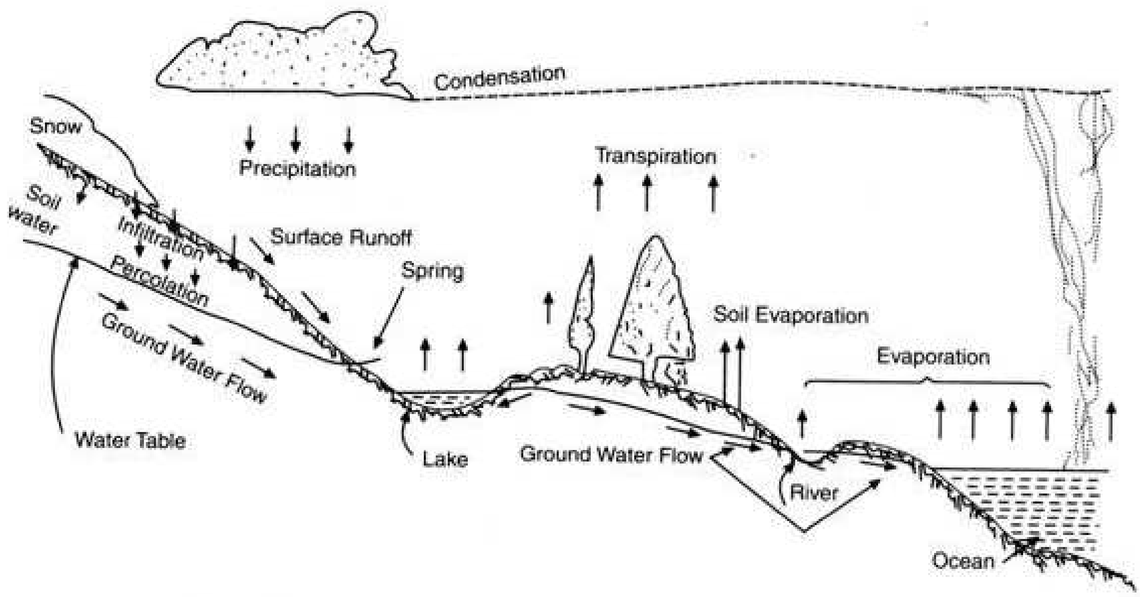
GROUND WATER

Water is an essential commodity to mankind, and the largest available source of fresh water lies underground. Knowledge of ground water, hydrology, once veiled in mystery, has expanded rapidly in recent decades.

Ground water hydrology may be defined as the science of the occurrence, distribution and movement of water below the surface of the earth. **Geohydrology** has an identical meaning, and **hydrogeology** differs only by its greater emphasis on geology. **Geology** governs the occurrence and distribution of ground water, **hydrology** determines the supply of water to the ground, and **fluid mechanics** explains its movement.

Ground water constitutes one portion of the Earth‘s water circulatory system known as the hydrologic cycle. Figure 1.2 illustrates schematically this cycle. Water-bearing formations of the Earth‘s crust act as conduits for transmission and as reservoirs for storage of water. Water enters these formations from the ground surface or from bodies of surface water, after which it travels slowly for varying distances until it returns to the surface by action of natural flow, plants or man. Ground water emerging into surface stream channels helps in sustaining streamflow when surface runoff is low or non-existent. Similarly, water pumped from wells represents the sole water source in many regions during much of every year.



The hydrological cycle

Ground water is commonly understood to mean water occupying all the voids within a geologic stratum. This *saturated zone* is to be distinguished from an unsaturated, or *aeration zone* where voids are filled with water and air. Water contained in saturated zones is important for engineering works, geologic studies, and water supply developments.

Practically all ground water originates as surface water. Principal sources of natural *recharge* include *precipitation*, *streamflow*, *lakes*, and *reservoirs*. Other contributions, known as artificial recharge, occur from *excess irrigation, seepage from canals*, and water purposely applied to increasing ground water supplies. Even *sea water* can enter underground along coasts where hydraulic gradients slope downward in an inland direction. Water within the ground moves downward through the unsaturated zone under the action of gravity, whereas in the saturated zone it moves in a direction determined by the surrounding hydraulic situation.

Discharge of ground water occurs when water emerges from underground. Most natural discharge occurs as flow into surface water bodies, such as streams, lakes and oceans; flow to the surface appears as a spring. Ground water near the surface may return directly to the atmosphere by evaporation from within the soil and by transpiration from vegetation. Pumpage from wells constitutes the major artificial discharge of ground water.

