic of angular planes (Figure 56). This is a much more detailed representation of the water table than what you can construct with only 4 piezometers (Figure 54). Next, use the stream surface elevations to further infer the gradient of the water table. Look at the mosaic of planes near piezometers 10, 13, 7, and 1. How do they connect to the stream? The hydraulic head at each of these piezometers is higher than that of the stream, so the planes are dipping eastward.

Figure 56 is a more accurate representation of the water table, but there are still abrupt transitions between the triangles. The rigid planes simply don't represent the water table realistically. In reality, the hydraulic head gradient between piezometers isn't uniform (i.e., a rigid plane or series of rigid planes); rather the water table continuously bends and curves like a wave (Figure 57).

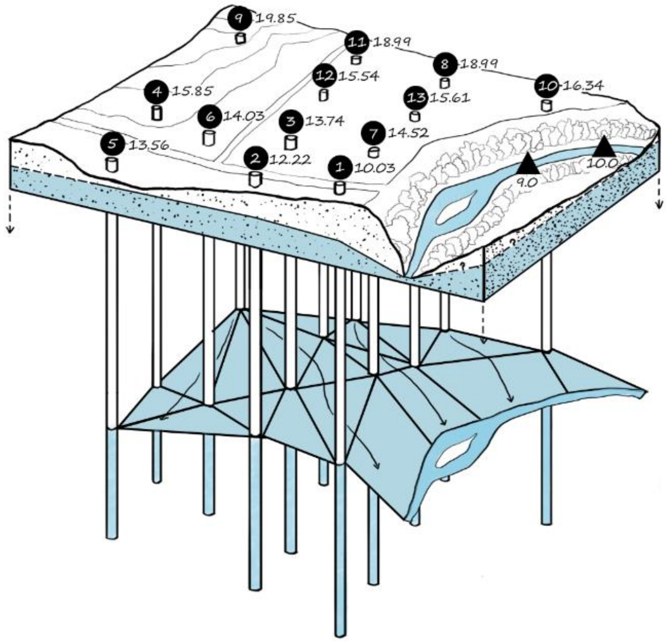

Figure 56. Agricultural field with 13 piezometers (black circles) and two stream gages (black triangles) installed. The datum elevation is the same for the stream gages and piezometers. Therefore, for this example, the stage is equal to the water table elevation. A mosaic of constant triangular planes demonstrates the complex nature of determining the direction of groundwater flow. Note: the planes representing the water table are projected downwards so that they are visible.

In all cases, when visualizing groundwater in the subsurface, you are limited by the number of piezometers you have (or how much information you have). If you only have a few piezometers throughout a field, you can't know detailed hydraulic head distributions or what is happening between piezometers – you must make simplifying assumptions. For example, in Figure 57, if you only had the information from the outer piezometers (e.g., 10, 9 and 5), you would need to assume that the water table between the piezometers is planar (i.e., one rigid triangle). This limits your knowledge significantly. Using a single plane, you can't know the hydraulic head distribution or where the water table gradient varies throughout the field.

The more piezometers you have, the more locations you can be certain about. Unfortunately, in many cases, you must estimate the hydraulic head

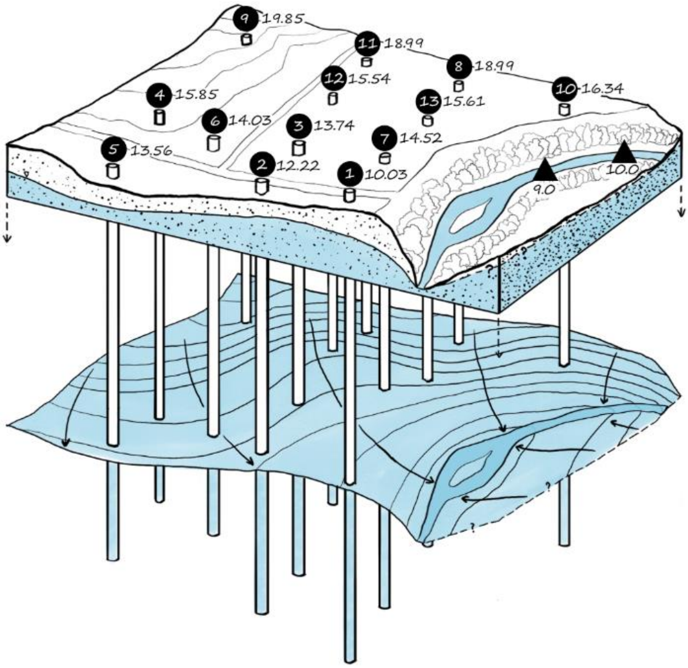

Figure 57. Example of the continuously curving water table beneath an agricultural field with 13 piezometers (black circles) and two stream gages (black triangles) installed. The hydraulic head distribution is more complicated than a mosaic of triangular planes (e.g., Figure 56). Note: the water table is projected downwards so that it is visible.

gradient based on what information is available. For example, there aren't any piezometers installed on the right side of our stream – so how can we estimate what the water table looks like? For now, we must make a make a reasonable guess that the right side roughly mirrors the left. In most cases, factors such as funding and legal permitting limit the number of piezometers. Granted that (in many cases) we will have limited knowledge available to use, we need additional tools to visualize the hydraulic head distribution between piezometers.

Equipotential Lines

A helpful way of visualizing and inferring a distribution of head is to use *equipotential lines*. "Equi" means equal, and "potential" is the energy potential, in the case of groundwater, hydraulic head. Therefore, equipotential lines are lines of equal hydraulic head.

When hiking, elevation is the energy potential that matters. So, contour lines (i.e., contours) on a map are equipotential lines, or lines of equal elevation. For contour lines, any point on a 40-foot contour has an elevation of 40 feet (above a datum elevation). By convention, lines are drawn at evenly spaced *contour intervals* so that there is an equal change in elevation between each line. For example, if the contour interval is 10 feet, there are contour lines at 40 feet, 50 feet, 60 feet, etc. If a point falls between contour lines, the elevation at that point is between the contour values. For example, 42, 46, or 48 all fall between the 40-foot contour and the 50-foot contour. Contour intervals allow the viewer to immediately see a gradient based on line spacing. If the contour lines are close together, the elevation changes significantly over a short distance, so the slope is steeper. Similarly, if the lines are further apart, the slope is less steep (less change in elevation over the same distance).

Figure 58a shows a topographic map with a 10-foot contour interval. Note that Figure 58a is a topographic map and Figure 58b is a cross section. On the topographic map there is a cross-section line from A to B. The cross section is a straight cut through the topography along the A-B line. Think of a birthday cake; the topographic map is the top of the cake where “Happy Birthday” is written, and the cross section is a slice of the cake (viewed from the side).

A cross section is a useful way to visualize changes in gradient. On the topographic map, follow the A-B cross-section line from left to right (Figure 58a). First, the contour lines increase from 40 m to 70 m and are spaced relatively far apart. Next, the lines are closer together from 70 m to 30 m. Finally, they are further apart again from 30 m to 10 m. As a comparison, look at the cross section (Figure 58b); the y-axis is elevation (m), and the x-axis is the same scale as the topographic map (Figure 58a). Note: sometimes cartographers (people who produce maps) will make axes different scales to highlight features. Most often, the vertical axis is made larger (i.e., exaggerated) to highlight relatively small slopes that are common for certain gradients such as topography or water tables. In our case, because the horizontal and the vertical axes are the same scale, Figure 58 shows no vertical exaggeration. The thin dashed lines down from the contours (Figure 58a) mark the elevations at each point along the cross section (Figure 58b). By connecting the dots, you can visualize the landscape. There is a hill! The left side of the hill has a gentler slope (where contour lines are farther apart) and the right side has a steeper slope (where contour lines are closer together).

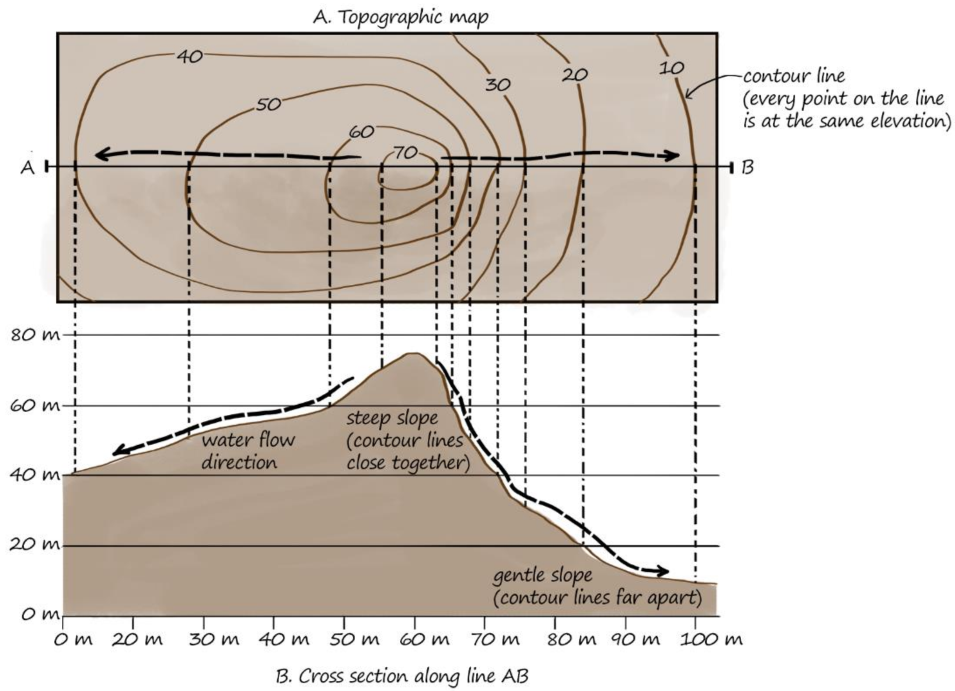

Figure 58. A topographic map (a) with a cross-section line from A to B and a cross section (b) along line A-B. The dashed lines are the direction of water flow (flow lines). There is no vertical exaggeration (the vertical and horizontal scales are the same).

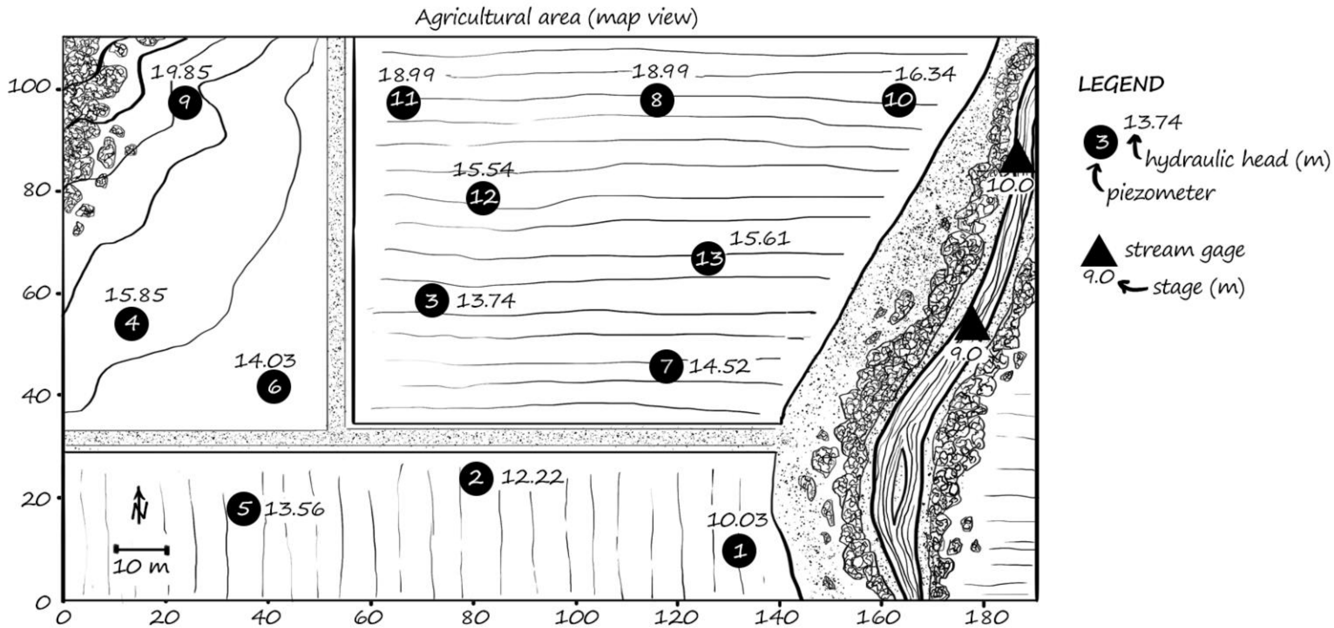

Figure 59. Agricultural field with 13 piezometers installed. This is the same as Figure 55.

Cross sections (Figure 58b) are often useful for answering specific questions. The orientation of a cross section depends on the geographic location of interest. For example, if you were interested in knowing the slope of a road, you could take a cross section of a topographic map along the road of interest. Similarly, if you wanted to know the geologic units and structures relevant for oil and gas exploration, you could take a cross section of a specific area to gather information about the geology of the subsurface. In both scenarios, the cross sections only tell you about the slice of the earth that you have chosen. The cross sections do not tell you how the land slopes perpendicular to the road or about the orientation of other geologic formations surrounding the cross section.

Topographic maps (Figure 58a) and cross sections (Figure 58b) can have a hydrologic meaning. For example, they could be used to help determine runoff direction. If rain falls on a hillside (ignoring infiltration), it will run downslope (down gradient). The thick dashed lines in Figure 58 are flow lines that show the direction of runoff along the cross-section line. The water flows from high to low contours, and the flow path runs perpendicular to the contour lines (Figure 58a). Other flow lines are not shown but would run perpendicular to contour lines throughout. Unbelievably, mapping groundwater flow beneath an agricultural field is like mapping runoff! Only, rather than using land surface elevation information, you must use hydraulic head.

Practice drawing equipotential lines for Figure 59 with integer values as the equipotential interval (e.g., 16, 17, 18 m). The idea is to connect locations of equal hydraulic head, just as you would connect points of equal elevation for a topographic map. Start by choosing an integer head value, say 11 m. Notice that none of the piezometers have a measured value of exactly 11 m. However, I bet you can find two adjacent piezometers that have values more than and less than 11 (e.g., piezometers 1 and 2). Based on the hydraulic head in adjacent piezometers, estimate where the head is equal to 11 m between them. Note: you will assume a linear gradient as we did with the rigid rectangles and triangles planes above (e.g., Figure 56). Draw a point there. Infer all the other locations on the map where the hydraulic head is equal to 11 m and add additional points (e.g., between piezometers 1 and 7). Using the available information, connect your points to draw an equipotential line for 11 m. Repeat for each interval value (e.g., 12, 13, 14 m, etc.). Once you finish, you will have an equipotential map! Keep in mind that drawing equipotential lines will require some imagination (and a pencil with an eraser). You will surely need to infer the hydraulic head at many locations. Remember that the stream is connected to the water table. Therefore, the stream gage information (i.e., the elevation of the water surface above a common datum) is a good approximation of the hydraulic head at locations in the streambed.

Figure 60 shows the agricultural field with equipotential lines. On the map, notice where the lines are closest together (e.g., in the northeast corner); this is where the hydraulic head gradient is greatest. Also notice where the lines are furthest apart (e.g., southwest corner); this is where the hydraulic head gradient is lowest. You can begin to see why a single rigid plane doesn't work well for this system; the gradient is not constant throughout. A flat plane can only describe a simple system. For groundwater systems, we need multiple points to estimate the value between adjacent wells, because the equipotential surface varies just like a contour map. The good news is that we already know how to interpret gradients using the skills that you have developed for topographic maps. The next step is to understand why these gradients form and how they can be used to better understand groundwater flow.

Imagine that the soil beneath the agricultural field in Figure 60 is homogeneous. If this is the case, then the equipotential lines do not just tell you about the hydraulic head gradient, they also inform you about the relative groundwater flux. Remember, flux is the product of the hydraulic conductivity and the hydraulic head gradient (Equation 9). Therefore, in a homogeneous system (constant K), the flux is the greatest where equipotential lines are closest together (e.g., northwest corner of Figure 60). The flux is lowest where the equipotential lines are farthest apart, (e.g., southwest corner of Figure 60). Where equipotential lines are evenly spaced, flux is constant. Keep in mind that even for a homogeneous system, an equipotential contour map does not tell you about the *absolute magnitude* of the flux – you need the value of K to calculate that.

Equipotential lines can inform us about the values of hydraulic head and the relative flux in homogeneous systems. However, what if we want to know the *direction* of groundwater flow? Can equipotential lines tell us that? Start by observing the hydraulic head at piezometers 11 and 12 (Figure 60). Is water flowing from piezometer 11 to 12? Well, in a sense, yes. We know water flows from high hydraulic head (18.99 m) to low hydraulic head (15.54 m). We can even find the gradient between the two piezometers and multiply by K to get the flux. But does this mean that a water molecule released at piezometer 11 flows *directly* to piezometer 12? Not necessarily. Why? Because water flows in the direction of the *maximum* gradient (Figure 50) just as water flows downhill in the steepest direction (it doesn't flow sideways across a hill).

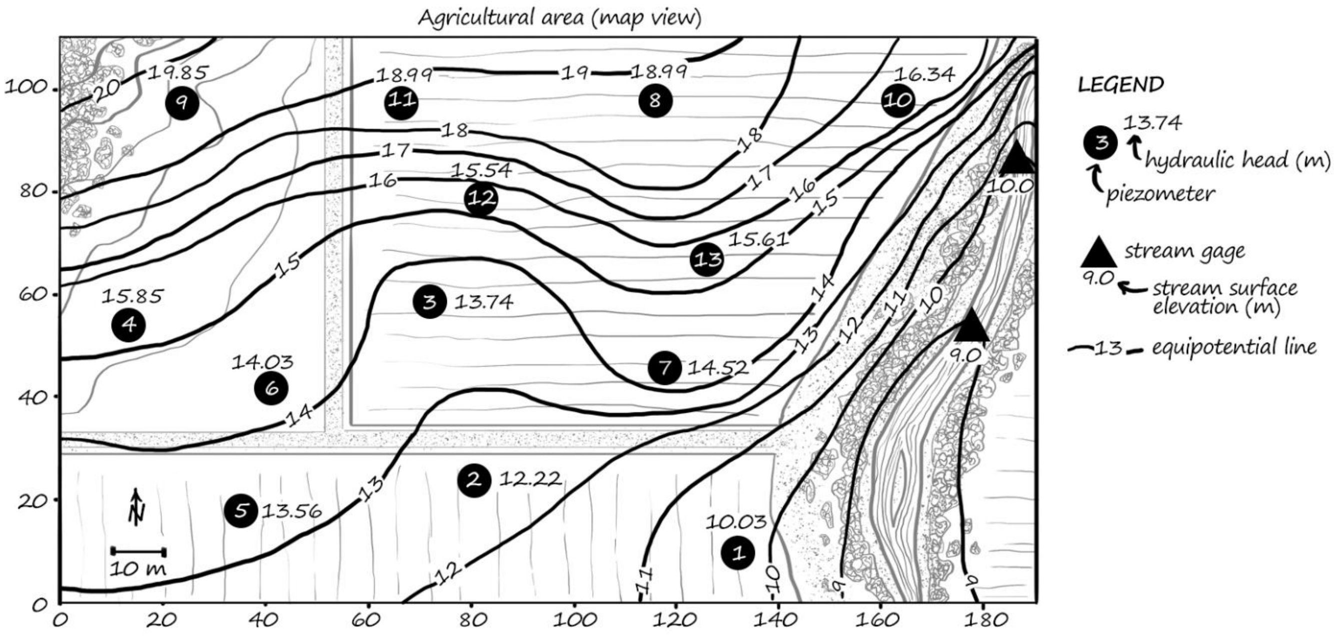

Figure 60. Agricultural field with equipotential lines (solid lines) at 1 m intervals.

To predict the exact path of groundwater molecules we must use another tool: *flow lines*. Flow lines help us visualize the direction of maximum gradient. For the topographic map example above (Figure 58), we drew flow lines (dashed lines) to represent the direction of runoff on the surface; runoff moved from higher elevations to lower elevations along the steepest path. We can do something similar with groundwater, but we must remember that groundwater moves from high hydraulic head to low hydraulic head. The flow lines (dashed lines) showing runoff in the topographic map (Figure 58) were perpendicular to the contour lines (solid lines). This was not a coincidence. For relatively simple cases (and most cases in practice), flow lines are assumed to be perpendicular (at a 90° angle) to equipotential lines. For equipotential lines, the energy potential (e.g., elevation or hydraulic head) is constant along the line. If the energy potential is constant, there is no energy gradient along the equipotential line, and there is no flow along the equipotential line. Therefore, the flow must be perpendicular to the equipotential line. If you draw a flow line starting at piezometer 11, you will find that the water generally flows just to the west of piezometer 12 (in the direction of the maximum gradient).

Figure 61 shows more flow lines for the agricultural field. Compare the flow path in Figure 61 to our initial estimate with four piezometers (Figure 54). Can you see the value of having more information? Equipotential lines help us more accurately visualize the hydraulic head distribution between piezometers and describe the flow paths in greater detail.

Next, let's visualize the hydraulic head distribution along a cross section (much like we did for the topographic example in Figure 56). Figure 62b shows the cross section along line A-B (line A-B is near the top of Figure 62a). Note: the orientation of our cross section is west to east in this example, but we could have taken a slice in any direction. On the A-B line, there are four piezometers (piezometers 9, 11, 8, and 10). The height of the water column above the datum elevation in each piezometer defines the hydraulic head at each location. Between piezometers, we must infer the hydraulic head. The dashed lines between piezometers (in the cross section, Figure 62b) represents the inferred hydraulic head values between the piezometers.

The hydraulic head *gradient* varies along the cross-section (Figure 62b). If the hydraulic head gradient were constant, the water table level in Figure 62b would be linear. However, the slope of the water table is steeper on the right side than it is on the left side, indicating that groundwater is moving more

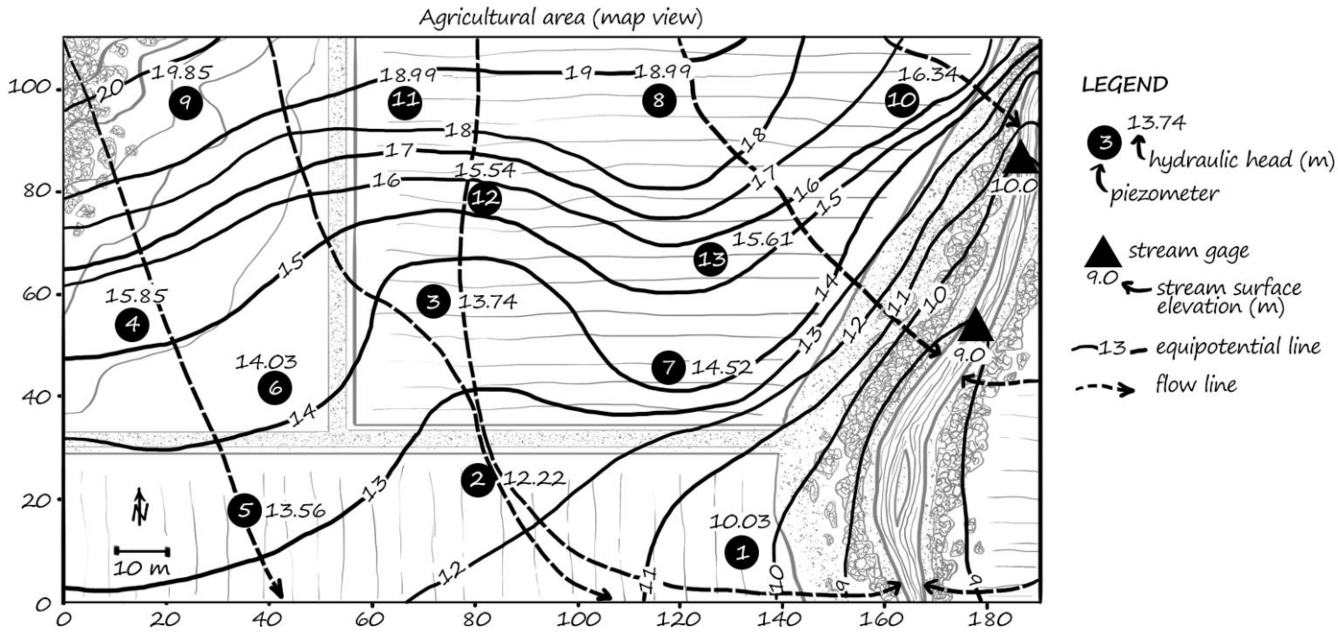

Figure 61. Agricultural field with equipotential lines (solid) and flow lines (dashed). Notice how the flow lines always intersect equipotential lines perpendicularly and never cross each other.

