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Hydrogeology of shallow and deep seated groundwater flow systems

Basic principles of regional groundwater flow

by

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1 Groundwater flow systems

1.1 Introduction

Groundwater flow is an effective agent for the transport of contamination in the subsurface. Groundwater flow systems can transport pollutants for short distances of 100 meter or less, or it can transport contaminations for distances exceeding several kilometers or 10ths of kilometers.

Groundwater flows through pores in unconsolidated deposits (sand, silt, clay) and through joints and fractures in consolidated bedrock (shale, sandstone, limestone, dolomite) (Figure 1). The openings in deposits and rock are called porosity and may reach from close to zero to more than 50 % of total rock volume (Figure 1). These pores may be more or less connected and thereby will be more of less permeable for water (Figure 2).

The topography of the groundwater surface is the determining boundary condition for groundwater flow. Groundwater flows always from highlands to lowlands, usually along curved flowlines such that water movement is downwards into the groundwater body in highlands (recharge areas) and upwards towards the groundwater table under lowlands (discharge areas). The driving force for groundwater flow is the gravitational force field. Modern gravitational groundwater dynamics revised a number of traditional concepts. In nature

(1) groundwater generally does not flow in the direction of the pressure gradient,
(2) groundwater generally does not flow parallel to the groundwater table, and
(3) groundwater flows in significant amounts through low permeability clay and mudstone layers.

The physics of groundwater flow, flow through low permeable layers, and regional groundwater flow is described in the following sections. Our goal is to show that groundwater, and thereby contamination, may seep vertically downwards from polluted areas and may not be monitored by routine installation of piezometers. The understanding of groundwater flow systems is a necessary condition for the design of effective monitoring systems, and thereby for any field investigation of subsurface contamination.

1.2 Basics of groundwater dynamics

During the 1960s, hydrogeology entered a new development phase in which the emphasis of hydrodynamic studies was widened from local springwater hydraulics to a regional scale. After the "discovery" of Darcy's law, the hydraulics of springs, dams and other engineering projects had been developed, primarily by engineers, to deal with practical day-to-day problems.

As is the case in many theoretical developments in engineering, the derived equations were based on simplifying and therefore limiting assumptions, such as the assumption that groundwater flows parallel to the groundwater table and in the direction of slope of the water table (Dupuit-Forchheimer assumption). In this approach, apparent deviations would be

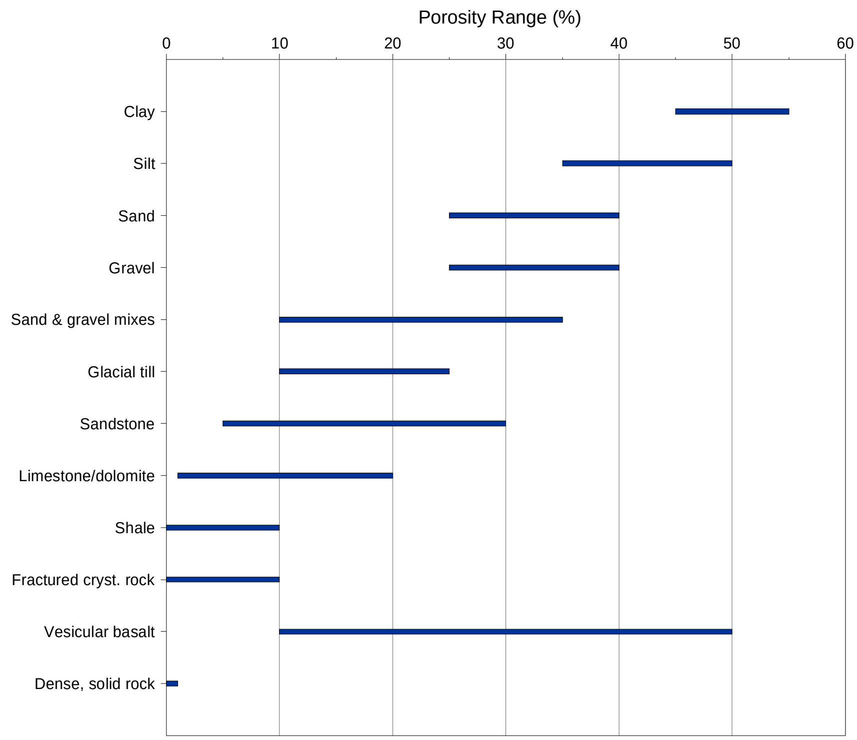

Figure 1. Porosities for common unconsolidated deposits and consolidated rocks (taken from Driscoll 1986, p.67, Table 5.1)