Rock Properties Affecting Ground Water

Ground water occurs in permeable geologic formations known as **aquifers**, that is, formations having structures that permit appreciable water to move through them under ordinary field conditions. *(The word aquifer can be traced to its Latin origin. Aqui- is a combination from aqua, meaning water, and –fer comes from ferre, to bear. Hence, an aquifer, literally, is a water bearer.)* Ground water reservoir and water-bearing formation are commonly used as synonyms. In contradistinction, an aquiclude is an impermeable formation which may contain water but is incapable of transmitting significant water quantities; clay is an example. *(The suffix –clude of aquiclude is derived from the Latin claudere, to shut or close).* An aquifuge is an impermeable formation neither containing nor transmitting water; solid granite belongs in this category. (*The suffix –fuge of aquifuge comes from fugere, to drive away).*

That portion of a rock or soil not occupied by solid mineral matter may be occupied by ground water. These spaces are known as voids, interstices, pores, or pore space. Original interstices were created by geologic processes governing the origin of the geologic formation and are found in sedimentary and igneous rocks. Secondary interstices developed after the rock was formed; examples include joints, fractures solution openings, and openings formed by plants and animals. Depending upon the connection of interstices with others, the may be classed as communicating or isolated.

The porosity of a rock or soil is a measure of the contained interstices. It is expressed as the percentage of void space to the total volume of the mass. If a is the porosity, then $\alpha = \frac{100w}{V}$, where ***w*** is the volume of water required to fill, or saturate, all of the pore space, and ***V*** is the total volume of the rock or soil.

In terms of ground water supply, granular sedimentary deposits are of major importance. Porosities in these deposits depend on the shape and arrangement of individual particles, distribution by size, and degree of cementation and compaction. In consolidated formations, removal of mineral matter by solution and degree of fracture are also important. Porosities range from near zero to more than 50 per cent.

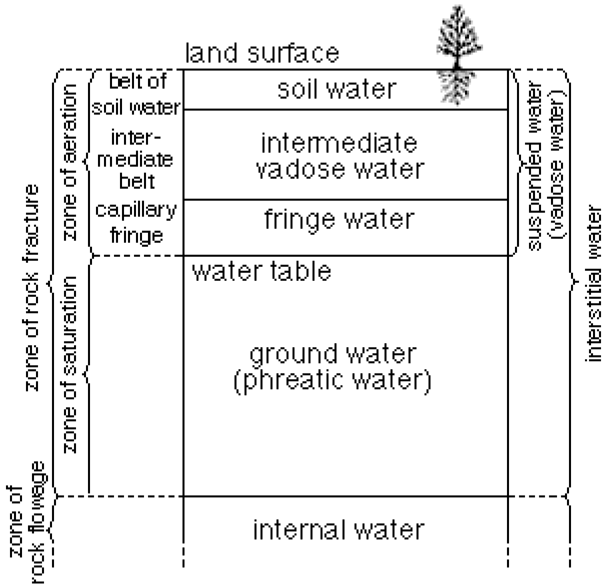
Representative Porosity Ranges for Sedimentary Materials

Material	**Porosity, [%]**
Soils	50-60
Clay	45-55
Slit	40-50
Medium to coarse mixed sand	35-40
Uniform sand	30-35
Gravel	30-40
Gravel and sand	20-35
Sandstone	10-20
Shale	1-10
Limestone	1-10

The subsurface occurrence of ground water may be divided into **zones of saturation and aeration.** In the zone of saturation all interstices are filled with water under hydrostatic pressure. The zone of aeration consists of interstices accupied partially by water and partially by air. Over most of the land masses of the earth a single zone of aeration overlies a single zone of saturation and extends upward to the ground surface, as shown in figure. The saturated zone is bounded at the top by either a limiting surface of saturation or overlying impermeable strata such as clay beds or bedrock.

In the absence of overlaying impermeable strata, the upper surface of the zone of saturation is the water table, or phreatic surface*. (Phreatic is derived from the Greek phrear, -atos, meaning a well).* This is defined as the surface of atmospheric pressure and would be revealed by the level at which water stands in a well penetrating the aquifer. Actually, saturation extends slightly above the water table owing to capillary attraction; however, water is held here at less than atmospheric pressure.

Gravitational water is excess soil water which drains through the soil under the influence of gravity.



Division of subsurface water

Saturated Zone. Ground water fills all of the interstices in the saturated zone, hence the porosity is a direct measure of the water contained per unit volume.

Not all of this water may be removed from the ground by drainage or pumping from a well, however, as molecular and surface tension forces will hold a portion of the water in place. Thus, retained water is that held in place against gravity. The specific retention of a rock or soil is the ration expressed as a percentage of the volume of water it will retain after saturation against the force of gravity to its own volume. If S_r is the specific retention, then

$$S_r = \frac{100w_r}{V}$$

where w_r is the volume occupied by retained water, and V is the gross volume of the rock or soil.

On the other hand, the water which can be drained is expressed as the specific yield S_y. The term, effective porosity, has a synonymous meaning. It may be defined as the ratio expressed as a percentag of the volume of water which, after being saturated, can be drained by gravity to its own volume.

