Groundwater Theory

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1 INTRODUCTION

Hydrogeology is a branch of earth sciences that investigates the quantitative and qualitative properties of groundwater and its movement trough the subsurface geological layers. Groundwater stocks are largely invisible to us, because the upper zone of the subsoil is usually dry. But with depth, the groundwater table is reached, which forms the upper surface of sometimes hundreds of meters thick water saturated rock. Groundwater is an important raw material for many purposes, such as drinking water and industrial applications. On the other hand, in the building construction and in agriculture, for example, the presence of groundwater can be a problem that must be managed by dewatering or drainage.

Groundwater has been explored for thousands of years, to meet the (drinking) water needs of humans. Many villages, castles, farms, ... had their own well. The easy availability of groundwater sometimes was a decisive criterion for locating a settlement or for developing specific craft industries. With industrialization and population growth in the nineteenth century, groundwater became an increasingly important natural resource.

This increased demand for groundwater enhanced the need for in-depth knowledge of the local geological structure of the subsoil, in order to locate the good aquifers. But more insight was also needed into the dynamics of groundwater movement and therefore also of the physical parameters of the groundwater aquifers. This is how hydrogeology originated.

In addition, new problems with groundwater management arose over the years. One of these was the threat of exhaustion, or at least the overexploitation of some aquifers. This was often the undesired side effect of a concentration of industrial activities in one area, on one (deep) aquifer. Another problematic impact of industrialization was soil and groundwater pollution. Until the 1970s, the credo in industry, and among the general population, was, that the soil is “the best filter there is”. If a company wanted to empty a tank with chemicals, it was simple: they built an earthen dike, discharged the contents of the tank within that dike, and the substance was allowed to infiltrate into the subsurface. The filtering by the soil did the rest. Problem (cheaply) solved - at least, so people thought.

The fact that this filter not only worked very selectively, and that some substances simply passed through, and that, moreover, it could also become saturated around the sources of pollution, only became apparent when, for example, drinking water extraction sites were suddenly threatened by nearing industrial pollution plumes in the groundwater.

The 1970s were also the years of growing environmental awareness among the population. The rivers were dead, the air was heavily polluted by industry. Landfills and waste dumps for household or industrial waste had been located all around. Usually, they lacked any kind of protection of the underlying soil and groundwater ("the soil is the best filter", isn't it?). Gradually, a legislative framework was developed, to protect groundwater and to remediate the contaminated sites and prevent further pollution. As part of the same environmental policy, more instruments have been created, such as the environmental impact assessment (EIA), which aim to proactively identify the possible impacts on the environment of an intended plan or project, and to propose remedial measures to mitigate or avoid such undesired impacts.

2 THE THEORY OF GROUNDWATER

Hydrogeology is a part of the earth sciences that relies heavily on the quantitative: measurements, analyses, calculations and simulations. This is in contrast to many other branches of the earth sciences, which often tend to be more descriptive and interpretive, and where the use of computational methodologies is comparatively rather limited.

Hydrogeology is also one of the more practice-oriented branches of the earth sciences, closely related to the field technology it supports. It is therefore often considered to be part of "applied geology". Some syllabi and manuals on hydrogeology often go rather quickly into "hands-on" mode and sometimes pay little attention to underlying theoretical foundations of the field technology. After a brief presentation of the occurrence of groundwater, and the introduction of a number of basic concepts and formulas, many hydrogeology textbooks quickly move on to the practical aspects of classical field hydrogeology, in particular the pumping test.

The pumping test is a fairly labour-intensive field method for determining the numerical parameters of an aquifer. The ultimate intention is to investigate, by means of calculations, whether the aquifer has sufficient potential to be exploited. The theory behind the pumping test is of course interesting by itself, and will be explained further on, but the weight sometimes given to the chapter "pumping tests" and related technologies in syllabi and in the literature (sometimes up to two thirds of the number of pages) indicates how practically oriented classical hydrogeology really is.

