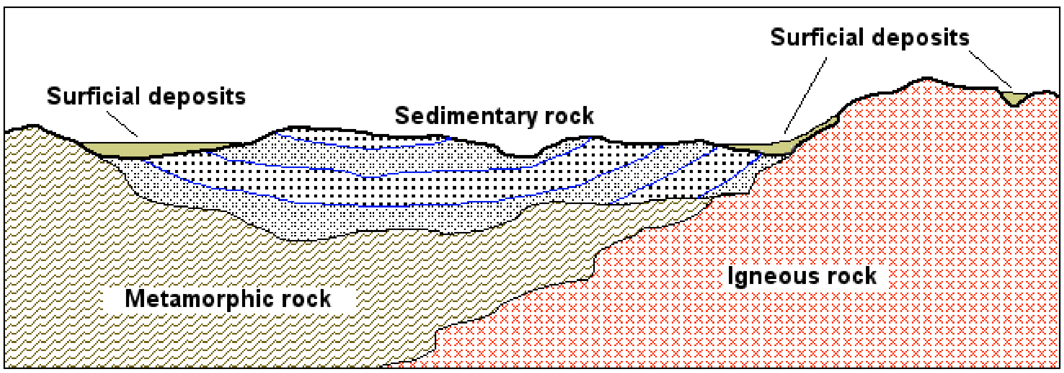
1 Properties of aquifers

1.1 Aquifer materials

Both consolidated and unconsolidated geological materials are important as aquifers. Of the consolidated materials (ie. bedrock), sedimentary rocks are the most important because they tend to have the highest porosities and permeabilities.

Sedimentary rock formations are exposed over approximately 70% of the earth's land surface. These sedimentary formations are typically hundreds to thousands of metres thick, and they are underlain by the igneous and metamorphic rocks that make up the rest of the crust.

Bedrock aquifers



Although most bedrock aquifers are within sedimentary rock, in some areas igneous or metamorphic rock can be important as aquifers.

Much of the bedrock is also covered with tens to hundreds of metres of unconsolidated sediments (a.k.a. **surficial deposits** or **drift**). These include colluvial materials (deposited from mass wasting) alluvial materials (deposited by flowing water), glacial deposits and eolian (wind-blown) deposits. They are "unconsolidated" because they have not been around long enough, and have not been buried deep enough to have become lithified.

Surficial aquifers

Surficial deposits that are more than a few metres thick can be very important sources of groundwater, partly because they tend to have quite high porosities and permeabilities, and also because they are amenable for the development of wells.

We will discuss the aquifer properties of specific types of rock and surficial materials in more detail later in the course.

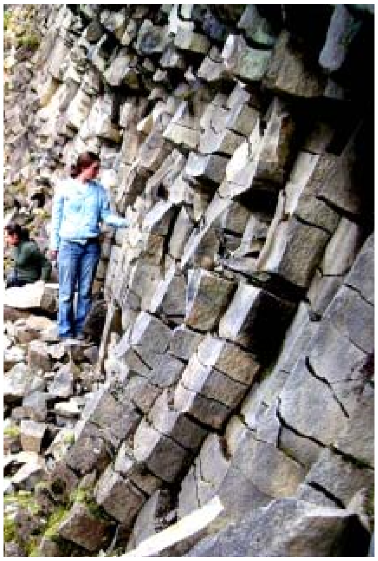
1.2 Rock properties

Rock properties

From the perspective of hydrogeology the important characteristics of rocks are how much open space they have, how well connected those open spaces are, how strong the rocks are, and how soluble they are.

Igneous rocks

Igneous rocks (i.e. rocks formed from the cooling of magma) are comprised of tightly interlocking crystals - primarily crystals of silicate minerals such as quartz, feldspar and amphibole - which tend to be relatively insoluble. In most cases the spaces between crystals are very small. In intrusive igneous rocks these crystals can be quite large (> 1 mm), whereas in volcanic igneous rocks they tend to be much smaller. Some volcanic rocks have vesicular textures caused by the exsolution of gases. Most intrusive igneous rocks (e.g. granitic rocks) are hard and strong, and are less likely to become fractured than other types of rocks. On the other hand, many volcanic rocks are quite well fractured because of their relatively rapid cooling or violent formation.



Fractured columnar basalt near Whistler B.C.