Therefore,

$$S_y = \frac{100w_y}{V}$$

where w_y is the volume of water drained. Because $w_r + w_y = w$, it is apparent that

$$S_y + S_r = \alpha$$

Thus, specific yield is a fraction of the porosity of an aquifer. Values depend upon grain size , shape and distribution of pores, and compaction of the stratum. For uniform sand, specific yield may equal up to 30%, but most alluvial aquifers give values in the range of 10 to 20 %.

Specific Yields of Water-Bearing Deposits in Sacramento Valley, California (After Poland and others)

Material	*Specific Yield, [%]*
Gravel	25
Sand, including sand and gravel, and gravel and sand	20
Fine sand, hard sand, tight sand, sandstone, and related	10
Clay and gravel, gravel and clay, cemented gravel, and	5
Clay, slit, sandy clay, lava rock, and related fine-	3

Geologic Formations as Aquifers

A rock formation or material which willyield significant quantities of water has been defined as an aquifer. Probably 90% of all developed aquifers consists of unconsolidated rocks, chiefly gravel and sand. These aquifers may be divided into four categories, based on manner of occurence; water courses, abandoned or buried valleys, plains, and intermontane valleys.

Water courses consists of the alluvium that forms and underlies stream channels, as well as forming the adjacent flood plains. Wells located in highly permeable strata bordering streams produce large quantaties of water.

Abandonded or buried valley are valleys no longer occupied by streams that formed them. Although such valleys may look like water courses in permeability and quantity of ground water storage, the recharge and capabilities for perennial yield are usually less.

Extensive plains underlain by unconsolidated sediments exist. In some places gravel and sand beds form important aquifers under these plains; in other places they are relatively thin and have limited productivity.

The ground water reservoirs are recharged chiefly in areas accessible to downward percolation of water from percipitaion and from occasional streams.

Intermonteine valleys are underlain by tremendous volumes of unconsolidated rock materials derived by erosion of bordering mountains. The sand and gravel beds of these aquifers produce large quantities of water, most of which is refilled by seepage from streams into alluvial fans at mouths of mountain canyons.

Limestones vary widely in density, porosity, and permeability, depending upon the degree of consolidation and the development of permeable zones after deposition. Those most important as aquifers contain sizable proportions of the original rock which have been dissolved and removed. The solution of calcium carbonate by water causes prevailing hard ground water to be found in limestone aquifers; also, by dissolving the rock, water tends to increase the pore space and permeability with time. Ultimate development of a limestone terrane forms a karst region, where subterranean drainage through the limestone creates large ground water reservoirs. Although uncommon, gypsum is another soluble rock.

Volcanic rocks may form permeable aquifers. Basalt flows are very permeable, corresponding to limestones in this regard. The permeable zones in volcanic rocks include flow breccias, porous zones between lava beds, shrinkage cracks, and joints. Rhyolites are less permeable than basalt.

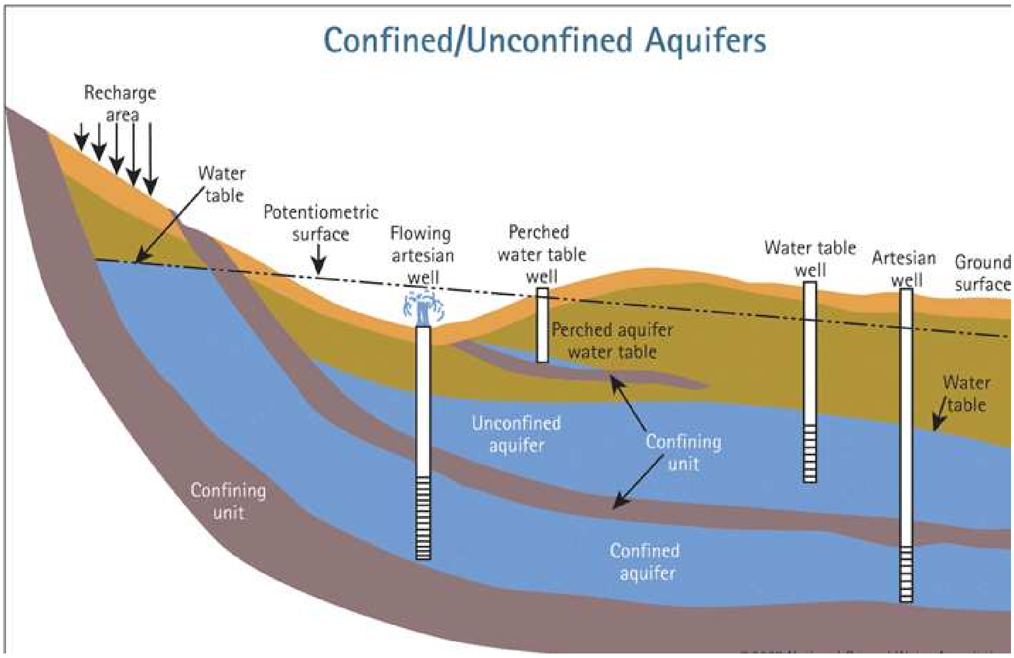
Sandstone and conglomerate are cemented forms of sand and gravel. As such, their and porosity yield have been reduced by the cement. The best sandstone aquifers are those which are only partially cemented.

