Hydrogeology Lecture Notes

Edition 1.0

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Hydrogeology lecture notes for the Hydraulics Bachelor's program.

Contents

CHAPTER I: GENERAL INFORMATION

I.1 INTRODUCTION

Hydrogeology is both a descriptive and quantitative science that studies the relationships between groundwater and geological materials and processes. In many regions of the world, groundwater is either the primary source of drinking water for populations, or even the only source of water. It is also limited in quantity and nowadays faces serious dangers of contamination from human activities. The development and management of water resources are important aspects of hydrogeology.

The breakdown of this subject's title suggests that it involves the study of water movement in soil:

- Hydraulics: Study of flows of incompressible Newtonian fluids such as water, but from an engineer's perspective.
- Underground: Term that designates in the current context the soil in its different forms provided that it presents, through its constituent particles, accessibility to the fluid in movement.

The study of groundwater movement is of great importance whenever the infiltration phenomenon is present.

This importance can be illustrated through several examples:

- Effects of interstitial flow on the stress state of a porous structure subject to this flow
- Study of an aquifer's regime
- Simulation of seawater intrusion into an aquifer when near the sea
- Study of aquifer pollution

I.2 ORIGIN OF GROUNDWATER

The supply of a hydrological basin is ensured by a portion of precipitation which represents effective precipitation obtained by subtracting losses from evapotranspiration.

The above supply is distributed into:

- Runoff, which will feed surface flow in the hydrographic network

- Infiltration, which will feed the groundwater aquifer

The infiltration height (quantity of water infiltrated through the soil surface per unit of time) or the infiltration rate (ratio of infiltration height to effective precipitation height) are influenced by several factors including:

- Basin geomorphology: topography and geometry of the hydrographic network
- Soil lithology
- Nature of surface developments such as dams, straightening of watercourses, etc.

I.3 CONCEPT OF HYDROGEOLOGY

Hydrogeology, the science of groundwater, is a discipline of earth sciences oriented toward applications. Its objectives are:

- Acquisition of numerical data through prospecting or field experimentation
- Study of the role of materials constituting the subsoil (distribution and characteristics)
- Study of flow modalities
- Study of physical and chemical properties of groundwater
- Realization of exploitation boreholes
- Management and planning of water exploitation
- Protection of groundwater resources

For this purpose, Hydrogeology, which is a multidisciplinary science, uses methods and means of geophysical prospecting, drilling and capture techniques, geochemistry of rocks and waters, underground hydrodynamics, statistics and the use of computers for data processing and for mathematical models of aquifer simulation.

I.4 GLOBAL WATER CYCLE

The movement of water particles in two main states (vapor and liquid) at the Earth's surface constitutes the global water cycle. This cycle can be summarized by the following equation:

P (precipitation) = E (evaporation) + R (runoff) + I (infiltration)

The water cycle begins with the annual transformation of a volume of water (577,000 km^3) into water vapor under the action of evapotranspiration (ET). This latter term groups two phenomena: the first physical (evaporation E), takes place from free water surfaces (oceans, seas, lakes, rivers...) and the second biological corresponding to plant transpiration.

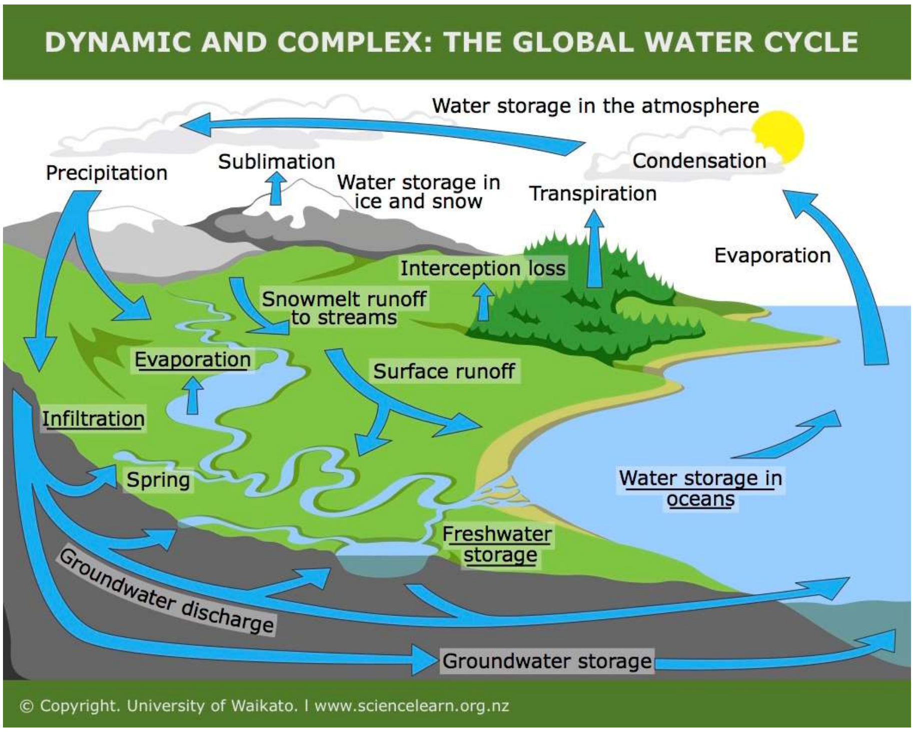


Figure 1: Dynamic and complex global water cycle.

In a second step, this water vapor condenses (transformation of vapor into liquid) in the form of clouds that give rise to precipitation (P): rain and snow. The annual volume of precipitation equals that of evapotranspiration: the global water cycle is therefore balanced.

In a third step, precipitation separates into three parts :

- A first part evaporates before even reaching the soil surface and rejoins the water cycle
- A second part flows toward the hydrographic network and free water surfaces: this is surface runoff (R). This volume is estimated at 43,800 km^3 of water/year
- A last part infiltrates into the subsoil: this is infiltration (I). The quantity of water that reaches the aquifers is called: effective infiltration

I.5 WATER IN THE WORLD

The water stock on Earth is unevenly distributed in six large reservoirs that total a volume of approximately 1.39×10^9 billion m^3. Fresh water represents only 2.9% of this volume, or 0.4×10^8 billion m^3 of water. These six reservoirs are:

- Ocean: constitutes the main driver of the water cycle. It plays a very important role in the circulation and evaporation of water, and in the homogenization of the globe's temperature. It constitutes the main reservoir, distributed over an area of approximately 361 million km^2

- Ice: located at the level of ice cap reservoirs and eternal snows (polar regions: Arctic Sea in the North and Antarctic Sea in the South)
- Groundwater: the capacity of the underground reservoir is estimated at 24×10^6 billion m^3 of water between depths of 0 and 2000 m
- Surface water: represented by lakes and surface watercourses
- Atmospheric water: contained in the gaseous envelope that surrounds the terrestrial globe
- Biological water: water contained in living beings, animals and plants

CHAPTER II – HYDROLOGY AND CLIMATOLOGY

II.1 HYDROLOGICAL SYSTEMS

II.1.1 Introduction

The water cycle is planetary and perpetual (continuous). Conducting hydrogeological studies requires dividing it into domains limited in space and durations accessible to observations and measurements (year, month, days). These dynamic domains are called "hydrological systems." Each hydrological system is a sequence of the water cycle, meaning it includes an input (impulse, example: infiltration), an internal circuit (water transfer for example between upstream and downstream) and an output (response to the impulse, example: spring flow rate).

II.1.2 Different types of hydrological systems

Three types of independent and nested hydrological systems can be distinguished:

a) Hydrological basin: It is limited by topographic ridge lines (summits of reliefs), delimiting the watershed of a watercourse and its tributaries. The single source of supply for the hydrological basin, assumed closed, comes from effective precipitation, that is to say precipitation that has escaped evaporation.

b) Hydrogeological basin: This is the fraction of the hydrological basin space located below the ground surface. It is the domain of groundwater. Its limits are imposed by geological structure. Its supply occurs through infiltration of part of the effective rainfall, having escaped surface runoff.

c) The aquifer: Is identified by the geological nature of the formations that constitute it (limestone, sandstone, sand...). It is fed by effective infiltration, and it corresponds to the study domain of groundwater. A hydrogeological basin can contain several aquifers.

II.1.3 Water balance concept

a) Water balance of a hydrological system

The water balance of a hydrological system is the accounting balance of inputs (receipts) equal to the average flow rate of inputs and outputs (expenses) represented by the average flow rate of flows. The balance refers to a domain limited in space and to a specific average duration (hydrological year for example).

The difference in water volume between inputs and outputs of the hydrological system generates a difference in water reserves (W). This difference can be null (balanced budget), positive (increase in reserves), or negative (decrease in reserves).

$$\text{Input flow rate} = \text{Outflow rate} + \Delta W$$

b) Aquifer balance

The balance in the influenced regime of an aquifer is written:

$$Ie + Qim = Qw + Qex + \Delta W$$

Where:

- Ie = Effective infiltration in m^3/year
- Qim = Imported flow rates in m^3/year
- Qw = Groundwater flow rates in m^3/year
- Qex = Exported flow rates in m^3/year
- ΔW = Reserve variations in m^3/year

II.2 CLIMATOLOGY

II.2.1 Introduction

Each aquifer system requires for its balance study a synthesis of climatic data, in order to place it in a hydroclimatic context and to evaluate groundwater recharge. For this, we rely on one or more meteorological stations representative of the study area and containing a large number of observations (rain, temperature, sunshine...).

II.2.2 Review of the water cycle

Precipitation (rain and snow) arriving at the earth's surface constitutes almost all water inputs to the soil. When rain reaches the soil, three processes begin:

- Soil humidification and infiltration
- Surface runoff
- Evaporation

a) Humidification and infiltration

In almost all countries where it rains, the subsoil normally contains water. A typical profile of the amount of water contained as a function of elevation is as follows:

This water content is of course a function of soil porosity and permeability. Below a certain elevation N, the water content no longer increases with depth. The soil is said to be saturated; all voids (pores) in the soil contain water. This water is said to belong to the water table. On the other hand, above elevation N, the soil is said to be unsaturated, the soil voids simultaneously contain water and air. In the saturated zone, water is subjected essentially to gravitational forces, while in the unsaturated zone capillary forces are added, which quickly become predominant.