There is, of course, nothing wrong with that. But here's a status question: is the methodology, applied to pumping tests and other field methods, which are primarily used to evaluate an aquifer on its quantitative potential, particularly for its possible extraction yields, simply applicable to other approaches to the groundwater discipline, such as the more desktop oriented environmental impact assessment? The classic guideline book for environmental impact studies for the groundwater discipline is actually often a selection of traditional field and desktop formulas from the hydrogeological handbook. This raises a question: can, for example, this formulary for calculating a field test, simply be adopted as a general methodology which is also applicable to calculations in the context of environmental impact assessment? Which leads to the next question: is there a general “one-fits-all” integrated groundwater theory, which allows to describe all phenomena, and to solve all problems?

There is another aspect of this question, which is related to the basics of the traditional field methodology: the diverse origin of the equations used therein. There is a mixed use of equations which have been mathematically derived from a basic differential equation (so called “analytical equations”), and of fieldwork derived, empirical equations, not at all mathematically related to the former (and to each other), but which are often needed as a stopgap to solve these analytical formulas, because the latter cannot be mathematically solved on their own. Hence, the formulary of hydrogeology sometimes looks to the user as some second-hand bought toolbox, in which a tool of a set may be missing occasionally, or in which a set of tools comes from two sources, but which, in general, still contains enough equipment to allow solving any kind of problems. After all, when the case becomes too complicated, the hydrogeologist can rely on groundwater models, which have a reputation of better dealing with field complexity, but in which the calculations take place behind a keyboard.

Of course this is not unique, there are other technology-oriented scientific disciplines where this is the case. For the classic, practice oriented hydrogeologist, aiming for investigating aquifers on their quantitative, exploitation related characteristics, this toolbox is certainly sufficient, he knows how to make clever use of the various “equipment”. But confronted in environmental impact assessment with a different approach to groundwater problems than the classical methodology, aimed at “mining” or discharging groundwater projects, one sometimes encounters unbridgeable gaps between the various tools, giving hydrogeology a perception of “muddling along with the available equipment”.

Applied to the environmental impact assessment, the classic “mixed” hydrogeological methods do not always allow to provide conclusive answers to the questions asked, at least not with the available theoretical approach. In the early years of environmental impact assessment, this issue was offset by a culture of fieldwork, mainly introduced by the first generation of experts. These were often attached to, or had just transferred from, academic institutions, where fieldwork culture was, and still is standard practice. Unfortunately, that fieldwork culture in environmental impact research has largely died out, partly because of market pressure on bidding prices for environmental impact assessments. As a result, hydrogeological field work is only carried out for really large groundwater related projects. This evolution became enhanced when soil and groundwater pollution aspects often got shifted to separate environmental legislation with proper procedures, and where fieldwork is still essential for data acquisition, reducing the importance of that quantitative aspect of groundwater in environmental impact assessment studies.

Partly due to the decline of the fieldwork culture in environmental impact assessment, the methodology for groundwater was left behind with a compendium of selected, stand-alone formulas, as a step-up alternative for the much more expensive groundwater modelling. Manual hydrogeological calculations, using a mixture of analytical and empirical equations, are usually applied to investigate simple problems, for example at the level of a single-well groundwater extraction, or for a shallow drainage for a single building. For more complex problems, numerical methods, "groundwater models", are used. But there is a gap of theoretical insights in between.

Generally, the inadequacy of the basic calculation methods to provide first hand complete solutions for simple cases, was either ignored by experts and others involved, or admittedly acknowledged, but accepted as something inevitable, for which there was neither a solution, nor a need to solve the posed problems. The only noticeable initiative was the urge by some hydrogeologists to discard the widespread use of Sichardt's Formula for drainage problems, but the proposed alternatives were not fully convincing. Simple models have been developed, but they work with two separate modules, one for steady-state and one for transient calculations. But both cannot be merged and they calculate independently from each other, e.g. the transient module does not allow to calculate how long it takes to reach the steady-state, calculated with the latter module. As a result of all this confusion, many experts simply use the Theis Formula as a stopgap for solving all problems, since "Theis calculators" are freely and easily available on the internet, thereby ignoring the limitations and original purposes of this equation.

