Chapter 1
Ground water

If a hole is dug, only the water that flows freely into the hole is groundwater. Since the air in the hole is at atmospheric pressure, the pressure in the groundwater must be higher if it is to flow freely into the hole. This is what distinguishes groundwater from the rest of the underground water.

Definition 1.1. Water, below the surface, at a pressure greater than atmospheric pressure, which thus flows freely into a hole through interconnected void spaces, is *groundwater*.

Groundwater and soil water together make up approximately 0.5% of all water in the hydrosphere. Figure 1.1 illustrates the various zones of water found beneath the surface. Water beneath the surface can essentially be divided into three zones: 1) the soil water zone, or *vadose zone*, 2) an intermediate zone, or *capillary fringe*, and 3) the ground water, or *saturated zone*. The top two zones, the vadose zone and capillary fringe, can be grouped into the zone of aeration, where during the year air occupies the pore spaces between earth materials. Sometimes, especially during times of high rainfall, those pore spaces are filled with water.

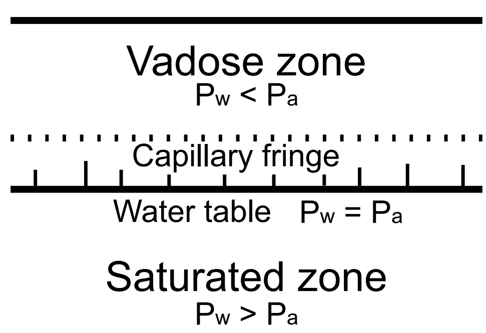

Fig. 1.1 Soil moisture and distribution of fluid pressure P_w relative to atmospheric P_a in the ground.

Beneath the zone of aeration lies the zone of saturation, or the groundwater zone. Here water constantly occupies all pore spaces. The *water table* divides the zone of aeration from the zone of saturation. The elevation of the water table is determined to be where the pore water pressure, P_w, is equal to atmospheric pressure, P_a. The height of the water table will fluctuate with precipitation, increasing in elevation during wet periods and decreasing during dry. In general, the water table has an undulating surface which generally follows the surface topography, but with smaller relief.

1.1 Porosity

Ground water and soil moisture occur in the cracks, voids, and pore spaces of the otherwise solid earth materials.

Porosity, n, is the percentage of the total volume that is void of material,

$$n = 100 \times \frac{V_v}{V}, \tag{1.1}$$

where V_v is the volume of void space, and V the total volume. We can write V_v as $V_v = V - V_s$, where the subscript s refers to the volume of the solid phase, $V_s = m_d/\rho_s$, the dry weight divided by the density of the soil/rock.

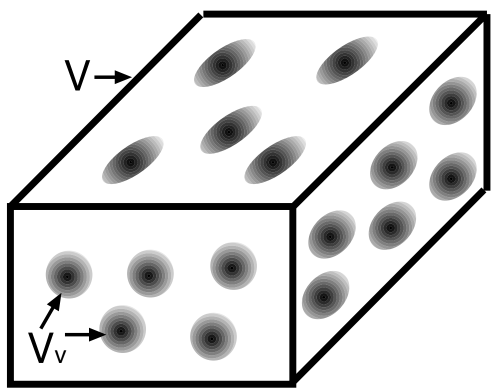

Fig. 1.2 Porosity $n = 100 \cdot V_v/V$, where V is the total volume and V_v is the volume of void space.

Porosity can be measured by drying the sample completely, at 105°C, and then submerging in a known volume of water. What goes into the rock is a measure of the *effective* void space, n_e.

Well-rounded, and sorted, sediments have porosity between 26 - 48 %, depending on the packing, but independent of particle size[1]. Figure 1.3 shows cubic (porosity 47.65 %) and rhombohedral (porosity 25.95 %) packing of spheres.

If there is a mixture of grain sizes, the porosity will be lower, since smaller particles tend to fill in the void spaces between larger ones.

[1] Strickly true only for all particles of the same size and spherical

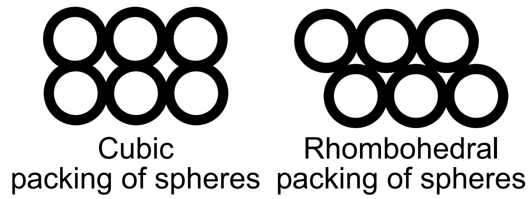

Fig. 1.3 Cubic (porosity 47.65 %) and rhombohedral (porosity 25.95 %) packing of spheres.

Porosity of sedimentary rocks is very variable, from 1% to 30%, see Table 1.1 [2]. Plutonic rocks have porosity of 1.5% if un-fractured, fracturing increases the porosity (by 2 - 5%) and weathering can also increase the porosity (+30 - 60%).