When water falls on the soil surface, it begins by humidifying the upper fraction of the soil. This increase in surface humidity does not necessarily lead to immediate deep infiltration, because as long as capillary forces are greater than gravitational forces, water is retained, like in a sponge.

When the water content exceeds a certain limit, called specific retention capacity, water propagates downward and humidifies a deeper zone of the soil. If the rain continues long enough, humidification will become increasingly important and will lead to infiltration. But this phenomenon is very slow: depending on the depth of the water table under the soil and its permeability, water arrival at the water table can occur within the week following the rain, within the month, or even within six months.

b) Surface runoff

If the rain intensity is strong, the soil cannot absorb the water input and an excess appears on the surface. The surface water film can then circulate on the soil, this is called runoff. This runoff occurs along the line of greatest soil slope and feeds the natural drainage network: ditches, streams, rivers... It carries solid particles by erosion, which generates solid transport in rivers.

c) Evaporation

Even during rain, a significant part of the water reaching the soil is immediately re-evaporated. Once the rain stops, this evaporation continues and gradually dries the water that is intercepted by vegetation or remains on the surface. Another phenomenon that works in the same direction as evaporation on the soil is plant transpiration. Plant roots are capable of taking up soil water from the unsaturated zone, or sometimes from the saturated zone if it is outcropping.

II.2.3 Climate types

The Martonne index (1923) is based on rainfall and temperature regimes to characterize the climate of a region. Thus, according to the index value, it defines distinct climates.

Example: Chlef Station, period 1990-2010 (20 years) we have: P = 332 mm, T = 21°C, so A = 10.7. We can say that we are in a semi-arid environment.

II.2.4 Precipitation

a) General information

The term "precipitation" encompasses all meteoric waters that fall on the earth's surface both in liquid and solid form: snow, hail... These precipitations result from the condensation of water vapors contained in the atmosphere. The various types of precipitation are measured by their "water equivalent" using standard pluviometers.

In many hydrological studies (flood flow prediction for example), it is essential to know not only the total height of precipitation for a given period, but also the temporal distribution of the latter. Recording pluviometers are used for this purpose, which give the curve of cumulated precipitation heights as a function of time.

b) Exploitation of rainfall data

Statistical treatment of data collected from national meteorology allows qualifying a given region at different time scales. Thus, on a monthly time step, we can distinguish wet months and dry months. While on an interannual basis, we can get a precise idea of the regularity or not of the regime, and we can highlight cycles of wet or dry years.

To determine the water depth precipitated on a basin, three main methods are used: arithmetic mean method, Thiessen method and isohyet method.

II.2.5 Temperatures

Temperature is an important parameter for characterizing a given region. The average temperature over several years of observation allows knowing the coldest and warmest month, as well as the average annual temperature of the region. The variation amplitude between minimum and maximum temperature is also a characteristic of each region.

II.2.6 Rainfall-Temperature coupling

It is interesting to couple rainfall and temperature on the same graph. The shape of the latter characterizes the climate of a given region. The Ombrothermic curve is a curve with three inputs: temperature, rainfall and time. It shows the dry and wet periods of an average year. The Climagram is also a graph whose shape is specific to a given region; it distinguishes between dry and wet months located on the graph on either side of the Gaussen line.

II.2.7 Evapotranspiration

This is an important parameter in the hydrological cycle, resulting from a physical phenomenon (evaporation) and a biological one (transpiration). It is a function of climate and its variations which are determined by air humidity, wind speed, water and air temperature, vegetation cover, solar radiation and atmospheric pressure.

a) Potential evapotranspiration (PET)

This is the water depth that a soil can theoretically lose. It only takes climate into account and does not involve precipitation. It can be measured by evaporometers. Its monthly estimation is often made using Thornthwaite's empirical formula.

The values thus calculated should be multiplied by a corrective term, according to the month and latitude. For Benairia station (1990-2010) we have detailed monthly calculations showing corrected PET values.

b) Actual evapotranspiration (AET)

This is the water depth actually lost by a soil under real conditions of soil water content. Its evaluation can be done on an annual scale using Turc and Coutagne formulas, or on a monthly scale using Thornthwaite's water balance.

Thornthwaite water balance: The soil will evaporate according to its degree of saturation. Effective infiltration (aquifer recharge) occurs when AET and soil saturation are satisfied. We arbitrarily set an easily usable reserve (RFU) of 100 mm. This water balance is more reliable than previous methods, since we work on a monthly scale that allows taking into account certain influences that can be masked on an annual scale.

The Benairia station example (1990-2010) (Chlef) shows that the water table has a recharge period of three months, from January to March. AET = 322 mm, or 35% of P.

II.2.8 Effective infiltration

If we consider that runoff is negligible, water infiltration is obtained by:

$$I = P - AET$$

CHAPTER III - AQUIFER SYSTEMS

III.1 - WATER IN TERRAINS

III.1.1- General information

All terrains contain a certain percentage of voids. Water from rain or surface circulations can penetrate into these voids, circulate there under the effect of gravity, and under certain conditions, accumulate there. This presence of water in soils and subsoils is of great importance either because it represents drinking or industrial water reserves, or because it always poses delicate problems for engineers called to build in depth.

III.1.2- Porous medium

A porous medium is called a body comprising a solid skeleton encompassing cavities called pores, generally interconnected, capable of containing one or more fluid phases. A soil is essentially formed of three types of rocks:

a) Loose granular rocks

The voids consist only of pores that characterize a continuous medium. For these rocks, we speak of interstitial porosity. For example, sands and sandstones have a total porosity that can reach up to 30% and even rocks that are assumed to have lower porosity can contain significant amounts of water.

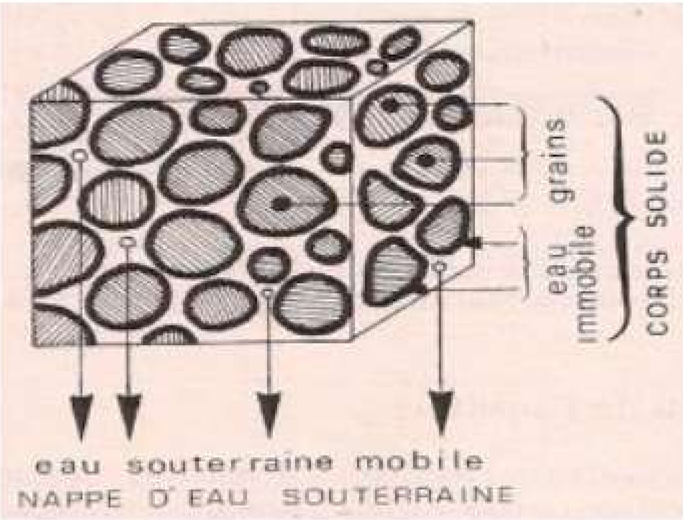


Figure 2: Sandy medium

Clays constitute a separate category; their lamellar constituent elements are organized in sheets. These are stacks of parallel layers separated by variable intervals where a fluid can lodge, which gives them, in particular, swelling properties in the presence of water. The percentage of voids can be very high, up to 90%.

b) Compact fractured rocks

A particular case of voids in compact rocks is fracturing, which characterizes the discontinuous medium. Through tectonics, almost all rocks of the Earth's crust are fractured (faults, fissures, joints). These fissures generally organize in at least two main fracturing directions that cut the rock into blocks. If the fissures are not clogged (clay, calcite, quartz...), voids are created and we then speak of fracture porosity.

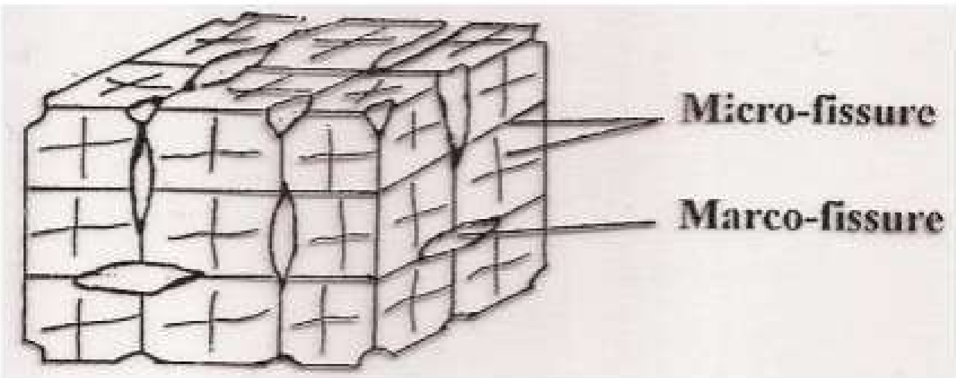


Figure 3: Fractured carbonate medium

Fissures are elongated cracks with more or less wide openings. They are classified into two types according to their dimension: micro-fissures whose hydrodynamic role is comparable to that of pores, and macro-fissures represented by faults, dislocations, and karst channels.

c) Mixed rocks

These are rocks whose voids consist of both pores and fissures. The two types of porosity (interstitial and fracture) coexist (example: sandstone, chalk, limestones).

III.1.3 Porosity

In a medium, there are three types of water: gravitational water that flows, retention water that remains around grains (humidity), and absorption water (bound to the grain surface by molecular attraction forces). The capacity to recover water in a loose or fractured rock is linked to the importance of its voids.