These problems with the groundwater aspect in the practice of environmental impact assessment, were the impetus to start the current investigation. The focus will however be not on the "know how", the field practice (or the modeling practice), but rather on the "know why". From this background, a number of calculation tools will then be proposed, either for the "simple cases" mentioned, where a groundwater model is too expensive, or cannot offer added value, or for an initial survey of the effects of the project, in order to decide whether or not the problem should be further investigated with a groundwater model. The aim is to capture at least orders of magnitude or extension of the effect, with a reliable margin of error and taking into account natural variations, so that it can be decided whether or not further investigation would be necessary.

In this book, therefore, an attempt is first and foremost made to seek a little more consistency in groundwater theory, so that hopefully, the insights in the methodology will change, and a number of gaps in the methodology can be filled in (or unambiguously identified as a gap), preferably without the need of stopgap empirical equations.

In the next chapters (Chapter 3 to Chapter 7), the existing theoretical knowledge is presented. From Chapter 8 to Chapter 11, the development of an analytical groundwater methodology will be worked out, and finally, the consequences discussed (Chapters 12 and 13). A last section contains references, lists of figures and tables, and appendices showing calculation examples from the methods proposed in Chapters 9 to 11.

3 OCCURRENCE AND DYNAMICS OF GROUNDWATER

3.1 Occurrence of groundwater.

Storage of groundwater in the subsoil is a stage in the water cycle. Groundwater is supplied by precipitation, evaporated seawater that has drifted as clouds to land, where it condenses and falls back. Like riverine flow, groundwater flow is a result of the conversion of the potential energy originating from solar radiation that makes sea water evaporate, into kinetic energy, which allows the water to flow back to the sea.

The presence of groundwater in the subsoil is of course linked to the presence of reservoir characteristics in the geological layers. No groundwater is present where the subsoil consists of impenetrable rock. Where the subsoil is porous, the groundwater level is sometimes found after digging a few decimeters, under an unsaturated zone. Such groundwater is stored under atmospheric pressure, and is referred to as a phreatic aquifer.

Aquifers can also occur under poorly permeable clay layers. A well-known phenomenon that such deeper layers show is, that they are under pressure ("confined" aquifers). As a result of that pressure, when a borehole is drilled through the boundary between the capping clay layer and the confined aquifer, the water level in a monitoring well in that deeper aquifer, shall rise significantly higher than the upper boundary of that aquifer. Sometimes, the groundwater level in such pressurized aquifers, rises even higher than ground level. Water then wells up from the monitoring well. This is named an artesian well, after the French region of Artois, where such wells have been exploited since the Middle Ages.

3.2 Groundwater dynamics.

The phenomenon of artesian wells is usually explained by the principle of connected vessels. The classic theory assumes that the confined aquifer is supplied in a topographically higher area, where the layer surfaces as a phreatic outcrop. Its phreatic level determines the pressure head in the aquifer. Deeper in the aquifer, water is trapped between two sealing rock layers, and stands under a pressure, by a potential generated by its phreatic supply head. Pressure will cause it to rise up to the level of its pressure head, when a borehole is drilled in the deeper realms of the aquifer.

Traditional graphic representations of an artesian aquifer then show either a monoclinal downward sloping layer, the artesian layer cropping out in a topographically elevated area, in a relatively upstream river valley, where the layer is intersected by the valley floor (Figure 3.1), or in any topographic high where the aquifer crops out. Such a supply model can be expected, for example, in situations where the aquifer is confined between two impermeable layers, like in folded layer sequences. However, from a dynamic point of view, the supply mechanism of confined aquifers is sometimes different, as will be explained below.