Sedimentary rocks

Sedimentary rocks are formed close to the surface of the earth at relatively low temperatures and pressures. Clastic sedimentary rocks are comprised of weathered and transported fragments of other rocks and minerals. Depending on the degree of sorting and rounding of those fragments, and the extent to which they are cemented together, clastic sedimentary rocks can be quite porous. Some clastic sedimentary rocks are also relatively soft and weak, and are easily susceptible to fracturing. Most sedimentary rocks also have some bedding features that can enhance porosity.

Bedded and cross-bedded Tapeats Fm. Sandstone, Chino Valley, Arizona

Chemical sedimentary rocks are comprised of minerals made up of material originally derived from other rocks, but transported in solution (such as the Ca^{2+} and HCO_3^- that combine to make calcite). As in igneous rocks, these crystals can be tight and interlocking. On the other hand, some chemical sedimentary rocks, such as limestone and evaporates, can develop significant porosity because of their solubility.

Weathered Quatsino Fm. limestone near to Cowichan Lake, B.C.

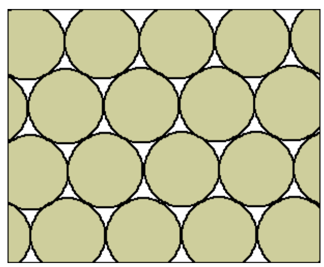
Metamorphic rocks

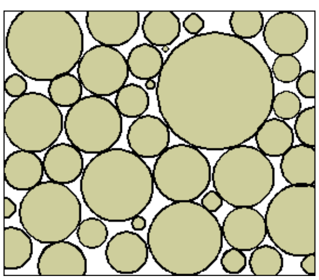
Metamorphic rocks form when existing sedimentary or igneous rocks get heated to the extent that the existing minerals start to recrystallize into new minerals. Almost invariably this leads to the development of interlocking textures with low porosities. Most metamorphic rocks are comprised of the relatively insoluble silicate minerals.

1.3 Porosity

Porosity

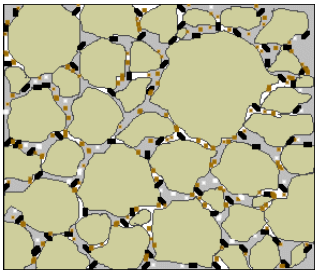
The empty spaces in between the crystals or fragments that make up a rock represent **porosity** that can hold water. Porosity is a measure of how much water a body of rock can hold, expressed as a percentage of the rock's volume. The spaces between the grains or crystals of a rock are referred to as **intergranular porosity**. Fractures in a rock also represent porosity, and this is known as **fracture porosity**.



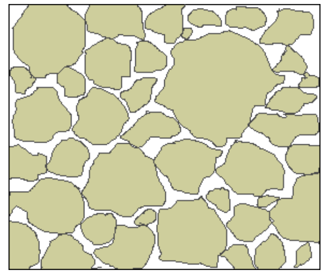


Intergranular porosity

A collection of perfect uniform spheres packed in a rhombic arrangement (as above) will have a porosity of approximately 26%. While sedimentary fragments are never exactly spherical, nor of uniform size, they are also never packed in a perfect arrangement, and porosities of natural sediments can approach or exceed 26%. The porosity of clastic sediments is proportional to sorting and roundness, with the greatest porosities developed in sediments with best sorting (most uniform grain size) and best rounding. Sediments with unrounded and unsorted grains tend to have lower porosities than those that are rounded and sorted, however the more important factor in porosity is the type and extent of matrix and cement.



Clastic sediments with a high proportion of silt- and clay-sized matrix (e.g. wacke sandstones) tend to have lower porosities than cleaner sediments (e.g. arenites). Other types of rocks (chemical sediments and igneous and metamorphic rocks), when not fractured, typically have very low porosity.



Fracture porosity

All rocks are susceptible to fracturing, either as a result of tectonic forces, or as part of weathering (e.g. related to pressure release). Strong rocks, like granite or gneiss, may not fracture as easily as weaker rocks, but they can still be sufficiently well broken up to induce significant fracture-related porosity – especially near to surface. Most sedimentary rocks fracture relatively easily. In addition, fracture-like porosity is developed along bedding planes in many cases.

Exfoliation fracturing of granite in the Coquihalla canyon, B.C.

While it is normally possible to estimate or measure the intergranular porosity of a body of rock - using samples of the rock - it can be very difficult to estimate its degree of fracture porosity. Intergranular porosity may be quite consistent over distances of hundreds of metres, but fracture porosity can vary significantly on scales of metres or less. Information on fracture densities and orientations, and on bedding-plane partings, can be derived from outcrops, and then extrapolated to the subsurface, and this type of information can be important to the understanding of groundwater storage and flow potential.