Crystalline and metamorphic rocks are relatively impermeable and are poor aquifers.

Clay ana coarses materials mixed with clayare generally porous but their pores are so small that they may be regarded as relatively impermeable.

Most aquifers are of large areal extent and may be visualized as underground storage reservoirs. Water enters a reservoir from natural or artificial recharge; it flows out under the action of gravity or is extracted by wells.

Aquifers may be classed as unconfined or confined, depending upon the presence or absence of a water table.



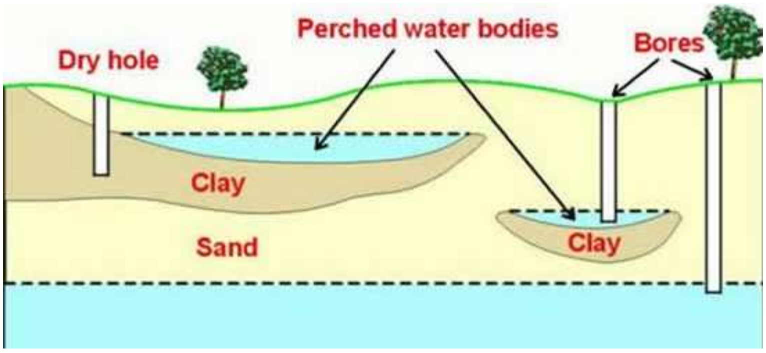
An unconfined aquifer is one in which a water table serves as the upper surface of the zone of saturation. The water table varies in undulating form and in slope, depending upon areas of recharge and discharge, pumpage from wells, and permeability. Rises and falls in the water table correspond to changes in the volume of water in storage within an aquifer. The upper figure is an idealized section through an unconfined aquifer; the upper aquifer in figure is unconfined also.

Confined aquifers, also known as artesian or pressure aquifers, occur where ground water is confined under pressure greater than atmospheric by overlying relatively impermeable strata. *(The word artesian has an interesting origin. It is derived from the French artesien, meaning of the pertaining Artois, the northernmost province of France. Here the first deep wells to tap confined aquifers, dating from about 1750, were drilled and investigated. Originally the term referred to a well with freely flowing water, but at present it is applied to any well penetrating a confined aquifer or simply the aquifer itself.)* In a well penetrating such an aquifer, the water level will rise above the bottom of the confining bed, as shown in the artesian and flowing wells at the figure above. Water enters a confined aquifer in an area where the confining bed rises to the surface or ends underground and the aquifer becomes unconfined. A region supplying water to a confined aquifer is known as a recharge area. Rises and falls of water in wells penetrating confined aquifers results primarily from changes in pressure rather than changes in storage volumes. Hence, confined aquifers have only small changes in storage and serve mainly as conduits for conveying water from recharge areas to locations of natural or artificial discharge.

The piezometric surface of a confined aquifer is an imaginary surface coinciding with the hydrostatic pressure level of the water in the aquifer. The water level in a well penetrating a confined aquifer defines the elevation of the piezometric surface at that point.

It should be noted that a confined aquifer becomes an unconfined aquifer when the piezometric surface falls below the bottom of the upper confining bed. Also, quite commonly an unconfined aquifer exists above a confined one, as shown in upper figure.

A special case of an unconfined aquifer is the perched aquifer, which is illustrated by figure bellow. This occurs wherever a ground water body is separated from the main ground water by a relatively impermeable stratum of small areal extent and by the zone of aeration above the main body of ground water. Clay lenses in sedimentary deposits often have shallow perched water bodies overlying them.



Perched aquifers

Ground Water Basins

A ground water basin may be defined as a physiographic unit containing one large aquifer or several connected and interrelated aquifers.

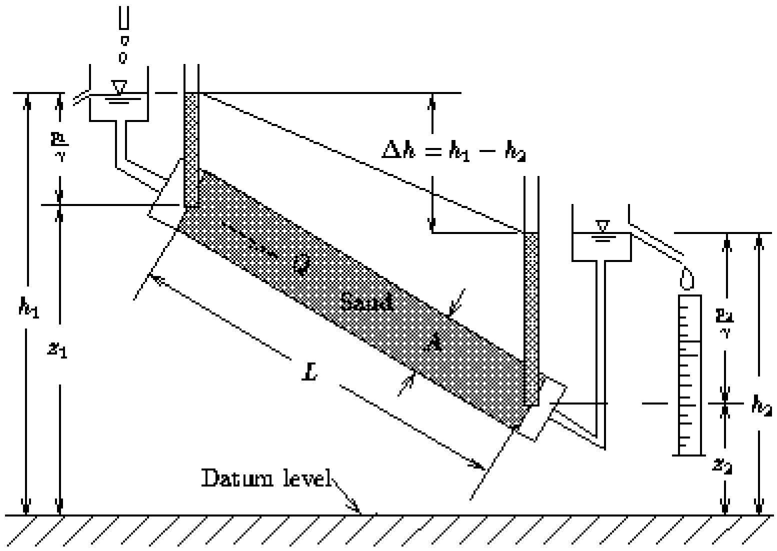
GROUND WATER MOVEMENT

Ground water in its natural state is constantly moving. This movement governed by established hydraulic principles. The flow through aquifers, most of which are natural porous media, can be expressed by what is known as Darcy's law. Permeability, which is a measure of the ease of flow through the media, is an important constant in the flow equation.