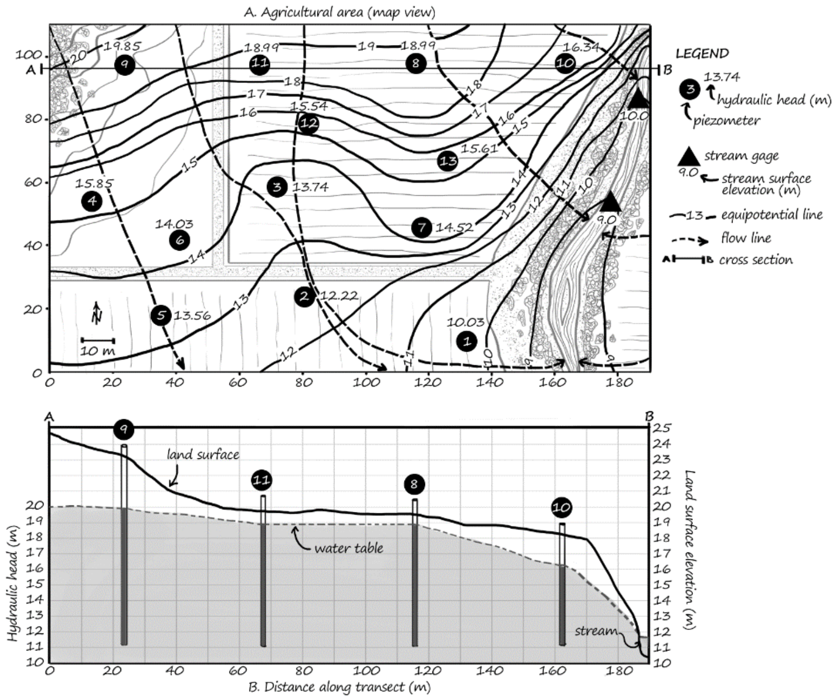

Figure 62. a) An aerial map with equipotential lines (solid) and flow lines (dashed) and b) a cross section along line A-B showing the equipotential surface (dashed) and the land surface (solid). In the cross section, there is a vertical exaggeration to highlight the changes in the land surface and hydraulic head.

quickly on the right side of the cross section than on the left (remember we have assumed that the sediments are homogeneous). Also, notice the land surface topography in Figure 62b (the solid line near the top of the piezometers). On large scales, the water table can sometimes (generally) mirror the surface topography. However, due to variations in pressure head in the subsurface, you must never assume that this is the case.

Compare the direction of flow in the cross section (Figure 62b) to the flow lines in Figure 62a. Do they match? Well, no. The flow lines are drawn in the direction of the maximum gradient, which is (generally) to the southeast. However, the flow direction in the cross section (Figure 62b) is limited by the orientation of the A-B line (west to east). This highlights a key point. If you take a cross section along two or more piezometers, you can only estimate the flow along the cross section. You can't tell the direction or magnitude of groundwater flow in the overall system. In other words, flow is a 3D phenomenon.

Many of the statements above regarding the hydraulic head and flux across the agricultural field are *qualitative* observations. However, we can also complete a *quantitative* analysis for the field. Begin by calculating the hydraulic head gradient (ΔH/L) between piezometers in the cross section (Figure 62b) using the change in hydraulic head between the piezometers (H_2-H_1) and the distance between the piezometers (L).

Calculate the gradient (ΔH/L) from piezometer 9 to piezometer 11:

- Hydraulic head at piezometer 9 (H_1): 19.85 m
- Hydraulic head at piezometer 11 (H_2): 18.99 m
- Location of piezometer 9 along the transect: 22 m
- Location of piezometer 11 along the transect: 68 m
- Distance between piezometers (L) = 68 m - 22 m = 46 m

$$\frac{\Delta H}{L} = \frac{H_2 - H_1}{L} = \frac{(18.99\, m - 19.85\, m)}{46\, m} = -\frac{0.86\, m}{46\, m} = -0.019$$

Notice how the hydraulic head gradient is negative (-0.019). Remember the head change is always the destination (piezometer 11) minus the starting point (piezometer 9), but the flow goes from high head (piezometer 9) to low head (piezometer 11), hence the negative sign in the Darcy equation (Equation 6). For practice, calculate the hydraulic head gradient from piezometer 11 to piezometer 9. For flow in this direction, the hydraulic head gradient is positive (0.019).

On your own, calculate the hydraulic head gradient between the other piezometers (11 to 8 and 8 to 10).

Answers:

- Hydraulic head gradient from 11 to 8: 0
- Hydraulic head gradient from 8 to 10: -0.060

Next, calculate the flux (q) between piezometers (8, 9, 10, and 11) if the hydraulic conductivity of the field is 20 cm/day.

Flux from 9 to 11:

$$q = \frac{Q}{A} = -K\,\frac{\Delta H}{L}$$
$$q = -\left(20\,\frac{cm}{day}\right)(-0.019) = 0.38\,\frac{cm}{day}$$

Flux from 11 to 9:

$$q = \frac{Q}{A} = -K\,\frac{\Delta H}{L}$$
$$q = -\left(20\,\frac{cm}{day}\right)(0.019) = -0.38\,\frac{cm}{day}$$

Flux from 11 to 8:

$$q = \frac{Q}{A} = -K\,\frac{\Delta H}{L}$$
$$q = -\left(20\,\frac{cm}{day}\right)(0) = 0\,\frac{cm}{day}$$

Flux from 8 to 10:

$$q = \frac{Q}{A} = -K\,\frac{\Delta H}{L}$$
$$q = -\left(20\,\frac{cm}{day}\right)(-0.060) = 1.2\,\frac{cm}{day}$$

Take a moment and reflect on the magnitude of the flux values between the piezometers in the cross section (Figure 62b) – they are relatively small! In fact, the largest flux along the cross-section line is around 1 cm/day (flux between piezometers 8 and 10). This demonstrates how it takes time for water to move around the sediment and through the pores!

In the aerial map (Figure 62a) above, many of the flow lines are pointing toward the river (e.g., flow lines in the northeast and southeast corners). Although you can't see what is happening at the boundaries of the river, you know that the banks of the river are *permeable*. Imagine that you are watching a water molecule that begins (in the subsurface) in the northwest quadrant of the map. The particle moves downgradient along a flow line, perpendicular to the equipotential lines, eventually reaching the bank of the river. Once the molecule reaches the river, imagine that it reaches near the right side of the A-B line – now use the cross section (Figure 62b) to visualize the subsurface.

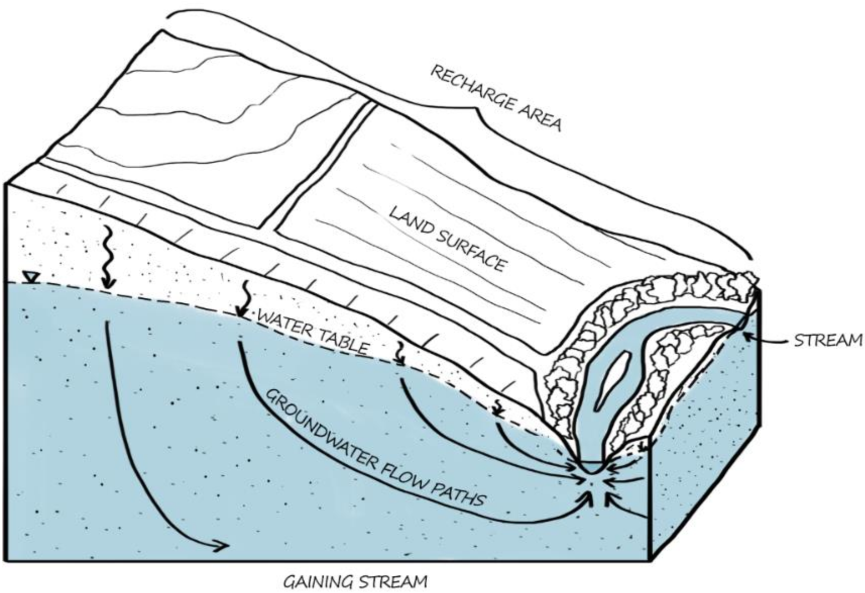

Figure 63. The stream is an extension of the potential surface. In this case, the flow lines are all pointing toward the stream and the stream is gaining water from the groundwater.

On the right side of the cross section (Figure 62b), the dashed water table line intersects the river surface elevation. This is because the surface of the river is a line of constant potential that connects to the energy potential in the subsurface. The water surface elevation in the river is an equipotential line. In the case of Figure 62b, a water molecule at piezometer 10, would flow in the subsurface (from high head to low head) and eventually reach the river. This is a typical process, when water is moving from the ground into the stream and the stream is gaining water from the bank. We call streams that are gaining water from the ground *gaining streams* (Figure 63). However, not all streams are gaining. Water can also move from the stream into the ground in a losing stream (if the hydraulic head in the stream is higher than the head in the banks). We will talk more about losing streams when we talk about groundwater pumping.

Flow Nets

Equipotential lines and flow lines can help you visualize groundwater flow for complex systems; Figure 62a is an example of an equipotential map with flow lines. But we do not have to stop there. We can also *calculate* groundwater flow using a more advanced visual tool called a *flow net*. A flow net quantifies 2D flow through a steady state system. While an equipotential map with flow lines is only a visualization tool, a flow net is a mathematical tool.

Simple Flow Net

Let's create a simple flow net to better understand this mathematical tool. Figure 64 shows a simple system: a vertical soil column that is 5 cm in height and 4 cm in width with steady state flow. The thickness of the vertical soil column (i.e., the distance going into the page) is 0.5 cm. The soil is homogeneous with a hydraulic head of 5 cm at the top and 0 cm at the bottom.

Before diving in, take a moment and reflect on the hydraulic head gradient in Figure 64 and imagine how water would flow through the column. What direction is the water flowing? The water is flowing from the top to the bottom, from high hydraulic head (5 cm) to low (0 cm). Hydraulic head gradients are starting to become more intuitive! Additionally, even if you didn't consciously put it into words, when you imagined water flowing it was moving *directly* downward through the column. This is because you imagined that the water could not flow through the sides of the tube; it flowed parallel to the sides of the tube. This is one rule of flow nets: flow lines always parallel impermeable boundaries (i.e., no-flow boundaries).

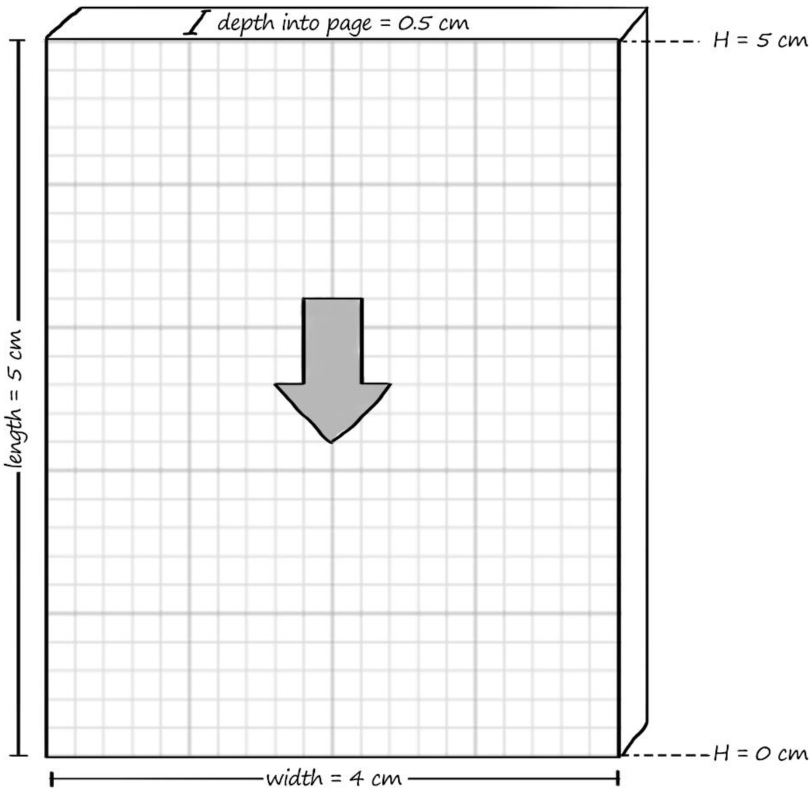

Figure 64. A simple, homogeneous soil column with downward flow. The hydraulic head at the top is 5 cm and at the bottom is 0 cm. The depth into the page is 0.5 cm.

Next, let's draw equipotential lines. Because the soil is homogeneous, equipotential lines will be equally spaced throughout the column. Think about that for a moment. If the system is steady state and the soil is homogeneous, then the equipotential lines (which represent hydraulic head) *must be* evenly spaced. Why is that? Refer to Darcy's law (Equation 6). If all other variables are constant (Q, K, and A), the hydraulic head gradient must also be constant. For this example, we will choose an equipotential interval of 1 cm. Therefore, there are five equally spaced equipotential lines throughout the column (Figure 65). Notice how the equipotential lines in Figure 65 are perpendicular to the sides of the soil column. This is another rule of flow nets! Equipotential lines are perpendicular to no-flow boundaries (e.g., the sides of the tube).

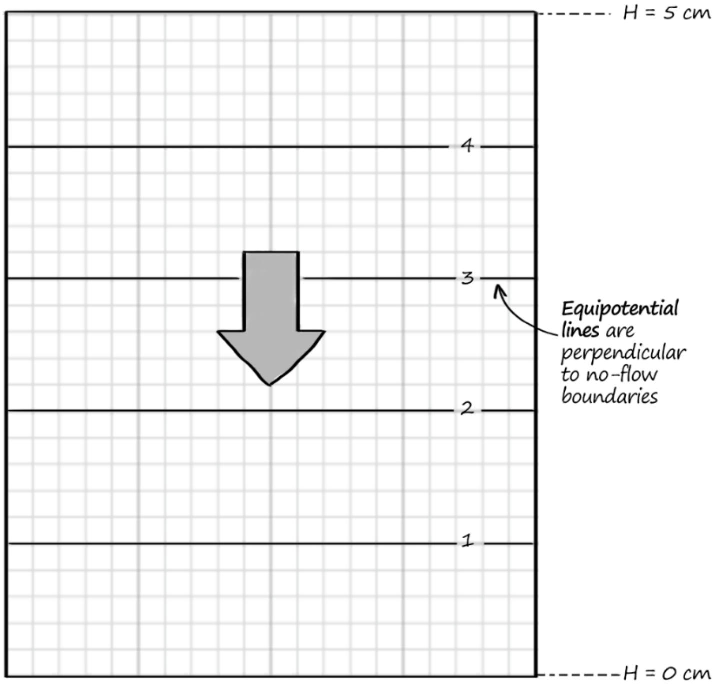

Figure 65. A simple homogeneous system with steady state flow. Equipotential lines are evenly distributed and are perpendicular to no-flow boundaries.

Lastly, we will draw flow lines. Remember, flow lines are perpendicular to equipotential lines. The area between adjacent flow lines is called a *flow tube*. The creator of a flow net gets to decide the number of flow tubes (just as they get to decide the interval for equipotential lines). In this case, we will choose four flow tubes. One key component of a flow net is that each flow tube must have the same amount of flow going through it. Therefore, for our example, we will draw three equally spaced flow lines so that the flow through each tube is equal (dashed lines, Figure 66).

We now have a flow net (Figure 66)! It may not look like much (yet), but this is a valuable tool. To an untrained eye, Figure 66 might just look like a grid. But this grid is more than *just* perpendicular lines – it is telling us something. The goal when drawing flow nets is to form a mesh with blocks as close to square as possible. One way to test this is to draw an inscribed circle in the middle of each square. An *inscribed circle* is the largest circle you can draw that touches all four sides of the square. If you must stretch the circle into an oval (in order to touch all four sides), you should adjust the number of equipotential lines and flow lines until you can form blocks that are as close to squares as possible. For this example, we chose the correct number of equipotential lines and flow lines to form perfect squares with inscribed circles.

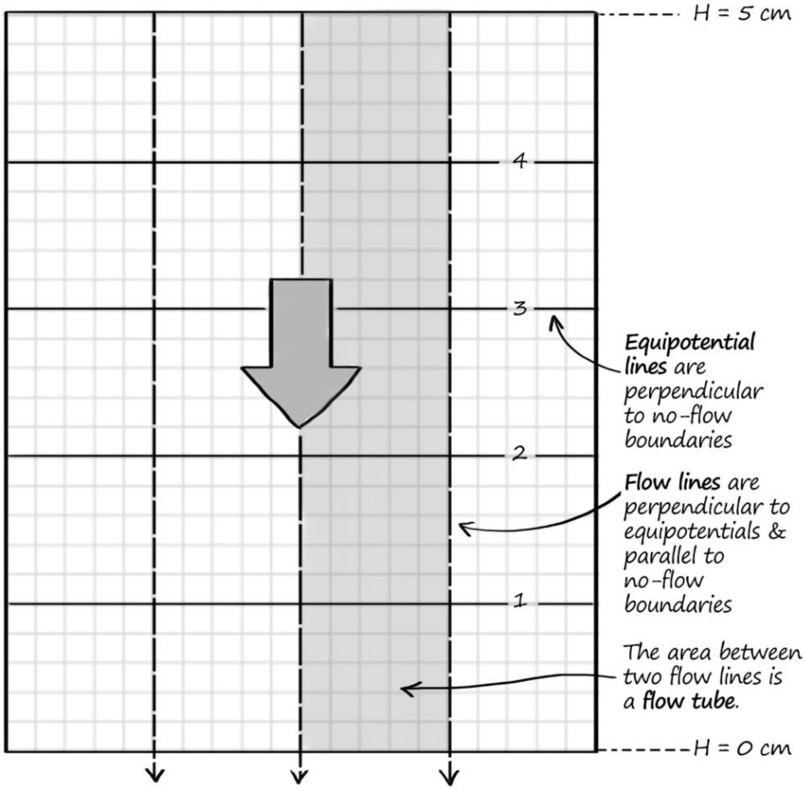

Figure 66. Flow net for a simple homogeneous system. Dashed lines are flow lines. There are four flow tubes – one is highlighted as an example.

Each square in the flow net (the space between the intersections of the equipotential lines and the flow lines) has the same amount of water flowing through it. Therefore, if we know the hydraulic conductivity (K) for the material (and the material is constant throughout), then we can calculate the total flow through the soil column. There are a couple of ways to do this. Most simply we can calculate the flow (Equation 6) going through one flow tube and multiply it by the number of flow tubes. To do this, first, calculate the flow going through one flow tube (Q_{tube}) using Darcy's law (Equation 6) (Figure 67). The hydraulic head at the top of one of the flow tubes is 5 cm, and at the bottom of the tube it is 0 cm. The length of the flow tube is 5 cm. The hydraulic conductivity of the system is 0.5 cm/hr. For the cross-sectional area (A), remember that the A in the Darcy equation must be *perpendicular to flow*. Therefore, A is not the area you see from the side in Figure 67. The flow is downward, so the cross-sectional area for the tube is the width of the tube (1 cm) multiplied by the depth going into the page (0.5 cm).

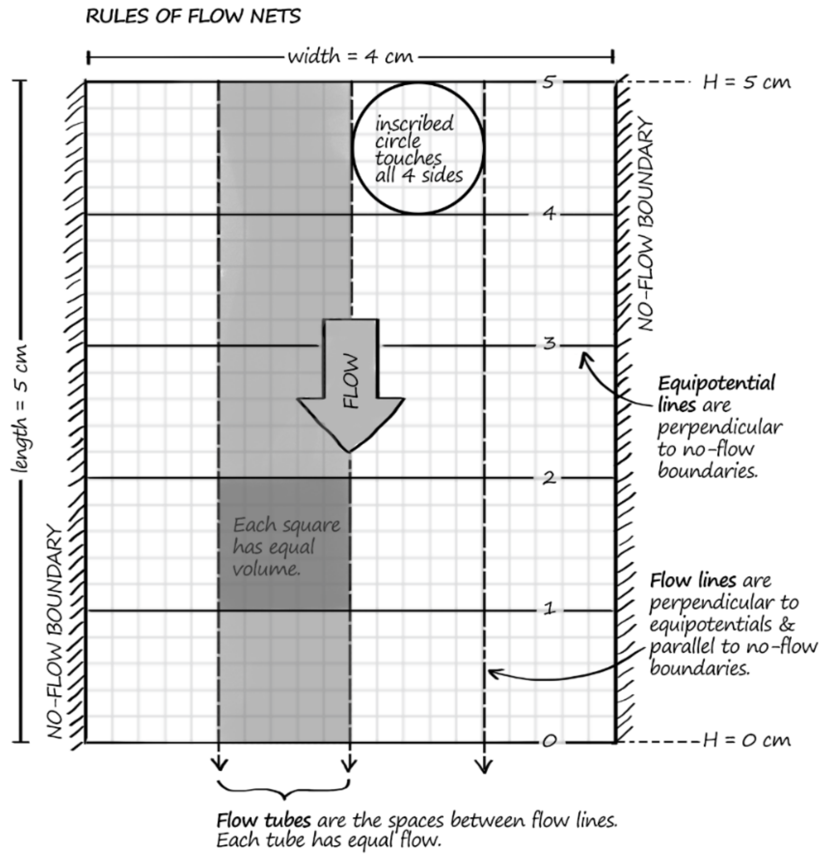

Figure 67. Flow net for a simple system. Dashed lines are flow lines. Solid lines are equipotential lines. Hatched lines are impermeable (no-flow) boundaries. Each square has the same amount of water and flux is constant throughout. The depth of the column into the page is 0.5 cm.

Known variables:

- $\nabla H = \frac{\Delta H}{L} = \frac{H_{bottom} - H_{top}}{L} = \frac{(0\ cm - 5\ cm)}{5\ cm} = -1$
- $K = 0.5\ \frac{cm}{hr}$
- $A = width\ of\ the\ tube * depth\ of\ the\ tube = 1\ cm * 0.5\ cm = 0.5\ cm^2$

Flow going through one flow tube (Q_{tube}):

$$Q = -K(A)(\nabla H)$$

$$Q_{tube} = -\left(0.5\frac{cm}{hr}\right)(0.5\ cm^2)(-1) = 0.25\ \frac{cm^3}{hr}\ downwards$$

The flow through one tube (Q_{tube}) is equal to 0.25 cm^3/hr downwards. Next, calculate the total flow going through the column by multiplying the flow through one tube (Q_{tube}) by the total number of flow tubes. In this case, the total number of flow tubes is 4. Therefore, the flow through the entire soil column (Q_{column}) is 1 cm^3/hr downwards.

If you are curious about the accuracy of this calculation, double check your value using Darcy's law for the entire column. For the entire soil column, the dimensions perpendicular to flow are 4 cm (width) by 0.5 cm (depth). Therefore, the cross-sectional area of the entire column is 2 cm^2. The hydraulic head at the top of the column is 5 cm and the hydraulic head at the bottom of the column is 0 cm. The length of the column is 5 cm. The hydraulic conductivity is 0.5 cm/hr.

Known variables:

- $A = width\ of\ column * depth\ of\ the\ column = 4 * 0.5 = 2\ cm^2$
- $\nabla H = \frac{\Delta H}{L} = \frac{H_{bottom} - H_{top}}{L} = \frac{(0\ cm - 5\ cm)}{5\ cm} = -1$
- $K = 0.5\ \frac{cm}{hr}$

Flow going through the entire soil column (Q_{column}):

$$Q_{column} = -KA\nabla H = -\left(0.5\frac{cm}{hr}\right)(2\ cm^2)(-1) = 1\ \frac{cm^3}{hr}\ downwards$$

Before we move forward, let's pretend that we chose a different number of equipotential and flow lines for the example above (Figure 67). We wish to demonstrate that it doesn't matter how many equipotential lines and flow lines you have (or what interval you choose) the flow net is still accurate. For Figure 68, we still aim to have inscribed circles, but this time we have equipotential lines every 0.5 cm and 8 flow tubes (rather than 1 cm equipotential lines and 4 flow tubes). We still have inscribed circles, but each circle is smaller. The depth into the page is still 0.5 cm. Take a moment to practice calculating the flow going through one flow tube as we did above (Answer: 0.125 cm^3/hr downwards). Next, multiply the flow going through one flow tube by the number of flow tubes (8 flow tubes). The flow going through the entire soil column is 1 cm^3/hr downwards just as we calculated before!

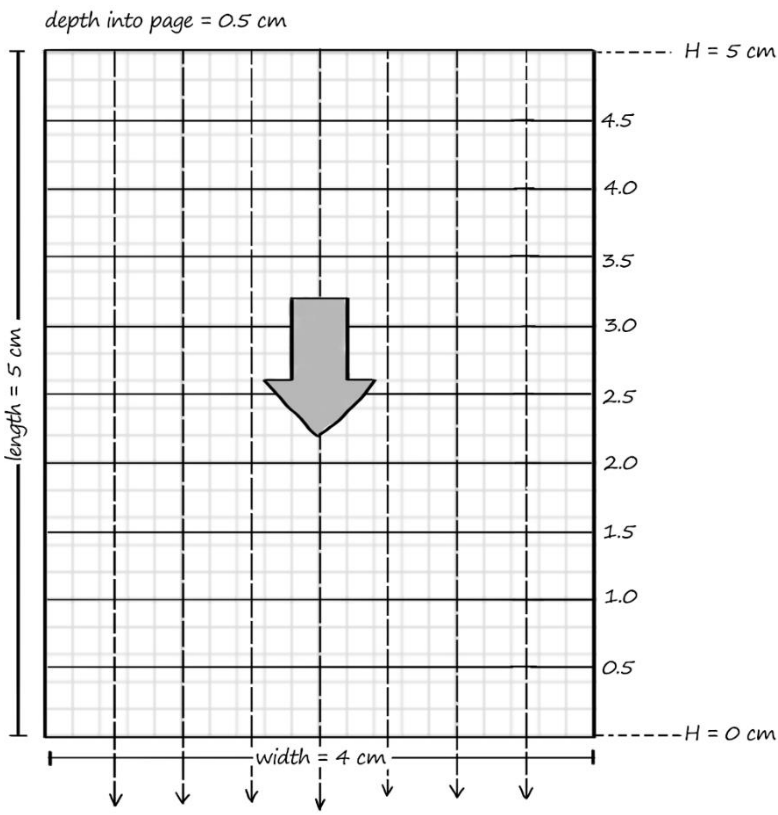

Figure 68. Same column as above (Figure 67) but with a different number of flow lines and equipotential lines.

Advanced Flow Nets

You might ask, why would we build a flow net when we could just use Darcy's law to calculate the flow through the soil column? And you wouldn't be wrong in asking this. But real-world problems aren't as simple as a straight-sided soil column (e.g., Figure 68). It is not always obvious where the water will flow, how quickly it will flow, or what the hydraulic head gradient is. To build on this knowledge, let's analyze a couple of more complex flow nets. In our examples above (Figure 67, Figure 68), each flow tube was the same size. However, in more complex systems the size of the squares in the flow net often varies.