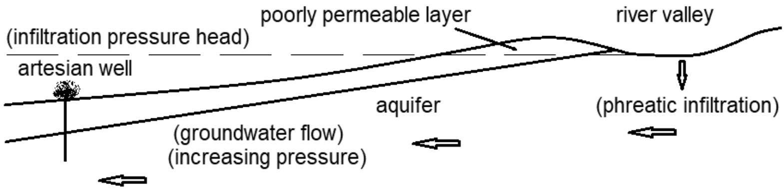

Figure 3-1: Schematic representation of supply of a confined aquifer by phreatic infiltration.

An alternative supply model for confined aquifers was proposed by Walraevens (1987), for a number of confined aquifers in the subsurface of Northwest Belgium (the “Ledo-Paniseliaan” and also the “Ypresian” aquifers). The circulation model was based on a groundwater modeling on one hand, but supported by groundwater level measurements and a study of the hydrochemical evolution of the groundwater over time. It shed a different light on the possible supply of deeper, confined aquifers. Important in this circulation model is, that the aquifers are not considered as separate units, vertically isolated from each other by impermeable clay layers, but that the latter are part of the flow system too.

In the study area, artesian wells exist, or have existed before overexploitation of deep confined aquifers has lowered the pressure levels below the ground surface. According to Walraevens (1987), the artesian pressure in the confined aquifers studied, is caused by the pressure emanating from the water mass built up in the surrounding ridges, a process in which the poorly permeable clay layers also have a part. The phreatic groundwater level more or less follows the topography. The water masses in the topographical heights act as a kind of "water tower", as a supply area, generating the pressure for the groundwater circulation. From there, the groundwater flow is driven to the draining river valley. The topographically lower river valleys are the discharge areas, where the groundwater is drained through the watercourses (Figure 3.2).

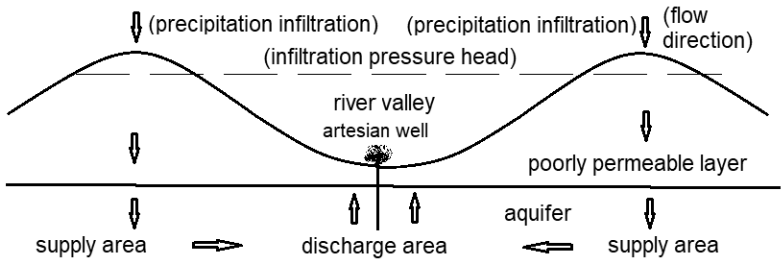

Figure 3-2: Model of supply of confined aquifers and groundwater circulation by precipitation infiltration in topographically higher areas (after Walraevens, 1987).

Due to this dynamic equilibrium, which is thus maintained by precipitation, keeping the water mass stored in the ridges at level, a hydrostatic pressure difference is created between the phreatic water masses in the higher ridges and those in the lower valleys. Because the mass of rocks, clay and sand forms one large continuous water storage, a downward flow is created below the topographical heights. This circulation also supplies the deeper aquifers in the area, even those covered by a hundred meter thick clay. The clay layers, with their much lower hydraulic conductivity, do of course exert resistance in this flow system. It is therefore understood that supplying such deep aquifers is a long-term process, and that such deep groundwater systems are prone to a risk of overexploitation.

This downward flow under the ridges, through the retarding clay layers, is confirmed by the observation that, the deeper an aquifer occurs, the lower its absolute hydraulic head. And water simply flows from a higher to a lower level. In discharge areas, the absolute hydraulic head of the deepest aquifer is the highest, so the groundwater flows upwards through the layers.