Darcy's Law

More than a century ago Henry Darcy, a French hydraulic engineer, investigated the flow of water through horizontal beds of sand to be used for water filtration. He has attempted by precise experiments to determine the law of the flow of water through filters. The experiments demonstrate positively that the volume of water which passes through a bed of sand of a thickness of the bed traversed; thus, in calling ***s*** the surface area of a filter, ***k*** a coefficient depending on the nature of the sand, ***e*** the thickness of the sand bed, ***P+H*** the atmospheric pressure added to the depth of water on the filter; one has for the flow of this last condition

$Q = (ks / e)(H + e + H_0)$, which reduces to $Q = (ks / e)(H + e)$ when $H_o=0$, or when the pressure below the filter is equal to the weight of the atmosphere.



Darcy`s experiment

This statement, that the flow rate through porous media is proportional to the head loss and inversely proportional to the length of the flow path, is known universally as Darcy's law. It, more than any other contribution, serves as the basis for present-day knowledge of ground water flow. Analysis and solution of problems relating to ground water movement and well hydraulics began after Darcy's work.

The experimental verification of Darcy's law can be performed with water flowing at a rate of ***Q*** through a cylinder of cross-sectional area ***A*** packed with sand and having piezometer taps a distance ***L*** apart, as shown in the scheme above. Total energy heads, or fluid potentials, above a datum plane may be expressed by the Bernoulli equation

$$\frac{p_1}{\gamma} + \frac{v_1^2}{2g} + z_1 = \frac{p_2}{\gamma} + \frac{v_2^2}{2g} + z_2 + h_l$$

where ***p*** is pressure, ***Y*** is the specific weight of water, ***v*** is the velocity of flow, ***g*** is the acceleration of gravity, and h_l is head loss.

Because velocities in porous media are usually low, velocity heads may be neglected without appreciable error. Hence, by rewriting, the head loss becomes

$$h_l = \left(\frac{p_1}{\gamma} + z_1\right) - \left(\frac{p_2}{\gamma} + z_2\right)$$

Therefore, the resulting head loss is defined as the potential loss within the sand cylinder, this energy being lost by frictional resistance dissipated as Reynolds number has been employed to establish the limit of flows described by Darcy's law, corresponding to the value where the linear relationship is no longer valid.

Reynolds number is expressed as

$$N_R = \frac{\rho \upsilon D}{\mu}$$

where ***r*** is the fluid density, ν the velocity, ***D*** the diameter (of a pipe), ***m*** and the viscosity of the fluid.

For almost all natural ground water motion, $\boldsymbol{N_R<1}$; therefore, Darcy's law is applicable. Deviations from Darcy's law may be found in rock aquifers, in unconsolidated aquifers with steep hydraulic gradients, or in those containing large diameter solution openings.

It is assumed that turbulent flow begins at the upper limit of Darcy's law. Recent experimental investigations by Schneebeli and Hubbert indicate otherwise.

Schneebeli found from visual observations of flow in porous media that the first appearance of turbulence occurred at a Reynolds number of about 60, whereas Hubbert found it in the range of 600 to 700.

The transitions form laminar flow to turbulent flow are gradual as evidenced by the mild curvature in scheme of Darcy's experiment, between $\boldsymbol{N_R=1}$ and $\boldsymbol{N_R=10}$.

The most important hydraulic properties of rocks relate to void ratios, the amount of water which can be drained from them and the ease of flow of water through them. This latter is expressed as ***K***, the coefficient of permeability and also by ***T***, transmissibility, which is the rate of flow of groundwater in m3 day^{-1} m^{-1} through a vertical strip of aquifer 1m wide extending the entire length of the saturated aquifer under an hydraulic gradient of 1 in 1.

Steady Flow – for steady flow there is no change in conditions with respect to time.

Unsteady Flow – the conditions depend on the time.

General Flow Equations

From Darcy's law $\nu = K\frac{\partial h}{\partial s}$ where $\boldsymbol{\upsilon}$, $\mathbf{K}$ and $\mathbf{h}$ are defined previously and **s** is the distance along the average direction of flow in anisotropic permeability conditions the velocity components in rectangular coordinate system are:

$$\nu_x = K_x\frac{\partial h}{\partial x},\ \ \nu_y = K_y\frac{\partial h}{\partial y},\ \ \nu_z = K_z\frac{\partial h}{\partial z},$$

where K_x, K_y, K_z are coefficients of permeability in X;Y and Z directions.