Some of the best aquifers are in rocks that have both intergranular porosity and well-developed fracture porosity.

Rusty coloration indicates water flow along both fractures and coarse-grained layers in Nanaimo Gp. sandstone

Solution porosity

As noted above, chemical sedimentary rocks in particular, but also some clastic rocks, typically include some minerals that are soluble in water or weak acids. The most obvious example is limestone, which is readily dissolved by rainwater that has reacted with carbon dioxide in the air and the soil (to become weak carbonic acid). Bodies of limestone can have very high porosity, although the porosity is commonly quite localized along fractures and bedding planes.

Dissolution and erosion of calcareous beds in Nanaimo Gp. sandstone, Malaspina Cut

1.4 Specific yield and specific retention

Non-connected pores

Porosity is a measure of how much water a rock can hold, but it is not necessarily a measure of how much water an aquifer can yield or produce. Firstly, some of the pores may not connect with other pores, or they may be so tenuously connected that water cannot move readily from one to another.

Surface tension

Secondly, while water molecules are very small, and can make their way into even the tiniest spaces in a rock, they are strongly attracted and held to solids by surface tension. The strength of the attraction between the water and the mineral grains is proportional to the distance between the water molecule and the surface. Water molecules within a few microns of a surface are very strongly attracted to the surface, while those a few tens or hundreds of microns away are less strongly attracted. If a coarse-grained sediment is allowed to drain (by gravity) much of the water will flow out of it, but some will be retained on the mineral surfaces. In a fine-grained sediment, where many of the pores may only be a few microns in diameter, most of the interstitial water will be tightly held to surfaces, and will not flow out.

Specific yield

The proportion of the water in the rock that does drain out readily is known as the specific yield. Fine-grained materials or rock, such as clay or shale, are likely to have specific yields of 5% or less, even where the porosity is in the order of 20 to 30%. Coarse-grained materials or rocks such as coarse sandstone can have specific yields that are closer to their actual porosity – in the range 20 to 35%.

Specific retention is the opposite of specific yield – being the amount of pore water that does not drain readily under gravity. The sum of the specific yield and specific retention is the porosity. A sandstone might have a specific yield of 20% and a specific retention of 6% for a total porosity of 26%. A shale might have a specific yield of 2% and a specific retention of 31%, for a total porosity of 33%.

Specific retention

1.5 Hydraulic conductivity

Hydraulic conductivity - also known as **permeability** - is the ability of the rock or unconsolidated material to transmit water. Although hydraulic conductivity is partly controlled by intergranular porosity, it is important to remember that very fine grained rocks and materials or rocks with disconnected porosity tend to have low conductivities. Hydraulic conductivity is also controlled by the extent and nature of fracturing and development of bedding-plane partings.

Hydraulic conductivity

Hydraulic conductivity – which is denoted using the symbol K - is expressed in velocity units (e.g. cm/s) but it is not a measure of velocity by itself. The hydraulic conductivity can be used to estimate the velocity at which water will flow through rock or sediment, assuming that there is some gravitational force to push the water. Such a force is provided by the **hydraulic gradient**, or the difference in height of the **potentiometric surface** (e.g. water table) between two points. The slope of the potentiometric surface is expressed as: dh/dl (change in height over change in length), and flow velocity is estimated using the formula:

Flow velocity

Hydraulic gradient

$$v = K(dh/dl)$$

In 1856 Henri Darcy estimated hydraulic conductivity by determining the rate of flow of water as a function of the hydraulic gradient, and he carried out a number of measurements using different sediments and varying hydraulic gradients.

Darcy's experiment

The hydraulic conductivity is used to estimate the flow of dilute water – a fluid with a viscosity of about 1.1 cp (centipoise) at normal temperatures, and a density very close to 1 g/cm^3. In fact, groundwaters can have quite variable viscosities and densities, due to differing temperatures and dissolved contents. Geologists might also be interested in understanding flow rates of other fluids, such as oil. For this reason, we need to consider the concept of **intrinsic permeability** - Ki - which is a measure of the properties of the rock or sediment only.

Intrinsic permeability (Ki)

In this course we will confine ourselves to consideration of the movement of dilute water only, and so we will use the parameter K (and the relationship v=K(dh/dl) to calculate flow rates.

Estimating K from grain size data

Hydraulic conductivity (or permeability) can be estimated in several ways. For unconsolidated coarse (i.e. sandy) sediments permeability is generally proportional to the grain size. Two methods for estimating K from grain-size data are summarized on p. 86 and 87 of Fetter.