The role of the clay layers is striking in this dynamic circulation model. In traditional hydrogeological views, it is rather quickly assumed that clay layers are "impermeable". An old, stratigraphic clay layer, however, contains as much pore space as a sandy aquifer, about one third of its volume, and that pore space is equally filled with groundwater. The problem is, that the water in clay is difficult to get in motion because of the pore structure. Clay does not consist of round grains, with a diameter larger than 50 micrometers, but of plates between 2 and 10 micrometers in size, what strongly inhibits the mobility of groundwater. Moreover, the edges of the sheet-like clay minerals have concentrated negative electric charges, which attract water molecules due to the latter's positive dipole. But a clay layer is basically aquiferous, and water can still move inside of it.

It is therefore necessary to differentiate between a very poorly permeable aquifer, such as a clay layer, and a really impermeable layer, which in no way carries pore space at all. A solid sandstone without any pore space, for example, is completely impermeable. This does not alter the fact that in some calculations or groundwater models, an underlying thick clay layer can be regarded as the lower limit of the study or model area, and can therefore be pragmatically assumed to be "impermeable", even though that clay layer is in reality porous and water saturated. This assumption can be justified if no aquifers would exist below it, or if the underlying aquifers would not be relevant or would not be significantly influenced by the problem under investigation.

The above described groundwater flow mechanism was confirmed by a groundwater model by Walraevens (1987), but also supported by an extensive hydrochemical study. Because present-day Lower and Central Belgium was under the sea level until about 2.5 million years ago, the aquifers were originally characterized by water with a pronounced marine signature. The circulation model predicted that, after the sea retreated, infiltration of precipitation water would gradually expel the water with marine signature from the pores of the host rock.

This freshening process of the groundwater would also involve replacement of cations adsorbed on clay minerals in the clayey formations. The hydrochemical study revealed an evolution of ion and stable isotope distribution, in a zoning that exactly matched the pattern predicted by the groundwater circulation model. More recent studies even revealed the footprint of the melting of the permafrost in the area, at the end of the Pleistocene glaciation in the area (Walraevens et al., 2021).

Hill ridges are hence supply areas, valleys are discharge areas. The topographical location of a pumping well can therefore be a differentiating factor when assessing its impact. One of the consequences of this view, relevant for environmental impact assessment is, for instance, that the vulnerability of aquifers is completely different from a dynamic than from a static point of view. In a static view, the hydraulic conductivity of the aquifer is the major criterion, regardless of the topographical location of the outcrop of the layer. In the same view, a thick unsaturated layer, which is usually found in topographically higher areas, is considered as a protection of the deeper aquiferous zone. In the dynamic interpretation, however, these highs are the supply areas, while a low-lying aquifer in a valley may be subject to a discharge regime. The assessment of the vulnerability of groundwater layers must therefore be done with the necessary precautions.

3.3 The ideal groundwater reservoir.

So far qualitative considerations about the occurrence and dynamics of groundwater. However, as soon as the need for quantifying groundwater processes enters, mathematics comes with. It will impose its proper conditions, in order to allow such calculations.

Groundwater calculations are always based on a number of preconditions, to which the medium to which the calculation applies, the aquifer, is supposed to meet. These preconditions involve a certain (mathematical) abstraction and conceptual simplification, and therefore, they certainly do not always correspond to the field reality. This does not mean that it would not be possible to calculate situations in field circumstances that do not strictly correspond to those preconditions, provided it is ensured that these deviations from the mathematical abstractions should not have a meaningful influence on the final result.

The boundary conditions and mathematical abstractions can be brought together in a theoretical groundwater reservoir model, with a characteristic geometry, which will be further described and discussed. A first distinction can be made between two main types of internal reservoir structure, namely the homogenous and the non-homogenous reservoir model.