Figure 69 shows groundwater flowing into and out of a lake system. The solid lines are equipotential lines, and the dashed lines are flow lines. The equipotential lines have the highest head on the west side and the lowest head on the east side. The hydraulic head in the lake is 75 m and is defined by the water level of the lake. This brings us back to a key point – the exchange of water between the surface and the subsurface. Remember our agricultural field example, where the stream level was an extension of the energy potential in the surrounding landscape (Figure 62b)? This system is similar. The lake level is an extension of the energy potential in the surrounding landscape, so the shoreline of the lake is in fact an equipotential line. The lake receives inflow from groundwater through its western shore and loses water to groundwater through its eastern shore. Notice that even when the equipotential lines are curved, the flow lines are always perpendicular to the equipotential lines. Also, the number of flow lines and equipotential lines are carefully chosen to form approximately square blocks such that you could draw inscribed circles in each square. This is a flow net!

Let's talk about the flow tubes in Figure 69. On the western shore of the lake, the flow lines near the center are closer together. In other locations, such as the southwestern quadrant of the map and the eastern shore of the lake, the flow lines are further apart. The spacing of the flow lines impacts the width of the flow tubes. What does this mean? When flow tubes are narrower, there is a smaller area available to convey the same amount of water. Remember, each square in the flow net has the same amount of water flowing through it. Therefore, narrower flow tubes have an increased *flux* because there is the same flow through a smaller area. In locations with wider tubes, the flux is lower, because a larger area is available for water to pass through.

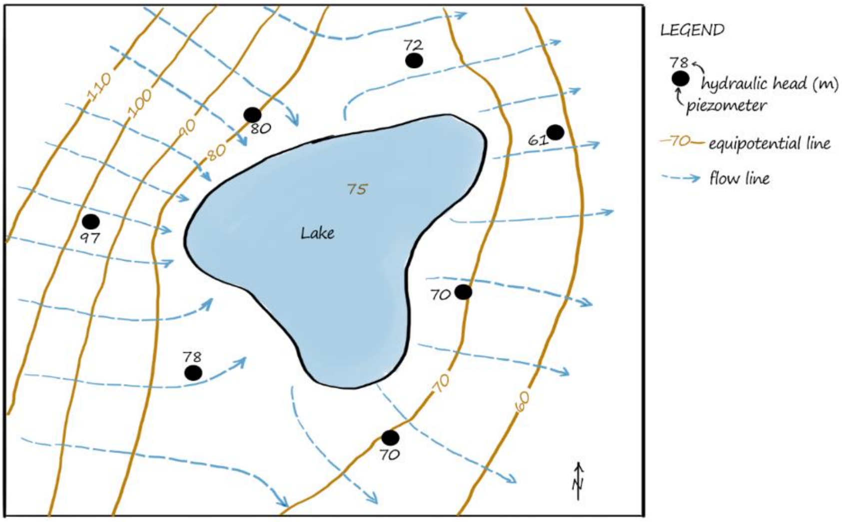

Figure 69. A flow net illustrating the flow of water into and out from a lake. Groundwater flow direction is shown using flow lines (dashed). Equipotential lines of equal hydraulic head are shown as solid lines. Adapted from Rosenberry et al., 2008.

Flow nets are especially helpful for understanding complex systems where the geometry of the Earth makes calculations impractical (i.e., the Earth's surface is hardly as simple as a soil tube). Along the earth's surface there are flow divides between watersheds. For example, the Great Continental Divide in North America is a surface hydrologic divide that separates the watersheds that drain to the Atlantic Ocean from those that drain to the Pacific Ocean. Just as there are flow divides on the surface of the earth, there are also subsurface divides, which separate certain parts of the subsurface from each other (Figure 70). Subsurface divides are sometimes controlled by geology but can also be controlled by topography. Figure 70 shows an example of subsurface flow boundaries that are isolating certain sections of a basin from each other. The flow system divides are shown as thinner dashed lines. The location where all three divide lines meet is called the *stagnation point* (represented by a dot in the figure); the local velocity of water is zero at this location. Therefore, the flow net in Figure 70 could help us calculate the flow in each of the three isolated parts of the subsurface. Locations where the flow tubes are narrow have faster flux and those deeper in the subsurface (that are wider) have slower flux.

Flow nets are a graphical solution to Darcy's law and are a useful quantitative tool! Drawing flow nets by hand can help you gain an intuitive understanding of any flow system. But it is also becoming increasingly common for hydrogeologists to construct complicated flow nets using computer software. Whether it is created by hand or by computer software, it is important for all hydrogeologists to have the necessary skills that are needed to interpret flow nets.

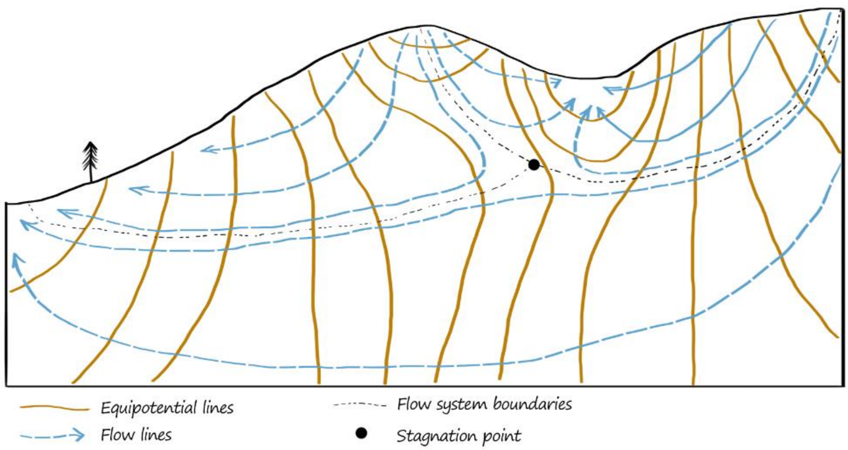

Figure 70. Advanced flow net for a system with several flow divides. Dotted lines show the dividing lines (after Tóth, 1963).

Conclusion

There is an art to the practice of hydrogeology. The discipline requires you to look at a landscape – its hills, soils, plants, lakes, and rivers – and to imagine what is beneath the surface. Piezometers are one tool that hydrogeologists use to understand groundwater flow, but they only disclose information about the hydraulic heads at *specific locations*. In locations where piezometer data is missing, the act of visualizing groundwater flow requires imagination that is bounded by solid scientific principles. Hydrogeologists use equipotential lines, flow lines, and flow nets to visualize groundwater movement. This chapter introduced you to each of these concepts and was simply the beginning of your journey to "seeing" water in the subsurface!

What to Remember

Important Terms	
contour interval	gaining stream
equipotential lines	stage
flow lines	stagnation point
flow net	stream gage
flow tube	

Chapter 5

Basics of Solute Transport

Introduction

In the previous chapter we discussed the idea of tracking a water molecule. We visualized the path a molecule would take using equipotential lines and flow lines. This is useful information! Now imagine that the molecule you are tracking is a harmful chemical. As water moves through the soil, it transports any substance that is dissolved in it (also known as a *solute*). This doesn't necessarily have to be a harmful chemical, but those are often the solutes we wish to track. Common solutes that we pay attention to are fertilizers, pesticides, salt, metals, and waste leached from septic systems, all of which can be harmful to humans and ecological communities. Additionally, solutes in groundwater can react chemically with each other and/or with the surrounding rock, altering the strength of underground geologic materials and potentially causing the failure of mining excavations, dams, and artificial slopes. It is also becoming increasingly common for waste (e.g., industrial, agricultural, domestic) to be stored or disposed of underground. Once underground, waste products can spread from their original source into lakes, streams, and oceans (remember, groundwater and surface water are connected). In this chapter we will walk you through the basics of solute transport and discuss the importance of understanding such processes.

Advection

You can think of solutes that are dissolved in water as being mixed, at the molecular level, with water. If we could instantly replace all the water in an aquifer with water that contains 0.001 mg/l of NaCl (salt), then little would change from our previous chapters. The water density and viscosity would change slightly, but all the equations that we've used previously to describe groundwater flow would work. This is because the solute molecules move along with the water.

Advection is the transport of a solute by the water that it is dissolved in. The term advection comes from the Latin root "advehere" which means "to carry". Imagine that the water molecules are carrying the solute; wherever the water moves, the solute moves with it. More specifically, suppose you have 100 solute molecules released in a blob. Each solute molecule is surrounded by (really, dissolved in) water molecules, almost like passengers in a car. The water and solutes (car and passengers) are moving at the same rate and along the same path. People within one car wouldn't spread out – similarly, a blob of solute molecules stays as a blob as it moves by advection. We refer to this type of transport as *plug flow*. Essentially, during plug flow, every water molecule is moving at the same velocity. Therefore, groundwater velocity alone can give you an idea of the rate and direction of solute movement through the subsurface.

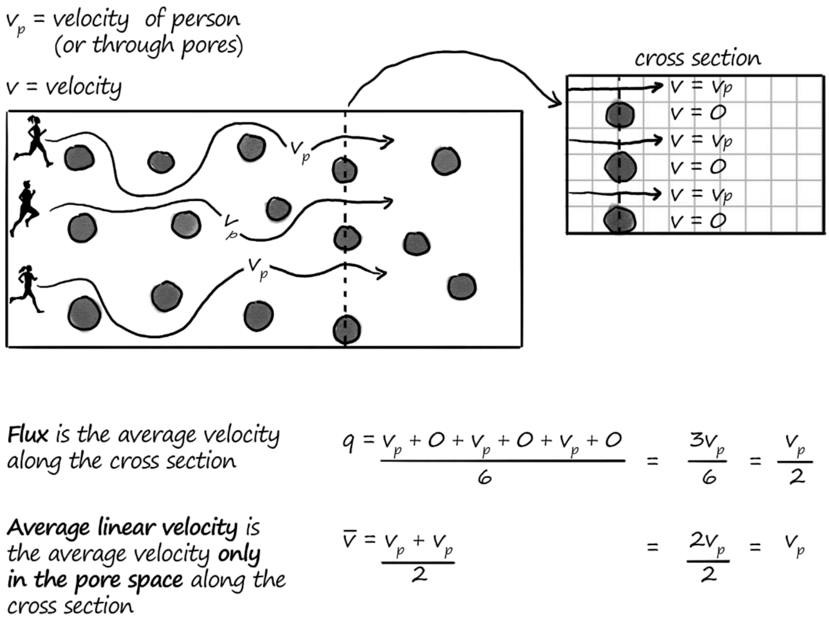

Figure 71. A group of people running across a room at $v = v_p$. The dashed line is a cross-section line perpendicular to the flow of people. The flux (q) is the rate that people move through the cross section and includes the velocity of the runners through the pore space (v_p) and through the obstacles ($v = 0$). The average linear velocity ($\bar{v}$) only accounts for the velocity through the pore space (v_p). If the obstacles account for half of the cross section, the flux is half the average linear velocity.

The driving force of solute transport via advection is water transport. However, you cannot simply use Darcy's flux to describe advection. There is a special term that describes the displacement of solute by advection — *the average linear groundwater velocity* ($\bar{v}$). But why the distinction between q and $\bar{v}$? The Darcy flux represents the rate at which water passes through a cross-sectional area. This cross-sectional area includes both pore space and solid particles, but water only flows through the pore area, not the solid area. Therefore, if you want to know the rate of water through the pore space alone, you need another term: the average linear groundwater velocity.

To better understand the difference between q and $\bar{v}$, imagine a room full of chairs (Figure 71). A group of people are asked to run from one side of the room to the other side, avoiding obstacles (the chairs). Look at the cross-section line in Figure 71. Can you assign a "runner velocity" (v) to each location along the line? The runner velocity between the obstacles will be equal to the velocity of the runners moving through it; for this scenario, all the runners are moving at the same speed through the obstacles ($v = v_p$). At other locations where there are obstacles (chairs), the runner velocity is zero ($v = 0$). Now, calculate the flux for this scenario by averaging all the runner velocities (v) along the cross-section line (remember to include all locations where $v = 0$ in your average). Can you begin to see why the flux is not representative of the average velocity of the runners ($\bar{v}$)? Although all the runners are moving at the same speed (v_p) the flux is less than this speed because the chairs create "dead space" (where, $v = 0$), and this dead space is included in the flux calculation (Figure 71). If we wanted to measure the average speed of the runners ($\bar{v}$), we would need to count only movement in the void space between the chairs.

The average linear velocity ($\bar{v}$) is the rate that the runners move between the obstacles. The runners move through a network of interconnected empty spaces, like a maze, just as groundwater molecules move in a porous medium. A fluid can only flow between the soil particles, and the runners can only move between chairs. Look at a cross section of your room (dashed line, Figure 71) – notice how the space between the chairs is only a fraction of the room.

As the runners move through the room, the average linear velocity is dependent on the density and configuration of the obstacles. The emptier the room, the faster the runners can get across it on average. Now imagine that you can look down on a cross section of a soil (Figure 72) – the pore space in the soil is only a fraction of the total area. In saturated porous media, the fraction of fluid-filled regions between particles is equal to the porosity (n)

(Equation 1). Therefore, we can use porosity to help calculate the average linear groundwater velocity ($\bar{v}$) (Figure 72).

$$\bar{v} = \frac{q}{n}$$

Equation 10: The relationship between the Darcy flux q (L^3/T), the porosity n (-), and the average linear groundwater velocity $\bar{v}$ (L/T)

Note: a fluid can only flow through the fluid-filled regions between soil particles. In our discussion of groundwater flow, we primarily focus on saturated media (i.e., when all the pore space is filled with water). Therefore, we will not distinguish between completely filled and partially filled pores in this book. However, it is important to note that water moves differently in an unsaturated or partially saturated medium.

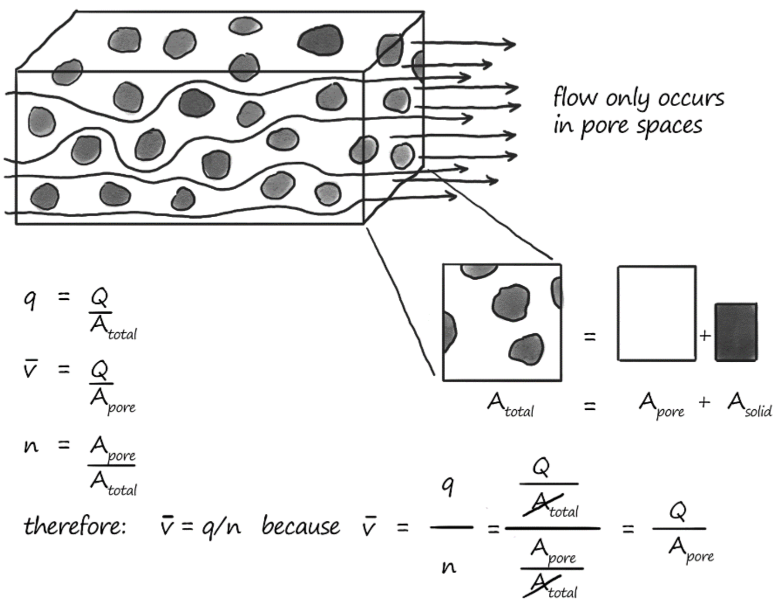

Figure 72. The cross-sectional area of a soil (A) is the summation of the area of the pore space (A_{pore}) and the area of the solid particles (A_{solid}). The flux is the flow (Q) divided by the total area (A) and the average linear groundwater velocity ($\bar{v}$) is the flow (Q) divided by the area of the pore space (A_{pore}). Using the definition of porosity, $\bar{v} = q/n$.

Average linear groundwater velocity is always greater than flux in a porous medium; water must flow faster in the pores to make up for the lack of flow where there are solids. To understand this, rather than imagining a porous medium, think of water flowing out of a straight section of hose. The flow rate out of the hose (Q) is found by measuring the time that it takes to fill a bucket of a known volume. The flux is the flow rate divided by the cross-sectional area of the hose. Now, imagine that you can watch 100 individual water molecules as they move through the hose. On average, all 100 molecules would be moving at the same rate (equal to the flux). This is because the porosity of the hose is 1 (the entire hose is void space). Now, imagine that you could block off half of the cross section of the hose by inserting a long metal rod into it. In other words, the porosity of your hose is now 0.5 (50% void space). To maintain the same flow rate as before, the water would have to move twice as fast around the rod! Or in other words, if you watched 100 water molecules as they move through the hose, they would need to move at a rate equal to twice the flux ($\bar{v} = 2q$). This is the same as $\bar{v} = q/0.5$ (or $\bar{v} = q/n$).

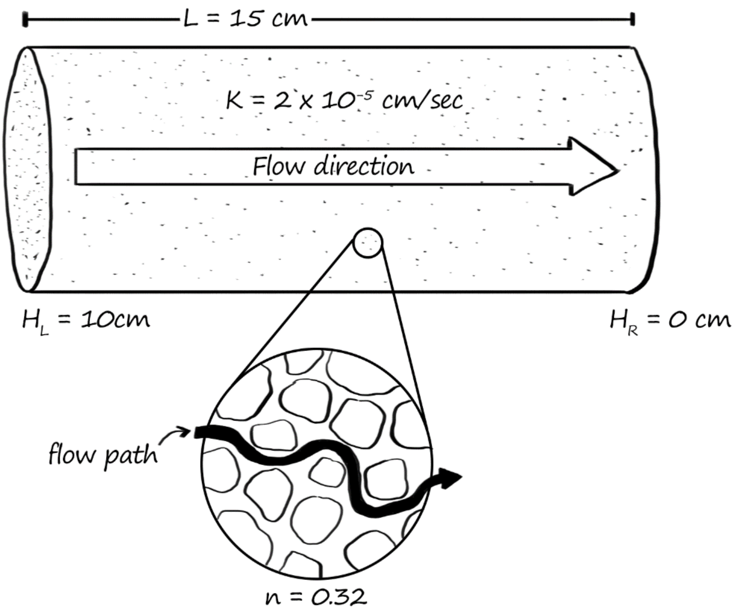

Figure 73. Horizontal tube with flow from left to right. The linear groundwater velocity is impacted by the flow path of the particles, which is dependent on the porosity of the medium.

To better understand the relationship between q and v̄, let's do some calculations. First, imagine that you have a 15 cm-long tube of saturated fine-grained sand (K = $2x10^{-5}$ cm/s) with water flowing through it (Figure 73). The water has a solute dissolved in it. The hydraulic head on the left side of the tube is 10 cm and the hydraulic head on the right is 0 cm. Thus, we know that water is flowing from left to right (high H to low H).

Calculate the water flux (q) through the tube (Equation 9):

$$q = -K\nabla H$$
$$q = -2x10^{-5}\frac{cm}{s}\left(\frac{(0\ cm - 10\ cm)}{15\ cm}\right) = 1.33x10^{-5}\frac{cm}{s}$$

Next, use the porosity of the soil (n = 0.32) to calculate the average linear groundwater velocity (v̄):

$$\bar{v} = \frac{q}{n}$$
$$\bar{v} = \frac{1.33x10^{-5}\frac{cm}{s}}{0.32} = 4.17x10^{-5}\frac{cm}{s}$$

As we expect, the average linear groundwater velocity (v̄) of the solute is much faster than the water flux (q). If water is only flowing through about a third of the total space, then the v̄ must be about three times the q (more specifically, q = 3.123v̄). The average linear groundwater velocity is in the overall direction of flow, and it does not consider flow tangential to this direction.

Displacement is the linear distance that something travels from one place to another and is equal to the velocity multiplied by the time spent traveling. The average linear groundwater velocity is the speed with which an average water molecule is moving in the direction of flow. Therefore, we can calculate the linear displacement of our solute in the direction of flow (via advection) by multiplying the average linear groundwater velocity by the travel time.

$$\Delta x = \bar{v}t$$

Equation 11: Displacement equation where Δx is the difference in length (L), v̄ is the average linear groundwater velocity (L/T), and t is the travel time (T).

Using Equation 11, how far can a solute molecule travel (by advection alone) through our 15 cm tube? Imagine that the solute molecule begins on the left side of the tube and travels at an average linear groundwater velocity of 4.17×10^{-5} cm/s. What is the linear distance traveled in the direction of flow after 50 hours (about 2 days)? Note: 50 hours = 1.8×10^5 seconds

$$\Delta x = \bar{v}t$$

$$\Delta x = \left(4.17\ x\ 10^{-5}\,\frac{cm}{s}\right)(1.8\ x\ 10^{5}\ s) = 7.5\ cm$$

The molecule travels halfway across the tube (7.5 cm) in 50 hours. Reminder, this calculation is only for one solute molecule. However, we can make the same calculation for a group of molecules. During advective transport (plug flow), all solute molecules travel at the exact same speed. Therefore, all the molecules in the tube travel the same distance after 50 hours (7.5 cm). That is, the molecules travel like a blob that doesn't spread out.

Next, let's visualize a pulse of solute in a 15 cm long (horizontal) column filled with sand. A pulse of solute will look like a burst of solute that occurs over a short period of time. The solute will be released at the left boundary of our horizontal column (over a short period of time). After the solute is released, water will flow through the left boundary for the remainder of the observation period. Figure 74a shows the horizontal column with flow in the x-direction. The plots show the location of the solute (as it transports via advection) at several different travel times. The x-axis is the horizontal distance along the column and the y-axis is the ratio of the concentration of the solute at that location (C) to its initial concentration at the left side of the column (C_0). The concentration (C) is a ratio of the mass of the solute to the volume of water it is dissolved in (Equation 12).

$$C = \frac{m}{V}$$

Equation 12: Equation for concentration, where m is the mass of the solute (M), V is the volume of the water (L^3), and C is the concentration (M/L^3).

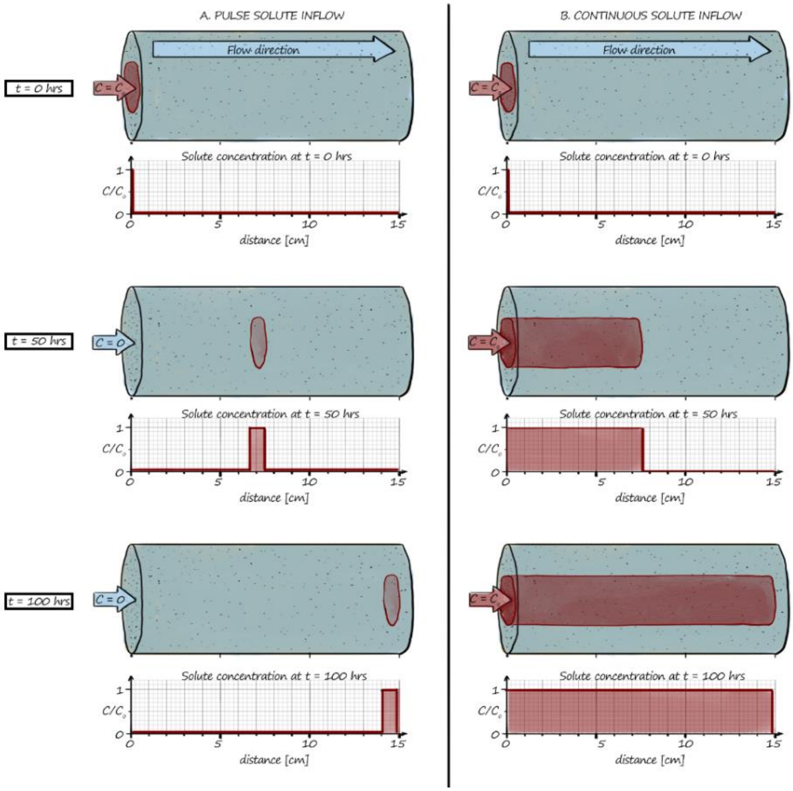

Figure 74. Comparison of a) pulse flow and b) continuous flow of solute in a 15 cm column. The graphs below show the concentration of the solute over time.

For Figure 74a, if the value on the y-axis is equal to 1 ($C/C_0 = 1$), the concentration of the solute is equal to the initial concentration (C_0) that was released at the left boundary. If the value on the y-axis is equal to 0, then there is no solute in that location. The top plot represents the solute at early time (t = 0); the solute is still located at the left side of the column. The middle plot is the solute at t = 50 hours, and the bottom plot is the solute at late time (t = ~100 hours). For all three travel times, the distance traveled by the solute is determined by the travel time and the average linear groundwater velocity (using the displacement equation, Equation 11). The concentrations of the solute at all travel distances are equal to the *initial concentration* ($C/C_0 = 1$) because all the solute molecules travel at exactly the same average linear groundwater velocity.

Thus far, we have discussed a *pulse flow* of solute; that is, a release of solute that happens over a brief period (Figure 74a). After a pulse release of solute, clean water enters the system and travels at the same speed as the average linear groundwater velocity. In contrast to this, a solute can also be released continuously (where the water flowing into the column always contains the solute). Let's use the 15 cm tube as an example. Compare Figure 74a and Figure 74b. For pulse flow, the concentration at the left boundary equals C_0 for a period of time, but then equals 0 (as clean water enters). For continuous flow, once the solute starts being added, the concentration at the left boundary stays at C_0. How does the solute concentration vary throughout the tube for each of these cases? We know that during pulse flow after 50 hours the solute moves 7.5 cm (see displacement calculation above). Therefore, the concentration is equal to C_0 at 7.5 cm. For continuous flow, the solute travels at the same linear groundwater velocity ($4.17x\text{-}^5$ cm/s) as it did for the pulse flow. Nothing has changed in the column to impact the velocity. However, because the solute is continuously released from the left boundary, the concentration is equal to C_0 for the entire distance from 0 cm to 7.5 cm. Figure 74 highlights these differences.

Advection and Mass Balance

To better understand what is going on during solute transport, let's revisit the concept of mass balance from earlier in the text. Mass balance equations are used to understand how a state variable is changing over time. For a lake system, we developed a mass balance using the volume of water in the lake as the state variable (i.e., how was the lake volume changing over time) (Figure 22). For solute transport, we will develop a mass balance equation using the mass of solute dissolved in the water – can you hypothesize why a mass balance equation for solute transport could be helpful?