In homogenous aquifers with isotropic permeability

$$v_x = K\frac{\partial h}{\partial x},\ \ v_y = K\frac{\partial h}{\partial y},\ \ v_z = K\frac{\partial h}{\partial z}$$

Velocity potential */ ϕ /* is a scalar function of space and time such that its negative derivative with respect to any direction is the fluid velocity in that direction.

$$\phi = -Kh,\ \text{ then } v_x = -\frac{\partial \phi}{\partial x},\ \ v_y = -\frac{\partial \phi}{\partial y},\ \ v_z = -\frac{\partial \phi}{\partial z}$$

It's apparent that a velocity potential exists for ground water flow.

Steady Flow – for steady flow there is no change in conditions with respect to time.

All ground water flow must satisfy the equation of continuity:

$$-\left[\frac{\partial(\rho v_x)}{\partial x} + \frac{\partial(\rho v_y)}{\partial y} + \frac{\partial(\rho v_z)}{\partial z}\right] = \frac{\partial \rho}{\partial t}$$

ρ is fluid density; constant / water is incompressible fluid /
t is time; no changes in the respect of time / steady flow /

$$\frac{\partial v_x}{\partial x} + \frac{\partial v_y}{\partial y} + \frac{\partial v_z}{\partial z} = 0$$

If we use the velocity potential:

$$\frac{\partial^2 \phi}{\partial x^2} + \frac{\partial^2 \phi}{\partial y^2} + \frac{\partial^2 \phi}{\partial z^2} = 0$$

Replacing $\phi = -Kh$

$$\frac{\partial^2 h}{\partial x^2} + \frac{\partial^2 h}{\partial y^2} + \frac{\partial^2 h}{\partial z^2} = 0$$

General partial differential equation for steady flow in homogenous and isotropic media.

Unsteady Flow – the conditions depend on the time.

To derive the equation for unsteady flow we have to use the **storage coefficient S**.

For unconfined aquifer this is the specific yield, but for confined aquifer it is a measure of the aquifer compressibility.

β is a change within a column of unit cross-sectional area extending upward through a confined aquifer.

$\beta = \dfrac{-\partial V / V}{\partial p}$, where V is volume and p is pressure.

Compressive force acts in a vertical direction (normal to the plane of aquifer) over a large area extend so that changes in horizontal directions are negligible.

When the piezometric surface is lowered a unit distance, the quantity of water released from the column by the pressure change is ***S***.

$$S = \partial V$$

The volume of the column is ***V=1b=b***, where b is the aquifer thickness.

The change in pressure is $\partial p = -\gamma(1) = -\gamma \ \left(\gamma \text{ - specific weight}\right).$

$$\beta = \frac{S}{\gamma b}$$

In elastic material

$$\frac{\partial V}{V} = -\frac{\partial p}{P}, \ \text{then } \partial P = \rho\beta\partial p$$

By replacing β and inserting this expression in the continuity equation

$$-\left[\frac{\partial(\rho v_x)}{\partial x} + \frac{\partial(\rho v_y)}{\partial y} + \frac{\partial(\rho v_z)}{\partial z}\right] = \frac{\partial \rho}{\partial t}$$

$$-\left[\frac{\partial(\rho v_x)}{\partial x} + \frac{\partial(\rho v_y)}{\partial y} + \frac{\partial(\rho v_z)}{\partial z}\right] = \frac{\rho S}{b\gamma}\frac{\partial p}{\partial t}$$

Using the values of velocity components assuming that ρ is a constant

$$K\left(\frac{\partial^2 h}{\partial x^2} + \frac{\partial^2 h}{\partial y^2} + \frac{\partial^2 h}{\partial z^2}\right) = \frac{\rho S}{b\gamma}\frac{\partial p}{\partial t}$$

Where subsidizing $p = \gamma \mathrm{h}$

$$\frac{\partial^2 h}{\partial x^2} + \frac{\partial^2 h}{\partial y^2} + \frac{\partial^2 h}{\partial z^2} = \frac{\rho S}{Kb}\frac{\partial h}{\partial t}$$

Partial differential equation governing the unsteady flow of water in a compressible confined aquifer of uniform thickness b.

It follows that the head loss is independent of the inclination of the cylinder.

Now Darcy's measurements showed that the proportionalities exists.

$$Q \propto h_L \ \text{and} \ Q \propto \frac{1}{L}$$

Introducing a proportionality constant ***K*** leads to the equation

$$Q = KA\frac{h_L}{L}$$

Expressed in general terms

$$Q = KA\frac{dh}{dL} \quad \text{or simply}\ v = \frac{Q}{A} = K\frac{dh}{dL}$$

Where ***dh/dL*** is the hydraulic gradient. Equation 3.5 states that Darcy's law in its simplest form namely that the flow velocity v is equal to the product of the constant ***K***, known as the coefficient of permeability, and the hydraulic gradient. This velocity is an apparent one, defined as the discharge divided by the cross-sectional area of the porous media through which it is flowing. The actual velocity varies from point to point throughout the media.