The ideal homogenous reservoir model assumes an aquifer that is perfectly horizontal and that lies in a perfectly flat landscape. It is homogeneous and isotropic in all directions. This means that the homogeneity and the isotropy apply in a circle of 360° around a pumping well. Hence, because any extraction will create axial symmetric impacts around the well, the representation of that impact can be reduced to one dimension, along a radius of that circle. Homogeneity applies to the host rock at the size level of the granular material (analogous to “texture” as used in petrography and sedimentology; basically that what is observable with a microscope). Pore spaces, which determine the hydraulic conductivity as an aquifer, are homogeneously distributed over the volume of the host rock. The latter doesn't necessarily have to be non-consolidated sandy material. The concept also applies to, for instance, limestones or sandstones that are supported by a homogeneous grain skeleton in a permeable porous matrix.

Isotropy is rather analogous to the concept of "structure" in sedimentology: the macroscopic arrangement of rock bodies. Isotropy means that the host layer is everywhere of equal thickness and has the same hydraulic conductivity. Anisotropy can arise from lateral variations in layer thickness or in hydraulic conductivity. An example is a reservoir where a poorly permeable formation is cut by a former erosion gully, once a channel in a sedimentary system, which had subsequently been filled with much more permeable sand than the formation where it was originally cut into. Each of these rock bodies can be internally homogenous, but together, they create an anisotropic aquifer.

Such perfect preconditions obviously rarely exist in the field reality. However, the scale of a local survey may be small enough to make use of these assumptions of homogeneity and isotropy, in a calculation. In case of significant deviations from the ideal axial-symmetric, homogeneous and isotropic reservoir model, groundwater models are the suitable calculation tool.

The heterogenous reservoir model obviously shows a lack of homogeneity and, (usually) a clear lateral anisotropy. Heterogenous reservoirs are mostly developed in impermeable rock such as limestone, sandstone, shale, granite,…. Often these are consolidated rocks that were once exposed on the earth's surface, and thereby underwent weathering and erosion. These processes created networks of cracks and fissures, even of cavities. Structural weaknesses in the rock, such as fractures, may have imposed a preferred orientation of the cracks and fissures. Ancient limestone cave systems, which were subsequently pulled down by tectonic soil movements and ultimately covered by younger formations over the course of geological times, also belong to this category.

In such a heterogenous reservoir model, the groundwater is not stored inside pore spaces, but in between (relatively) impermeable rock masses. This causes a strong lateral inhomogeneity between, on the one hand, the impermeable rock masses themselves, and on the other hand, the open spaces between those rock masses, which are filled with groundwater. Such a reservoir can be isotropic, for example when a rock is homogeneously and equidimensionally jointed (cracks in the three perpendicular directions, causing the rock to be fragmented, and the cracks have been opened by weathering and erosion), or very heterogeneous, when the cracks and fractures, run according to a preferred tectonic direction (e.g. fault systems).

In the aquifers of those reservoirs, the hydraulic conductivity is often very high, because the water in the cracks, crevices and cavities can flow relatively easy, rather unhindered by a granular structure. Such aquifers can easily supply large amounts of groundwater, making them very attractive for industry and the drinking water supply. The inhomogeneity and the degree of anisotropy of such reservoirs strongly deviate from the ideal reservoir. Calculations will preferably require groundwater modelling.

But even when calculating with the "perfect" homogenous reservoir, mathematics requires more stringent rules. Regardless of the situation they are applied to, equations always assume specific boundary conditions, within an ideal reservoir model. This ideal model, as outlined above, is homogenous, has a perfectly horizontal groundwater surface, in a perfectly horizontal terrain. The saturated thickness is constant everywhere, meaning the hydraulic head also is. Most of all, "everywhere" means "everywhere" in its most wide meaning, since mathematics imposes that the ideal homogenous aquifer extends infinitely.

These boundary conditions of the ideal aquifer fill many field hydrogeologists with criticism, since they obviously do not exist, first of all the infinity of the reservoir. The contrast between the ideal boundary conditions in which the groundwater equations are valid on one hand, and the roughness of the terrain on which they have to be applied on the other hand, has instigated some dislike against using them, out of fear for inaccuracies in the calculations. This is one of the reasons why calculations using analytical methods are often the subject of criticism (e.g. Seward et al., 2015, Barlow et al., 2018, Zhai et al., 2021, Louwyck et al., 2022a).