Table 1.1 Porosity of sediments

Material	**n (%)**
Well-sorted sand and gravel	25 - 50
Sand and gravel, mixed	20 - 35
Glacial till	10 - 20
Silt	35 - 50
Clay	33 - 60

For *grain size* there are various names given to specific size ranges depending on whether looking in the engineering, geologic, oil, and so on, literature, but they are broadly similar. The *uniformity coefficient* $C_u = d_{60}/d_{10}$, where d_{60} and d_{10} are the grain sizes where the total mass of finer grains is 60% and 10%, respectively, is a measure of how well sorted the sediments are. If $C_u < 4$ the sample is well sorted (most of the grains of similar size), and if $C_u > 6$ it is poorly sorted.

Figure 1.4 shows the grain size distribution of the till beneath Ice Stream B in Antarctica [5], which is poorly sorted.

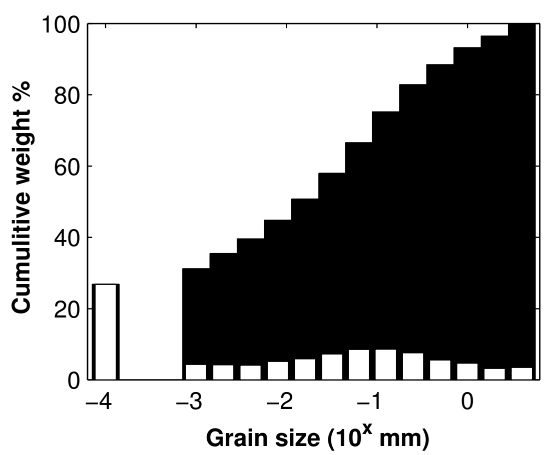

Fig. 1.4 Particle diameter (mm) as a function of weight % for the till beneath Ice Stream B, Antarctica. Grains larger than 4 mm have a weight of 3.55%, and smaller than 10^{-3} mm have a weight of 26.8%.

Void ratio, e, is the ratio between the volume of voids and the volume of solids,

$$e = V_v/V_s. \tag{1.2}$$

Porosity and void ratio are related, since the total volume is obviously $V = V_v + V_s$,

$$n = \frac{e}{1+e}, \quad e = \frac{n}{1-n}. \tag{1.3}$$

Finally *bulk density*, $\rho_b = (1-n) \times \rho_d$, where ρ_d is the particle density of the aquifer material.

1.2 Conductivity and permeability

Some rocks are porous, but the voids are not, or poorly, interconnected. These rocks cannot convey water from one void to another. In the mid-1800s, Henry Darcy, a French engineer, made the first systematic study of the movement of water through a porous medium [2].

Darcy's law is illustrated in Figure 1.5, and states that,

$$q = \frac{Q}{A} = -K\frac{dh}{dl}, \tag{1.4}$$

where q [L T^{-1}] is specific discharge (sometimes called Darcian velocity), A [L^2] is cross-sectional area, Q [L^3 T^{-1}] is discharge, dh/dl is the hydraulic gradient, and K *hydraulic conductivity* [L T^{-1}].

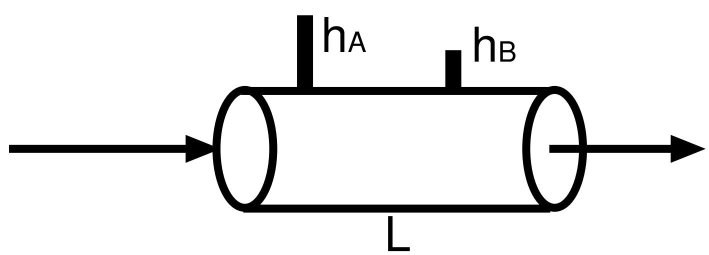

Fig. 1.5 A horizontal pipe filled with sand to demonstrate Darcy's experiment. Water is applied under pressure at point A and discharges at point B.

If q is interpreted as velocity it is important to keep in mind that it is the average velocity through an area, A.

The hydraulic conductivity K is also referred to as the *coefficient of permeability*.

The discharge will depend on the properties of the fluid, as well as the properties of the porous medium. The hydraulic conductivity, K, therefore is a function of both fluid and particle properties. Darcy's law can be written,

$$q = -C \times d^2 \times \frac{\gamma}{\mu} \times \frac{dh}{dl}, \tag{1.5}$$

where, $\gamma = \rho g$ is the specific weight of the fluid, μ is the dynamic viscosity of the fluid, C is called shape factor, and d is the diameter (of equal size beads).