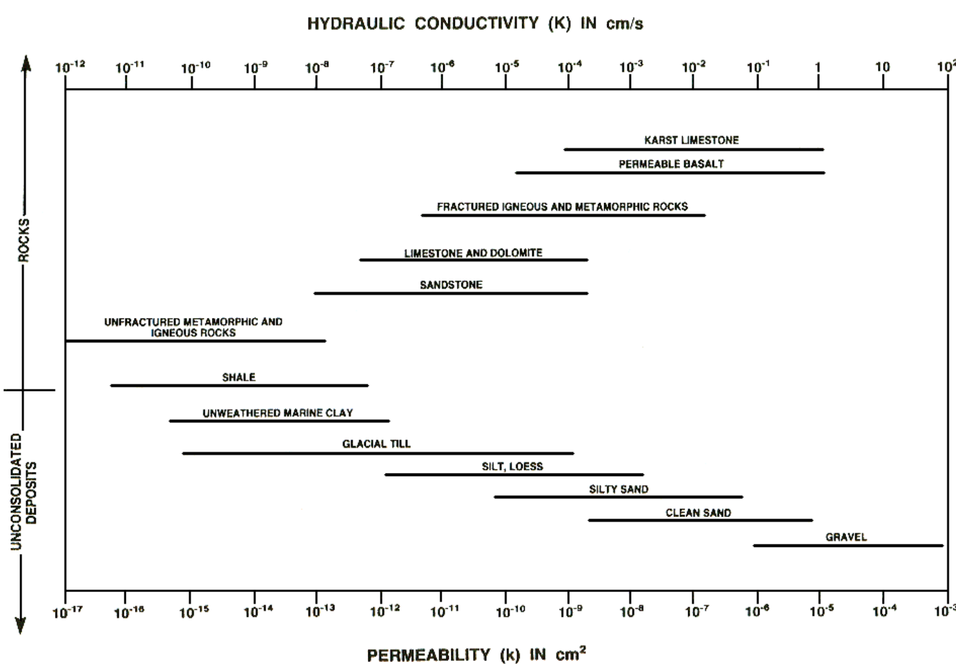

Figure 2. Range of values of hydraulic conductivities [cm/s] and permeabilities [cm^2] in unconsolidated deposits and rocks (from CCME, 1992)

removed by applying correction factors.

As a further assumption, the pressure gradient grad p was erroneously regarded as the single cause of groundwater flow, thus excluding the influence of gravity. Gravity, however, is the actual driving force for both regional and local groundwater movement.

The underlying physical principles were demonstrated more than sixty years ago by Hubbert (1940). In his fundamental "Theory of Groundwater Motion" he demonstrated that a physically consistent, energy-based treatment of groundwater flow, incorporating the energy fields of both the gravitational potential and the pressure potential (potential = energy/mass), made it possible to understand phenomena and relationships which could not be explained by traditional well hydraulics. This applies also to the occurrence of artesian flows, and the movement of groundwater through clay layers between aquifers.

The current report does not contain the detailed derivation of the physical relationships involved, although the methods used here have not yet been generally adopted in engineering and scientific hydrodynamics. A detailed discussion of the hydrodynamic principles can be found in Weyer (1978). Only a short explanation of the relationships is given in the following section. The parameters involved, and symbols representing them, are summarized in Table 1.