However, a conceptual representation that brings the ideal reservoir model closer to the field reality, is the "extensive aquifer", coined by Edelman (Edelman, 1972). The latter concept originally applied to an aquifer, which has a comparatively small thickness with respect to its lateral extension. This concept is however very useful for local problems with a very small area of impact, relative to the lateral extent of the aquifer. The latter can hence pragmatically be assumed as infinite. In the limited study area, therefore (unless of course field conditions manifestly indicate otherwise), the topography can be assumed as horizontal, and the hydraulic head, the thickness of the saturated zone, the hydraulic conductivity, and, in general, the homogeneity and the isotropy of the aquifer, can be considered constant everywhere. With this assumption, using analytical equations can be envisaged without concern about unacceptable inaccuracies. For example, an aquifer in a very gently sloping monoclinal sand layer, can be locally assumed to be horizontal within the restricted limits of the study area.

4 ANALYTICAL STEADY-STATE GROUNDWATER FORMULAS

4.1 Bernouilli's Law, groundwater potential and hydraulic head.

Hydrogeology essentially deals with flowing water, so such processes fall under the laws of hydraulics. A basic law of hydraulics is Bernoulli's Law:

$$\frac{1}{2}\rho v^2 + \rho g h + p = cte$$

With:

ρ: the density of the liquid [M/L³].
v: the velocity of the current [L/T].
h: the hydraulic head [L].
g: the gravity acceleration (L/T²), in the SI system equal to 9.81 m/s².
p: the pressure [M/T²/L].

The left-hand side of Bernoulli's law consists of three pressure terms: a velocity term, a potential energy term and a static pressure term. The sum of those three terms is always constant. Bernoulli's Law applies to situations involving flowing liquids. Two important conditions for the validity of Bernoulli's Law are steady flow and a homogeneous density of the liquid. Steady flow thus implies the existence of a constant gradient, a constant difference in potential.

In the agents that regulate the flow as expressed by Bernoulli's Law, one natural parameter, the gravitational acceleration g, is immutable, and hence a limiting boundary condition.

In the velocity pressure term (or dynamic pressure) and the potential pressure term, one spots a resemblance with the equations for kinetic and potential energy respectively:

$$E_{kin} = \frac{1}{2}.m.v^2$$

And:

$$E_{pot} = m.g.h$$

With:

E_{kin} and E_{pot}: respectively kinetic and potential energy [$M.L^2/T^2$].
m: mass [M].
h: hydraulic head [L]
g: the gravity acceleration (L/T^2), in the SI system equal to 9.81 m/s^2.

The constant in the right-hand side of Bernouilli's Law represents a constant pressure. Pressure can be expressed in a number of ways, such as the force per area, or the energy per volume, or the energy density. Bernoulli's Law is therefore actually a conservation law of pressure, of energy density, and thus, indirectly, of energy.

For standing water (v=0) Pascal's Law applies:

$$p = p_0 + \rho.g.h$$

Here p is the hydrostatic pressure at depth h and p_0 is the pressure at the upper limit of the water surface.

Bernoulli's Law can also be represented in a simplified form, by dividing all terms by the product of the density with the acceleration of gravity. All terms will then be expressed in units of length:

$$E = \frac{v^2}{2.g} + h + \frac{p}{\rho.g}$$

The quantity E is a measure of the energy level of the flowing water, but in units of length instead of in units of pressure.

In standing water (v = 0) holds:

$$H = h + \frac{p}{\rho.g}$$

H is called the hydraulic head of the aquifer. The hydraulic head corresponds to the vertical distance between the lower boundary of the aquifer and the water surface in this aquifer. This level is called the piezometric level. Basically, H represents an altitude relative to a reference level. In a horizontal aquifer this is everywhere the lower boundary of that