Permeameter

Permeability can also be estimated using a permeameter in a manner similar to that used by Darcy in 1856. Water, with an applied hydraulic head, is allowed to flow through a sample of rock or sediment, and the rate of flow is measured. The technique is illustrated and described on pages 90 to 93 of Fetter.

Grain-size data and permeameter readings only provide estimations of the permeability of a sample of the actual material, under conditions that are likely to be very different from those that exist in reality.

Pump and slug tests

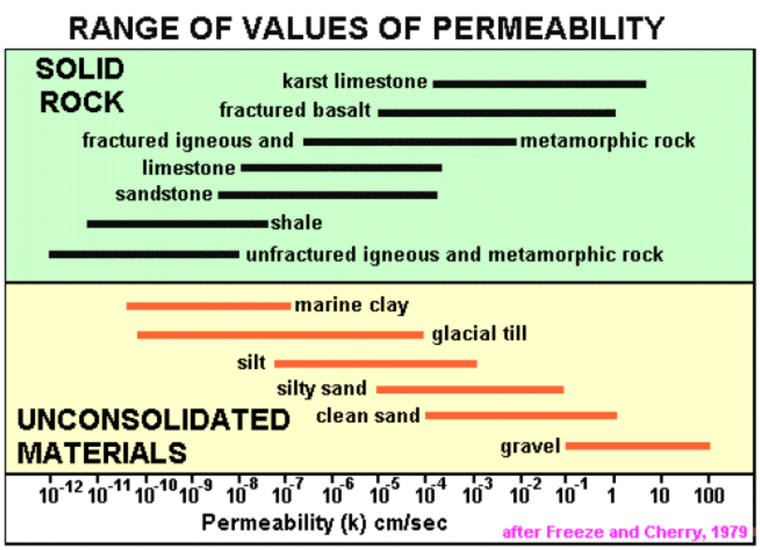
Finally, permeability can be measured in the field by observing the rates at which water levels in wells change in response to pump tests (which involve lowering the water level in a well) or slug tests (which involve raising the water level in a well). These methods, which give a more realistic measure of the actual permeability, are described in Chapter 5 of Fetter.

1.6 Aquifers

Aquifers

Range of permeabilities

As shown on the diagram to the right, or table 3.7 of Fetter, geological materials have hydraulic conductivities that can range over many orders of magnitude.



Different rock types can be classified as aquifers or non-aquifers depending primarily on their permeability, but also on the amount of water that is required. An aquifer is any geological unit that can transmit water at a rate that is sufficient to supply a well – although in some cases that may be a very low-flow well. For some purposes, such as for domestic use, the production could be sufficient with a permeability as low as 10^{-5} cm/s, while for others it might have to be much higher. As noted by Fetter, an aquifer may consist of unconsolidated sand or gravel, sandstone, carbonates, basalt or strongly fractured granite or metamorphic rock. The sandstones and mudstones of the Nanaimo Group are reasonably good aquifers for domestic wells, but only because they are quite well fractured.

Definition of an aquifer

Non-aquifers are generally referred to as: **confining layers**, **aquitards**, **aquicludes** and **aquifuges**. They might include clay deposits or mudstone, or unfractured igneous or metamorphic rock. Confining layers that are relatively permeable are known as **leaky confining layers**.

Non-aquifers

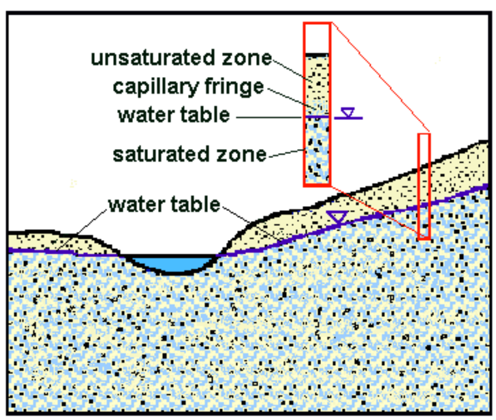
An aquifer in one situation may be considered an aquiclude in another. For example where a siltstone is interbedded with mudrock it is possible that the siltstone could be considered to be an aquifer. The same siltstone interbedded with conglomerate might be thought of as an aquiclude. An aquifuge is supposed to be completely impermeable, but there is no rock that won't let some water through.