Two Types of Porosity

a) Total Porosity (n):

- $n = \frac{\text{Volume of voids}}{\text{Total volume}} \times 100$ (in %)

- Represents all empty space in the rock

b) Effective Porosity (ne):

- $ne = \frac{\text{VoVolume of gravitational waterlume of voids}}{\text{Total volume}} \times 100$ (in %)

- This is the more important measure for hydrogeology
- It only counts the voids where water can actually flow

This effective porosity interests the hydrogeologist. It is useful to relate it in the case of loose rocks to the physical characteristics of reservoirs. The main factors are:

- The respective diameters of grains: for uniform particle size distribution, it does not decrease when grain diameter decreases
- The homogeneity of particle size distribution: if the terrain is formed of grains of very different sizes, the smallest among them can occupy the interstices between the largest and porosity is considerably reduced
- The arrangement of grains: expresses their spatial disposition. Porosity is strongly influenced by grain arrangement. It decreases from 47.6% for cubic arrangement to 25.9% for rhombohedral arrangement
- The specific surface area of grains: this is the water-grain contact surface. Effective porosity increases with the specific surface area of grains. One consequence is the decrease in porosity with depth.

c) Orders of magnitude of porosity

These values can vary depending on grain size, consolidation, and medium compaction. Porosity is measured directly in the laboratory (poses the problem of sample representativity) and indirectly in the field by various geophysical methods (resistivity, sound velocity, well logging...).

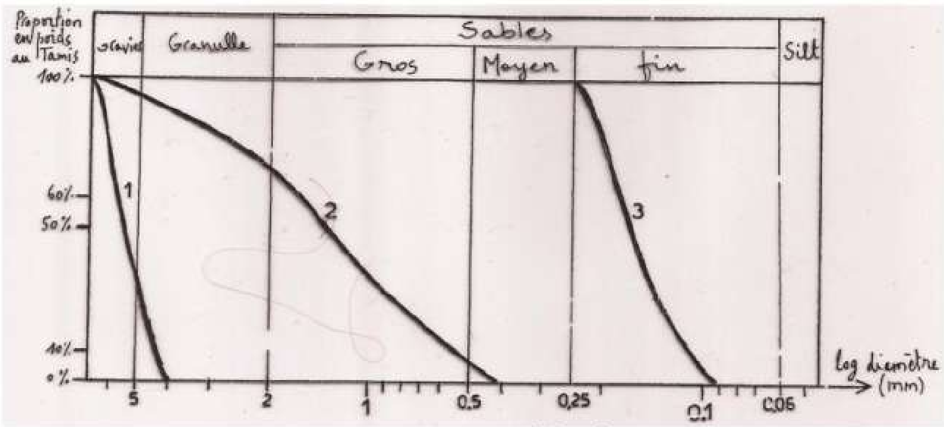
III.1.4 Permeability

It is appropriate to complete the notion of porosity with that of permeability, as it should not be forgotten that porosity value is not proportional to void dimensions. Example: clay, whose voids are microscopic, is much more porous than most other terrains. It must therefore be specified that a porous formation is not necessarily permeable. However, a permeable formation is, by definition, porous.

Permeability is a quantity that characterizes the ease with which water circulates in a terrain under the effect of a hydraulic gradient. Permeable materials offer more or less resistance to fluid passage. There exists a whole range of permeabilities, from practically impermeable media to those in which fluid circulation occurs almost without energy loss.

III.1.5 Grain Size Analysis

For unconsolidated media (sands, gravels...), we seek to know the grain size distribution of the medium and sediment classification. To perform grain size analysis, a terrain sample (about 500 g) is collected and dried in an oven before weighing. Then, the sediment is passed through a series of sieves of decreasing size, all being shaken by an electric agitator located at the base. The refuse (sediment fraction collected) from each sieve is weighed separately, and the weight is transformed into a percentage of the total weight of the initial sample. A very clear representation of the results is the cumulative frequency curve (in %), plotted as a function of the logarithm of sieve mesh size.



1 – Gravier uniforme. 2 – Gravier sableux hétérogène. 3 – Sable fin uniforme

Figure 4: Grain size analysis diagram

The grain size curve allows to:

- Study the statistical distribution of grain diameters
- Classify the sample and designate it by a precise lithological term (gravel, sand, clay...)

- Have a precise idea of sample homogeneity or heterogeneity from the curve slope, which is represented by the uniformity coefficient:

$$CU = d_{60}/d_{10}$$

(if CU < 2, the grain size distribution is uniform)

- Estimate sample permeability by different empirical formulas based on the characteristic index d_{10} (diameter corresponding to the 10% ordinate of the cumulative curve), which most conditions the medium's permeability properties. The most used formulas are:

- **Hazen Formula**:

$$\mathbf{K\ (cm/s) = A.\ (d_{10})^2\ (cm)}$$

A being a coefficient that varies according to sediment grain size distribution. The most used value in hydrogeology is A = 100 (for grains: 0.1 < d < 3 mm and for CU < 5).

- **Schneebeli Formula**:

$$\mathbf{Log_{10}\ K\ (cm/s) = 2\ log_{10}\ d_{10}\ (cm) + 2}$$

This formula applies to clean sands with round grains.

- Size screen openings and choose additional gravel dimensions during water exploitation drilling equipment.

III.1.6 Notions of Isotropy and Homogeneity

A homogeneous terrain is terrain that presents at every point in a given direction, the same resistance to fluid flow. If in addition this resistance is the same regardless of direction, the terrain is isotropic.

III. 2 AQUIFER SYSTEMS

III.2.1 Definitions

- Aquifer: layer of permeable rocks comprising a zone sufficiently conductive of groundwater to allow significant flow of a groundwater table and capture of appreciable quantities of water by economical means.

The aquifer is homogeneous when it has interstitial permeability (sand, gravels); percolation velocity is slow. It is heterogeneous with fracture permeability (granite, limestone); percolation velocity is faster.

- Aquifer table: ensemble of water contained in the saturated zone of an aquifer, all parts of which are in hydraulic connection.

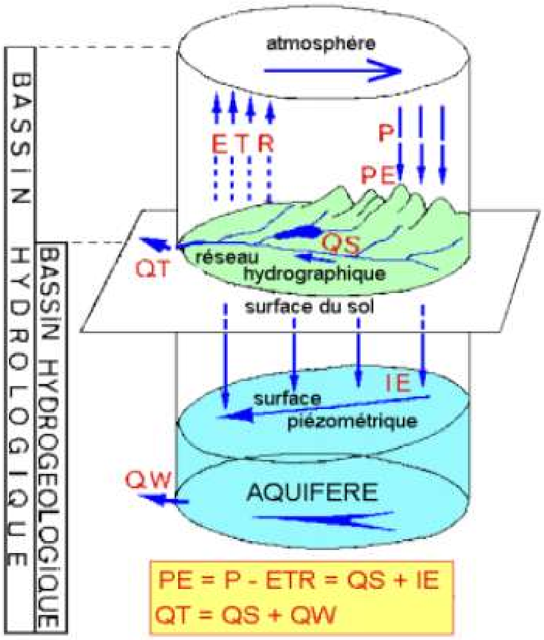


Figure 5: Hydrological basin, hydrogeological basin and aquifer

III.2.2 General Process of Water Table Formation

Through gravity, part of rainwater infiltrates into the soil, either directly or after circulation on its surface. Depending on the permeability of encountered terrains, it descends more or less deeply. This approximately vertical circulation is interrupted by encountering terrain of low permeability. At the base, this impermeable formation represents the watertight floor of the water table. Water accumulates there by saturating all voids of the overlying more permeable terrains. Thus, an aquifer table forms in these formations.

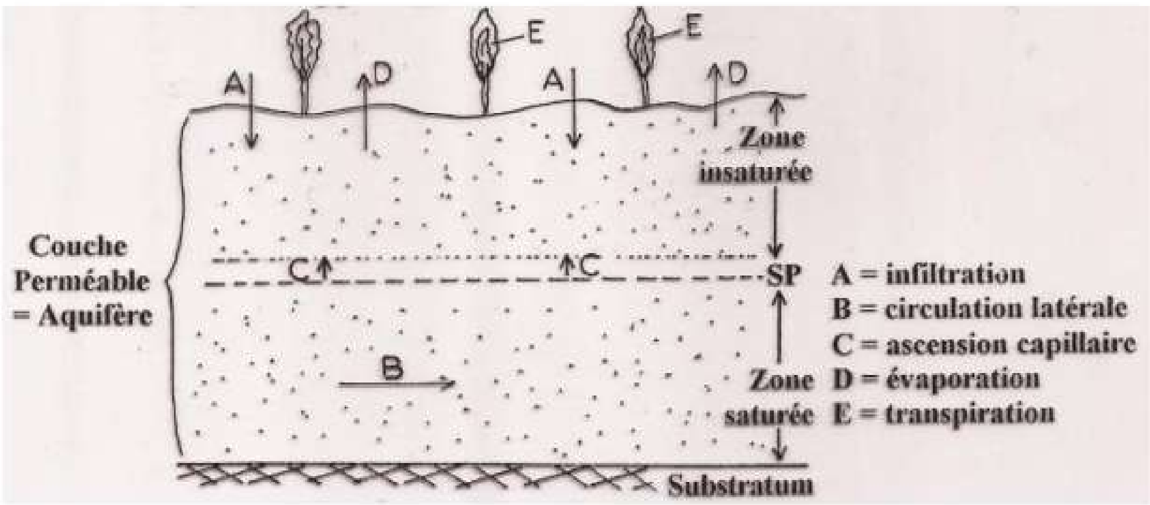


Figure 6: Aquifer table

III. 2.3 - Different Types of Water Tables

III.2.3.1 - Unconfined aquifer: An unconfined aquifer is a water table contained in a partially saturated permeable layer resting on an impermeable or semi-permeable layer. The free surface is always at atmospheric pressure (direct communication with free air through interstices).