Intrinsic permeability depends only on the properties of the porous medium,

$$K_i = C \times d^2. \tag{1.6}$$

So, hydraulic conductivity and intrinsic permeability are related by, $K = K_i \times \gamma/\mu$. Intrinsic permeability, K_i, is sometimes measured in darcy, where 1 darcy = 9.87×10^{-9} cm^2 $\simeq 10^{-8}$ cm^2.

Estimating water content

Specific yield, S_y, is the ratio of the volume of water that drains from a rock, owing to gravity only, to the total volume.

Specific retention, S_r, is the ratio of the volume of water a rock can retain against gravity drainage, to the total volume of the rock.

If two rocks have equal porosity, but different grain sizes, more moisture will be left in fine-grained rocks, due to surface tension.

The sum of specific yield and retention is equal to the porosity, $n = S_y + S_r$.

The *specific storage*, S_s, is the amount of water per unit volume of a saturated formation that is stored or expelled from storage owing to compressibility of the mineral skeleton and the pore water per unit change in head. The specific storage is given by,

$$S_s = \rho_w \times g \times (\alpha + n\beta), \tag{1.7}$$

where ρ_w is the density of water, g is the acceleration due to gravity, n is porosity, α is the compressibility of the mineral skeleton [L T^2 M^{-1}], and β is the compressibility of water [L T^2 M^{-1}] ($\beta = 4.4 \times 10^{-10}$ m^2 N^{-1} at 15.5°C).

The *storativity* of a confined aquifer is the product of specific storage, S_s, and the aquifer thickness, b,

$$S = b \times S_s. \tag{1.8}$$

1.3 Capillary forces

The attraction that molecules in liquid have for each other results in surface tension. One of the effects of surface tension is the tendency to reduce surface area. The

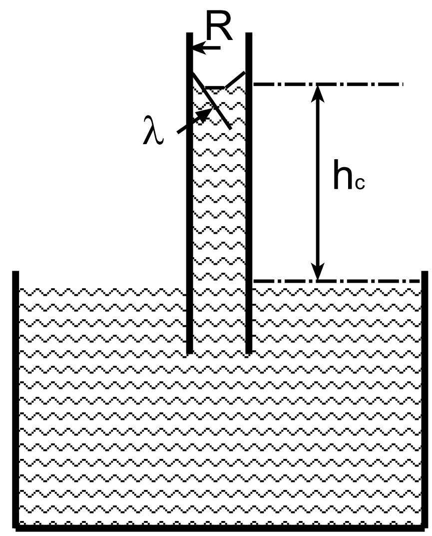

Fig. 1.6 Rise of fluid column due to capillary force.

geometrical shape that has the smallest surface area for a given volume is a sphere [3].

Because the attraction of, for example, glass molecules for the water molecules is greater than that of water molecules for each other, the water is pulled up along the walls of a tube.

Figure 1.6 shows the rise of a fluid in a tube due to capillary forces. The pressure on the concave side of the interface is greater than that on the convex side (in the liquid near surface). The height of the water column must be such that the pressure at a point inside the tube, near the surface of big container is equal to the pressure at the surface outside the tube. The rise of a fluid in a capillary tube is given by,

$$h_c = \frac{2\gamma}{\rho g}\frac{\cos\lambda}{R}, \tag{1.9}$$

where,

λ is the contact angle,
γ is the surface tension of the fluid [M T^{-2}],
R is the radius of the tube.

For water at 18°C, $\rho \sim 1000$ kg m^{-3}, $\gamma = 0.073$ kg s^{-2}, and we can take $\lambda \sim 0°$, which gives (where R is in mm), $h_c = \frac{15}{R}$ mm. Figure 1.7 shows the height to which the fluid rises according to Eq. 1.9, using the values of ρ, γ for water at 18°C. The contact angle varies with the chemical composition of the liquid and wall, as well as with impurities on the wall, but approaches zero for pure water on clean glass ($\lambda \to 0$, $\cos\lambda \sim 1$).

The rise thus depends on the type of sand (material) and packing. An approximate relationship was found [4],

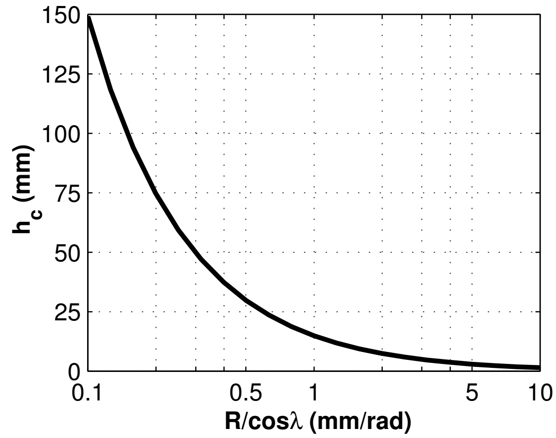

Fig. 1.7 The rise of fluid column due to capillary force, see Eq. 1.9.

$$h_c = \frac{2.2}{d_H}\left(\frac{1-n}{n}\right)^{2/3}, \tag{1.10}$$

where d_H is the harmonic mean grain diameter in millimeters and n is porosity. Figure 1.8 shows the rise of a water column, h_c, as a function of grain diameter and porosity.