It should be noted that ground water flows is in accordance with the hydraulic gradient.

Range of Validity of Darcy's Law

In applying Darcy's law it is important to know the range of validity within which it is applicable. Because velocity in laminar flow, such as water flowing in a capillary tube, is proportional to the first power of the hydraulic gradient (Poiseuille's law), it seems reasonable to believe that Darcy's law applies to laminar flow in porous media.

For flow in pipes and other large sections, the Reynolds number serves as a criterion to distinguish between laminar and turbulent flow.

For almost all natural ground water motion, $N_R<1$; therefore, Darcy's law is applicable.

Deviations from Darcy's law may be found in rock aquifers, in unconsolidated aquifers with steep hydraulic gradients, or in those containing large diameter solution openings.

It is assumed that turbulent flow begins at the upper limit of Darcy's law.

The transitions from laminar flow to turbulent flow are between $N_R=1$ and $N_R=10$.

For turbulent flow it is applicable the law

$$v = K\sqrt{\frac{dh}{dL}}$$

For combined flow

$$I = \alpha v + \beta v^2$$

Ground Water Flow Lines

Flow nets. For specified boundary conditions **flow lines** and **equipotential lines** can be mapped in two dimensions to form a flow net.

Since groundwater may be regarded as a continuous contact body of fluid, if a pressure change occurs at any point there will be an effect everywhere.

The actual flow of groundwater is controlled by head changes, the head at a point being a measure of the potential energy of the fluid relative to a specified state.

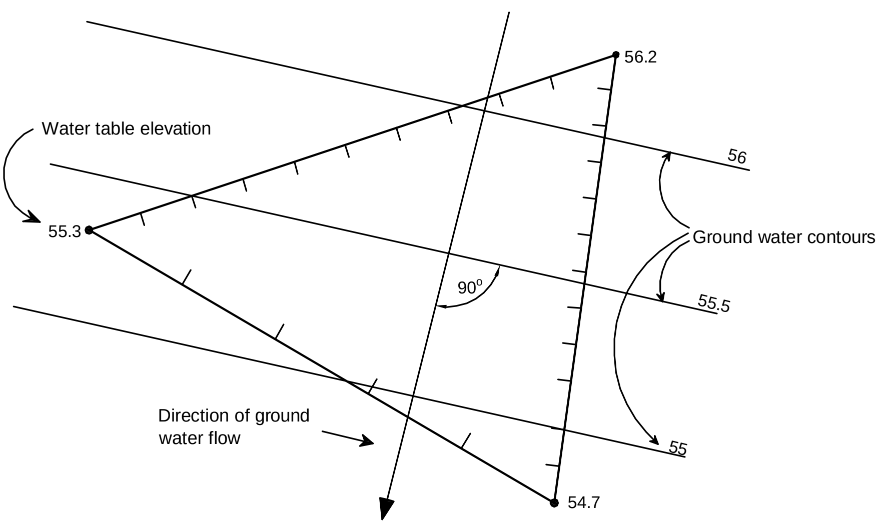
Points possessing equal fluid potential define **equipotential surfaces** with the aquifer and from these it is possible to construct **flow lines** for the groundwater.

A flow line is defined here as a line such that the macroscopic velocity vector is everywhere tangent to it.

It is possible to construct a **flow net** from these.

With only three ground water elevations known from wells, estimates of local ground water contours and flow directions can be determined.

Flow lines, sketched perpendicular to contours, show directions of movement



Estimate of ground water contours and flow direction from water tab elevations in three wells.

The two sets of lines from an orthogonal patter of small squares.

Graphical solution is based on trial-and-error approximation.

The hydraulic gradient

$$i = \frac{dh}{ds}$$

The constant flow between two adjacent flow lines is $dq = K\frac{dh}{ds}dm,$ ***m*** is the unit thickness.

For squares of flow net $ds \cong dm,$ then $dq = Kdh$.

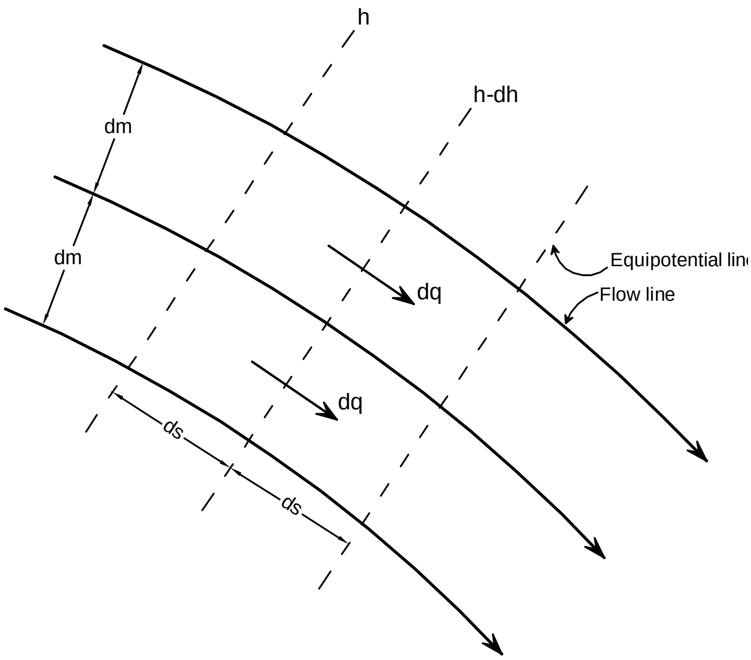
For the flow net where the total head loss ***h*** is divided into ***n*** squares between any two adjacent flow lines.