In physical mechanics, the potential is the energy per unit mass. In a groundwater body, the above mentioned energy fields of the gravitational potential Φ_g and the pressure potential Φ_p are combined, resulting in the fluid potential:

$$\Phi = \Phi_g + \Phi_p \qquad (1)$$

$$\Phi = g * z + \frac{p}{\rho} \qquad (2)$$

The fluid potential gradient, -grad Φ, is the force vector which is used in Darcy's equations to calculate the flow vector $\vec{q}$:

$$\vec{q} = - \text{grad } \Phi \qquad (3)$$

In accordance with the laws of thermodynamics, the groundwater flows from points with higher potential energy to points with lower potential energy, completely independent of the direction in space. It can therefore flow in any spatial direction, upwards, downwards, and laterally; however, it will always flow in the direction of decreasing potential Φ. *Even so, it can flow against the gradient of the pressure potential, against the gravitational gradient, and against the slope of the groundwater table*. The groundwater flow direction is modified if an anisotropic permeability tensor interferes, i.e. the flow deviates from the direction of the force vector. For this discussion we will limit ourselves to isotropic conditions, in which the flow follows the direction of the force vector, - grad Φ. As the isolines of the energy field Φ are

Φ Mechanical Potential (Length^2/Time^2) Φ_g Gravitational Potential (Length^2/Time^2) Φ_p Pressure Potential (Length^2/Time^2) -grad Φ Force vector in Darcy equation $\vec{q}$ Flow vector in Darcy equation $\vec{g}$ Gravitational force z Elevation above reference elevation (usually mean sea level)	p Pressure ρ Density of water η Dynamic viscosity h Head ∇^2 Laplace-operator $\nabla^2 = \frac{\delta^2}{\delta x^2} + \frac{\delta^2}{\delta y^2} + \frac{\delta^2}{\delta z^2}$

k Intrinsic permeability (Length^2)

the permeability factor in the Darcy equation, which refers only to the geometric properties of the medium.

σ Fluid conductivity (Time)

the permeability factor in the Darcy equation, which incorporates the geometric properties of the medium (k) and the fluid (ρ,η) .

$$\sigma = \frac{k * \rho}{\eta}$$

K Hydraulic conductivity (Length/Time)

the permeability factor in the Darcy equation, which incorporates all constants related to the properties of the medium (k) , to the fluid (ρ,η) and to the gravitational force vector ($\vec{g}$) .

$$K = \sigma * \vec{g}$$

Table 1. Definition of groundwater dynamic variables

generally curved, parallel flow of groundwater occurs only under special conditions.

At this point we would like to mention that many derivations of traditional groundwater hydraulics are based on the simplifying assumptions which Hubbert (1940) has showed to be valid only in special cases: namely that flow is in the direction of the pressure gradient, and that flow is parallel to the groundwater table. It should therefore not be surprising that the application of modern groundwater dynamics leads to results that will appear unusual to many readers.

It is now easy to understand why the energy potential Φ occupies a central position in modern groundwater dynamics. However, how can one measure the fluid potential Φ in the field? It can be done, if one determines how much work the potential can perform against the ubiquitous gravity field, according to the equation

$$\Phi = h * g \qquad (4)$$

For this purpose one installs an observation well or a piezometer in the area of interest, with a short screen at the lower end of its pipe located at the desired depth of observation. The height to which the water level rises in the pipe indicates the hydraulic head h at the level of the screen. Figure 3 shows the meaning of the various components that make up the total head. As Figure 3 shows, the hydraulic head consists of two components, the pressure head

$$h_p = \frac{p}{\rho * g} \qquad (5)$$

and the gravitational head

$$h_g = z \qquad (6)$$

These heads are a measure of the corresponding potentials (energy per mass). The corresponding total potential can be calculated from the heads, using equation (4)

Next the terms hydrodynamics, hydrostatics, and groundwater dynamics are explained. Hydrodynamics deals with flowing water, whereas hydrostatics uses simplified hydrodynamic equations for the special case of non-moving water (Φ = const.).

Classical hydrodynamics, as for example taught in the 19th century by Sir Horace Lamb, developed during application to surface-water problems. Its definitions and derivations are therefore not generally applicable to groundwater problems where, under certain conditions, the permeability is a low-symmetry tensor. The widely used velocity potential of classical hydrodynamics is only mathematically defined and has no physical meaning. The use of the velocity potential is physically invalid, and leads in general cases to erroneous results. As the term hydrodynamics often implies the use of the velocity potential, we will designate the hydrodynamics of groundwater as groundwater dynamics. This stresses that we appropriately utilize the hydraulic potential Φ, not the velocity potential. Classical