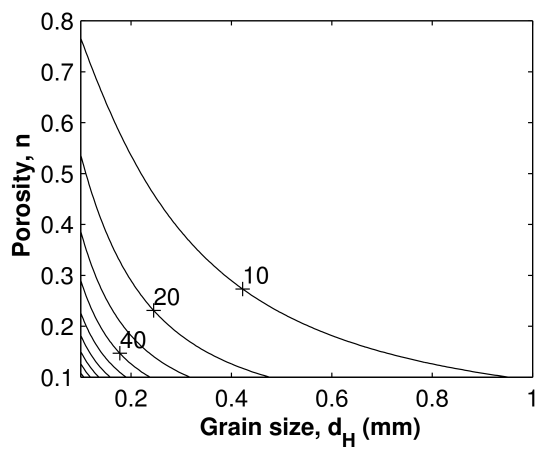

Fig. 1.8 The rise of a fluid column, h_c (mm), due to capillary force (Eq. 1.10) for different grain sizes d_H and porosity n.

Capillary fringe (Fig. 1.1) can saturate the soil above the water table, but the fluid pressure will be negative w.r.t. local atmospheric pressure, indicating that the capillary fringe is a part of the vadose zone.

1.4 Soil moisture

The zone between the ground surface and down to the water table is called the *vadose zone* (unsaturated zone). Water there is held to the soil particles by capillary forces. The vadose zone (see Figure 1.1) may contain a three-phase system:

- *Solid*. Mineral grains, and organic material.
- *Liquid*. Water with dissolved solutes.
- *Gaseous*. Water vapor, and other gases.

We can write the equation for porosity as,

$$n = 100 \times (1 - \rho_b/\rho_m), \tag{1.11}$$

where $\rho_b = m_s/V$ is the dry bulk density, subscript s for solid, and $\rho_m = m_s/V_s$ is the particle density.

The water that is immediately available to plants is a part of the *soil water*. Soil water can be sub-divided into three categories: 1) hygroscopic water, 2) capillary water, and 3) gravitational water.

Hygroscopic water is found as a microscopic film of water surrounding soil particles. This water is tightly bound to a soil particle by molecular forces so powerful that it cannot be removed by natural forces. Hygroscopic water is bound to soil particles by adhesive forces that exceed 31 bars and may be as great as 10 000 bars (recall that sea level pressure is equal to 1013.2 millibars which is just about 1 bar).

Capillary water is held by cohesive forces between the films of hygroscopic water. The binding pressure for capillary water is much less than hygroscopic water. This water can be removed by air drying, or by plant absorption, but cannot be removed by gravity. Plants extract this water through their roots until the soil capillary force (force holding water to the particle) is equal to the extractive force of the plant root. At this point the plant cannot pull water from the plant-rooting zone and it wilts, called the *wilting point*.

Definition 1.2. At the *wilting point*, soil moisture is too tightly bound to the soil particles for plant roots to be able to withdraw it.

Gravity water is the water moved through the soil by the force of gravity. The amount of water held in the soil at the point where gravity drainage ceases is called the field capacity of the soil. The amount of water in the soil is controlled by the soil texture (Figure 1.9).

1.5 Aquifers

Definition 1.3. Aquifers are groundwater-bearing formations that are sufficiently permeable to transmit and yield water in usable quantities.

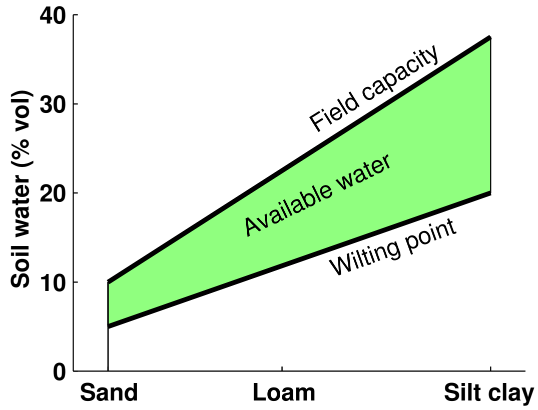

Fig. 1.9 Field capacity and wilting point for different types of soil.