Imagine a horizontal soil tube with water entering from the left side and leaving from the right side. To begin, imagine that your state variable is water volume. During the observation time, there is a volume of water that enters (and leaves) the system. The volume of water that is entering (or leaving) the column is dependent on how quickly water is flowing and over what time period. Therefore, the change in the volume of water within the system is equal to the flow rate (Q) multiplied by the observation time. For this soil tube, we can develop a mass balance equation using the volume of water that is entering (inputs) and leaving (outputs) the system over a specific time period.

Now, imagine that there is a pulse of solute that enters the soil tube (at a given flow rate). We are no longer interested in only tracking the volume of water but also tracking the mass of the solute. Once the solute enters the system, the pulse is transported via advection. While the pulse moves through the column, the total mass in the system is constant (i.e., conservation of mass). Eventually, the pulse of solute reaches the end of the column and exits. So, how can we develop a mass balance for this scenario? We need a new state variable that will tell us about the solute.

For solute mass balance equations, we often use a state variable called mass flux. The *mass flux* is the mass of the solute flowing through a region per unit time. The change in mass (due to advection) is dependent on the concentration of the solute and on how quickly water enters or leaves the column during the observation period. If you consider a continuous flow of water moving through an area with a solute dissolved in the fluid (at a certain concentration) – you can multiply the flux by the concentration to calculate the mass of the solute moving through the region per unit time (or the advective mass flux).

Therefore, the advective mass flux is:

$$J_a = \frac{QC}{A} = qC$$

Equation 13: Advective mass flux equation, where J_a is the advective mass flux (M/L^2T) (the subscript a refers to advection), q is the Darcy flux (L/T), and C is the solute concentration (M/L^3).

In a simple system, like a soil column with horizontal flow, we calculate the transport of solute in the same way that we calculate the flow of water. Going back to the car analogy, if one car passes a point every minute (the traffic flow), and each car has six people in it (the "concentration" of people in each car), then we could say that six people pass the location each minute. Similarly, if the rate of water flow is 0.1 cm^3/min and the solute concentration in the water is 1 g/cm^3, then the solute mass transport (mass flux) is 1 g/min. This is a practical explanation of why we define the mass flux as the water flux multiplied by the concentration.

Just as we did with the bear, Manhattan, and bathtub examples in the "System" chapter, let's run through different system states for the soil tube with advective transport. Imagine that you have a clean, steady state soil tube; clean water is flowing into the left side at the same rate that it is flowing out of the right side. Next, imagine that there is a continuous inflow of solute dissolved in water. After there is a continuous inflow of solute, (initially) the water coming into the tube has solute dissolved in it, but the water leaving the tube does not. The system is (still) in a steady state with respect to water flow (one state variable). However, with respect to solute mass (another potential state variable) it is gaining; the input of advective mass flux is greater than the output. Over time, the solute travels through the tube (Figure 74b) and eventually the solute will reach the other end. Once the water flowing into the tube is flowing at the same rate *and has the same concentration* as the water flowing out of the tube, then the system is at steady state with respect to both flow *and transport* (Figure 75). Not only is there no change in storage of water, but there is also no change in storage of solute; the water transports the solute, and the advective mass flux into the tube equals the advective mass flux out of the tube. Next, if you shut off the continuous inflow of solute (but not the continuous inflow of clean water), the advective mass flux at the input now equals zero. The system is at steady state with respect to flow but is now losing with respect to solute mass. It will take some time for all the solute to exit the tube, but because the output mass flux is greater than the input mass flux, eventually the tube will be clean again and will return to being in a steady state with respect to both water flow and solute mass.

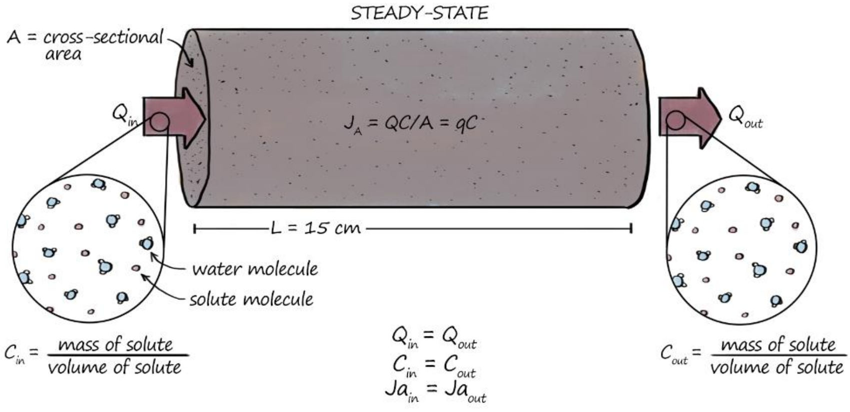

Figure 75. Mass balance of a 15-cm column with solute transport at steady state. the flow of water in equals the flow of water out, and the solute concentration in is equal to the solute concentration out.

Dispersion and Diffusion

Dispersion

There are more mechanisms that impact solute transport than advection alone. To explain, let's go back to our analogy of running through an obstacle course of chairs (Figure 71). If there is plug flow through the room (movement by advection alone), a group of runners travel as a blob, all moving at the same rate ($\bar{v}$). This is easy to understand and visualize. However, it is unlikely that all the runners *actually* move at the same rate. Think about the scenario more carefully. If you have more than one person running through an obstacle course, each person is likely to follow a slightly different path, experiencing different obstacles. So, it is unlikely that a blob of people will all stay together as they travel. Each person will have a slightly different velocity and path length. In some cases, runners will experience places where the chairs are closer together (resulting in slower movement) or perhaps some runners will experience more splitting or branching in their path. In other cases, some runners will physically be faster than others. This brings up a key point about $\bar{v}$ – it is the *average* velocity at which runners appear to be moving across the room. That is, if you take a cross section of the room (e.g., dashed line, Figure 71) and measure the rate at which each person is moving perpendicular to the cross section (only through the void space), then take the average of all those rates – you will find the average linear groundwater velocity. The average linear groundwater velocity does not represent flow on a microscopic scale, and it does not describe the variations in velocity of a molecule at different points along its path, or variations in velocity between different molecules (i.e., it assumes all molecules are flowing at the same rate).

For groundwater flow, hydrogeologists *conceptually* understand some of the processes that account for microscopic variations in groundwater velocity (e.g., pore shape, pore size, pore interconnectivity, and variability in those properties from pore to pore) — but it is difficult for hydrogeologists to create groundwater models to account for these small-scale variations. In fact, it is impossible to know all the small-scale variations even in the simplest of systems. Hydrogeologists use the term *dispersion* to compensate for all the variations in velocity that are impossible to account for individually. In the case of the runners through the obstacles of chairs, dispersion can explain why the blob of people do not stay as a blob, but instead spread out. The same is true for solutes. If we follow a 100-molecule blob through a soil tube, the initial concentration of a solute does not stay as a compact blob, but rather becomes dispersed or spread out, due to microscopic variations in flow velocity and flow paths (Figure 76).

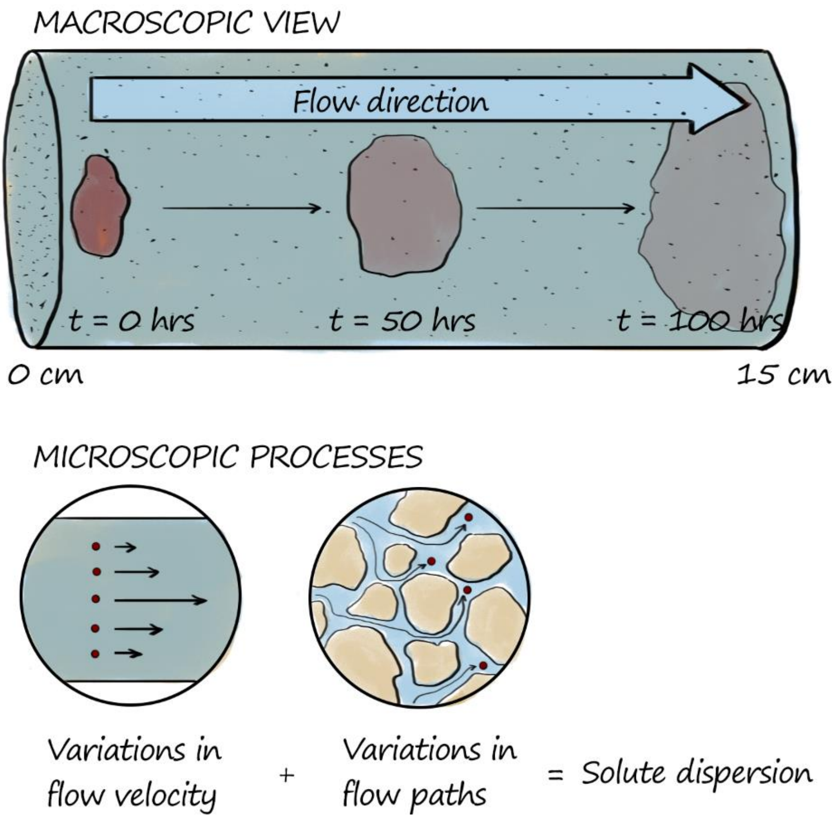

Figure 76. Dispersion happens during transport due to microscopic processes such as variations in flow velocity or flow paths.

What does dispersion look like in terms of concentration? Imagine a marching band in a parade; each row has 10 musicians. The band is quite good, so they march exactly in time. Therefore, as you stand and watch the band pass, there are always 10 people in a line in front of you (i.e., the concentration of musicians is 10 per street-width); each band member is moving at the same pace as the *average* band member, so the concentration of musicians is dictated by advection alone. However, what if the band is not so good? They start in a perfect formation of 20 rows with 10 musicians, but each musician marches at their own pace. At the very beginning of the parade, they look uniform. However, the longer the band marches, the more spread out they become. One piccolo player is 100 meters in front of the entire band, while a trombone player is 200 yards behind their nearest bandmate. In the middle of the band, there are still 10 musicians moving together. To draw an analogy with solute transport, we must recognize that the *average* band member is moving at a pace that is dictated by the water flow, but some band members (molecules) are moving faster, and some are moving more slowly.

Figure 77 shows our 15 cm column with both advective and dispersive transport. There are a few things to notice. Some of the molecules are moving more slowly than the average linear groundwater velocity and some are moving more quickly; these molecules are like the trombone and piccolo player. If we think of the different factors that influence the speed of each solute molecule, it is highly unlikely that all of them will cause it to move as fast as possible. It is more likely that the factors will cancel each other out – leading to molecules travelling close to the average velocity. Most of the molecules move at the average velocity, some move a bit faster (or slower), and only a few travel much faster (or slower). This produces a normal distribution (i.e., a bell curve) of velocities throughout the column (Figure 77). This means that lower concentrations of solute arrive at the right side of the column at earlier times (and later times), but most of the solute arrives around the same time. Additionally, the concentration of the solute at all locations (with dispersion) is lower than the initial concentration ($C/C_0 < 1$), because the concentration is spreading out. Just like the marching band! If a piccolo player has moved ahead of their line, then there are fewer marchers in that line!

Remember the concept of a mass balance? It works here, too. If the blob becomes more spread out, then the concentration must be reduced at each location to maintain mass balance. Keep in mind, the areas under both the advection-only line and the advection/dispersion line are the same (Figure 77), because in all scenarios there is a conservation of mass. Mass balance equations are effective for these problems because there is no solute lost during transport. Note: there are some cases where mass is not conserved during solute transport (e.g., nonconservative tracers), but in these examples, we will assume conservation of mass.

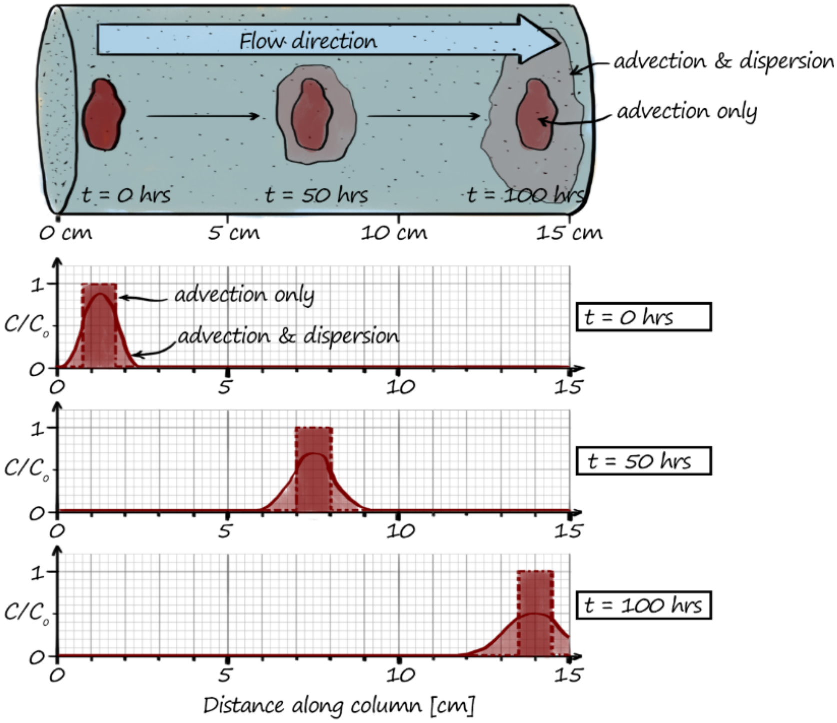

Figure 77. A 15 cm column with solute transport. Darker red represents advective transport, lighter red shows advective and dispersive transport.

Remember there are two ways we describe solute inflow: pulse flow and continuous flow (Figure 74). Consider the example of people running through a room. A pulse flow would be if 25 people all started running at the same time at the left boundary. In contrast, a continuous flow would be an endless stream of people arriving at the entrance and running through the room. Figure 77 represents a pulse flow of solute. Take a moment and think about how the system changes if the solute is applied continuously (Figure 78b). A continuous inflow is analogous to new rows of marching band members joining the parade each time the previous row leaves the starting line. When each new row is added, the band members look as they should, 10 of them in a perfect line across the street. We assume that all the marching band members are moving at the same pace (advection only) until they cross the boundary (dispersion only happens within the system boundaries). As the members increase in distance from the start, there is more spreading. The initial concentration of band members remains constant at the start of the parade (C

$= C_0$), but the concentration decreases as you move further from the start; only a few band members have a faster velocity than the average (Figure 78b).

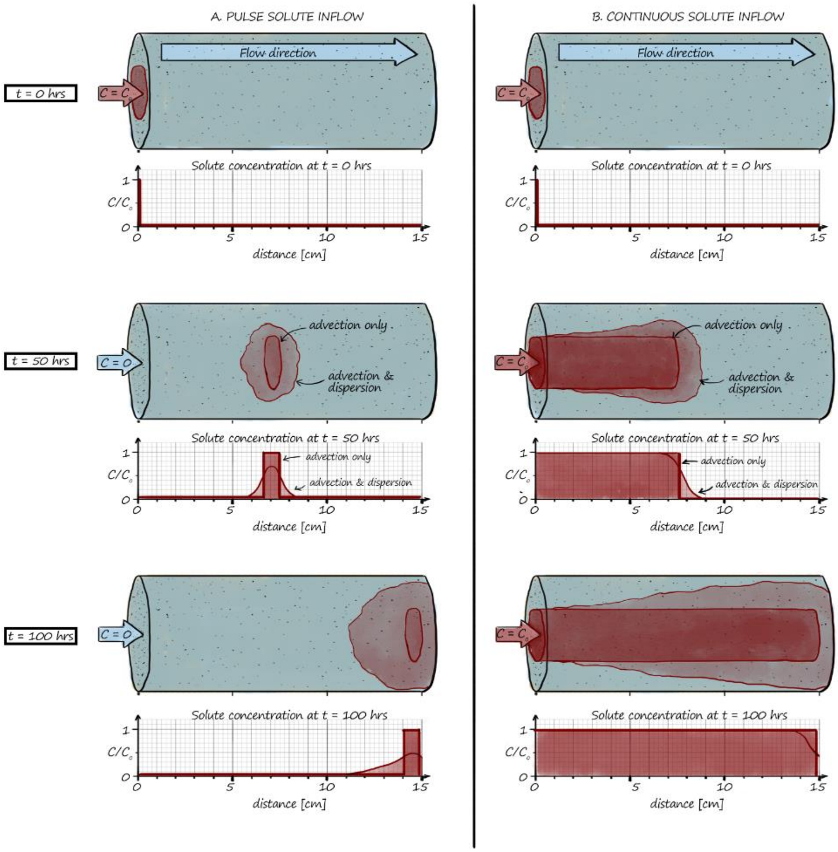

Figure 78. A 15 cm column with solute transport via advection (darker red) and advection and dispersion (lighter red) for (a) pulse flow and (b) continuous flow.

Take another look at Figure 78. Notice that dispersion is occurring in the direction of flow (i.e., horizontally) as we've described above. Some of the molecules are arriving earlier (or later) than the average linear groundwater velocity. We call this type of spreading *longitudinal dispersion*. There are three things that lead to dispersion in the direction of flow: variations in velocity within each pore (Figure 76), variations in velocity from place to place within a medium (Figure 76), and variations in the travel distance that each molecule takes. However, you also might notice that the plume in Figure

78 is also spreading in directions transverse to flow. This is another effect of dispersion. For pulse flow, the solute spreads like a ring around the plume (Figure 78a) and for continuous flow it spreads like a cone (Figure 78b). We call this additional spreading *transverse dispersion*. If there were only longitudinal dispersion for these plumes, there would only be spreading in the direction of flow (in this case horizontally), but the spreading extends out laterally as well. In this book, we will focus on longitudinal dispersion in detail – but it is important to know that transverse dispersion also exists.

Dispersion and Mass Balance

For our mass balance equation, let's reconsider the mass flux. Remember that for the *advective mass flux* (J_A) the concentration was dependent on the water flux (Equation 13). Advective flux is like a well-trained marching band: you would not expect any musicians to travel faster than the average velocity. However, if we wish to consider the effects of dispersion on mass flux, we need to consider the *variability* of velocities. The dispersive mass flux is analogous to the bad marching band, in which the variability of velocities leads to changes in mass flux. The piccolo player is moving more quickly, and the slow-marching trombone player is moving more slowly. Both players are changing the timing of the mass flux. The piccolo player increases the mass flux at an earlier time by decreasing the mass flux at the time that they would have arrived (if they had marched at the average rate). The trombone player does the opposite. For reasons that are not critical to understand, we represent the variability of velocity as if it were caused by variability of concentration. That is, the dispersive flux (J_m) is highly dependent on the change in concentration over distance.

$$J_m = -D_m \frac{dC}{dx}$$

Equation 14: Dispersive flux equation where J_m is the dispersive flux (M/L^2T), D_m is the mechanical dispersion coefficient (L^2/T) (Equation 15), dC is the change in concentration (M/L^3) in the x-direction (dx) (L).

Note: In Equation 14, the subscript m (e.g., J_m) refers to mechanical dispersion, and dC/dx represents the change in concentration (dC) with the travel distance (dx) or in this case the change in concentration in the direction of flow. There can be changes in concentration in more than one direction as we discussed above (e.g., transverse dispersion). However, for simplicity, we will only do calculations with one-dimensional dispersion or longitudinal dispersion.

Notice the negative sign in Equation 14. The negative sign is necessary because the flux occurs from high to low concentration – similar to the reasons why we need a negative sign in Darcy's law (Equation 6). The mathematics of this is not crucial, but you'll notice that if C is constant then J_m is zero!

The additional variable in Equation 14 is the mechanical dispersion coefficient (D_m).

$$D_m = a\bar{v}$$

Equation 15: Mechanical dispersion coefficient (D_m, units L^2/T) where $\bar{v}$ is the average linear groundwater velocity (L/T) and α is a medium-specific coefficient known as the dispersivity (L).

The *mechanical dispersion coefficient* is one parameter that describes the effects of all the small-scale processes that combine for dispersion. Some choose to call it a "fudge factor". The coefficient does not strictly have a physical meaning because it represents the *average* of many processes. The dispersion coefficient increases linearly with the average linear groundwater velocity. To understand why, imagine cars traveling at 35 mph. At 35 mph, they will spread out less than if they were traveling at 75 mph for the same amount of time. Similarly, if everyone in a marching band were to move at twice their original speed ($\bar{v}$) for the same amount of time, the band members would end up more spread out (dispersed).

The dispersion coefficient relates to a property of the medium known as the *dispersivity* (Equation 15). Dispersivity accounts for the *pore velocity* to determine the amount of dispersion that will occur and varies with the scale of observation. If you were to look at the age variation in one high school classroom, it would be less than if you looked at the entire school (which would be less than if you looked at the entire town). For our calculations we will focus on the mechanical dispersion coefficient for longitudinal dispersion (Equation 15). However, because transverse spreading is influenced by different factors than the factors that influence longitudinal spreading, you will sometimes see different dispersion coefficients applied in the longitudinal and transverse directions.

As a final note on dispersion, let's go back to one of our analogies: runners crossing a room full of chairs (Figure 71). If the configuration of the chairs in the room is different in one half of the room compared to the other half, the runners will spread out more. Imagine that there are 25 people on the left side of the room. At time t = 0, the people begin running to the other side of the room. Compare what would happen if 1) there are chairs randomly placed in the room but evenly distributed and 2) most of the chairs are randomly spread along the upper half of the room, while the lower half only has a few chairs. How will the obstacles influence the flow through the room in the two scenarios? For the first scenario, the presence of obstacles slows flow like soil particles slow flow (hydraulic conductivity). For the second scenario, the obstacles also slow flow, *and* the variability of hydraulic conductivity leads to dispersion. The runners in the upper half of the room travel more slowly than those in the lower half. Dispersion is dependent both on the average linear groundwater velocity (which depends on the path runners take around the chairs) and the properties of the medium (how closely spaced the chairs are in different parts of the room).

What if we wanted to account for dispersion in a solute mass balance equation? We can do this by combining the mass flux (J) due to advection (J_a) and the mass flux due to dispersion (J_m):

$$J = J_a + J_m = qc - D_m \frac{dC}{dx}$$

We represent the variability of mass flux as if it is due to variations in concentration with distance, and we assume that the velocity is constant and equal to the average linear groundwater velocity. This allows us to have a constant advective mass flux that is increased or decreased by the dispersive mass flux. Another way to think about it is that the dispersive flux is a *correction* to the assumption of a constant advective flux.

Diffusion

The process of dispersion is tied to advection; advection is driven by the groundwater velocity, and small variations in the velocity cause dispersion (spreading of the solute). But we can't stop there, we have one more mechanism of solute transport to discuss. Even if advection wasn't occurring and there wasn't groundwater flow, there would still be spreading of the solute. Wait, what!? Imagine that you have a bowl of water and the water in the bowl is completely static ($\bar{v} = 0$). However, when you place a drop of food coloring into the water, the highly concentrated droplet does not stay in one place (Figure 79). Rather the food coloring spreads out (you can try this at home). Over time, the food coloring eventually mixes into the water completely. Although there is no flow in the bowl (no advection), displacement of the food coloring still occurs due to the innate thermal motion of the food coloring molecules and the water (Figure 79). This process is called *diffusion*. Specifically, diffusion is the mixing of mass by random motion which causes dissolved solutes to move from areas of high concentration to low concentration. Diffusion tends to dominate the movement of solutes in slow moving or static systems where advection and dispersion are negligible.

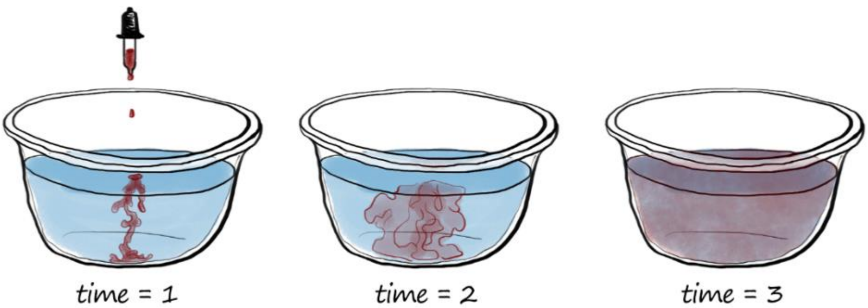

Figure 79. A drop of food coloring is added to a static bowl of water at time 1. Over time, the solute spreads due to diffusion, the mixing of mass due to random motion.

Just as flow is driven by changes in hydraulic head or the hydraulic head gradient, you will hear people say that diffusion is driven by differences in solute concentration or the concentration gradient. This description isn't totally accurate. Really, diffusion is only driven by random motion. However, if you imagine a line drawn through a bowl of water with a high concentration on one side and a low concentration on the other, random motion will cause more molecules to move from the high concentration to low concentration rather than the other way around. So, concentration differences aren't really causing diffusion, but they can be used to predict the results of diffusion processes.

To visualize diffusion, imagine a room full of people. Initially, everyone gathers near the snack table. But as the snacks run out, what happens? By nature, everyone moves, even slightly, in random ways. Someone moves randomly to the left. Then someone offsets this movement by randomly moving to the right. Then someone else makes up for that relocation by moving into the previously occupied space. The people near the wall are confined to the walls of the room, so the crowd doesn't spread in that direction. The effect is that the crowd slowly spreads out throughout the room. The spreading of the crowd continues until the concentration of people (the number of people per unit area) becomes more or less constant.

The people in the room never stop moving. In fact, they are continually mixing, even after their concentration is constant. Confusingly, in everyday life we would say that the initial crowd around the snack table "disperses". However, this isn't really dispersion as we've explained it above. Dispersion is spreading out due to *variations in velocity*. In this case, the people are diffusing – mixing because of random movement (e.g., Figure 79, Figure 80). Another difference between dispersion and diffusion is that diffusion is irreversible (you can't unmix the food coloring out of the water), whereas dispersion, at least in theory, is reversible. These subtle differences are why we refer to diffusion as a *mixing* process and dispersion as a *spreading* process.