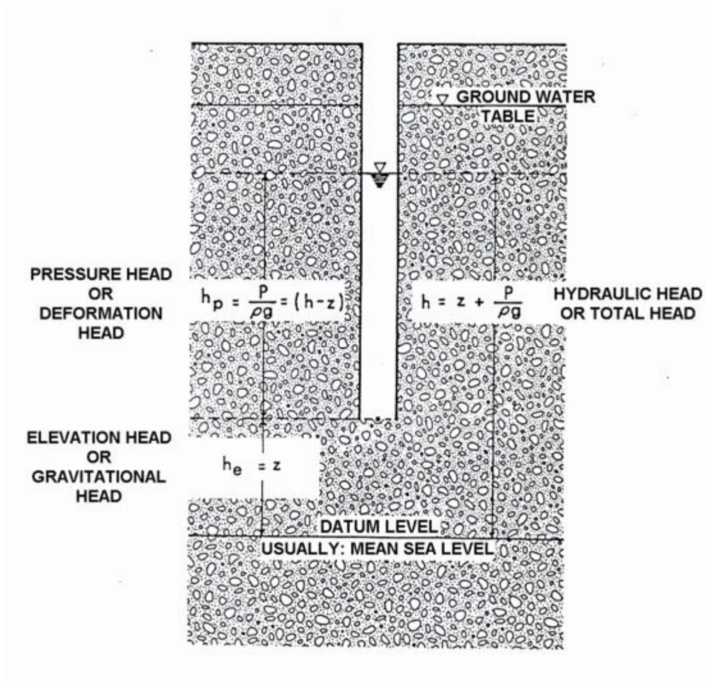

Figure 3. Definition of heads in a piezometer.

hydrodynamics, moreover, often uses grad h as the force vector, instead of grad Φ. This use does not represent a physically correct approach.

1.3 Groundwater flow through layers of low hydraulic conductivity

During the solution of field equations for the energy potential and for the flowlines derived therefrom as gradients of the energy field, hydraulic conductivity appears as a parameter in the field equations. The solution consists of minimalization (integration) of the differential equations and thus of the energy consumption in the total field under consideration.

The energy and flowlines thus calculated run continuously through both aquifers (e.g. sandstone, limestone, marl, sand and gravel) and aquitards (e.g. shales, slates and clay). It is therefore normal that aquifers communicate with one another through intervening aquitards. This phenomenon is currently often called 'leakage'.

The widespread assumption that groundwater does not flow in significant quantities through clay layers must be proven for each individual case. Detailed studies of clay-aquitards in North America have shown that the hydraulic conductivity provided by fracture systems in clays can be as much as three orders of magnitude higher than the hydraulic conductivity of the clay blocks as determined in the laboratory.

As the continuity equation applies to groundwater flow systems - no groundwater is lost or gained inside the system - it is possible that, in undisturbed groundwater flow systems, more water moves through the overlying clay layers than through the underlying aquifers. How can this be? That condition can be explained in a simple way if one expands equation (3) with the cross-sectional area A and the hydraulic conductivity K, to obtain a mass (M) for the amount of flowing groundwater:

$$q * A = - \mathrm{grad}\ \Phi * A * K$$

$$M = - \mathrm{grad}\ \Phi * A * K \qquad (7)$$

For the purpose of this explanation, the quantity of flow in the aquifer (M(a)) shall equal the quantity of flow in a clay layer (M(c));

$$M(a) = M(c) \qquad (8)$$

The following differences, however, exist in the subsurface:

1. The hydraulic conductivity of the aquifer K(a) is significantly higher than that of the clay K(c)

$$K(a) >> K(c)$$

To compensate for this, nature arranges the force fields, and thus the groundwater flow systems, in such a way that

2. The total cross-sectional area through the clay becomes much larger than that in the aquifer

$$A(a) \ll A(c),$$

and

3. The hydraulic gradients in the aquifer become much smaller than those in the clay

$$- \mathrm{grad}\ \Phi(a) \ll - \mathrm{grad}\ \Phi(c).$$

This means that the groundwater flows through clay layers by the shortest path - usually normal to the layering, whereas it moves in aquifers often parallel to the layering. Through corresponding variations in the sizes of the cross-sectional areas (A) and the force vector (-grad Φ), nature insures that equation (8) is satisfied. This happens, because the continuity condition and two basic requirements of thermodynamics have to be satisfied:

1. The entire groundwater system strives towards the condition of lowest energy in the global gravity field; it is for this reason that groundwater moves in the system at large.