Atmospheric precipitation is the main source of fresh groundwater. The water may have infiltrated directly into the ground where it landed, or been collected via surface runoff and then seeped into the ground.

An *aquifer* is a geologic unit that can store and transmit water at rates fast enough to supply reasonable amounts to wells. This generally means that the intrinsic permeability is greater than 10^{-10} cm^2 (10^{-2} darcy); the most common aquifers are unconsolidated sands and gravels.

Aquifers are divided into two types: (1) unconfined and (2) confined. Figure 1.10 shows a schematic of confined and unconfined aquifer.

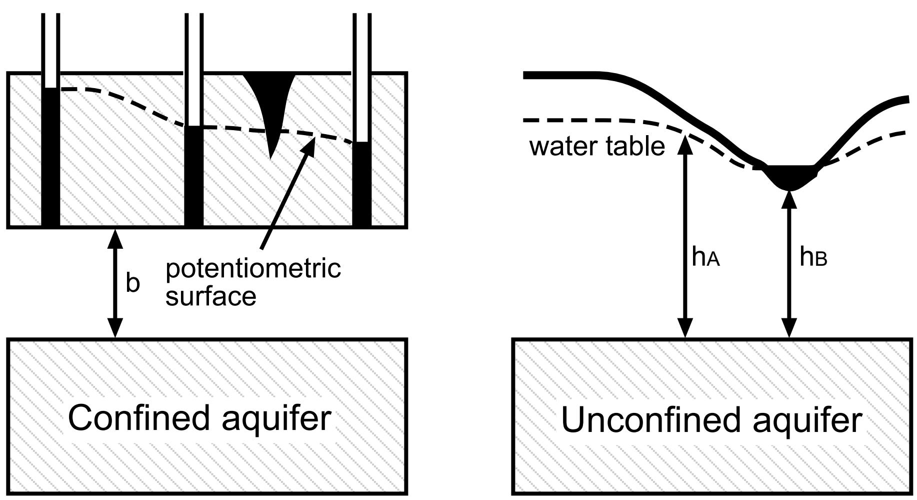

Fig. 1.10 Confined and unconfined aquifers.

Unconfined aquifers are underground lakes in porous materials. The top of the unconfined aquifer is the *water table* (also called *phreatic surface*, after the Greek word phrear, "well"), which is the plane where the groundwater pressure is equal to the atmospheric pressure [1]. Figure 1.11 shows water flowing from the lower part of an mountain, indicating that the water table is at that elevation. Unconfined aquifers (sometimes called water-table aquifers) have continuous layers of high intrinsic permeability extending from the land surface to the base of the aquifer.

Fig. 1.11 Water flowing from the middle of the mountain.

A confined aquifer is a layer of water-bearing material that is sandwiched between two layers of much less pervious material. The pressure condition in a confined aquifer is characterized by the *potentiometric* surface, which is the surface obtained by connecting equilibrium water levels in tubes, or piezometers, penetrating the confined aquifer.

If the potentiometric surface is above the upper confining layer, the static water level in a well will be above the aquifer. Such a well is called an *artesian* well, named after wells bored in Artois (N-France) in the eighteenth century [1].

A *confining layer* has an intrinsic permeability lower than about 10^{-10} cm^2 (10^{-2} darcy).

Aquifer transmissivity, is a measure of the amount of water that can be transmitted horizontally through a unit width by the fully saturated thickness of the aquifer under a unit hydraulic gradient. It is,

$$T = b \times K, \tag{1.12}$$

where

T is transmissivity [$L^2 T^{-1}$],
b is saturated thickness of the aquifer [L],
K is hydraulic conductivity [$L T^{-1}$].

1.6 Groundwater flow

Potentiometric height is the height to which water will rise in tightly cased wells. For unconfined aquifers, the potentiometric height (or surface) is equal to the hydraulic head (Fig. 1.10). The potentiometric height is given by,

$$\phi = \phi_0 + z + \frac{P_w}{\rho_w g}, \tag{1.13}$$

where P_w is the water pressure, z is the elevation above datum, ϕ_0 is a constant, ρ_w is the density of water, and g the acceleration due to gravity.

Darcy's law

If we use ϕ for the potentiometric height (or hydraulic head for an unconfined aquifer), we have,

$$q = -K\frac{d\phi}{dl}, \tag{1.14}$$

where $q = Q/A$ is the flux divided by the area. This is a one-dimensional version of Darcy's law.