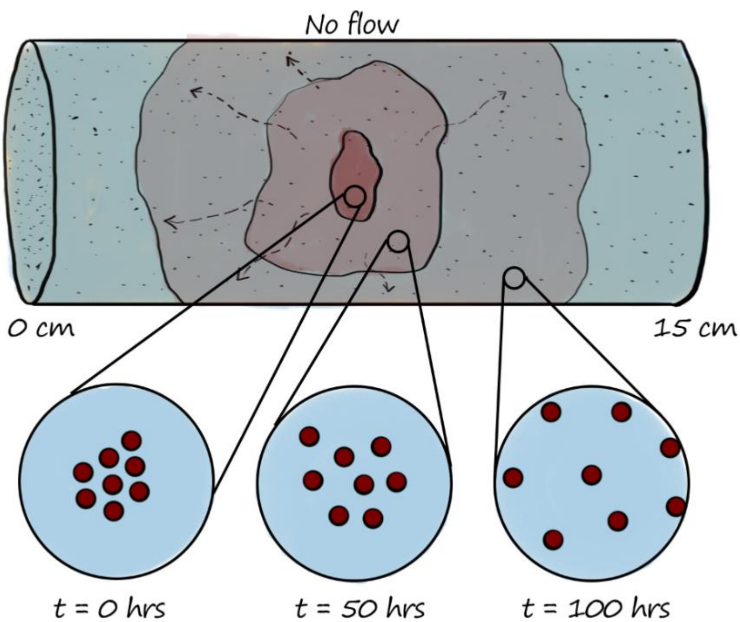

Figure 80. Diffusion in a 15 cm column with no flow. The solute begins as a blob in the middle and slowly mixes with the surrounding water over time.

Diffusion and Mass Balance

The net movement of people in a room can be predicted by the concentration gradient (i.e., the change in the number of people per unit area) in any given direction. People move through the room due to random motion. However, we can *predict* that they will move from areas of high concentration to areas of lower concentration. Solute does the same thing! Although diffusion isn't *controlled by* the concentration gradient, we use the concentration gradient to predict the effects of diffusion. Returning to mass flux, we can describe the mass flux due to diffusion (J_d) as

$$J_d = -D_d \frac{dC}{dx}$$

Equation 16: The diffusion mass flux equation (J_d, units ML^2/T), where D_d is the diffusion coefficient (L^2/T), dC is the change in concentration (M/L^3), and dx is the change in distance (L). The subscript d represents diffusion through a porous medium.

Notice that the equation for diffusive mass flux (J_d) (Equation 16) looks *exactly like* the equation for dispersive mass flux (J_m) (Equation 14). This is by design. The only difference between the two equations is the diffusion coefficient (D_d) which does not depend on the average linear groundwater velocity. Rather it depends on the medium (specifically, the porosity, n) and the nature of the solute (some things diffuse faster than others). Because diffusion often looks like dispersion, they are often treated as mathematically identical, and we combine these terms into one coefficient called the *hydrodynamic dispersion coefficient* (D_h). Confusingly, you will often hear the hydrodynamic dispersion coefficient referred to just as the "dispersion coefficient" ... but remember D_h is really the combination of *both* processes, dispersion and diffusion.

Now we have all the pieces we need to form an equation to describe the *total solute mass flux* (J) of a system.

First, we can add up the mass fluxes for each of the solute transport processes (advective, dispersive, and diffusive):

$$J = J_a + J_m + J_d$$

Then, we can replace the variables (J_a, J_m, and J_d) with their individual equations:

$$J = qc - D_m \frac{dC}{dx} - D_d \frac{dC}{dx}$$

Lastly, we can add in the hydrodynamic dispersion coefficient ($D_h = D_m + D_d$), and the final advection-dispersion equation is:

$$J = qc - D_h \frac{dC}{dx}$$

Equation 17: The final advection-dispersion equation (J, M/L^2T) where q is the Darcy flux (L/T), C is concentration (M/L^3), D_h is the hydrodynamic dispersion coefficient ($D_m + D_d$, L^2/T), dC is the change in concentration (M/L^3), and dx is the change in distance (L).

The advection-dispersion equation (Equation 17) is a valuable tool for calculating changes in solute transport. For example, it can be used to predict the movement of a contaminant plume, and the predicted concentration distribution after the source of contamination has been removed. Example calculations using the advection-dispersion equation require mathematical solutions that are more complex than the scope of this text. Therefore, we only present this equation to explain concepts and their implications.

Solute Transport: Case Study

Water Contamination

Porous media in the subsurface are surprisingly effective at filtering water and supplying clean water for drinking, agriculture, and other human and environmental uses. Unfortunately, despite the natural processes that filter groundwater, it is still relatively easy to contaminate it. Humans use products such as detergents, fertilizers, gasoline, and road salts that carry harmful solutes. Therefore, human activities often contaminate groundwater resources. In addition to human-caused contamination, there are natural sources of contamination, such as seawater or contaminant-rich rock formations that introduce solutes to systems.

In some cases of contamination, we can know that a solute came from a specific location (e.g., an industrial waste pipe, a localized oil spill, a leaky underground storage tank, etc.). In these cases, we can "point" to the source of contamination. Scientists refer to this type of pollution as *point source* pollution. In other cases, contamination occurs over a widely distributed area, such as pesticides applied over agricultural fields, or runoff from a network of roads. For these types of contamination, it is more difficult to point to a specific source. Scientists call these *nonpoint sources* of pollution. However, the processes that control solute transport in the subsurface are the same regardless of the source of contamination.

As hydrogeologists, we must try to understand the movement of solutes through the ground to prevent or remediate contamination. In some cases, we will need to determine who caused contamination so that they can be held responsible for cleanup. In other cases, we will need to determine if an existing plume of contamination will ever reach a particular location such as a municipal water supply or a spring. The models that we build to characterize underground transport are the same for each case. However, what changes is the availability of data and the level of uncertainty in our assessments.

Solute Transport Case Study

Transport of a Single Molecule via Advection

To understand how solutes might enter a system and how to analyze their movement, let's go back to our agricultural field example from Figure 59. Imagine that the owner of this agricultural field is using equipment to spray pesticide. She plans to start treatment around piezometer 10 (Figure 81) and work her way diagonally across the field. However, prior to starting operations a malfunction occurs. The highly concentrated pesticide (50 g/L) is not properly diluted (yet), and a spill occurs near piezometer 10 allowing the pesticide to infiltrate into the groundwater. What will happen to this solute over time?

Using what we already understand about groundwater movement and solute transport, let's analyze this scenario. In contrast to an application of pesticide picked up by runoff (where contamination is discrete and/or discontinuous), this highly concentrated pesticide spill is an example of a *point source pollution*. Additionally, this pesticide spill is a *pulse release* that occurs instantaneously (or over a short period of time).

To begin, we will track *one solute molecule* through the system. Imagine that after the solute molecule infiltrates into the ground, it moves exactly as fast as the average linear groundwater velocity (in other words, we will concentrate on advective transport of the molecule first). Look at Figure 81.

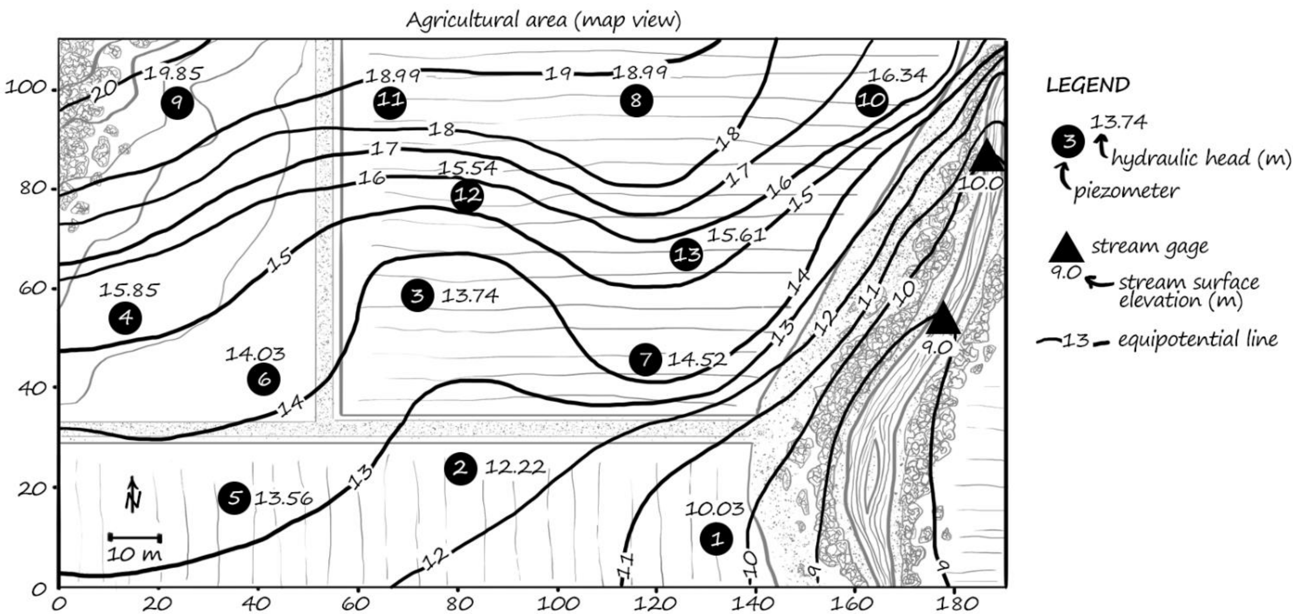

Figure 81. Agricultural field with equipotential lines.

If the solute molecule infiltrates into the ground at piezometer 10, where does it travel? Well, where does the groundwater travel? Remember your tools from previous chapters! First, draw flow lines. Pay close attention to the flow line near piezometer 10 – this will give you a general idea of where the solute molecule travels (by advection). After you finish drawing, remember that an equipotential map with flow lines is only a tool for *visualizing* groundwater movement. If we want to understand the water flux (mathematically), we need to draw a flow tube (which is a key component of a flow net) (Figure 82b).

The flow tube in Figure 82a is shown in detail in Figure 82b. The hydraulic head upgradient of piezometer 10 is 19 m and the hydraulic head downgradient of piezometer 10 is 9 m: the groundwater flows towards the stream. Within the flow tube the soil is a homogeneous loam with a porosity of 0.43 and a hydraulic conductivity of 0.25 m/day. The length of the flow tube is 52 m along the centerline of the tube. Using this information, can you estimate how far the solute would travel (by advection alone) after two days?

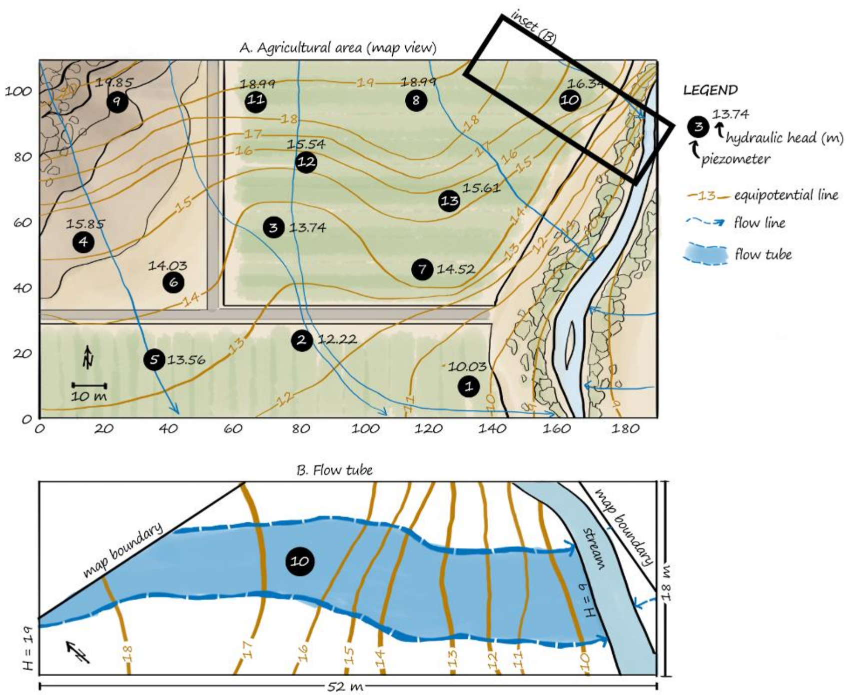

Figure 82. Agricultural field with (a) one flow tube around piezometer 10 highlighted, and (b) a detailed view of the same flow tube.

To do this you must first calculate the water flux:

$$q = -K\nabla H$$
$$q = -\left(0.25\frac{m}{day}\right)\frac{(9\ m - 19\ m)}{52\ m} = 0.05\frac{m}{day}$$

Next, you must calculate the average linear groundwater velocity to track the solute through the porous media. Note: $\bar{v}$ is the velocity along the centerline of the flow tube:

$$\bar{v} = \frac{q}{n}$$
$$\bar{v} = \frac{0.05\frac{m}{day}}{0.43} = 0.12\frac{m}{day}$$

Lastly, you must use the displacement equation to understand how far the solute travels in 80 days:

$$\Delta x = \bar{v}t$$
$$\Delta x = \left(0.12\frac{m}{day}\right)(80\ days) = 9.6\ m$$

The calculations above tell you that a pesticide molecule moves 9.6 meters in 80 days via advection. To better visualize this, go back to the map and mark the solute's location along the flow line. Measure from piezometer 10 along the center line of the flow tube. Note: there is a scale on the map to help you determine distances. Once you mark the point, you'll notice that the spill is nearly half the distance to the stream (Figure 83). Yikes!

If you look at the hydraulic head distribution near piezometer 10, you'll notice the stream is *gaining*. Therefore, any pesticide molecule that migrates downward to the water table will follow the same path as the groundwater, and eventually end up in the stream. With this, the farmer is curious how long it will take for the solute to enter the stream. If the stream is 25 m away from the contamination location, can you use the displacement equation (Equation 11) to help her answer this question?

$$\Delta x = \bar{v}t$$
$$t = \frac{\Delta x}{\bar{v}}$$
$$t = \frac{25\ m}{0.12\frac{m}{day}} = 208.3\ days$$

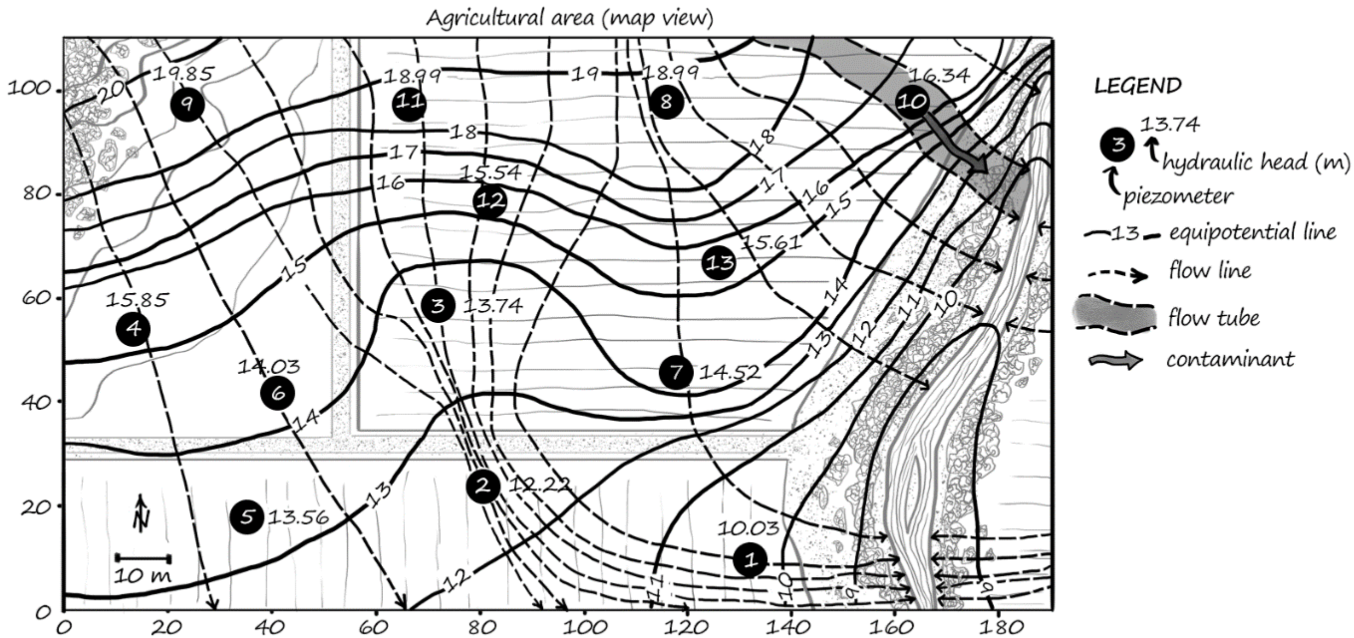

Figure 83. Agricultural field with a flow tube and location of contaminant after two days.

When we covered flow nets, we discussed how the groundwater velocity is not necessarily constant all along a flow tube. When there is variability in groundwater velocity, there is variability in the width of the flow tube. For example, if the flow tube gets wider, the water is slowing down (assuming that the porosity is constant). If the tube is narrower, the water is speeding up. If you want to be more exact in your flow tube calculation above, you must calculate how long it takes for a water molecule to flow along each section of the flow tube (where the width varies) and then add up the travel times for each section. We won't do that here, because you would probably use a computer model to do this for a real-life scenario!

Transport of a Collection of Molecules

The sequence of visualization techniques used above were for a *single solute molecule*, but we could do something similar with a *collection of molecules.* Imagine that near piezometer 10, a group of molecules are released instantaneously. Assuming transport via advection alone, the molecules arrive at the same time and move as one blob. Additionally, the concentration that arrives at the stream (after 208.3 days) is the maximum (or initial) concentration (50 g/L). However, advection is only part of the story. Just as with an imperfect marching band, not all contaminant molecules move together in perfect harmony. In fact, each molecule travels at slightly different speeds along different paths due to dispersion and diffusion. This has implications for the maximum concentration, arrival time of the first solute molecule, and the duration of the contamination.

After the molecules are spilled, some molecules arrive earlier (like the piccolo player) and some later (like the trombone player). Therefore, the spreading of the solute pulse can cause contamination for a much longer time than it took for the spill to occur (Figure 84b). Additionally, the shape of the bell curve in Figure 84b is determined by the hydrodynamic dispersion coefficient (which combines the effects of dispersion and diffusion). The higher the coefficient the more spreading there will be. The more the solute spreads, the lower the maximum concentration at the peak of the bell curve. Keep in mind that the same total amount of solute (represented by the area under the curve) is maintained for each scenario.

Solute Transport Complications

Unfortunately, the real world is more complicated than our simple agricultural field scenario above (Figure 84). Heterogeneous soils, sorption, decay, and time-varying flow can all complicate our ability to model solute transport. For clarity, we wish to discuss each of these.

Soil heterogeneity (Figure 85a), such as different porosities or grain sizes in different layers, causes more solute spreading. Heterogeneity increases dispersivity, or the differences in pore velocity among all possible flow paths. It also creates scenarios where the concentration at a given location can be quite different than at other nearby locations, which can make solute transport in heterogeneous soils difficult to predict.

Sorption, which is when a solute temporarily sticks to solid particles as water flows past, also complicates things (Figure 85b). We often model sorption scenarios using a *retardation coefficient* (R). The retardation coefficient affects the velocity, slowing down travel times for the solute compared to the water; retardation factors are determined by the chemistry of the soil and the solute. Sorption causes the peak solute concentration to arrive later than expected; as solute molecules are sorbed, fewer molecules are in the dissolved phase. Sorption can be thought of as solutes temporarily "stepping" out of the groundwater flow. We usually consider reversible sorption, when solutes move onto the soil surfaces (when the solute concentration in the water is high), and then move back off the solids (when the concentration in the water is low). Sorption also causes individual molecule velocities to vary, which increases spreading.

Decay describes a range of processes that can cause contaminants to break down in the subsurface, including reacting chemically with soil solids or being consumed by microbes (Figure 85c). If we understand the nature of the degradation, we can model the breakdown as a decrease in concentration as a function of travel time. Decay causes a lower maximum concentration due to the processes that consume mass. Note: some types of degradation depend on the concentration – so they may be related to dispersion – and some types of degradation are (or are not) affected by sorption, so they may (or may not) be affected by retardation as well.

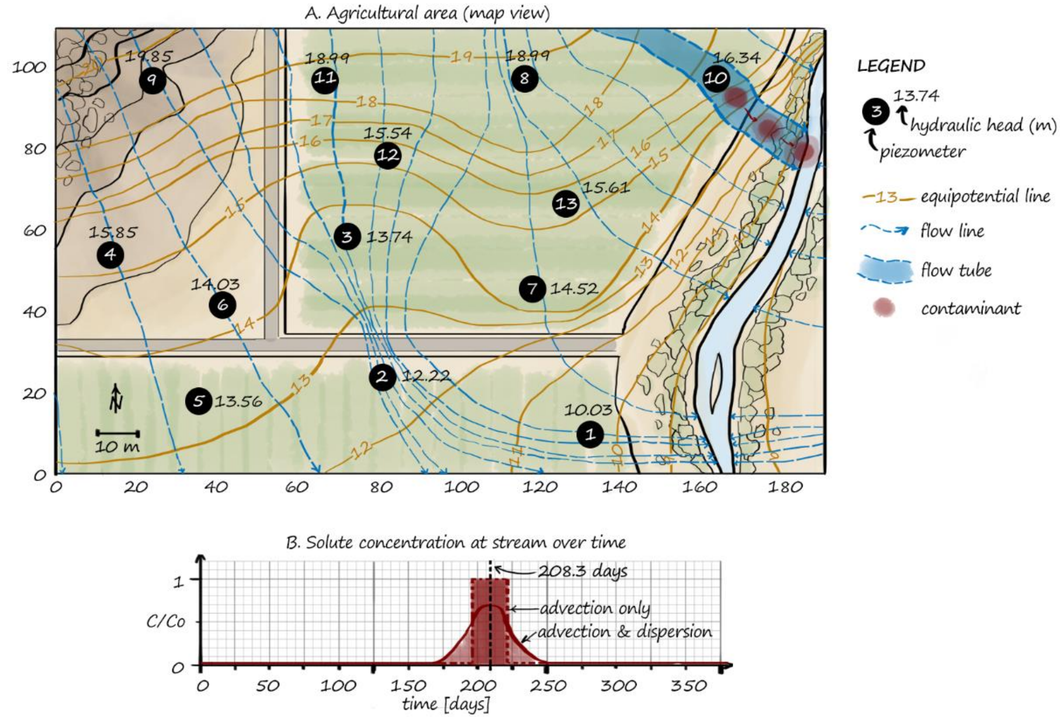

Figure 84. Agricultural field (a) with a solute with advection only (darker red) and advection and dispersion (with diffusion) (lighter red). The travel time and concentration are shown in (b) the graph below the figure.

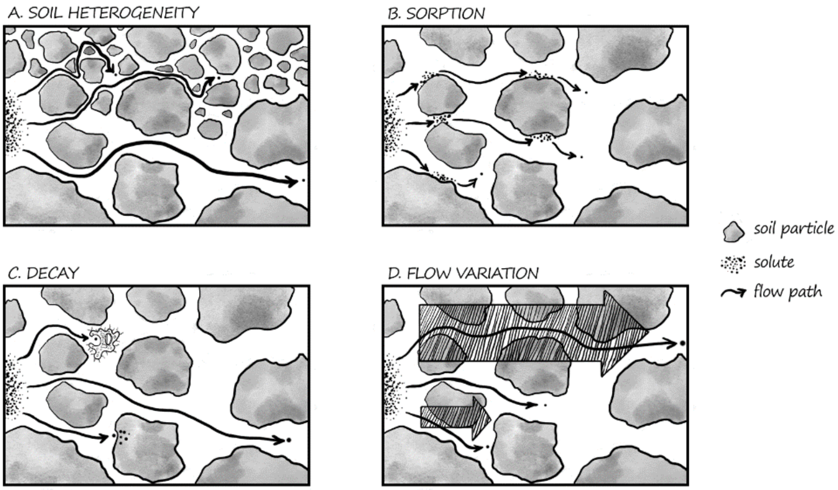

Figure 85. Solute transport processes that complicate dispersion processes such as a) soil heterogeneity, b) solute sorption to the particles, c) solute decay, or d) variation in water flow.

Lastly, remember that advection is a key driver of solute transport. Therefore, if the direction or velocity of groundwater flow changes with time, then the resulting solute plume will be much more complicated (Figure 85d). It will be displaced and smeared even more than usual. This effect is often overlooked by hydrogeologists who assume (for simplicity) that flow is steady state.

Conclusion

At this point you have learned not only how to track a water molecule, but also how to visualize what occurs if that water molecule has a solute dissolved in it. Solute transport is an important part of hydrogeology. In fact, there are whole textbooks dedicated to the topic! Solutes in the subsurface are not always harmful, but when they are, they have the potential to have negative consequences on humans and ecological communities. Therefore, in many cases, we want to know how to track solutes. The three mechanisms introduced in this chapter (advection, dispersion, and diffusion) are the basics of solute transport. The basics of solute transport are foundational for studying real-world questions regarding contaminate spills, storage, and wastewater management. You must learn algebra before you can do calculus. We hope that this chapter has sparked your curiosity and encouraged you to dive deeper into the complexities of this important topic.