Aquifer: a body of rock that contains sufficient saturated permeable material to yield a significant quantity of water to wells and springs
Aquiclude: a body of rock that is incapable of transmitting a significant quantity of water (equivalent to "**confining layer**")
Aquitard: equivalent to acquiclude
Aquifuge: (obsolete) a body of rock with no interconnected openings

Definitions from the AGI Glossary of Geology

An aquifer that is exposed at surface, with no overlying confining layer, is described as an **unconfined aquifer**. The upper limit of the saturated zone in such a case is known as the **water table** (see Figures 3.18 and 3.20 in Fetter). The wet but not saturated region above the water table is known as the **capillary fringe**, while the relatively dry area above that is the **vadose zone** or **unsaturated zone**.

Unconfined or water table aquifers



An aquifer that is covered by a confining layer (i.e., a geological unit – either consolidated or unconsolidated - with low permeability) is a **confined aquifer**. If a well is drilled into a confined aquifer the water in the well may rise above the upper limit of the aquifer. This would be known as an **artesian well**, and the level to which the water rises defines the potentiometric surface (Figure 3.22 in Fetter). Under some circumstances the **potentiometric surface** may extend above the ground surface, producing a **flowing artesian well**.

Confined aquifers

Potentiometric surface

It is possible that an unconfined aquifer may exist over top of a separate and distinct confined aquifer. In such a case there would be a water table surface and a potentiometric surface in the same area, and they would be different.

If a body of permeable rock has isolated impermeable lenses, such as clay lenses in a sand deposit, some groundwater may accumulate above the clay lenses. The upper limit of such a body of water is known as a **perched water table**.

Perched aquifers

The distinction between <u>water table</u> and <u>potentiometric surface</u> is critically important, especially in regional hydrogeological studies where a map of water levels is being prepared in order to understand groundwater flow patterns. For any wells that are being used to measure the water level the hydrogeologist must clearly understand which aquifer is represented by that level, and must be careful not to mix data from different aquifers.

Within an aquifer the water will flow from areas where the elevation of the potentiometric surface is high to areas where the elevation is lower. For example, if the water table is at an elevation of 100 m in well A and 90 m in nearby well B, we can assume that, in general, the groundwater will flow from A towards B. For an unconfined aquifer made up of unconsolidated sediments this principal may apply literally, but there are many situations where the aquifer is either heterogeneous or anisotropic, and the groundwater will be constrained to flow in a specific direction.

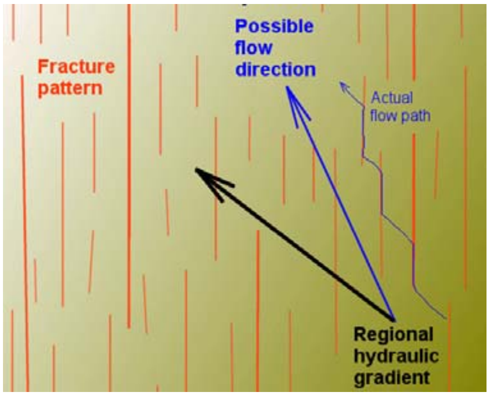
Groundwater flow direction

Examples of heterogeneous aquifers are shown in Fetter's Figure 3.26, and they include aquifers that become thicker or thinner in a certain direction, aquifers that have a change in lithological character, such as a down-dip change grain size or sorting, and aquifers that exhibit distinctive changes in permeability related to sedimentary layering.

Heterogeneous aquifers

Anisotropy within an aquifer can be related to a preferred orientation of grains, as shown in Figure 3.27, or to the concentration of faulting, fracturing or bedding planes in a particular orientation.

Anisotropic aquifers



For example, if a unit is strongly and persistently fractured in a north-south direction and the hydraulic gradient is from southeast to northwest (as shown to the left with the heavy black arrow), there could be a tendency for water to be channelled along the fracture planes such that the overall flow direction is towards the north-northwest.

1.7 Water flow within the unsaturated zone

Flow of recharging water

Within the unsaturated or vadose zone (above the water table) water does not flow according to Darcy's law. Providing that the material is isotropic and homogeneous, recharging water – such as that from a precipitation event - will flow vertically downward in response to gravity. The rate of flow will be dependant on the hydraulic conductivity of the material, but its rate cannot be estimated using v = K(dh/dl).

Because the distribution of precipitation during a large storm may be variable, and the rates of infiltration through the vadose zone may also be variable, temporary high spots may develop on the water table. This can lead to localized flow systems that are different from the normal flow patterns for that area.

The topic of water flow within the vadose zone is discussed in detail in Chapter 6 of Fetter, particularly in section 6.7.