The mean velocity through a cross-sectional area is $v = Q/A$, sometimes called Darcy velocity. Note that this is not the velocity of the water itself. That is given by the seepage velocity, or average linear velocity, which is the *average velocity of the water itself*,

$$v_x = \frac{Q}{n_e \times A}, \tag{1.15}$$

where n_e is the effective porosity.

Applicability of Darcy's law

Slowly moving fluids are dominated by viscous forces. In *laminar flow*, molecules of water follow smooth lines, called *streamlines*.

As the velocity increases, the moving fluid gains kinetic energy, and when the inertial forces are more influential than the viscous forces, the fluid particles begin to rush past each other in an erratic fashion, resulting in *turbulent flow*, see Figure 1.12.

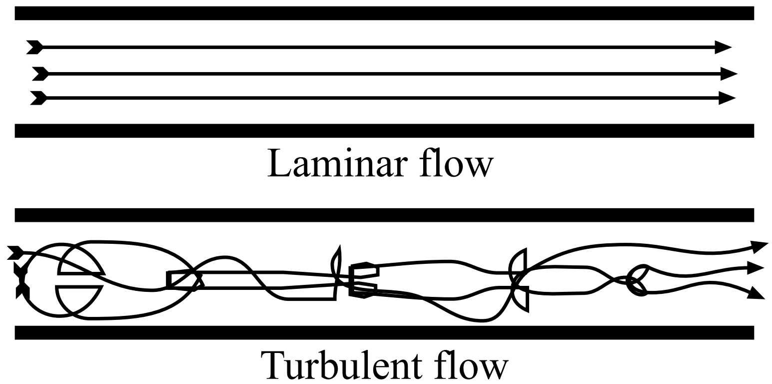

Fig. 1.12 Laminar and turbulent flow.

The *Reynolds number*, *Re*, is used to determine whether the flow will be laminar or turbulent,

$$Re = \frac{\rho \times q \times d}{\mu}, \tag{1.16}$$

where,

ρ is the fluid density [M L^{-3}],
q is the discharge velocity [L T^{-1}],
d is the diameter of the passageway the fluid moves [L],
μ is the viscosity [M T^{-1} L^{-1}].

In an open channel, or pipe flow, the transition between laminar and turbulent flow occurs at Re $\sim$ 2000. It is difficult to estimate d for groundwater flow since the pathways of water are often very irregular. In most cases groundwater flow is slow enough to be laminar.

Steady flow in confined aquifers

Consider a confined aquifer of uniform thickness, and a linear gradient of the potentiometric surface. Then we have,

$$Q = -K \times b \times \frac{d\phi}{dl}\hat{y}, \tag{1.17}$$

where,

$\hat{y}$ is a unit with perpendicular to vertical,

b is the thickness of the confined aquifer (vertical),
$d\phi/dl$ is the slope of the potentiometric surface.

Steady flow in an unconfined aquifer

Now the water table is also the upper boundary of the region of flow (Fig. 1.10), and from simple geometric arguments the hydraulic gradient will increase in the direction of flow.

Two assumptions, called *Dupuit assumptions*, allow us to solve this problem. They are (1) the hydraulic gradient is equal to the slope of the water table, and (2) for small water-table gradients, and the streamlines are horizontal.

Starting with Darcy's equation,

$$Q = -K \times h \times \frac{dh}{dx}\hat{y}, \tag{1.18}$$

where h is the saturated thickness of the aquifer. Integrating,

$$\int_0^L Q dx = -K\hat{y}\int_{H_1}^{H_2} h\, dh. \tag{1.19}$$

After rearrangement, we get,

$$Q = \frac{\hat{y}}{2}K\left(\frac{H_1^2 - H_2^2}{L}\right), \tag{1.20}$$

where

H_1 is the head at the origin,
H_2 is the head at L,
L is the flow length.

The total discharge at any location must be a constant, no water is coming in or going out of the system in this case.

To derive the equation for the hydraulic head, when adding or loosing water at some rate w, we consider the fluxes through an "box"-element. Since the pressure is zero, we write $\phi = h$. The components of the specific discharge from Darcy's equation are,

$$q_i = -K\frac{\partial h}{\partial x_i},$$

where $i = (x, y)$. Then continuity requires (no storage),

$$\frac{\partial}{\partial x}(q_x h) + \frac{\partial}{\partial y}(q_y h) = w. \tag{1.21}$$

We can substitute for q_i,

$$\frac{\partial}{\partial x}\left(Kh\frac{\partial h}{\partial x}\right),$$

and use the fact that,

$$h\frac{\partial h}{\partial x} = \frac{1}{2}\frac{\partial (h^2)}{\partial x},$$

which leads to,

$$-K\left(\frac{\partial^2 h^2}{\partial x^2} + \frac{\partial^2 h^2}{\partial y^2}\right) = 2w, \tag{1.22}$$

where w is net addition or loss rate.