What to Remember

Important Terms	
advection	mass flux
average linear groundwater velocity	mechanical dispersion coefficient
continuous flow	nonpoint source
decay	plug flow
diffusion	point source
dispersion	pulse flow
dispersivity	retardation
displacement	solute
hydrodynamic dispersion coefficient	sorption

Important Equations
$v = \frac{q}{n}$
$\Delta x = vt$
$J = qc - D_h \frac{dC}{dx}$

Important Variables		
Symbol	**Definition**	**Units**
q	flux	L^3/T
n	porosity	-
v	average linear groundwater velocity	L/T
x	length	L
t	time	T
J	mass flux	M/L^2T
c	concentration	M/L^3
D_h	hydrodynamic dispersion coefficient	L^2/T

Chapter 6

Pumping Wells

Introduction

You are likely familiar with the term "well" as something you see in old movies and cartoons, where a bucket is lowered down a hole and filled with water. Have you ever thought about how a well works? The type of well you are imagining from the movies is called a dug well; if the ground is soft and the water table is shallow, then a large diameter dug well (lined with solid material to prevent it from collapsing) can be used to extract water. The concept is the same as if you were at the beach and dug in the sand until the hole started to fill with water. If you could keep the sides of your hole from collapsing, you could scoop water out of the hole. When you scoop water out of the hole, where does the water come from? It flows in from the surrounding soil. Why does this happen? It is because by taking water out, you have lowered the hydraulic head causing water to flow from the surrounding soil into the hole. Dug wells work the same way!

For deeper water tables, drilled wells are common. Drilled wells can be more than 1,000 feet (about the height of the Empire State Building) deep. Most modern wells are drilled wells and require a pump to lift water to the surface. When water is withdrawn, it can temporarily lower the water level in and around the well. If you take water out relatively slowly, compared to the rate of groundwater flow, the well might seem to refill almost immediately. But, if you take out water quickly (which is common with pumping), you can decrease the groundwater level in and around the well significantly. Think about the impacts of this. How might a depression of the water table impact the *hydraulic head* at and around the well? Can you hypothesize how these changes in head might affect the hydraulic gradient and the direction that groundwater flows in the surrounding area?

Agricultural Field with a Pumping Well

At this point, you likely have a good idea of how groundwater moves in the agricultural field example. If you don't, take a moment to visualize groundwater flow by physically or visually drawing flow lines on Figure 86. Next, after you have visualized groundwater flow, imagine what would happen if there was a pumping well at piezometer 10 (in the northeast corner of the field)? Thus far, you understand that groundwater pumping withdraws water. So, what would the withdrawal of water do to the groundwater levels in piezometer 10 and the surrounding area?

Let's explore groundwater pumping on the agricultural field by looking at the aerial map in Figure 87a. Notice that there is a pumping well (shown with a crossed circle) that was previously piezometer 10. Compare the hydraulic head value of the pumping well in Figure 87a to the hydraulic head value in piezometer 10 in Figure 86. Has it increased or decreased? The hydraulic head previously was 16.34 m and now it is 11.50 m. Therefore, the hydraulic head has decreased due to pumping. What about the hydraulic head in the surrounding piezometers? Previously, the hydraulic head values were 18.99 m, 18.99 m, 15.54 m, and 15.61 m at piezometers 8, 11, 12, and 13, respectively (Figure 86). If you compare this to Figure 87a, the hydraulic head in all these piezometers has also decreased. Brainstorm: why did the hydraulic head decrease in all the surrounding piezometers, if water was pumped at only one location (piezometer 10)? Lastly, compare the equipotential lines in Figure 86 to the lines in Figure 87a. Make observations. How have they changed? Are they closer or further apart? What are some hypotheses that could explain these changes in the equipotential lines?

To gain better insight, look at a cross section of the pumping area both before and after pumping (Figure 87b). Notice how the hydraulic head changes around the well; the changes can most easily be seen as the difference between the two lines (the before pumping line and the after pumping line). The greatest change in hydraulic head happens at the pumping well (piezometer 10). However, piezometer 8 has the second largest drop, and piezometer 11 the next largest. There is no change in the hydraulic head at piezometer 9. The greatest changes in hydraulic head occur near the well and they decrease as you get further away from the well. When a drilled well extracts water, it pulls water out from groundwater storage. First, it extracts water from the pores that

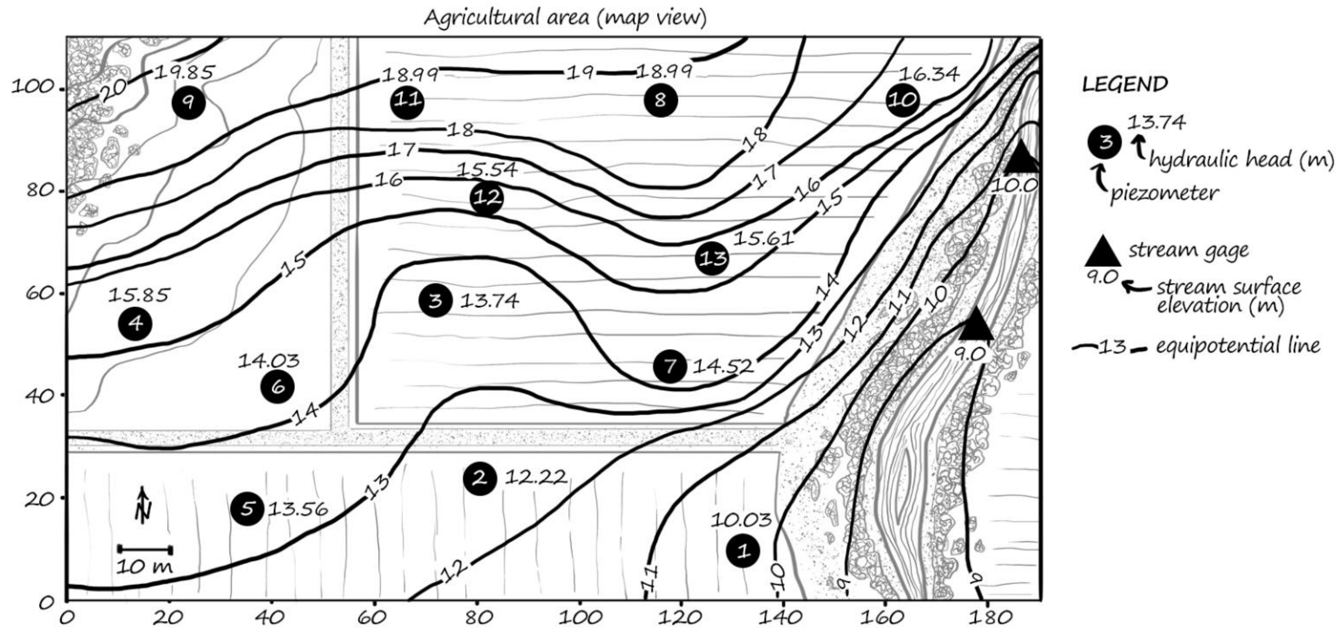

Figure 86. Agricultural field with piezometers installed. The solid lines are equipotential lines.

are in closest proximity to the bottom of the well. However, as it depletes the water around the well, changes in the hydraulic head gradient cause nearby water to move toward the pumping well. The water surrounding piezometer 10 is not isolated from the rest of the groundwater; it is all connected.

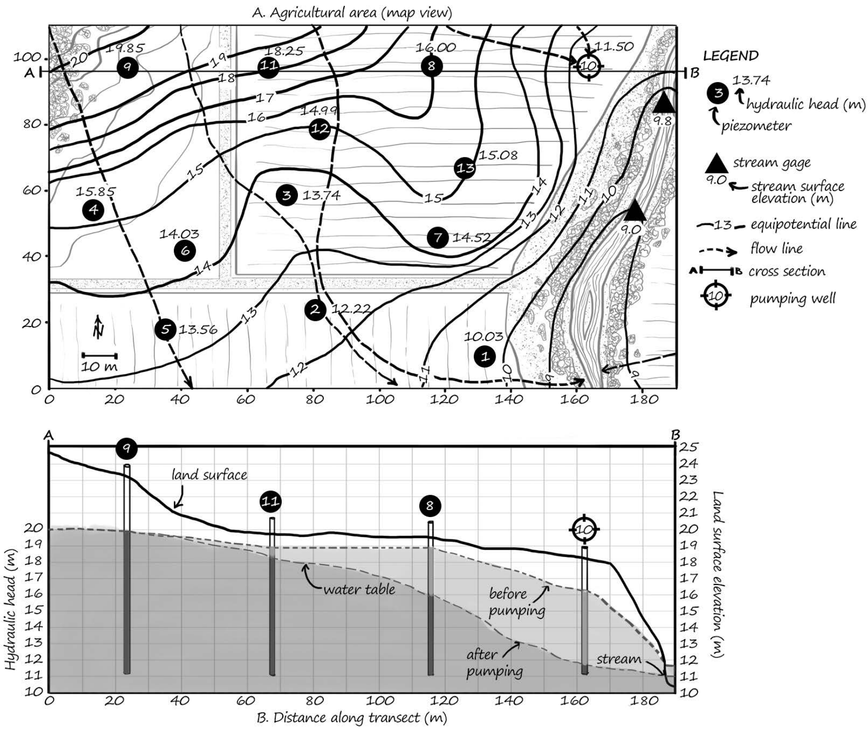

Figure 87. An aerial map a) of an agricultural field with pumping at piezometer 10 and a cross section b) with hydraulic head values both before and after pumping.

Piezometers 8, 10, and 11 (those in the cross section in Figure 87b) are not the only locations in the agricultural field that have changes in head due to pumping. In fact, after pumping, there is a change in hydraulic head that occurs in all directions extending laterally from the well. The collective change in the water table due to pumping is something hydrologists call a *cone of depression*. This is because the drop in water level surrounding a pumping well often looks like a cone. To visualize this cone, imagine that the cross-sectional view of the water table after pumping (Figure 87b) extended in all directions surrounding the well; the tip of the pumping well would be the lowest point of the cone (piezometer 10 on the cross section), and changes in head extend out in all directions. Therefore, if you took a cross section of the cone, you would see a v-shape as in Figure 88b (now, just imagine that in 3D!).

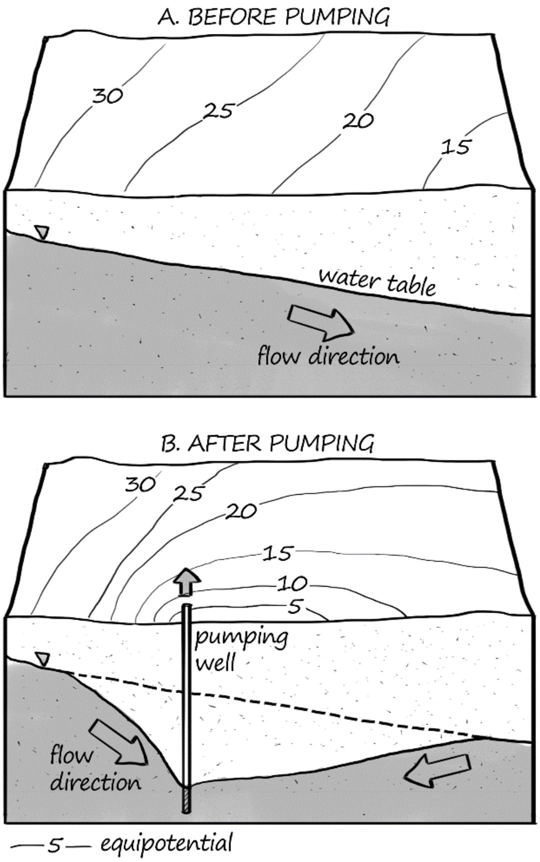

Figure 88. A simple groundwater system a) before pumping and b) after pumping. Equipotential lines are shown at the surface. After pumping, the flow direction changes and a cone of depression forms around the well.

Figure 88 shows a conceptual example of changes that can happen to the hydraulic head distribution when groundwater pumping occurs. Before pumping (Figure 88a) the hydraulic head (as shown by the equipotential lines) varies from 30 m to 15 m, and the groundwater generally flows from left to right. Perhaps prior to pumping, the water is flowing downhill, parallel to the land surface. However, what about the hydraulic head distribution after pumping? Figure 88b shows how the hydraulic head distribution changes after pumping. The equipotential lines upgradient from the well are closer together, showing a steeper gradient than the gradient that existed previously. Additionally, downgradient of the well, the water is now flowing toward the well rather than away from it (i.e., in the opposite direction from the original gradient). Had the water been flowing downhill prior to pumping, this an example of a scenario where groundwater could be flowing *uphill* due to an increase in pressure head. Alternatively, pumping could also have decreased the gradient without reversing it.

Wow this is significant! Groundwater pumping can reverse a hydraulic head gradient and cause groundwater to change direction and flow towards a well. To understand the implications of these changes in hydraulic head, let's go back to our agricultural field. After a period of pumping, the landowner notices that the level of water in the stream is decreasing. She is curious if the changes in the stream are related to pumping. Is that even possible? Remember, surface water and groundwater are connected. So, changes in groundwater can surely affect surface waters. Previously we discussed *gaining stream*s (Figure 63). For gaining stream systems, the water table has a greater hydraulic head than the stream (Figure 89a). Therefore, the hydraulic head gradient causes water to flow from the ground into the stream. Perennial systems, which rely on groundwater for baseflow at times when there is no precipitation or snowmelt, are often gaining streams. In contrast, when the hydraulic head gradient near a stream is reversed (causing water to flow away from the river) – water can flow from the stream into the ground. This is called a *losing stream* (Figure 89b). Hydraulic head gradients are often reversed due to groundwater pumping (water is moving toward the pump rather than toward the stream). Additionally, even if a well just reduces the rate at which a stream gains water from groundwater, it can reduce the streamflow. This is a process known as well capture.

You will sometimes hear news stories about how groundwater pumping is drying up rivers! This can happen when the hydraulic head gradient is reversed, and pumping wells are draining water from streams (Figure 89b). However, keep in mind that groundwater pumping is not the only scenario for a losing stream. The water table could be lower than the stream because of pumping, but the stream level could also be higher than the water table because of high flow events (e.g., precipitation or spring runoff), causing a temporary increase in recharge from the stream to groundwater. Any time the water level in a river is higher than the surrounding water table, the stream is losing. Therefore, losing streams are important sources of recharge for nearby groundwater.

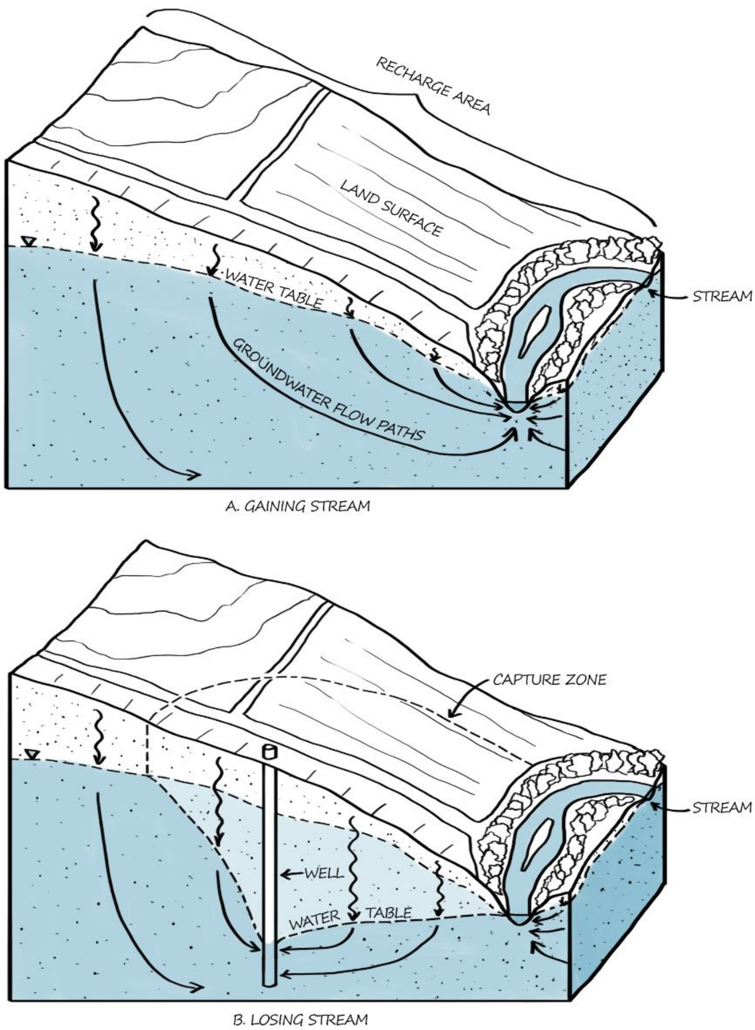

Figure 89. Agricultural field with a a) gaining stream and b) losing stream (with well capture).

As stated above, during pumping, water is removed from the subsurface. Initially the water is removed from locations closest to the well, but over time water is removed from regions further from the well. *Storativity* and *transmissivity* control where water is removed and to what extent. These can be tricky concepts to grasp. Imagine that you are standing at a hot dog stand at the front of a lengthy line of people. You are told that the hot dog costs $2, but you realize you only have $1. You turn to the person behind you and ask if you can borrow $1. If you have $1 and they give you $1, you are golden. However, what if they say that they are only willing to give you $1 if the person behind them gives them 50 cents? And then the person behind them demands 25 cents from the next person in line? How long does it take for you to get your hot dog? It depends on how much each person is willing to contribute (e.g., it will take longer if the first person gives you the dollar in exchange for 90 cents from the next person) and how long it takes each person to convince the person behind them to join in the funding scheme. Pumping water from a well is similar. The area around the well "gives up" water to satisfy the pumping demand. But, once the area gives up water, it also takes water from a bit farther away. As you move further from the well, more distant zones give up water from storage, but they also receive less water from farther away (just as the person near the back of the line gives up 50 cents but only receives 25 cents from the next person).

Storativity is a measure of how much the medium must "give" to the funding scheme, or to be precise, how much water is produced per unit drop in head due to pumping. An aquifer with a low storativity will need a higher change in head and/or the effects of pumping will have to extend over a larger area to provide a given volume of water. *Transmissivity* is a measure of how easily the "message" for more water is transmitted away from the well. It also controls how much of a gradient is required for a certain amount of water to flow towards the well. An aquifer with a low transmissivity requires a larger gradient to generate the same amount of flow, so the cone of depression is steeper.

The impact of pumping on nearby piezometers and stream systems is dependent on the shape and extent of the cone of depression. Each system is unique. If the farmer of our agricultural field wants to know how pumping will affect the stream – she needs to consider the rate and duration of pumping, properties of the aquifer (such as the transmissivity and storativity), and the distance from the pumping well to the stream.

Solute Transport and Pumping Wells

Pumping wells do not just affect water quantity, they can also mobilize and transport contaminants. For example, imagine that our farmer replaces piezometer 5 with a pumping well to serve as her regular water supply (Figure 90). The water quality at piezometer 5 is excellent – no complaints. However, she tests the groundwater regularly throughout the field for certain compounds. Over time, she notices that some piezometers on the field are detecting undesirable contaminants at exceptionally low concentrations. After a little more time, the concentration begins to increase. If this trend continues, the water will eventually be unusable! What can she do? Where is the contamination coming from? Is the pumping at piezometer 5 causing other locations on the field to be vulnerable to contamination? Will the contamination eventually reach the pumping well? Is there anything that can be done to solve this problem?

To begin tackling these questions, we must first understand how water flows on the field. Prior to diving into the nitty gritty of solute transport and pumping wells, can you remember how we visualized groundwater flow using piezometer data? By using equipotential and flow lines! Using the piezometer information in Figure 90, draw the appropriate equipotential and flow lines considering the newly installed pumping well at piezometer 5. Keep in mind some of the basic rules of equipotential lines (e.g., they never cross each other and are lines of equal head) and flow lines (e.g., they are always perpendicular to equipotential lines).

Now that you have an equipotential map with flow lines (Figure 91b), how would you describe the direction of groundwater movement? Which direction does a molecule move if it is initially located at piezometer 2? What about if it was located at piezometer 1? This is the same approach we used in previous chapters. Water flows in the direction of the greatest hydraulic head gradient. In this case, water in the southwest corner of the field is generally moving toward piezometer 5 (the pumping well). In some cases, water that was previously moving downgradient toward the stream (e.g., piezometer 2) (Figure 91a) is now moving toward piezometer 5 (Figure 91b). In other locations where water was already moving downgradient, pumping has increased the gradient, making it flow more quickly (e.g., piezometer 4). Take a moment and compare the flow lines from before (Figure 91a) and after (Figure 91b) pumping.

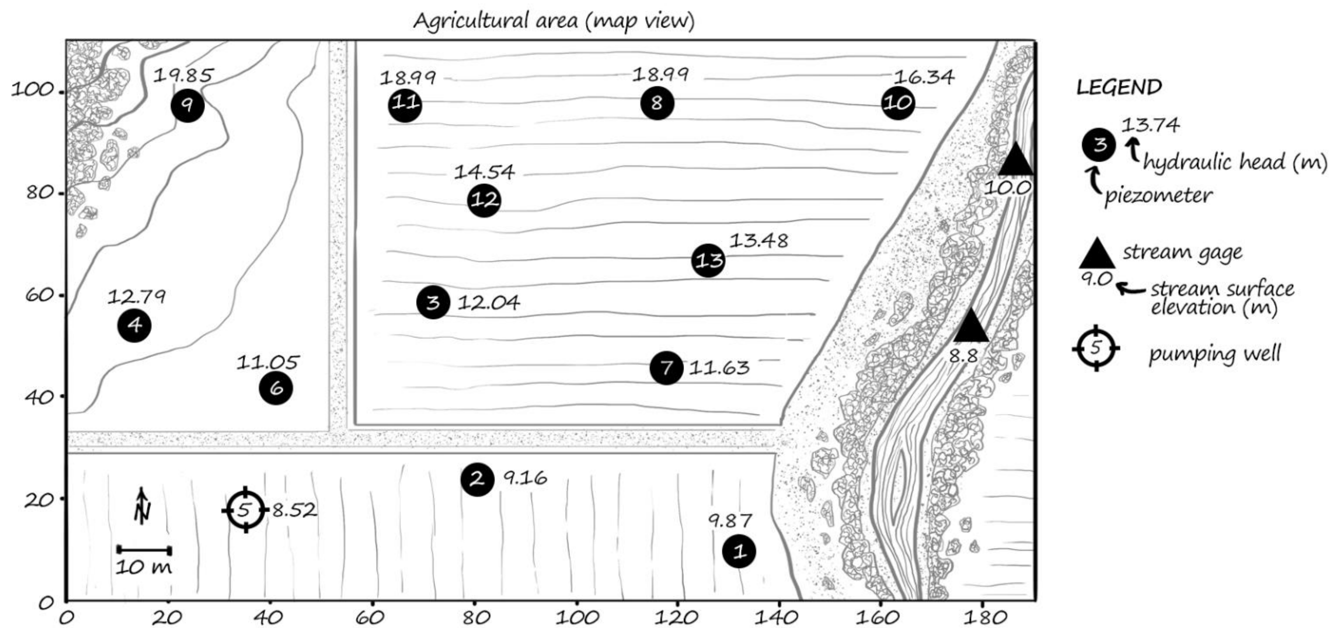

Figure 90. Agricultural field and new piezometer readings after pumping at well 5.

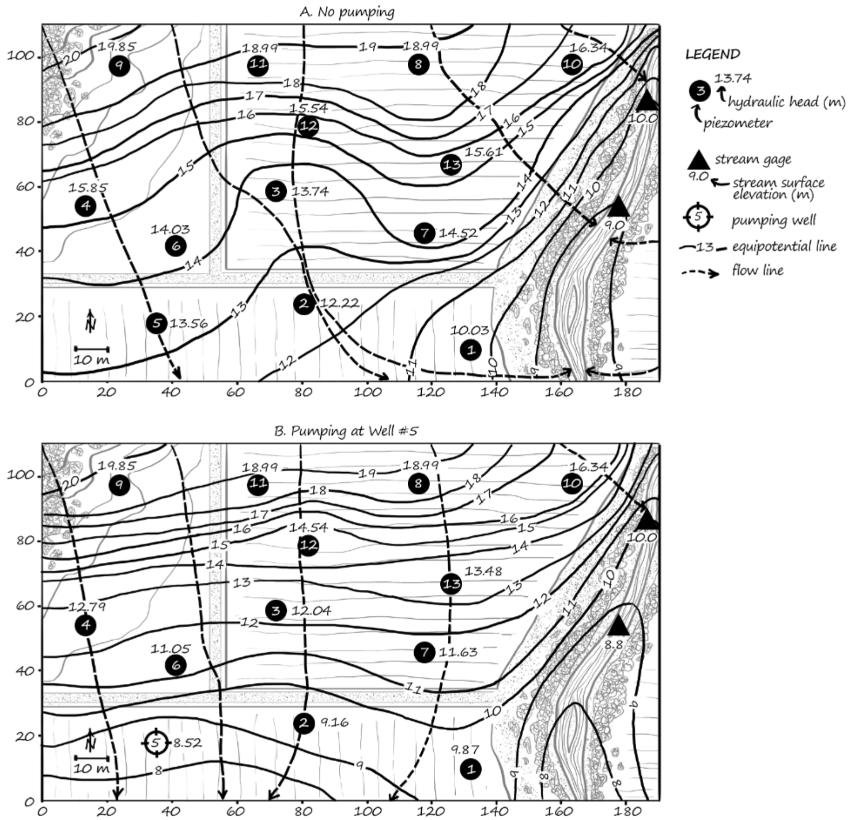

Figure 91. Hydraulic head distribution (a) before and (b) after pumping at well 5.

Groundwater movement is a key component of solute transport. Now that you have a fairly good idea of the direction of groundwater movement (Figure 91), let's begin tracking solute molecules. After some water quality testing in each of the piezometers, you find that the contaminant is only located in piezometers 8, 13, 7, and 1 (Figure 92). First, focus on advection; pretend that the compound is being transported only by the flow of water. Can you hypothesize where the molecules originated from using flow lines? Start at each of the contaminated piezometers and backtrack along existing flow lines. If there are not flow lines drawn exactly where you need them, you can also run your finger perpendicular to the equipotential lines to gain a general understanding (or lightly draw them on your own using a pencil).

After tracing the flow lines, you'll notice that they generally converge near piezometer 8 (Figure 92). This is a good initial guess of the source of the contamination! Next, consider the flow velocity (i.e., the average linear groundwater velocity) to gain *quantitative* information about the advective transport. For example, calculate the average linear groundwater velocity of the contaminant and make an approximation of how long it will take for the contaminant to travel from piezometer 8 to piezometer 1.