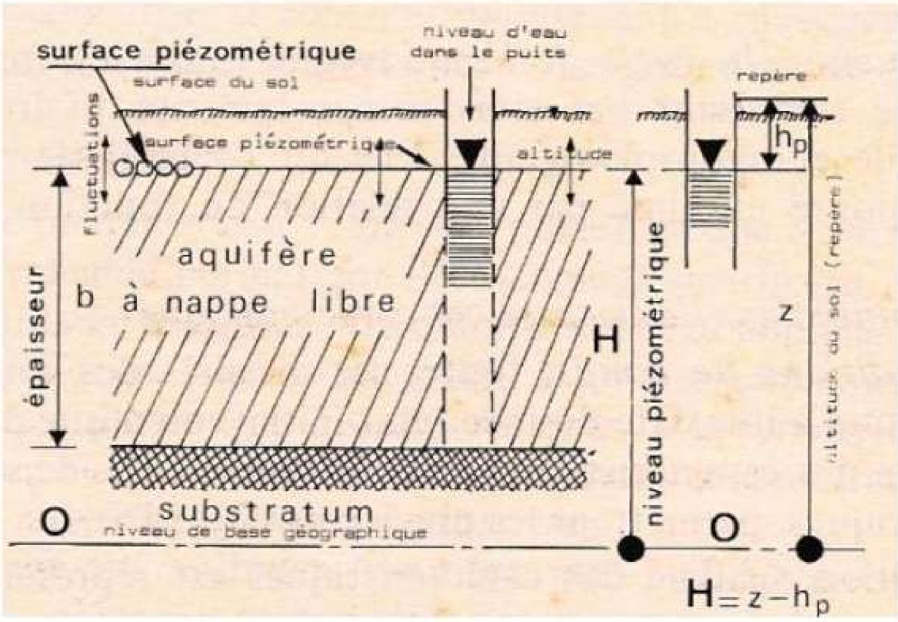


Figure 7: Unconfined aquifer schema

a) Valley aquifer: refers to a water table whose drainage occurs solely through valleys. Water flows toward outlets which are low points in topography (springs, rivers...). In arid countries, in valleys, floods of temporary wadis bring much water that can infiltrate and feed the water table; this is their main source of supply.

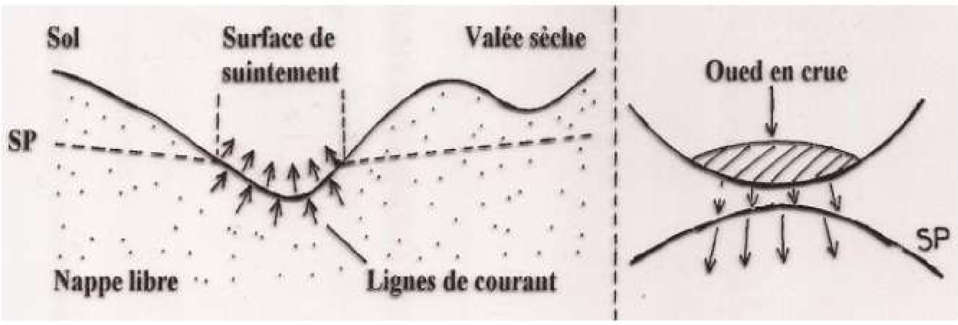


Figure 8: Valley aquifer

b) Alluvial aquifer: this is an unconfined aquifer located in alluvium that lines a river course. The thickness of alluvial filling can be important, with coarse materials (sands, gravels, pebbles) that are very permeable. These materials are saturated almost to the soil surface. Table water is generally in equilibrium with river water, being sometimes drained by the river, sometimes fed by it. This type of water table is also called supported. It is very vulnerable to pollution.

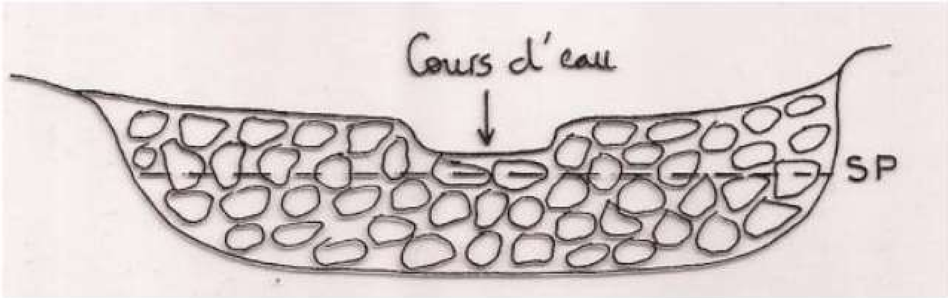


Figure 9: Alluvial aquifer

c) Coastal aquifer: the continental freshwater table is in hydrostatic equilibrium with the saltwater table from seawater. These 2 water tables mix little; their interface constitutes a salt wedge. Any drawdown of the freshwater table causes equilibrium rupture and progression of the salt wedge toward the interior of the land.

d) Karst aquifer: in limestone country, water charged with atmospheric carbon dioxide attacks the rock, continuously enlarges fissures, creates galleries, caverns and chasms, resulting in true underground rivers. Water circulation velocities in karst channels are high and springs can be abundant (resurgences).

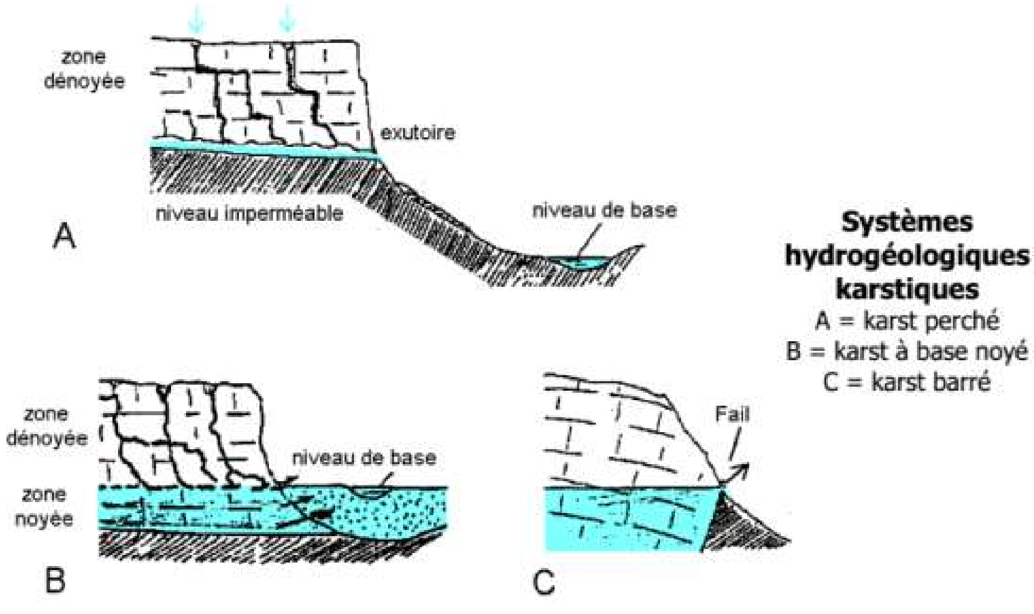


Figure 10: Karst hydrogeological systems

III. 2.3.2 - Confined aquifer

A confined, pressurized, or artesian aquifer is called a water table contained between two impermeable geological formations. The aquifer roof is thus maintained below the piezometric surface. If the aquifer roof is pierced, water rises and establishes at a level according to the pressure to which it is subjected. At the limit, we have artesian drilling. This artesianism can however disappear with time if the water table is exploited, by reduction of pressure in the aquifer.

Figure 11: Confined aquifer schema

III. 2.3.3 - Semi-confined aquifer

An intermediate case between the two types of water tables is the semi-confined aquifer. There is water exchange with the overlying or underlying aquifer: this is the leakance phenomenon. It requires two conditions: the existence of a semi-permeable formation and the existence of a head difference h. Water flows from the aquifer having the strongest hydraulic head toward the one with the weakest hydraulic head.

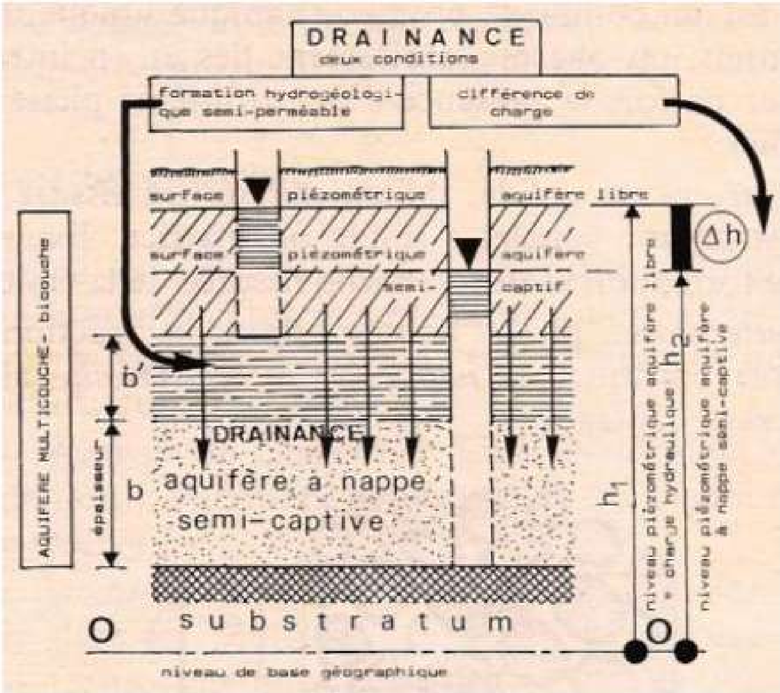


Figure 12: Semi-confined aquifer schema

III.2.4 Emergences (springs or outlets)

Aquifer tables are fed by infiltration water from the surface and empty through outlets, essentially springs. These emergences are imposed by the geological structure of the aquifer and site geography. A spring's flow depends on the type and richness of the water table that feeds it, and there is direct correspondence between this flow and the water table's hydraulic head. Springs can be classified according to their structural position.