If $w = 0$ we get Laplace equation. If the flow is only in the x-direction, we get

$$\frac{d^2 h^2}{dx^2} = -\frac{2w}{K}, \quad where \quad x = \begin{cases} 0, & h = H_1, \\ L, & h = H_2, \end{cases} \tag{1.23}$$

which gives,

$$h^2 = H_1^2 - \frac{(H_1^2 - H_2^2)x}{L} + \frac{w}{K}(L - x)x, \tag{1.24}$$

where

h is the head at x,
x is the distance from the origin,
L is the distance from the origin to the point where H_2 is measured,
w is the recharge rate [L T^{-1}].

Figure 1.13 shows an example of the hydraulic head obtained using Equation 1.24.

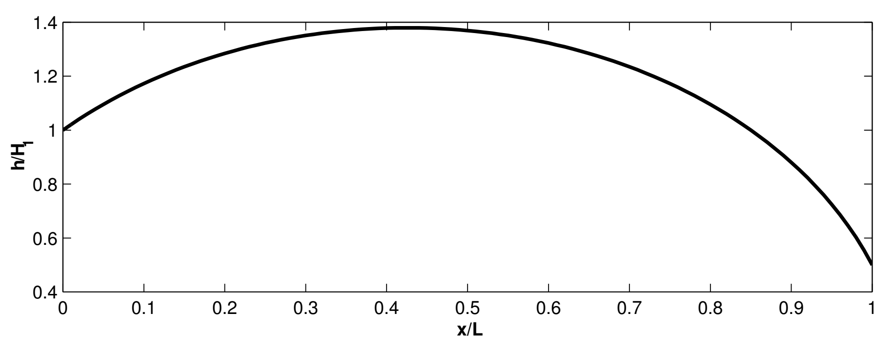

Fig. 1.13 The normalized hydraulic head, h/H_1, for an unconfined aquifer with infiltration, where $H_2/H_1 = 0.5$ and $w/(KH_1) = 5$.

The discharge at any section a distance x from the origin is,

$$\frac{Q}{\hat{y}} = \frac{K(H_1^2 - H_2^2)}{2L} - w\left(\frac{L}{2} - x\right). \tag{1.25}$$

1.7 Flow to wells

The decline in water level around a pumping well is called *drawdown*.

We can learn about aquifer properties when a well is pumped at a constant rate and either the stabilzed drawdown, or the rate of drawdown is measured.

These basic assumptions are made:

- The aquifers are homogenous and isotropic, and radially symmetry.
- The wells fully penetrate the aquifers.
- The aquifer is bounded on the bottom by a confining layer.
- Geometry is horizontal of infinite extend.
- The potentiometric surface is horizontal at, and not changing prior to, $t = 0$, and all subsequent changes are due to the pumping alone.
- Flow is horizontal and Darcy's law holds. Water has a constant density (ρ) and viscosity (μ).

Since we assume radial symmetry it is convenient to use polar coordinates. Two dimensional flow in a confined aquifer is then given by,

$$\frac{\partial^2\phi}{\partial r^2} + \frac{1}{r}\frac{\partial\phi}{\partial r} = \frac{S}{T}\frac{\partial\phi}{\partial t}, \tag{1.26}$$

where (as before), ϕ is the hydraulic head [L], S is storativity [-] (1.8), T is transmissivity [$\mathrm{L}^2\mathrm{T}^{-1}$] (1.12), t is time [T], and r is radial distance [L] from the pumping well.

If there is also a (vertical) recharge rate w [L T^{-1}], a term w/T is added to the left side of (1.26).

Steady State Radial Flow in a Confined Aquifer

Now we look for the steady state solution to (1.26), that is

$$\frac{d^2\phi}{dr^2} + \frac{1}{r}\frac{d\phi}{dr} = 0, \tag{1.27}$$

which has the general solution $\phi = A\ln r + B$.

As boundary conditions one could impose certain fixed values for the head for two values of r. A more realistic set of boundary conditions is to assume a constant head (ϕ_0) at the outer boundary $r = R$, and to impose that at the inner boundary a certain given discharge (Q_0) is extracted from the soil. Then the boundary conditions are,

$$r = \begin{cases} R, & \phi = \phi_0, \\ r_w, & 2\pi r H q_r = -2\pi T r (d\phi/dr) = -Q_0, \end{cases} \tag{1.28}$$

where Darcy's law has been used in the form $q_r = -Kd\phi/dr$, H is the aquifer thickness, and the negative sign before Q_0 means that the discharge flows in the negative r direction, that is, towards the well. The solution subject to these boundary conditions is then,

$$\phi - \phi_0 = \frac{Q_0}{2\pi T} \ln\left(\frac{r}{R}\right). \tag{1.29}$$

Since $r < R$ the logarithm is always negative.