For the transport calculations suggested above, you need site-specific information. The soil is a homogeneous loam with a porosity of 0.43 and a hydraulic conductivity of 0.25 m/day. The length of the flow tube from piezometer 8 to piezometer 1 is 110 m. The hydraulic head at the top of the flow tube is 18.5 m and at the bottom is 8 m.

First, find the groundwater flux using the hydraulic conductivity and flow tube information (the hydraulic head gradient):

$$q = -K\nabla H$$

$$q = -\left(0.25\frac{m}{day}\right)\frac{(8.5\,m - 18.5\,m)}{110\,m} = 0.0227\frac{m}{day}$$

Next, calculate the linear groundwater velocity to track the solute through the porous media. Note: $\bar{v}$ is the velocity along the centerline of the flow tube:

$$\bar{v} = \frac{q}{n}$$

$$\bar{v} = \frac{0.0227\frac{m}{day}}{0.43} = 0.053\frac{m}{day}$$

Lastly, use the displacement equation to understand how long it takes the solute to travel 110 m (the distance from piezometer 8 to piezometer 1):

$$\Delta x = \bar{v}t$$

$$110\,m = \left(0.053\frac{m}{day}\right)(t) = 2{,}075\,days$$

Using the series of equations presented above, you learn that it will take the solute approximately 2,075 days (or about 5.69 years) to travel across the field (from piezometer 8 to piezometer 1) (Figure 93). Although these calculations are only for advective transport, they provide a good *initial guess* of the contaminate transport on the field.

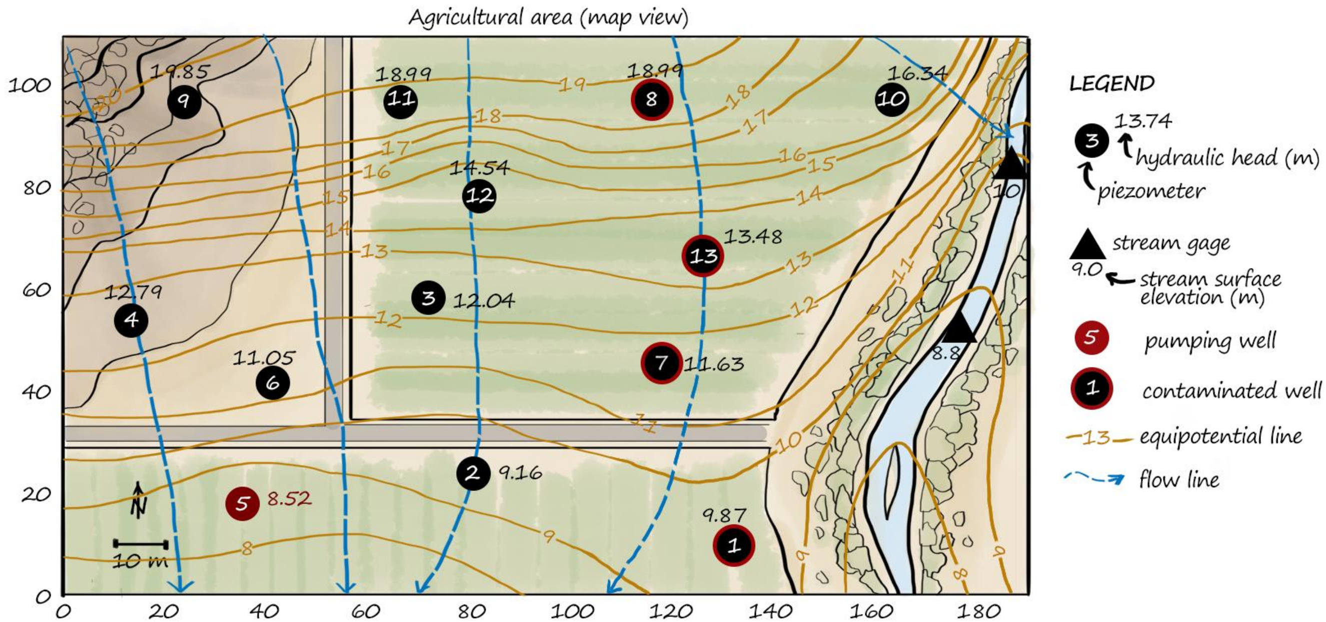

Figure 92. Equipotential and flow lines after pumping with the contaminated wells highlighted (circled in red).

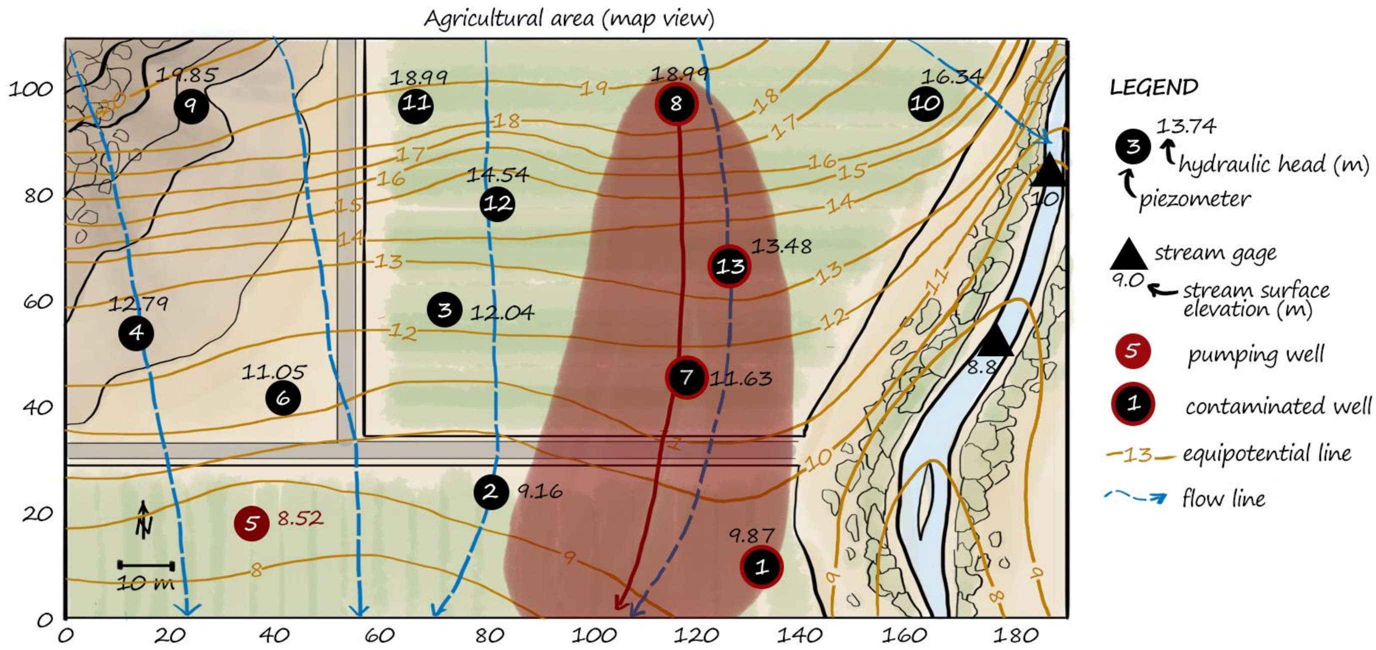

Figure 93. Estimated extent of contaminant plume based on contaminated wells and hydraulic head distribution.

As you learned previously in the previous chapter, advection is a key driver in solute transport, but it isn't everything. If you wanted a more precise model, you would need a more experienced hydrologist to construct a transport model that considers additional factors such as dispersion, diffusion, sorption, and degradation of the contaminant. This model would require even more information about the chemical properties of the contaminant and the subsurface! Additionally, a simple calculation using the displacement equation (Equation 11) gives you a general idea of how long it will take for the compound to reach the pumping well. But this travel time is only for the current pumping rate… What if the farmer reduces pumping? Could you predict what would happen in the future? Will the field require additional remedial action to prevent further contamination? Could the farmer install another pumping well to capture the contaminant plume? This is where computer models are handy. You can use models to run several different scenarios before deciding what action to take (keeping in mind the assumptions that went into the model predictions).

Piezometers vs. Groundwater Monitoring Wells

We have described how you can use piezometers for monitoring. However, you will often hear the term *groundwater monitoring well* and piezometer used interchangeably. Both are pipes that allow water to rise and fall as a way of measuring groundwater. But monitoring wells have perforations (or slits) over most of the length of the pipe, so the water level inside the pipe reflects the total water pressure integrated over the perforated length (Figure 94). A well-designed piezometer only has perforations at the bottom of the tube and measures the hydraulic head at the end of the pipe (remember, like a manometer) (Figure 94). Piezometers can detect the water pressure in the soil at a certain depth and are good for getting a picture of the groundwater system at a specific point. However, if you want to get an overall picture of the depth to groundwater, you might want to use a monitoring well. Other names for monitoring wells are observation wells, perforated pipes, and open-sided wells.

Unfortunately, because all wells (including pumping wells) are relatively expensive, they often must do double or triple duty. An existing well could be used for water quality sampling and water level monitoring – even if the well was only originally designed for water extraction. Monitoring wells work similarly to piezometers in some situations but differently in others. Figure 94 shows the outside of a piezometer and monitoring well. If we wish to understand how the perforations impact the water levels in the two different pipes, we must visualize the dynamics *inside* of the pipes.

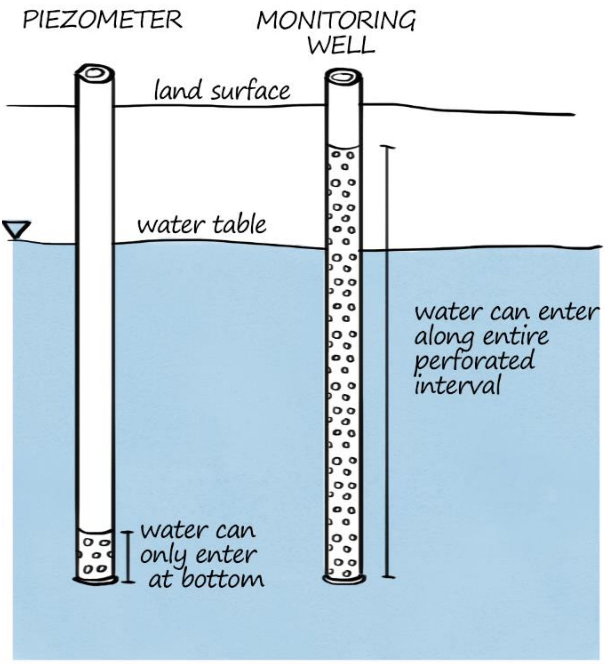

Figure 94. Difference between a piezometer (left) and a monitoring well (right). Notice that piezometers only have perforations at the bottom but monitoring wells have longer perforated intervals. Keep in mind, this figure only shows the outside of the wells and does not show the water level inside of the pipes.

Let's do a thought experiment. First, imagine that you could poke a hole above the intake perforations of a piezometer (Figure 95). What would happen? The answer is... it depends. If there were only horizontal groundwater flow in the system (Figure 95a), the hydraulic head would be the same at every elevation and there would be no gradient between the location of the original perforations and the new hole. Therefore, there would be no flow, and the water level would be the same as before you poked the hole (Figure 95a). But, what about in the case of downward flow (Figure 95b)? The hydraulic head at the new hole would be higher than the hydraulic head at the bottom of the piezometer. Therefore, water would flow into the top hole and out of the bottom of the piezometer (Figure 95b). The water level in the piezometer would be more or less controlled by the head at the hole rather than the head at the bottom. Lastly, what if there were upward flow? Water would flow in at the bottom of the piezometer and out at the hole (Figure 95c). Therefore, the water level in the pipe would likely reflect the conditions somewhere between the conditions at the hole and at the bottom of the piezometer, depending on the hole size.

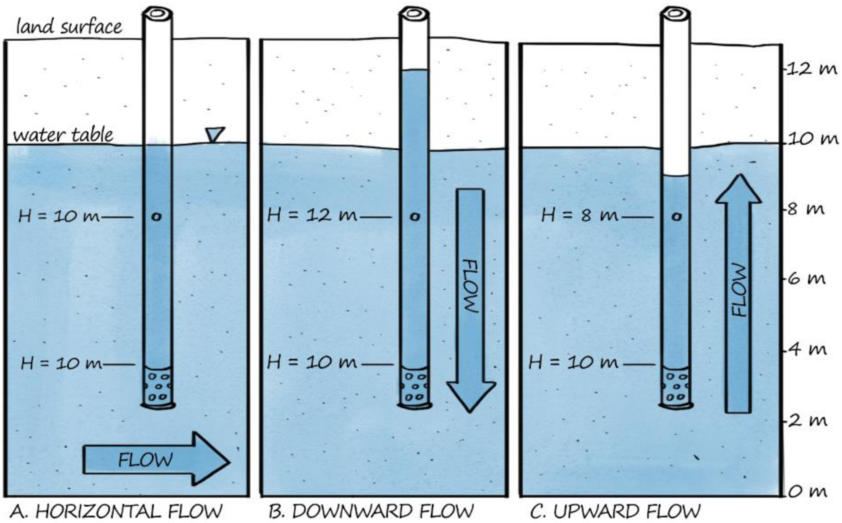

Figure 95. Influence of different groundwater flow patterns on a piezometer with a short, perforated interval at the bottom of the pipe and a small hole drilled in the middle of the pipe for (a) horizontal flow, (b) downward flow and c) upward flow. The flow direction changes in a-c based on the hydraulic head.

Now that you are more comfortable with the complications of poking a hole in the side of our piezometer (Figure 95) – what if you poked 10, or 100, or 1,000 holes? Better yet, what if there was a continuous slit on the side of the piezometer that stretched over 5 meters? Complicated. Well, this is "well territory"; and this is what you will likely run into on a low-budget groundwater monitoring effort. The wells available to you might be pumping wells that were designed to produce water (large-diameter pipes with a long-screened interval). Or you might be given a well or series of wells designed for chemical monitoring. The pipe could have a large-screened interval to detect non-hazardous contaminants at high concentrations (e.g., salinity), or it could consist of multiple smaller wells, each with a single opening to pinpoint the contaminant location. So, what are you to make of all of this? Recognize that *the screening interval matters*. But you often won't be able to do anything about it – you get what you get. Therefore, tuck the screening information away and use it to temper your interpretation of the results. For example, it doesn't make sense to report water levels to the nearest millimeter if you are measuring over a 10 m screened interval. This is a good reminder that the real world is always more complicated than the examples given in this textbook. Additionally, you won't always have access to the proper equipment or all the necessary information. You must make the best of what you are given!

Conclusion

We hope you can see how all the concepts we have introduced throughout the book are working together. In fact, they are working together intricately! We encourage you to skim through the figures in this chapter and observe how much more you understand about groundwater than when you first picked up this book. Soil properties and systems thinking are built into each example. Additionally, hydraulic head gradients and equipotential lines are not only something that you can apply to buckets of soil, but (now) they are part of your visualization process for understanding more complex scenarios. When you started this book, you were learning about soil properties that control pore space. Now, you understand not only how water moves through pores, but you can visualize how external factors (such as pumping wells) impact flow direction, contaminant transport, and stream systems. In this chapter we considered idealized systems to help you to understand how to "see" groundwater. Now, the challenge for the remainder of your career is to imagine what is happening in real-world scenarios.

What to Remember

Important Terms
cone of depression
losing stream
storativity
transmissivity
monitoring well

Chapter 7

Applications in Hydrogeology

Introduction

Members of our communities have an interest in our water resources because we use water for many critical activities, but we also must protect it for its environmental roles. Throughout this book, you have learned fundamental concepts that define this tradeoff. The only way that we can make wise, fair decisions about water resources is for more citizens to become informed – like you are becoming. The key is always to be willing to learn and, at the same time, to become aware of what else you need to learn.

You have spent quite a bit of time using this book and learning more about groundwater. Let's take a minute to reflect on what you have learned and why it might matter. Early on, you gained insight into key properties of porous media. These properties helped you understand how and why water moves. You were introduced to Darcy's law and the mechanisms of groundwater movement, discovering how hydraulic head controls flow. With a basic understanding of flow, you expanded your scope to an agricultural system – visualizing groundwater movement using equipotential and flow lines. Not only did you gain insight into how groundwater moves, but you learned how to map the advective transport of a dissolved solute. Lastly, you learned about pumping wells and their potential to mobilize contaminants and manipulate groundwater flow, such as the ability of pumping to potentially withdraw water from streams!

That's a lot! But importantly, you also know that there is a lot that you still don't know about groundwater systems. Some of these things are advanced mechanisms that require the use of computer models for proper visualization. Other limitations are conceptual complications (such as unsaturated flow) that we didn't cover in detail in this book. And that's fine – if you decide to become a hydrogeologist, you will have plenty of time to learn these things. When it comes to making specific decisions or trying to solve water resources problems, we need trained professionals; this book is your first step towards becoming one. For many careers in environmental science, the level of hydrogeology you now have will put you well above the crowd!

Case Study: Discovering Your Knowledge and Limitations

Questions regarding groundwater availability and contamination are all around us. If you decide to work as an environmental scientist, you will likely be asked a hydrogeologic question at some point in your career. For example, with a basic understanding of hydrogeology, someone might ask you to predict the outcome of a scenario. Will a contamination plume reach my municipal water supply? Does the pumping at my neighbor's house have the potential to change water availability at my house? Where are the dissolved compounds in my drinking water coming from? At what rate can the city pump water from the regional aquifer without negatively impacting local stream systems?

Keep in mind, real-world hydrologic questions are complex, and this book hasn't necessarily given you the skills to always *answer* these questions. However, you do have conceptual knowledge that can further the conversation. For example, you can help others understand a hydrogeologic problem and lead them to the expertise that they need, or you can make an educated guess. Whatever the circumstances, when you are approached with a hydrogeologic question, step back and think about what concepts you understand, and what advice you can provide given your assumptions and knowledge limitations. To practice this skill, let's end the book with a real-world case study that will integrate everything that you have learned.

Contamination in Woburn, MA

Imagine that it is 1985. Anne Anderson and seven other families in Woburn, Massachusetts have filed a lawsuit against three companies: W.R. Grace, Beatrice Foods, and UniFirst Corporation. During the 1960s and 1970s, W.R. Grace operated the Cryovac Plant, which manufactured food-processing equipment, UniFirst operated an industrial dry-cleaning company, and Beatrice Foods owned the John Riley Tannery. The suit alleges that toxic chemicals used at these industrial facilities (during the 1960s and 1970s) entered the groundwater system and traveled to two municipal water wells (G & H), which were in operation between 1964 and 1979. The families claim that the ingestion of the groundwater from these two wells caused severe illness and a cluster of childhood leukemia cases.

You are working as an entry-level environmental scientist in Woburn during this time (1985), for a firm that has been asked to provide expert testimony at the trial. What steps should you take to evaluate whether these three companies are responsible for the contamination? As an entry-level hydrogeologist, you don't necessarily have the expertise to fully defend or contest the accusation made against the companies. However, if your company gave you the responsibility to gather information for the initial investigation, what should you look at?

To begin your investigation, think about what information might be readily available. The two things that come to mind are topographic maps and pumping well records. Figure 96 shows a topographic map of the area with a contour interval of 5 ft. Based on your knowledge of the *potential relationship* between land surface topography and groundwater flow patterns (elevation head helps control hydraulic head, Equation 8), make an initial estimate of the groundwater flow patterns prior to pumping (pre-1964). Remember, topographic gradients are different than groundwater gradients. The water table surface does not necessarily mimic the surface topography. However, you can use the surface topography as an initial guess. Using the topography, how do you think groundwater and surface water interact in this system (Figure 96)? Could the river have also been a source of contamination?

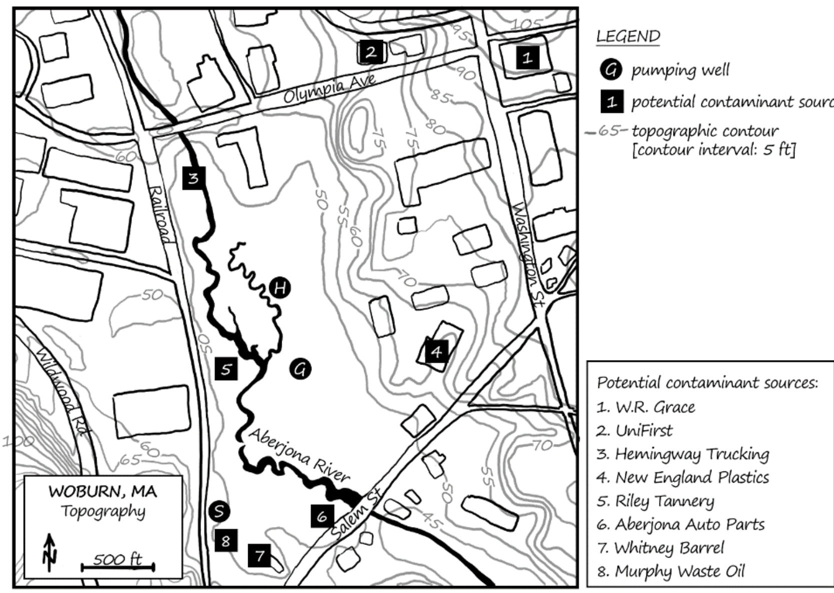

Figure 96. Map showing the topography of Woburn, MA near the potential contaminant sources. Adapted from Bair & Lahm, 2006.

Look at the spatial relationship between the different industrial facilities, the pumping wells, and the river. Are the facilities upgradient or downgradient? These are the questions you can begin asking yourself. From the pumping well records, you also learn that between 1964 and 1979, wells G and H were pumped periodically; together producing on average about 1,100 gallons per minute when in operation.

Next, search for regional groundwater monitoring information. Land surface topography (Figure 96) is a good first guess, but you cannot infer groundwater flow directions from surface topography alone. Are there monitoring data that could help you further construct historic groundwater flow? An equipotential map, or better yet a flow net, could help you estimate groundwater flow patterns and identify the source(s) of contamination. Unfortunately, in the real world these data are hard to come by, and you discover that there is not an extensive collection of historic groundwater monitoring information for the region.

So, what can you do? You need some sort of way to *visualize* solute transport and groundwater movement during the time of contamination, between 1964 and 1979. But all you can do is collect data *now*, in 1985. Take a minute and brainstorm some solutions. There are a few things about the system that have not changed. The subsurface material has stayed the same. Therefore, you could look at geologic maps or conduct lab analyses of the subsurface sediment to determine the soil properties (e.g., porosity, permeability, hydraulic conductivity, etc.). You could also look up the well designs of G and H, which probably have not changed over time: what are their depths, screened intervals, and typical pumping volumes? After rummaging through existing data, you'll still need to quantify groundwater movement before and after pumping. So, what can you do to obtain this information? How can you *reconstruct* pre-1964 and post-1964 flow conditions in the aquifer?

Reconstructing Historic Conditions

To begin your journey of study historic conditions, you decide to install several dozen groundwater monitoring wells at various depths to determine the hydraulic head throughout the study area. Your assumption is that, since the wells stopped pumping in 1979, the water table has returned to its pre-pumping levels. Therefore, you can measure hydraulic head in the present day (1985) to understand the pre-pumping conditions (pre-1964) of the past. You also notice that the municipal wells (G & H) are located on the east side of the Aberjona River, within its floodplain. In addition to deep monitoring wells, you also install three shallow piezometers in the Aberjona River (both upstream and downstream of G & H) to understand the relationship between the groundwater and the river.

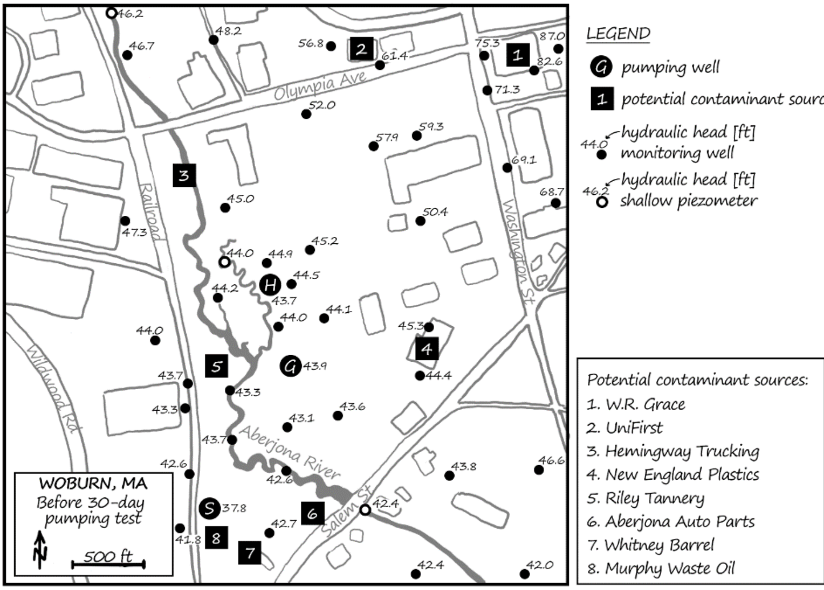

Figure 97. A map of Woburn, MA pre-pumping. The potential contaminant sources are numbered (1-8) and the pumping wells are lettered (S, G, and H). The shallow piezometers are small, white circles and the groundwater monitoring wells are small, dark circles. All the values next to the piezometers and monitoring wells are hydraulic head values (in feet) and are all calibrated to the same datum elevation.

Figure 97 shows the study area with shallow stream piezometers (small white circles) and monitoring wells (small black circles). The shallow piezometer measurements are located along the Aberjona River: at the top of the map, near Salem Street, and in a small tributary in the middle of the map. Major roads are labeled, and wells G and H are shown as large circles. The industries (potential contaminant sources) are shown as numbered squares. The values next to the shallow piezometers and monitoring wells are the hydraulic head values (in feet) and are all referenced to the same datum elevation.