2. The energy equilibrium is achieved in such a way that - integrated over the entire system - the least possible energy is expended.

These are, in generally understandable terms explained, the reasons why as much or more water can flow through low permeability layers as through the aquifers in undisturbed groundwater flow systems that are in hydraulic equilibrium.

1.4 Topography, orography and artesian wells

How can one describe such groundwater flow systems in the subsurface?

The first and easiest step in the determination of the boundary conditions of regional groundwater flow systems is the evaluation of topographic and orographic maps. Because in general the groundwater table follows the topographic surface.

Topographic highs are recharge areas for regional groundwater flow systems, whereas the stream valleys are the discharge areas. This means that groundwater flows in the recharge areas from the groundwater table downwards into the groundwater body, whereas in the discharge areas it flows upwards toward the groundwater table, where it either evaporates or discharges into a surface-water body. Therefore the valleys of large creeks, rivers and lakes are normally endpoints of regional groundwater flow systems. These are also the areas in which artesian wells can occur, because in discharge areas the energy level inside groundwater bodies is higher than at the groundwater table. Artesian wells thus indicate groundwater discharge areas.

1.5 Regional groundwater flow

1.5.1 Results of previous investigations

Hubbert (1940) already showed an example of the calculation of regional groundwater flow with very simple boundary conditions and parameters (Figure 4). Such calculations of static, and therefore time-independent, energy fields in a groundwater body can not be carried out with the simple Darcy equation. Instead the latter, together with the continuity equation, is transformed into the Laplace equation:

$$\nabla^2 \Phi = 0 \qquad (7)$$

This differential equation can be solved analytically for two- or even three-dimensional cases with simple boundary conditions, but for more complicated conditions this can only be done numerically, after discretization. The development of the modern computer enables the solution of much more complicated problems.

Figure 5 demonstrates that the waterlevels in piezometers along a flowline do not represent the watertable, but only the energy level (potential) at the point in the groundwater body where the piezometer screen is located. A watertable observation well must have a screen that covers the full range of variation of the watertable. The relationship between these are explained in more detail by Weyer (1978).

Since the beginning of the 1960s, models have been used increasingly for the investigation of regional groundwater dynamics. J. Tóth (1962) and Freeze & Witherspoon (1966, 1967) investigated the basic principles underlying the phenomena of large-scale groundwater flow systems.

Figure 6 shows the basic principle of regional groundwater dynamics. Depending on the topography of the groundwater table, the groundwater flows in watertable highs (recharge areas) into the groundwater body, and from there towards areas with a less elevated watertable (discharge areas). With regard to energy, the system of groundwater bodies tries to minimize the overall energy content. This process is being retarded by the hydraulic resistance of the materials traversed by the water. Renewed rainfall and the consequent replenishment of the groundwater interrupt the process, and they increase the sum of the potential energy contained in the system again.

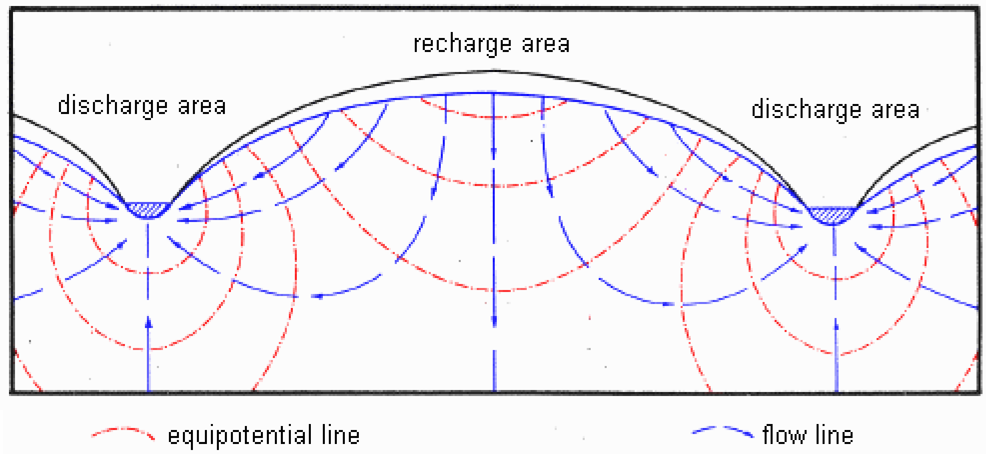

Figure 4. Schematic distribution of equipotential lines and flowlines in an isotropic aquifer between two valleys (after HUBBERT, 1940, p. 930)