The water level at the well is obtained by setting $r = r_w$ in Equation 1.29,

$$\phi_w = \phi_0 + \frac{Q_0}{2\pi T} \ln\left(\frac{r_w}{R}\right). \tag{1.30}$$

Figure 1.14 shows the drawdown in hydraulic head.

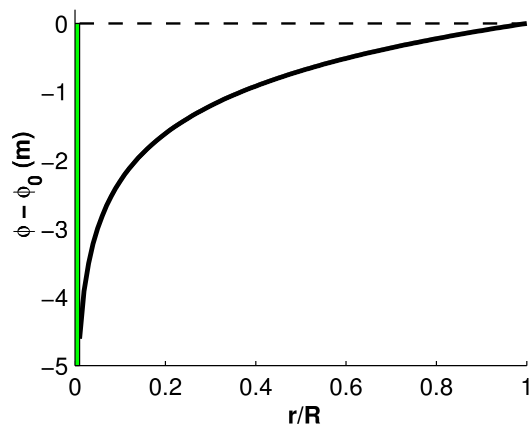

Fig. 1.14 Equilibrium drawdown (steady state) in a confined aquifer (Eq. 1.29). The ratio $Q_0/(2\pi T)$ is set to be equal 1.

Steady State Radial Flow in an Unconfined Aquifer

Since there is no water pressure, we can write $\phi = h$,

$$\frac{d^2h^2}{dr^2} + \frac{1}{r}\frac{dh^2}{dr} = 0, \tag{1.31}$$

with boundary conditions,

$$r = \begin{cases} R, & h = H, \\ r_w, & -Q_0 = -\pi K r \frac{dh^2}{dr}, \end{cases} \tag{1.32}$$

where hdh/dr has been written as $\frac{1}{2}d(h^2)/dr$, which gives,

$$h^2 = H^2 + \frac{Q_0}{\pi K} \ln\left(\frac{r}{R}\right). \tag{1.33}$$

1.8 Problems

1.1. Compare the surface area of a sphere, cylinder and box, for same volume of material (same mass, since same density in all cases). Which is smallest?

Hint:
$V_b = l \times b \times h$, and $A_b = 2 \times l \times b + 2 \times b \times h + 2 \times l \times h$,
$V_c = \pi r^2 \times h$, and $A_c = 2\pi r^2 + 2\pi rh$,
$V_s = 4\pi r^3/3$, and $A_s = 4\pi r^2$.

Put $V_b = V_c = V_s$ (use for instance $l, h, b = 1$), calculate V with the r you get from the volume.

1.2. Find the relation between porosity, n, and void ratio, e. That is, start with (1.1) for n and (1.2) for e, and show that (1.3) holds.

1.3. What is a) an aquifer, and b) the piezometric surface?

1.4. Sketch a) a confined aquifer, and b) an unconfined aquifer.

1.5. Derive Equation (1.11).

1.6. What does the Reynolds-number measure ?

1.7. What is the hydraulic head, ϕ, at an intermediate distance x between ϕ_1 and ϕ_2, for a steady flow in a confined aquifer (1.17)?

1.8. For an unconfined aquifer find: a) The distance, d, from the origin where $Q = 0$ (1.25), i.e. the top of the water table, and b) the maximum hydraulic head (1.24). Hint: $h_{max} = h(x = d)$.

Suggested reading

From Fetter [2]:

- ⋆ Chapters 1-3, for aquifers and soil.
- ⋆ Chapter 4.6, for Darcy's law.
- ⋆ Chapter 4.13, for confined aquifers.
- ⋆ Chapter 4.14, for unconfinded aquifers.
- ⋆ Chapter 5.1-5.4, for flow to wells.
- ⋆ Chapter 6, for soil moisture and ground-water recharge.

References

References

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2. Fetter, C.W.: Applied Hydrogeology, fourth edn. Prentice Hall, New Jersey (2001)
3. Price, M.: Introducing groundwater, first edn. Chapman and Hall, London (1985)
4. Todd, D.K.: Ground Water Hydrology, first edn. Wiley International Edition, New York (1959)
5. Tulaczyk, S., Kamb, B., Scherer, R., Engelhardt, H.F.: Sedimentary processes at the base of a West Antarctic ice stream: Constraints from textural and compositional properties of subglacial debris. Journal of Sedimentary Research **68**(3), 487–496 (1998)