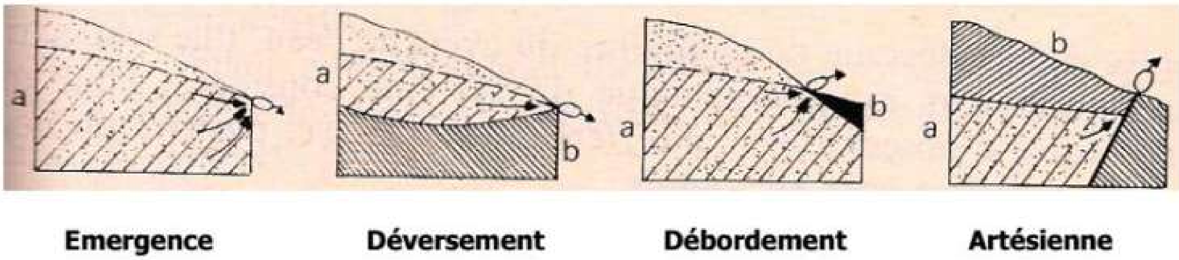


Figure 13: Emergence (springs or outlets)

This document provides a comprehensive overview of hydrogeological principles, covering water infiltration processes, aquifer systems, and the various types of groundwater formations found in different geological settings.

CHAPTER IV - AQUIFER MAPPING

Aquifer mapping aims to represent its geometry (or configuration), structure, and to schematize the functions of the reservoir (storage and conduction) and its hydrodynamic behavior. There are two types of maps: structural and piezometric.

IV.1 STRUCTURAL MAPS OF THE AQUIFER

Structural maps of the aquifer represent the morphology, the position of boundary surfaces, the thicknesses necessary for volume calculations, and the spatial distribution of hydrodynamic parameters. These maps are created by synthesizing data on geology, boundary conditions, and physical (particularly lithology and granulometry) and hydrodynamic parameters of aquifers (results from pumping tests).

IV.2 PIEZOMETRIC MAPS

IV.2.1 Introduction

Piezometric maps, created using water level values in wells, represent the spatial distribution of hydraulic heads in an aquifer at a given date. These maps are the basic documents for analyzing and schematizing the capacitive and conductive functions of the reservoir and the hydrodynamic behavior of the aquifer. The map of the piezometric surface of a water table is a study and exploitation tool for this water table. It directly defines the flow conditions of liquid filaments circulating on the surface of the water table. It also allows the study of fluctuations over time in piezometric levels and thus the evaluation of the water table's reserve and recharge conditions.

IV.2.2 Creation of Piezometric Maps

This is based on measuring piezometric levels, plotting them on topographic maps as contour lines, and interpreting them using isopiezic curves.

a) Measurement and Plotting of Levels on a Map

Measurements of water table depths are carried out using an electric probe in wells or piezometers under stable water table conditions, outside periods of heavy rain or pumping, and over the shortest possible period. Indeed, this document has a reference value at a given date due to seasonal and multi-year fluctuations.

The depths of water measured in water structures are converted into piezometric levels in NGM elevation (Piezometric Level = Elevation - Water Depth/Soil) and plotted on a topographic map of appropriate scale. The greater the density of points, the larger the scale, and vice versa.

b) Drawing Isopiezic Curves

For drawing piezometric or isopiezic curves, the graphical interpolation method of triangulation is used. Data is grouped in threes at the vertices of triangles and connected by straight segments. Each side of the triangle is divided into proportional segments.

Isopiezic curves are obtained by connecting points of equal piezometric level with straight segments. The curves are then smoothed to obtain regular curves.

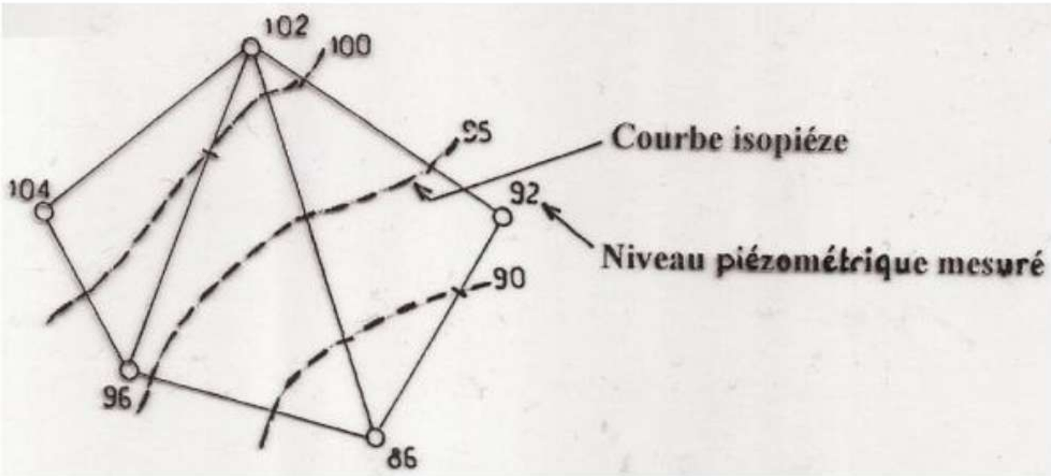


Figure 17: Drawing Isopiezic Curves

The distance between isopiezic curves (variation in level between two successive curves) must be adapted to the problem being studied and depends on the density of measurement points, the hydraulic gradient, and the map scale. Generally, the distance will decrease if the density of points or the map scale increases. Conversely, if the hydraulic gradient increases, the distance will also increase.

c) Drawing Flow Lines

Isopiezic curves are equipotential curves for liquid particles; they correspond to curves of equal elevation of the piezometric surface. Flow lines are perpendicular to equipotential curves that they intersect. They represent the trajectories taken by groundwater during its circulation.

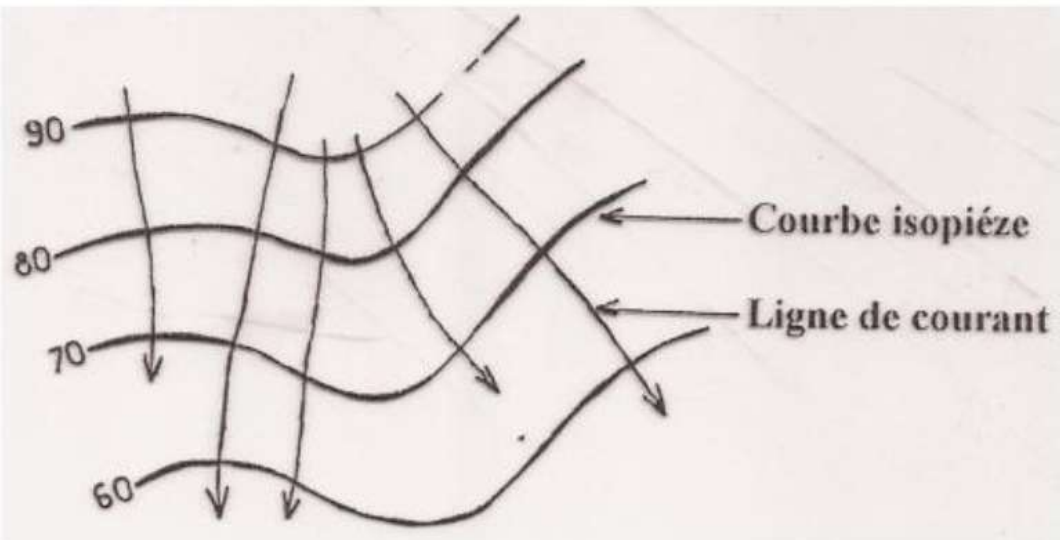


Figure 18: Drawing Flow Lines

IV.2.3 Interpretation of Piezometric Maps

The piezometric surface of free water tables is represented by contour lines called isopiezic curves (like any topographic surface), meaning curves of equal altitude (e.g., NGM elevation) of its top. For confined aquifers, the approach is identical; it should simply be noted that the piezometric surface no longer coincides, as in the free water table, with the top of the aquifer due to the pressurization of water by the impermeable cover. The piezometric surface is located above this impermeable cover.

a) Direction of Water Table Flow

Flow lines indicate the general direction of water table flow, which moves from upstream (high hydraulic potential) to downstream (low hydraulic potential). Drawing flow lines also allows the identification of two main axes of the piezometric surface:

- The drainage axis: an axis of convergence of flow lines; it represents a privileged flow sector of the water table (water-rich sector).
- The groundwater flow divide axis: an axis from which flow lines diverge; it represents an unfavorable sector of the water table. This axis also constitutes a limit of a hydrogeological sub-basin.