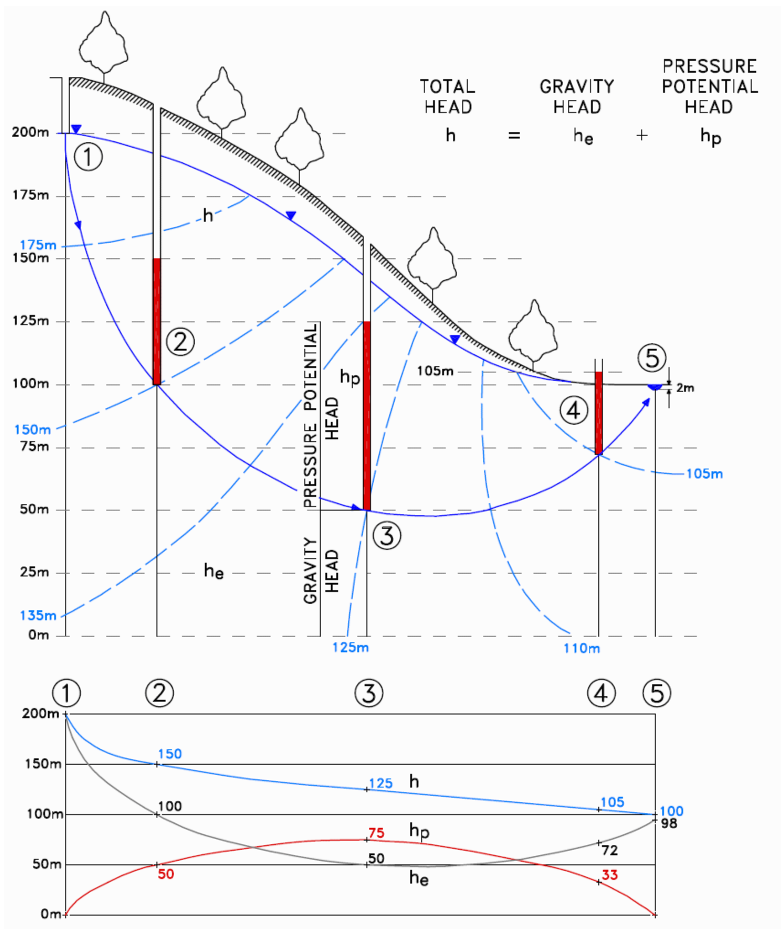

Figure 5. Sketch of water levels in piezometers (red column) along a schematic flowline (after Weyer, 1978).

J. Tóth (1962) used static models to investigate the dependence of the energy distribution in groundwater bodies on various boundary conditions and parameters such as depth of an impermeable layer, and topography of the watertable. An example of the cross-sections calculated by him is shown in Figure 6. He came to the surprising conclusion that different groundwater flow systems exist even in a homogeneous medium, and that these do not necessarily discharge into the nearest surface stream. Moreover, the flow direction in a somewhat lower level below the watertable can be independent of the slope of the watertable, or even in the opposite direction. As can be seen in Figure 6, in such cases most of the groundwater flows near the surface and in the general direction of the slope. Local, intermediate, and regional flow systems appeared.

The next step forward in understanding was the results of the 2-dimensional mathematical model studies by Freeze and Witherspoon (1966, 1967). They expanded Toth's (1962) studies about the influence of morphology and investigated first of all the influence of permeability distribution, and thus of geology, on the energy fields. As could be expected, the very extensive influence of topography and geology on the groundwater flow was confirmed. The comparison between two pairs of results from those model calculations, shown in Figures 7 and 8, confirms this further, as discussed below.