$$dh = \frac{h}{n}$$

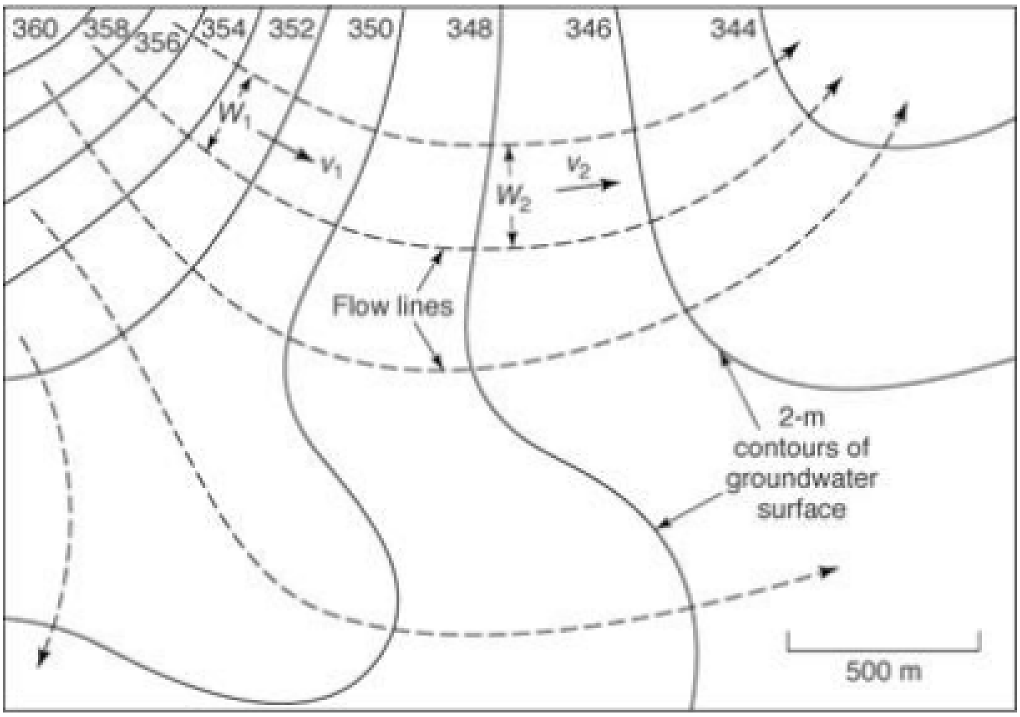
If the flow is divided into ***m*** channels by flow lines, the total flow is

$$Q = mdq = \frac{Kmh}{n}$$

Thus geometry of the flow net, together with the permeability and head loss, enables the total flow in the section to be computed directly.

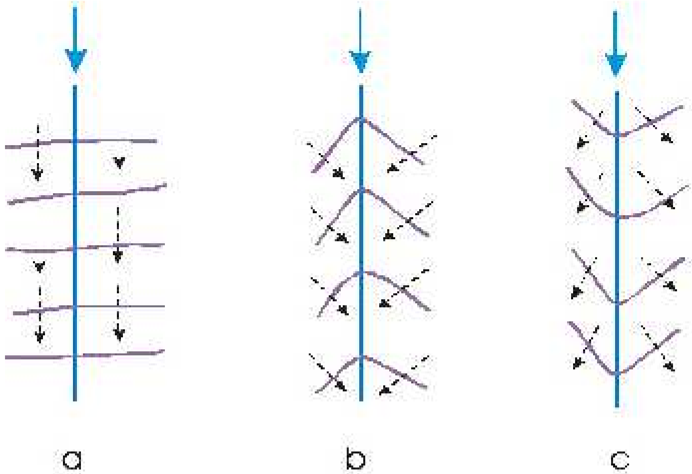


Direction of an orthogonal flow net formed by flow and equipotential lines.



Contour map of ground water levels

Contours of ground water levels (equipotential lines) and flow lines (perpendicular to equipotential lines) indicate area of recharge and discharge



River and a ground water flow

a) no hydraulic connection between river and aquifer;
b) the river drain the aquifer;
c) the aquifer drains water from the river