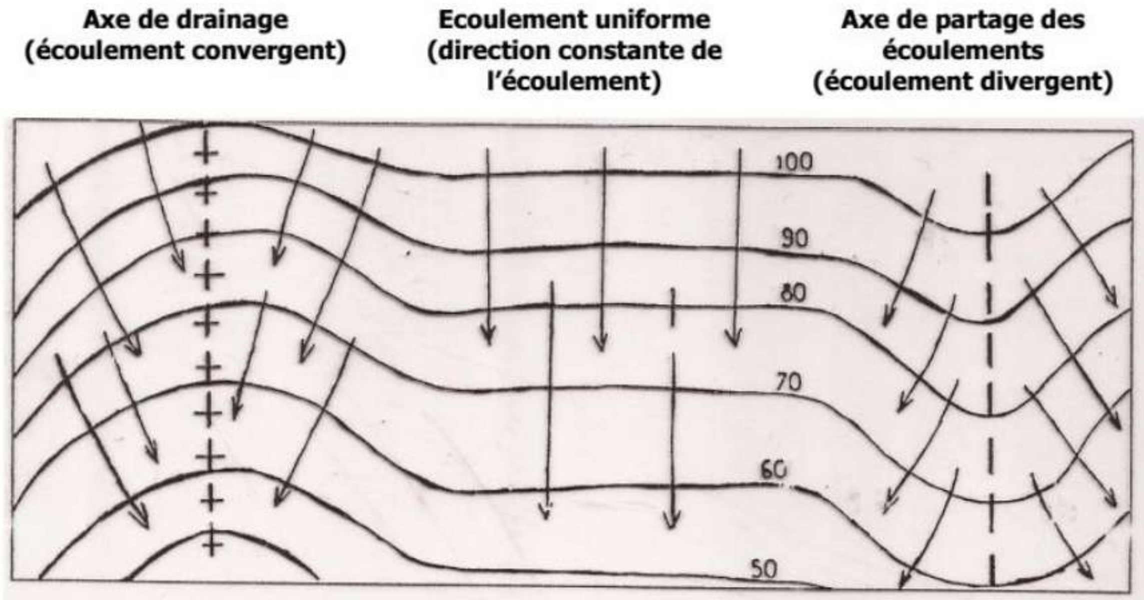


Figure 19: Direction of Water Table Flow

b) Calculation of the Hydraulic Gradient

Calculating the hydraulic gradient is very useful for understanding the functioning of the water table. It is calculated along a flow line and is equal to the ratio of the distance between isopiezic curves to the distance between isopiezic curves on the scale: **i = (h1 - h2 / L) * 100**. The spacing of isopiezic curves immediately and visually informs about the value of the gradient. The closer the curves, the higher the gradient, and vice versa. According to Darcy's law (Q = T, i, L), the hydraulic gradient is inversely proportional to the Transmissivity (K × e) of the water table. Areas of low hydraulic gradient on a piezometric map are the most interesting in terms of productivity and, consequently, the most favorable for the installation of exploitation structures.

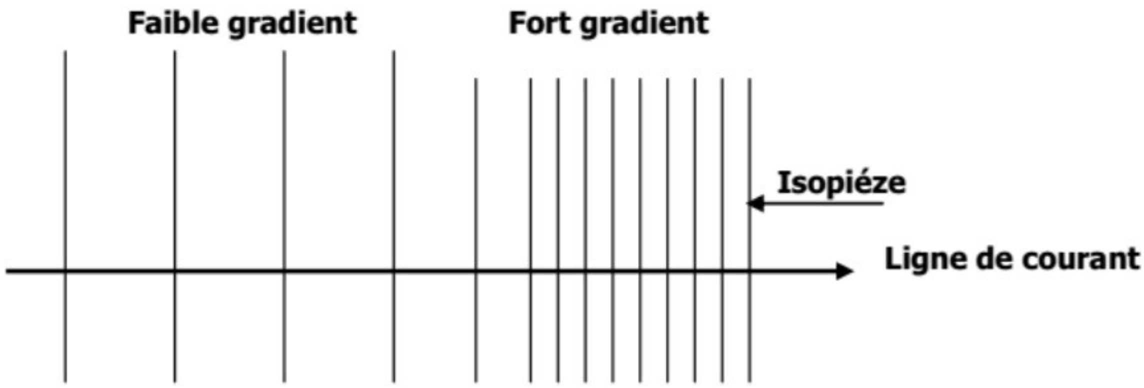


Figure 20: Hydraulic Gradient

c) Boundary Conditions

Analyzing the shape of piezometric curves and local geological conditions allows the identification of recharge and drainage zones of the water table. If isopiezic curves are perpendicular to the boundaries of the aquifer, they identify an impermeable boundary. Conversely, if the isopiezic curves are oblique or parallel to the boundaries of the aquifer, they identify a recharge or discharge boundary of the water table, depending on the direction of flow.

Far from the boundaries of the aquifer, if the isopiezic curves are closed, they identify, according to the direction of groundwater flow, localized recharge zones (piezometric domes) or drainage zones (piezometric depressions).

d) Hydraulic Relationships Between Water Table and Watercourse

Between an aquifer and the watercourse crossing it, there can be hydraulic relationships of drainage or recharge of the water table by the watercourse.

Drainage of the water table by the watercourse is frequent during low water periods. The water table flows towards the watercourse and emerges at the level of springs located in its bed. The piezometric surface of the water table is at a higher elevation than that of the watercourse. The isopiezic curves form arcs of circles with concavity oriented towards the hydraulic downstream of the water table. The flow lines converge towards the watercourse. The watercourse can, in turn, recharge the water table during flood periods. In this case, the flow lines diverge from the river towards the water table, and the concavity of the isopiezic curves is oriented towards the hydraulic upstream of the water table. The water level in the watercourse is at a higher elevation than that of the water table.

e) Piezometric Monitoring

In many hydrogeological basins, certain water points are regularly monitored (monthly to bi-monthly measurements). These measurements aim to provide a precise idea of the fluctuations in the piezometric level across the entire water table on a seasonal and multi-year scale. This monitoring allows, for example, continuous visualization of:

- The response time of the water table to rainfall inputs
- The long-term consequences of operating a capture field
- The consequences of a long period of drought
- The consequences of external input (irrigation by water from a dam)

CHAPTER V - FUNDAMENTAL LAW OF GROUNDWATER FLOW

The conduction function of an aquifer allows the transport of groundwater and the transmission of influences. It is imposed by the structure of the aquifer: geometric and hydrodynamic parameters. Only gravitational groundwater participates in the flow and is subject to the laws of groundwater hydrodynamics.

V.1 DARCY'S LAW

V.1.1 Laboratory Device with Lateral Flow

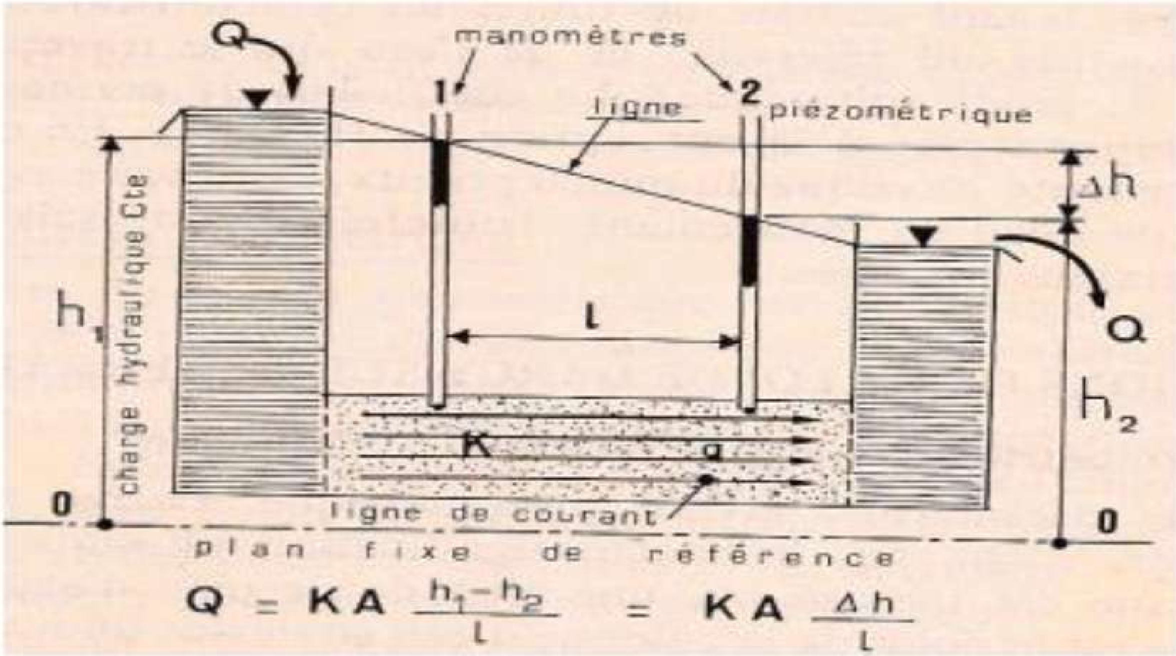


Figure 21: DARCY's Experimental Device

V.11.2 Statement of the Law

The fundamental basis for calculating quantities of groundwater or flow rate of an aquifer, through groundwater hydrodynamics, is DARCY's experimental law (1856) which showed that: the volume of water Q in m^3/s, filtering through the sand column of length l in m, through the cross-section A in m^2, is a function of a proportionality coefficient K in m/s, characteristic of the formation, and the head loss per unit length of the cylinder h/l (dimensionless). The term K is called the permeability coefficient. It has the dimension of a velocity. It materializes the groundwater circulation function.

V.1.1.3 Validity Conditions

DARCY's law is established by laboratory experiments meeting very strict conditions. Four conditions must be met for the law to be applicable: continuity, isotropy, homogeneity of the reservoir, and laminar flow.

- **Continuity** is the characteristic of a permeable medium having interconnected voids in the direction of flow. Example: sand, sandstone, alluvium, gravel, limestone with microfractures...
- **Isotropy** refers to a medium in which physical characteristics (particularly grain size) are constant in all three directions of space. Otherwise, the medium is said to be anisotropic.
- **Homogeneous** when its physical characteristics are constant at all points in the direction of flow. Otherwise, the medium is said to be heterogeneous.
- **Laminar** flow is characterized by continuous, rectilinear, individualized flow lines occupying the same relative position between them. The water flow velocity is constant and is lower than the critical velocity beyond which turbulent flow appears (head loss not proportional to flow rate).

Reminder: the limit of laminar flow is defined by the REYNOLDS number in porous media. Re is a ratio of inertial forces to viscous forces. In laminar flow, friction forces are very important compared to inertial forces, resulting in a small Re. Conversely for turbulent flows.

The validity conditions of Darcy's law may seem very restrictive if we consider the numerous lithological variations of hydrogeological formations (stratification, lateral facies changes, schistosity...). But in reality, cases where DARCY's law is not applicable are limited to very heterogeneous formations, karst networks, and when the flow velocity is very high.