Using the hydraulic head values throughout the study area, which direction is the groundwater flowing? Are the hydraulic heads in the shallow stream piezometers higher or lower than the surrounding monitoring wells? The values in the stream are lower than in the surrounding aquifer (Figure 97), which tells you that the Aberjona River is gaining inflow from groundwater (Figure 89).

The hydrogeology of the Aberjona River valley is complicated; there is a rich diversity of geologic material, and a river and a wetland are both present in the region. This system (Figure 97) is too complicated to do a simple flow calculation using Darcy's law (Equation 6). Therefore, you decide to reconstruct the historic (pre-pumping) groundwater levels using an equipotential map with flow lines. Note: do not include the shallow piezometers in your equipotential map – they are monitoring hydraulic head in the streambed sediments, not the surrounding aquifer. They are simply there to help you understand if the river is gaining or losing.

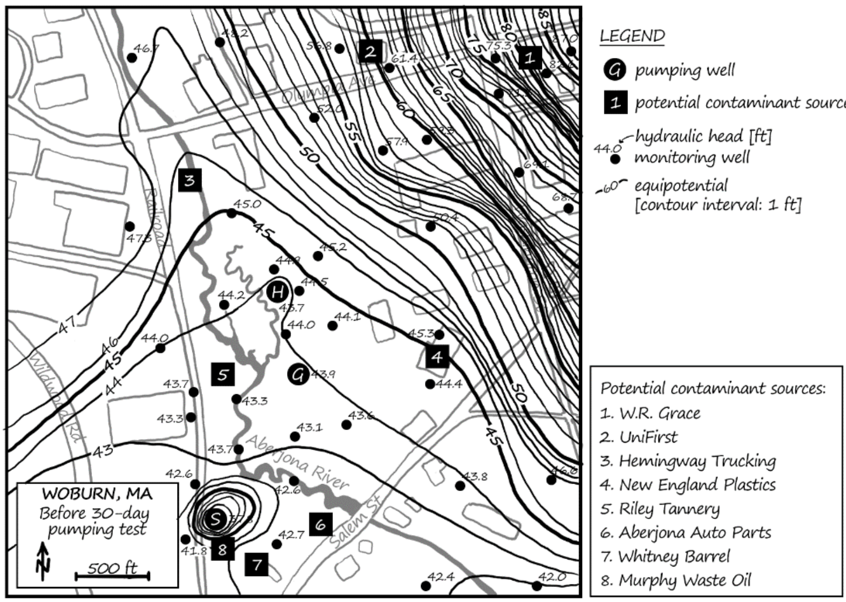

Figure 98. Map of Woburn, MA prior to the 30-day pumping test with equipotential lines (solid lines).

Figure 98 shows the equipotential lines for your system (pre-pumping). Now, remember, flow lines are perpendicular to equipotential lines. Take a moment and sketch some flow lines on Figure 98. Be sure to include flow lines that provide information regarding groundwater flow near the potential contaminant sources. Generally, what direction is the groundwater flowing? Where is the hydraulic head gradient the steepest? Least steep?

Well S is located at the bottom of the map near the scale. What do you notice about the equipotential lines surrounding this well (Figure 98)? They form a circle. This means that there is a cone of depression around Well S prior to pumping (in wells G and H). Well S is a pumping well that is currently (in 1985) in operation and is owned by Riley Tannery. Keep in mind that Well S was not operating during the time of the (historic) contamination. This piece of information will likely be a complicating factor in your reconstruction analysis. Unfortunately, this type of complication happens! The system isn't exactly as it was from 1964 to 1979.

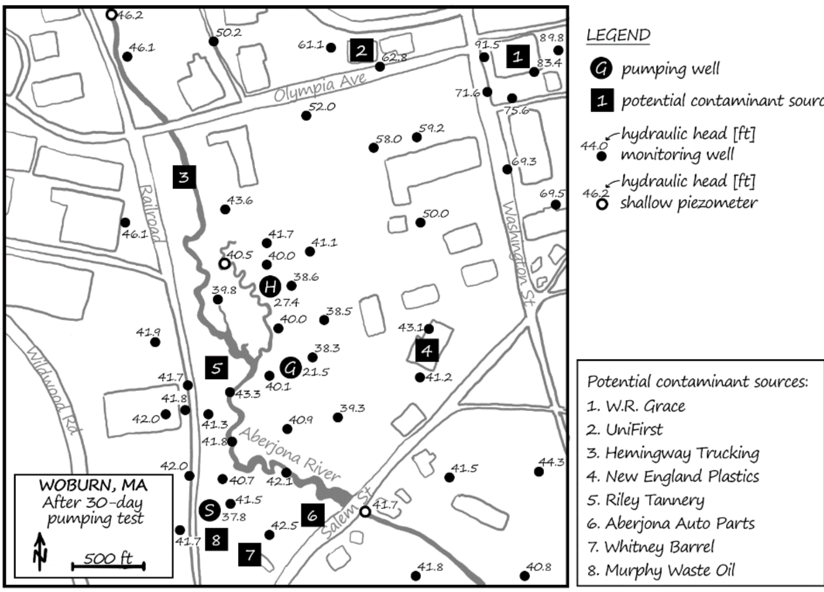

Figure 99. Map of Woburn, MA after pumping wells G and H for 30 days.

After thoroughly analyzing "pre-1964" conditions (Figure 98), you must now reconstruct the pumping conditions that existed from 1964 to 1979. Therefore, you pump wells G and H for 30 days. After 30 days, a new steady state is reached, and wells G and H have approximately the same water levels (now in 1985) as they did during the time of contamination. The new hydraulic head readings (for both the monitoring wells and the shallow stream piezometers) are shown in Figure 99. Take a moment and compare pre-pumping (Figure 97) and post-pumping (Figure 99) hydraulic head values. Look at the values in the shallow stream piezometers. Is the stream still gaining post-pumping? Based on these new measurements, it appears that it is not; the water in the river is flowing into the aquifer (and the stream is losing).

To more broadly understand the changes in hydraulic head due to pumping, draw a new equipotential map for the post-pumping scenario. Compare your equipotential lines pre-pumping (Figure 98) to those from post-pumping (Figure 100). Where is the hydraulic head gradient the steepest? What impact does the change in the hydraulic gradient have on the rate of groundwater flowing across the site? Notice, there is still a cone of depression around the Riley Well (Well S) (as there was before pumping (Figure 98)), but the surrounding water table has changed.

Focus on the location of the three industries involved in the lawsuit: W.R. Grace, UniFirst, and Riley Tannery. Does the possibility exist for groundwater to flow from each of these industrial sites to wells G or H under pre-pumping conditions (Figure 98)? What about during pumping conditions (Figure 100)? Take a moment and draw flow lines extending out from 1 (W.R. Grace), 2 (UniFirst), and 5 (Riley Tannery) for both conditions. Compare your flow lines to the potential contaminant paths shown in Figure 101 and Figure 102.

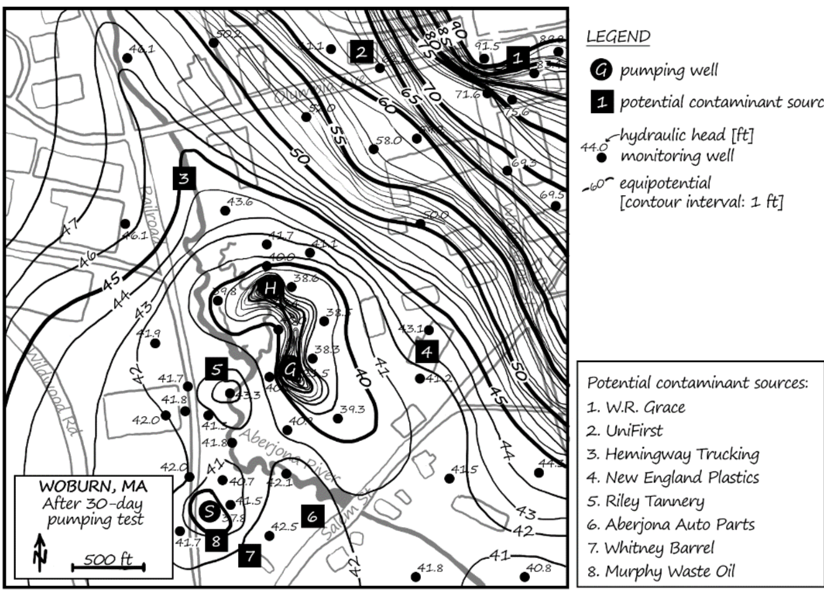

Figure 100. Map of Woburn, MA with post-pumping equipotential lines.

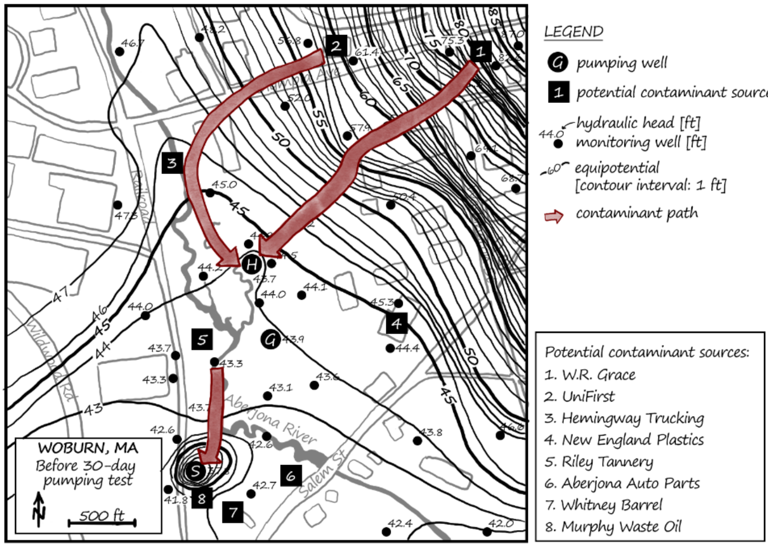

Figure 101. Map of Woburn, MA during pre-pumping conditions. The solid lines are equipotential lines, and the red lines show the potential contaminant path based on flow lines.

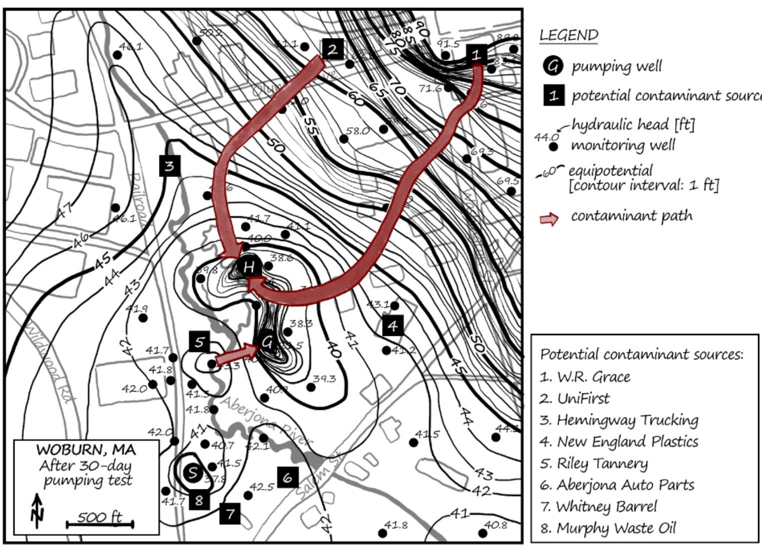

Figure 102. Map of Woburn, MA during post-pumping conditions. The solid lines are equipotential lines, and the red lines show the potential contaminant path based on flow lines.

Using your flow lines (and Figures 101 and 102) decide which industries could have potentially contaminated wells G and H. Then, estimate the travel time from these locations to wells G and H. The aquifer material has an average hydraulic conductivity of 250 ft/day and an average porosity of 0.25. To calculate the travel times from the potential sources to wells G and H, use the following steps:

1. Measure the length of the flow line from (the source to the wells) to the nearest 25 feet.
2. Measure the change in head along the flow line to the nearest tenth of a foot.
3. Calculate the average hydraulic head gradient along the flow line.
4. Calculate the average linear groundwater velocity (Equation 10).
5. Calculate the travel time using the displacement equation (Equation 11).

Based on your rough calculations, which company/companies would you recommend be held responsible for the contamination? Are there other industries that were not considered that could potentially also be responsible (e.g., Hemingway Trucking Co.)? Do you have enough information to say this with confidence? We will not give you the solution – at this point, you should have enough knowledge to work through the problem on your own.

In the actual trial, the 6-person jury determined one company (W.R. Grace and Co.) liable for the contamination. The jury did the best they could with the information available in the 1980s. However, groundwater and solute transport modeling is a rapidly evolving field, and new tools have been and continue to be developed, making it easier to rapidly explore a broader range of possibilities. If you want to learn more about this landmark trial, we recommend reading Jonathan Harr's gripping account titled "A Civil Action" (Harr, 1996).

Since the trial, organizations such as Ohio State University have undertaken more in-depth studies, simulating the movement of the contamination plume in Woburn, MA, using newly acquired information from contaminant cleanup efforts (Ohio State University, n.d.). Ohio State considered the three properties brought to trial as well as two additional properties they identified as potential sources. The results from Ohio State's model (Ohio State University, n.d.) cannot determine with full certainty how the Woburn wells became contaminated, but they do provide some plausible scenarios. Their studies suggest that chemicals from W.R. Grace probably never reached the wells, and if they did, they arrived near the time the wells were shut down. They recommend that the most plausible scenario is that much of the contamination came from Beatrice Foods (who owned the Riley Tannery) and Hemingway Trucking Co., and that a substantial amount of the groundwater came from the Aberjona River (another potential source of contamination).

One strength of computer modeling over creating a hand-drawn equipotential map with flow lines (Figure 101 and Figure 102) is that computer models can consider a wider range of factors. They can consider the influence of the chemical composition of the plume. Different chemical compositions will vary in their movements. Therefore, computer models can simulate various industrial solvents (e.g., TCE and PCE) with different densities and therefore different flow properties. Models can also consider specific processes that will affect solute transport, such as solvent binding onto organic matter in the aquifer. Or they can predict solute concentrations while considering variations in precipitation, unsaturated conditions, and variable pumping rates of wells G and H. Even with computer modeling, no one can be 100 percent sure what happened from 1964-1979. Groundwater concepts and advanced modeling can only help clarify the range of possibilities.

Conclusion

For many of you, this chapter was likely the first time that you have seen anything like a real-world hydrogeologic analysis. With the information and guidance presented here, we hope you can better understand how complicated real-world groundwater systems can be! Although, keep in mind, you don't need to know how to complete this type of analysis on your own – the reality is, many people (even very experienced people) spend a lifetime developing those skills.

The next time that a concerned citizen approaches you regarding a hydrogeologic question, you will have some information you can share with them. Remember, whatever conclusion you draw, you will need to make simplifying assumptions about the system. As mentioned in the Woburn modeling case above, even the most advanced hydrogeologists make assumptions. The reality is that we can't see *exactly* what is happening underground — we can only continue to improve our ability to visualize it.

Chapter 8

Summary

> "The existence, origin, movement, and course of [underground] waters, and the causes which govern and direct their movements, are so secret, occult, and concealed that an attempt to administer any set of legal rules in respect to them would be involved in hopeless uncertainty, and would, therefore, be practically impossible."

Frazier v. Brown, Ohio 1861

The quote above is an excerpt from an Ohio Supreme Court trial (*Frazier v. Brown*, 1861), concluding that the movement of groundwater was too complex to be regulated. The consequence of this trial was a principle known as the "rule of the biggest pump" – an approach to groundwater law that allowed landowners to withdraw unlimited amounts of water, regardless of the impact on neighboring wells or ecosystems. The rule of the biggest pump was based on the premise that it was impossible to prove that any harm was directly caused by pumping.

Since 1861, our understanding of hydrogeology has advanced dramatically, and numerous legal cases in state and federal courts have reshaped groundwater law into a complicated tangle. What do you think? Is groundwater still too "secret, occult, and concealed" to be regulated? Under what circumstances might a hydrogeologist be able to identify the impacts of groundwater pumping on a neighboring well? On rivers? On contaminant plumes? Under what circumstances is the uncertainty too high to make a decision? Our hope is that after reading this book, you can begin to answer these questions and shed light on the mystery of underground waters.

Glossary

advection: transport of a solute by the water it is dissolved in

angularity: the sharpness of the edges of a grain

aquifers: underground layers where water is stored within rock fractures or unconsolidated materials and can flow relatively easily (e.g., sand, gravel)

average linear groundwater velocity: the rate of movement through pore space alone in the direction of flow

boundaries: the limits of a system; the divide between a system and everything else

capillary zone (fringe): the transition zone between the saturated and unsaturated zone where the pores are completely filled (fully saturated) with water by capillarity and under negative water pressure

closed system: a system that allows the transfer of energy, but doesn't allow the transfer of mass

compaction: the degree to which a material decreases in volume when it is subjected to a load

cone of depression: spatial distribution of the collective change in the water table due to pumping

confined aquifer: an aquifer that is bound by confining layers that have a lower permeability (e.g., silt or clay)

continuous flow: solute transport where the water entering a system always contains solute during the observation time

contour interval: the difference in potential energy between equipotential lines

decay: a range of processes that can cause contaminants to break down in the subsurface, including reacting chemically with soil solids or being consumed by microbes

diffusion: the mixing of mass by random motion which causes dissolved solutes to move from areas of high concentration to low concentration

dispersion: the spreading of a contaminant due to variations in velocity

dispersivity: a property of the medium that relates the pore velocity to the dispersion coefficient to determine the amount of dispersion that will occur

displacement: the linear distance something travels in the direction of flow

dry bulk density: the dry weight of a soil divided by its volume

effective porosity: the interconnected pore volume that excludes isolated pores

elevation: the height above or below a given datum elevation

elevation head: the potential energy arising from elevation per unit weight

equipotential lines: lines of equal energy potential

evaporation: the process of converting liquid water into vapor due to an increase in temperature and/or a decrease in pressure.

evapotranspiration: the summation of evaporation and transpiration

final condition: a description of a system at the end of the observation time

flow lines: lines that help you visualize the direction of groundwater flow that are perpendicular to equipotential lines

flow net: a visualization tool that quantifies 2D flow through a steady state system

flow tube: the area between adjacent flow lines

fluid pressure: the pressure at any point and time within a fluid due to all forces acting on it

flux: the rate at which water is moving through a cross section perpendicular to the direction of flow

gaining stream: a reach of a stream that is gaining water from the ground based on the hydraulic head gradient

grain size: the size of a single grain in reference to its smallest side

head: total potential energy per unit weight

heterogeneous medium: a medium that is not the same throughout (e.g., a layered system).

homogeneous medium: a medium that is the same throughout; at all locations, the medium has the same properties

hydraulic conductivity (K): the ability of a medium to transmit water in response to a head gradient

hydraulic head: the combination of pressure head and elevation head

hydraulic head gradient: the change in hydraulic head over a length.

hydrodynamic dispersion coefficient: combination of the diffusion and mechanical dispersion coefficients

hydrologic cycle: the continuous movement of water in the Earth-atmosphere system

impermeable: a material that doesn’t allow liquid (or gases) to flow through it

infiltration: when water seeps into the subsurface from above the ground surface

initial condition: the description of a system immediately before the beginning of the observation time; state variables at time zero

inputs: anything that crosses the system boundaries from the outside to the inside

isolated system: a system where neither energy nor mass is transferred across the boundaries

losing stream: a reach of a stream that is flowing from a stream into the ground due to the hydraulic head gradient

manometer: an open tube that tells you about the pressure head in a system at the specific point where it is installed; the pressure head at the bottom of the manometer is equal to the height that the water has risen in the manometer

mass balance: an application of the conservation of mass that accounts for inputs and outputs to a system and the changes in mass stores with time

mass flux: the mass of the solute flowing through a region per unit time per unit cross sectional area perpendicular to the direction of mass transport

mechanical dispersion coefficient: a parameter that describes the effects of all the small-scale processes that occur during dispersion (e.g., variations in flow velocity or variations in flow paths)

monitoring well: pipes that allow water to rise and fall with perforations that extend over most of the length of the piping; the water level inside the pipe reflects the total water pressure head integrated over the perforated length

nonpoint source: contamination over a widely distributed area

observation time: the time period during which a system is observed

open system: a system that allows mass and/or energy to transfer across boundaries

outputs: anything that crosses from the inside to the outside of system boundaries

particle density: the density of the solid particles that make up a sample

percolation: when water moves through cracks and pores underground

permeability: the ability of a medium to transmit a fluid

permeable: a property of a material that allows liquids (or gases) to flow through it

piezometer: a field tool hydrologists use to measure hydraulic head; piping placed in the subsurface that are open to the atmosphere at the top, sealed along its length, and open to groundwater flow at the bottom.

plug flow: solute transport by advection alone

point source: a discrete contamination location

pores: the spaces between grains where there is void space

porosity: the ratio between the volume of the pore space relative to the total volume

porous medium: a material containing solid grains and void space (pores)

potential energy: the energy that is stored in an object

potential gradient: the difference in potential energy over a distance between two points

potentiometric surface: the height to which the water would rise if it could get through the overlying confining layer in a confined aquifer

precipitation: the water that condenses and falls from clouds due to gravity

pressure head: the height to which water rises in a column due to the pressure exerted at a certain point (point P); the weight of the water in the column is pushing down as hard as the weight of the water above point P is pushing up, so the height in the column is how far from the energy at point P can move water upwards against the force of gravity

pulse flow: the release of a solute that happens over a brief time period during the observation time

recharge: when percolating water in the subsurface reaches the water table

residual water content: the minimum water content that can be achieved by drainage

retardation coefficient: a property that affects the velocity, dispersion, and diffusion of a solute, slowing down travel times compared to the water; retardation factors are determined by the chemistry of the soil and the solute.

saturated water content: the maximum amount of water a porous medium can store in its pore space; the maximum water content of a porous medium

saturated zone: the region of the subsurface where are the pores are completely filled with water

sieves: assorted sizes of screened mesh that allow various sized particles to pass through; used to determine sorting and grain size in a sample

soil texture triangle: a method for determining the soil texture where you use relative percentages of different grain sizes (sand, silt, and clay)

soil texture: a name for a soil based on the grain size and sorting

soil: the upper layer of the earth that is a mixture of organic remains, clay, and rock particles.

solute: any substance that is dissolved in another substance

sorption: when a solute (temporarily) sticks to solid particles as water flows past

sorting: the distribution of grain sizes

stage: the height of the water surface in a stream above a datum elevation

stagnation point: where the local velocity of water is zero

state variables: variables (e.g., pressure, temperature, volume, mass) used to define the system state or the behavior of the system (e.g., gaining, losing, steady state, static)

static: a system state where absolutely nothing about a system is changing through time or in space

steady state: a system state where internal processes are occurring at the same rate at every point in the system through time

stream gage: an instrument that records the amount of water in a stream or river and its stage

storativity: a measure of how much water is produced per unit drop in head due to pumping

system: a group of interdependent items that form a unified whole; changes in one part of the system impact other components in the system

system state: the condition of the entire system. (e.g., static, steady state, gaining, losing, transient)

total bulk density: the total mass (of the solids, air, and water) in a porous medium divided by the total volume of the porous medium.

transient: a system state where internal processes are occurring, and their rate is changing anywhere in the system through time

transmissivity: a property of a porous medium that controls how much of a gradient is required to cause a certain amount of water to flow towards a pumping well; it is dependent on the average horizontal hydraulic conductivity of the aquifer and the aquifer thickness.

transpiration: the exhalation of water vapor through plant's leaves

unconfined aquifer: an aquifer that is not bound by impermeable units and the upper surface of the saturated zone is at the same elevation as the water table

unsaturated zone: a region of the subsurface where water content is less than porosity

water content: how much water is stored in a sample as a ratio between the volume of water and the total volume.

water saturation: the fraction of pore space that is filled with water

water storage: the water that is held in the void space between particles

watershed: a geographical area where all the streams and rainfall drain to a common outlet point (outflow of reservoir, spring, etc.); everything upstream drains downstream.

zone of effective saturation: a region where water saturation is almost complete except for entrapped air

zone of residual saturation: a region that is as dry as it can be through drainage

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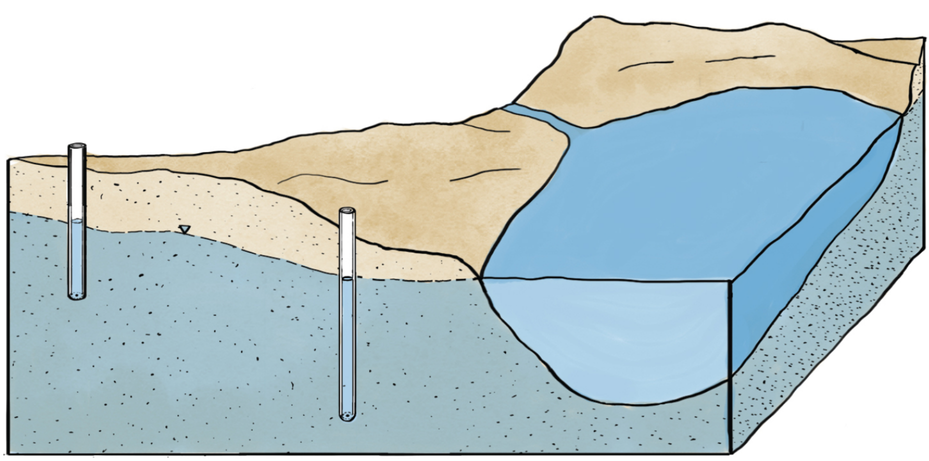
The Basics of Groundwater is an approachable and engaging introduction. The book is modeled after traditional groundwater texts but illustrates fundamental concepts using analogies and examples while eliminating much of the higher-order math. Each chapter includes original scientific illustrations and in-depth examples that give the reader opportunities to perform thought experiments and to practice learned knowledge through worked problems. This is a book that students will enjoy reading and that instructors will enjoy teaching from.



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