V.2 HYDRODYNAMIC PARAMETERS

V.2.2.1 Permeability

Permeability is the ability of a reservoir to be crossed by water under the effect of a hydraulic gradient. It expresses the resistance of the medium to the flow of water passing through it. It is measured by two parameters: the permeability coefficient and intrinsic permeability.

a) Permeability Coefficient

This coefficient, denoted K, is defined by Darcy's law: $K = Q / A . i$. It has the dimension of a velocity and is expressed in m/s. All materials conduct water to varying degrees. The values of the permeability coefficient range from 10 to 1×10^{-11} m/s and by convention we can distinguish three types of formations:

- **Permeable**: $K > 1 \times 10^{-4}$ m/s. Example: Gravel, coarse sand...
- **Semi-permeable**: $1 \times 10^{-4} > K > 1 \times 10^{-9}$ m²/s. Example: clayey sand, fine sand
- **Impermeable**: $K < 1 \times 10^{-9}$ m²/s. Example: clay.

b) Intrinsic Permeability

This is the parameter that characterizes the intrinsic permeability of the aquifer formation independent of the fluid characteristics. This geometric permeability, denoted k, depends on the granulometric characteristics of the terrain (effective diameter, grain surface, and effective porosity) and is expressed in m^2 or in darcy.

Intrinsic permeability is related to the permeability coefficient by the relation:

- Decreases rapidly with temperature.
- K is an inverse function of viscosity and increases with temperature.

One consequence is the increase of K with depth (effect of the geothermal gradient).

- The determination of the permeability coefficient K is done at the conventional temperature of 20°C.
- In hydrogeology, it is generally admitted that K = k.

V.2.2.2 Transmissivity

The production of a capture in an aquifer is a function of its permeability coefficient K and its thickness e. This is why a new parameter, transmissivity, denoted T, has been created. It evaluates the conduction function of the aquifer.

- Inversely proportional to the hydraulic gradient i of the aquifer.
- Including the thickness of the aquifer, transmissivity allows representation on maps of productivity zones.
- Long-duration pumping tests.
- Between 1×10^{-4} and 1×10^{-2} m^2/s for porous media and 1×10^{-2} and 1×10^{-1} m^2/s for fractured media.

V.2.2.3 Diffusivity

The diffusivity D of an aquifer is the ratio of transmissivity to the storage coefficient. It is expressed in m^2/s and governs the propagation of influences in the aquifer (variation of hydraulic head or pressure, pollution transmission...). It is much more important in confined aquifers (low S) than in unconfined aquifers (high S).

V.2.2.4 Water Flow Velocities in an Aquifer

a) Filtration Velocity Vf

The filtration velocity V calculated by DARCY relates to the total cross-section (A) of the flow. It has no physical reality.

b) Effective Velocity Ve

In an aquifer, only gravitational water moves between the grains of the formation. The effective flow surface is thus reduced to the voids created by the solid body (grains + retention water) and therefore depends on the effective porosity ne. The expression of DARCY's law corrected, related to the effective cross-section for calculating the effective velocity Ve is therefore:

The effective flow cross-section is smaller than the total cross-section A. Therefore, at constant flow rate, the effective velocity Ve is greater than the filtration velocity Vf. The effective velocity approaches the real velocity of water displacement measured in the field by tracer techniques.

NB: The calculation of effective velocity is very important for calculating the transfer time of pollution between two points in the aquifer.

V.2.2.5 Storage Coefficient

a - Definition

The storage coefficient, denoted S (dimensionless), is the volume of water released or stored per unit surface area of the aquifer (1 m²), following a unit variation of the hydraulic head h. This coefficient represents the capacitive function of the reservoir, which is characterized by the storage or release of groundwater. It is expressed in % and is determined in the field by pumping tests.

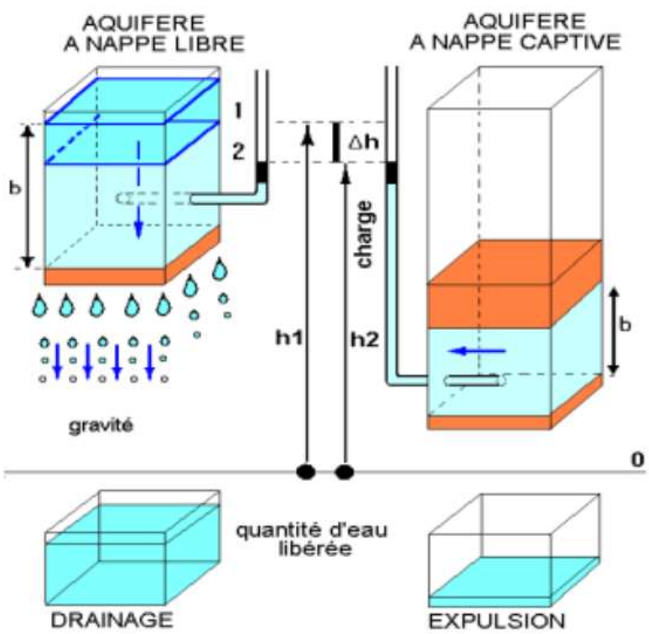


Figure 22: Head Variation and Volume of Water Released

b - Case of Unconfined Aquifers

In the case of an unconfined aquifer, the storage coefficient corresponds to the quantity of gravitational water released under the action of the force of gravity. It is comparable to effective porosity and generally varies between 1 and 25%. The release of water in an unconfined aquifer is explained by the replacement of part of the water contained between the grains by air.

c - Case of Confined Aquifers

In the case of a confined aquifer, air has no access to the aquifer. The storage coefficient corresponds to volumes of water extracted by decompression of the aquifer formation (expansion of water and deformation of the aquifer rock). The elastic moduli being low, the volume of water released is much smaller with equal characteristics than in unconfined aquifers. It generally varies between 1×10^{-4} and 1×10^{-3} (0.01% and 0.1%). The expansion of water is insufficient to justify the volume of water extracted from a confined aquifer, for a given variation in drawdown. The compaction of the aquifer must be added.

This S effect can have serious geotechnical consequences. Example: compaction corresponds to ground subsidence and even minimal elevation changes can cause cracks in a building, part of which would rest on incompressible rock. In addition to damage to buildings and other structures, reactivation of landslides can sometimes be observed.

V.3 DETERMINATION OF PERMEABILITY COEFFICIENT

In this section, we will only mention methods used in the laboratory. It must be kept in mind that these methods do not allow correct measurement of the permeability of all terrains. To estimate the permeability of a terrain as a whole, in situ methods (particularly pumping tests) are used.

To estimate or measure permeability in the laboratory, it is necessary to collect a terrain sample. This sample, of small size, will not be representative of the entire aquifer:

- The characteristics of the terrain will be modified due to sampling;
- The sample will not allow taking into account permeability variations due to faults;
- The aquifer will generally be sampled at the outcrop (where the terrain is modified by weathering). To constitute a characteristic sample, samples would need to be taken at different levels of the aquifer, which is difficult to achieve and would be costly;
- The sample will not be in the conditions of pressure, adjacent forces, and temperature that were originally its own and which are difficult to evaluate.

These techniques for measuring or estimating permeability in the laboratory are in fact more used by soil mechanics engineers than by hydrogeologists (indeed, if working with disturbed soils, as are the samples, the order of magnitude of permeability provided may be acceptable).

Estimation of Permeability

Laboratory permeability can be estimated from grain size (Hazen relation cited above or Casagrande relation) in the case of unconsolidated rock:

Casagrande Relation:

For coarse-element soils (> 1 mm) whose grains are assumed to be cubic, permeability can be expressed as a function of the void ratio e:

$$K = 1.4\, K_{0.85} \cdot e^2$$

$K_{0.85}$ is the permeability for e = 0.85. It is therefore sufficient to determine the permeability corresponding to an arbitrary value of e and the K values corresponding to other e values are obtained by means of the equation.

These relations do not take into account the shape of grains. They should only be used for the specific cases for which they have been defined. In practice, they are unusable for natural terrains that have different and more complex structures than the studied soils.

V.4 Permeameters

Permeability can also be measured using a permeameter on a terrain sample.

a) Constant Head Permeameter

Used especially for granular materials such as sands. In the constant head permeameter, a constant head difference is maintained between the inlet and outlet. It involves measuring the volume of water that flows in a time interval through the sample. Conductivity is determined by Darcy's law:

From which an expression for hydraulic conductivity is written:

If flow rate Q is expressed as the volume of water V that flows in the system in time t, if the head loss through a sample of length L equals Δh, and if A is the area of a cross-section normal to the flow, then hydraulic conductivity can be directly deduced from Darcy's law and is:

For the permeameter test, it is important that the hydraulic gradient be as close as possible to field values. The head difference Δh should not exceed half the length of the sample. With higher gradients, the test is shorter but there may be turbulent flow and Darcy's law is no longer valid.

Feeding the system from below is recommended to expel air bubbles that may have been entrained by the flow. Too strong an upward gradient can also create quicksand situations that are no longer representative of field conditions.

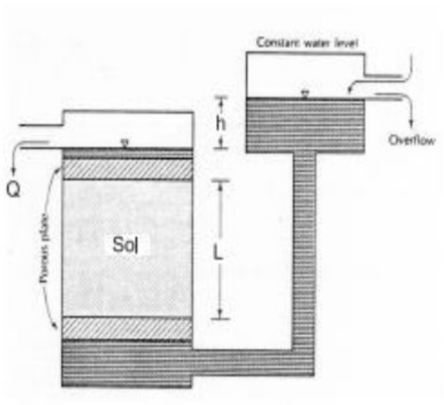


Figure 23: Constant Head Permeameter

b) Falling Head Permeameter

Used for cohesive materials of lower hydraulic conductivity (silt, clay, till). Because of low flow rates and very long test duration, precise determination of volume and level changes as a function of time in a smaller tube that feeds the permeameter is attempted. At the beginning of the test, the initial hydraulic head h_0 is noted. After a given time interval (which may be several hours), the water level h is measured again. The inside diameter, dt, of the supply tube, the diameter dc of the sample, and the length L of the sample must be measured. Hydraulic conductivity is calculated using the following formula:

In permeameter tests, it is important that the sample be completely saturated. The presence of air bubbles trapped in the porous medium reduces hydraulic conductivity. The sample must also be in close contact with the walls to avoid a preferential path that would artificially increase the value of K.

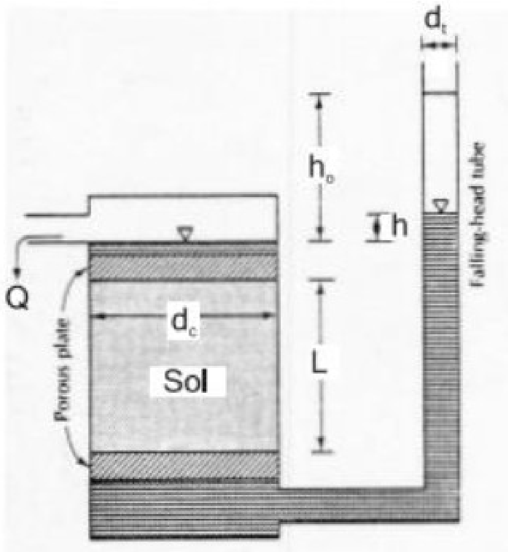


Figure 24: Falling Head Permeameter

c) Equivalent Permeability of Superposed Layers

Permeability K depends on the material. Suppose an anisotropic material formed by the superposition of horizontal layers of thickness ei and permeability Ki; the flow occurs at velocity V which can be decomposed into VH + VV.

- Horizontal Equivalent Permeability

Through each stratum flows a discharge:

The total discharge is therefore:

but the total discharge can also be written:

Therefore:

$$K_H = \frac{\sum_{i=1}^{n}(K_i\, e_i)}{\sum_{i=1}^{n} e_i}$$

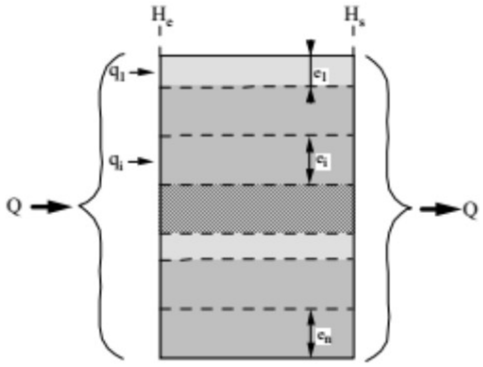


Figure 25: Horizontal Equivalent Permeability

- Vertical Equivalent Permeability

The vertical discharge that passes through all layers:

is also the discharge that passes through each layer. We can also write:

Hi-1 and Hi being the hydraulic head respectively at the top and base of layer i.

Now:

Therefore:

Thus:

$$K_v = \frac{\sum_{i=1}^{n} e_i}{\sum_{i=1}^{n} \frac{e_i}{K}}$$

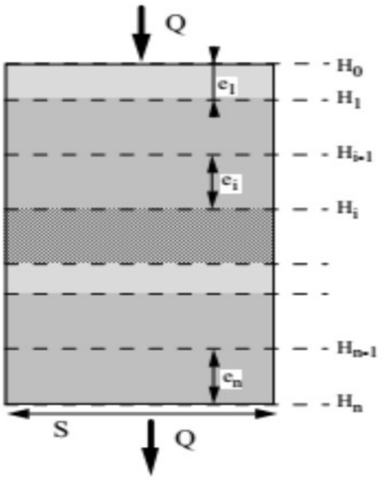


Figure 26: Vertical Equivalent Permeability

CHAPTER VI - FUNDAMENTAL DIFFUSIVITY EQUATION IN POROUS MEDIA

We propose to establish a relation that allows describing the movement of a water particle as a function of its position, time, physical characteristics of the fluid, and the host rock (the terrain).

We have 3 types of relations:

- Isothermal state equations of the fluid
- The equations of dynamics
- The continuity equation

The equations of dynamics translate for viscous fluids, whose viscosity coefficients are assumed constant, into Navier-Stokes equations.

From these 3 types of equations, the diffusivity equation can be established.

VI.1 Diffusivity Equation

Ss: specific storage coefficient S: storage coefficient (S: storage)

If horizontal permeability is constant over the thickness of the aquifer:

Q > 0 discharge withdrawn per unit surface area of the aquifer Q < 0 discharge injected per unit surface area of the aquifer (thus in m/s): for example, rain

If T is isotropic the equation becomes:

which is the form of the diffusivity equation we will use hereafter.

In steady-state regime the equation becomes:

In steady-state regime or for a confined aquifer in transient regime provided that S and T are constant and the boundary conditions, the diffusivity equation is linear in h.

In these cases, if D is a given integration domain of the diffusivity equation, equipped with boundary and initial conditions, it is demonstrated that if h satisfies these boundary conditions and if h satisfies the diffusivity equation, h is, in general, the unique solution of the problem (in steady-state or transient regime).

One of the most frequent works in groundwater hydraulics consists of a well reaching an aquifer. These wells allow withdrawing a constant or variable discharge Q from the aquifer (or part of the aquifer) thus reached.

To know the hydraulic characteristics of reservoirs, it is frequent to perform "pumping tests" which consist of recording variations of hydraulic head (or discharge) at the well and in piezometers close to the well as a function of time. These variations are interpreted according to different theoretical behavior models. In what follows we present some of these models.

VI.2 DIFFUSIVITY EQUATION IN RADIAL COORDINATES

In radial coordinates the diffusivity equation is written:

a) Theis Solution

Theis proposed a solution to this equation in the case of an infinite, homogeneous and isotropic medium of constant transmissivity in time and space (confined aquifer or slightly drawn down unconfined aquifer), a borehole capturing the aquifer over its entire thickness from injection or pumping at constant discharge, in an infinitely small well:

t: time r: radial distance from the center of the well s: aquifer drawdown (difference between the initial piezometric level of the aquifer and its level after a pumping time t, observed at distance r from the well) Q: pumping discharge T: Transmissivity (product of permeability by aquifer thickness) S: storage coefficient

b) Jacob Solution

An approximate solution of the previous equation can be proposed, Jacob's formula:

This approximation remains valid for u values less than 10^{-1}; that is, in practice, for long pumping times and at distances close to the pumping well.

Indeed, we can write that:

for small values of u we can limit the development to 2nd order and write:

Remarks:

1. Small values of correspond to long pumping times or points close to the well (small r)
2. The term has dimensions of length squared. If we set:

we can write the drawdown in the form: which is close to Dupuit's formula.

We also understand that this notion of radius of influence is questionable, since this radius depends on pumping time. However, since aquifers are never infinite, there often exists a certain distance at which recharge of the aquifer occurs.

VI.3 PUMPING TEST

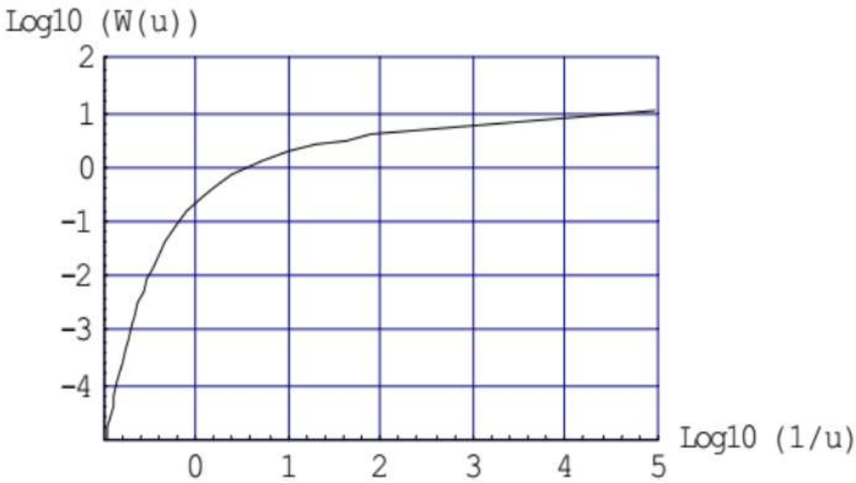
A classic pumping test consists of measuring (in the pumping well or in a piezometer) the drawdowns induced in the aquifer by pumping of a few hours (possibly longer) in order to determine the hydraulic parameters of the aquifer: S and T (the other parameters Q, t, and r being a priori measured).

The Theis solution is often represented in the form of a nomograph in log-log axes, to facilitate graphic interpretation of pumping tests.

More specific tests are used in terrains of low permeability. Tests are interpreted by comparing different theoretical models to actual tests and the properties of investigated terrains are obtained by fitting theoretical curves to experimental curves. The different theoretical models correspond to different geomodels.

1 - Graphic Interpretation of Theis Method

Graphic interpretation of pumping consists of adjusting the experimental curve to the theoretical curve.



Indeed:

Or in log:

The experimental curve log(s), -log(t) can therefore be deduced from the theoretical curve log[W(u)], log(u) with 2 translations:

- a translation of according to the x-axis or s or W(u)
- a translation of according to the t or u axis

The value of the translation along the x-axis: allows deduction of the transmissivity value.

The translation along the y-axis: by deduction will allow deduction of the storage coefficient S value.

After superposition of curves, an arbitrary point is chosen for which the coordinates W(u), u of the theoretical curve and the coordinates s, t of the experimental curve are noted. The introduction of these values into the equations and allows obtaining the transmissivity value and the storage coefficient.

Theis Theoretical Curve

2 - Graphic Interpretation of Jacob Method

Drawdowns can be plotted as a function of time on semi-log paper. The points align along a straight line of slope. Transmissivity can therefore be deduced.

The x-intercept allows deduction of the storage coefficient.

Remark: points corresponding to short times are poorly aligned because they correspond to values of. The x-intercept allows deduction of S.