Before we concern ourselves with Figures 7 and 8, the concepts artesian and subartesian should be explained in hydrodynamic terms. In connection with these concepts a lot of confusion was created by the American school around Theiss (1935, 1938) and Jacob (1940, 1950). There, every confined aquifer was called artesian. However, on the basis of groundwater dynamics we know that this relationship is normal for deeper aquifers, and that it first of all does not mean that confined aquifers can be viewed by themselves and disconnected from the collective system of groundwater bodies. It should be pointed out here, that this is assumed for example in many pumping-test equations. However, the definition of confined aquifers is meaningless from the point of view of groundwater dynamics and physics

Figure 9 presents a physically adequate definition of the concepts 'artesian' and 'subartesian', which corresponds with the traditional usage in European science. Subartesian describes a situation in which the water in a piezometer rises above the local watertable; in an artesian situation the water rises above the ground surface, and it could therefore discharge freely. Both cases in Figure 9 show that the energy content at the level of the piezometer screen is higher than that at the watertable. An upwards directed flow therefore usually exists in the groundwater body.

A schematic representation of surface topography is shown in Figures 7 and 8. In Figure 7, a hummocky surface topography was assumed. The relative horizontal and vertical dimensions were set as S (horizontal) and 0.1 S (total vertical thickness of the layers).

For a 6 km long section the total thickness would be 600 m; for a 600 m long section the thickness would be 60 m. The permeabilities in the lower part of Figure 7 were similarly defined relatively, as K=1 and K=100. This means that the assumed aquifer is 100 times as permeable as the overlying layers. The two relative assumptions for the dimensions and the

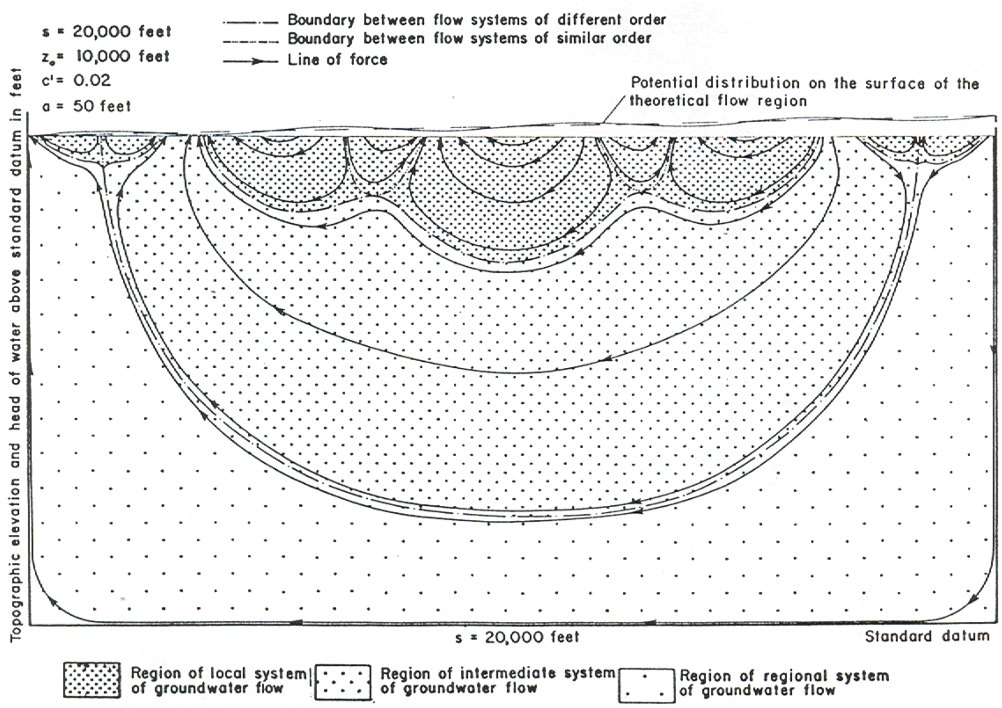

Figure 6. Groundwater flow systems in a homogeneous and isotropic medium with a slightly sloping and hummocky groundwater table (from J.Tóth, 1962). Topographic oscillation = 50 ft (16 m).

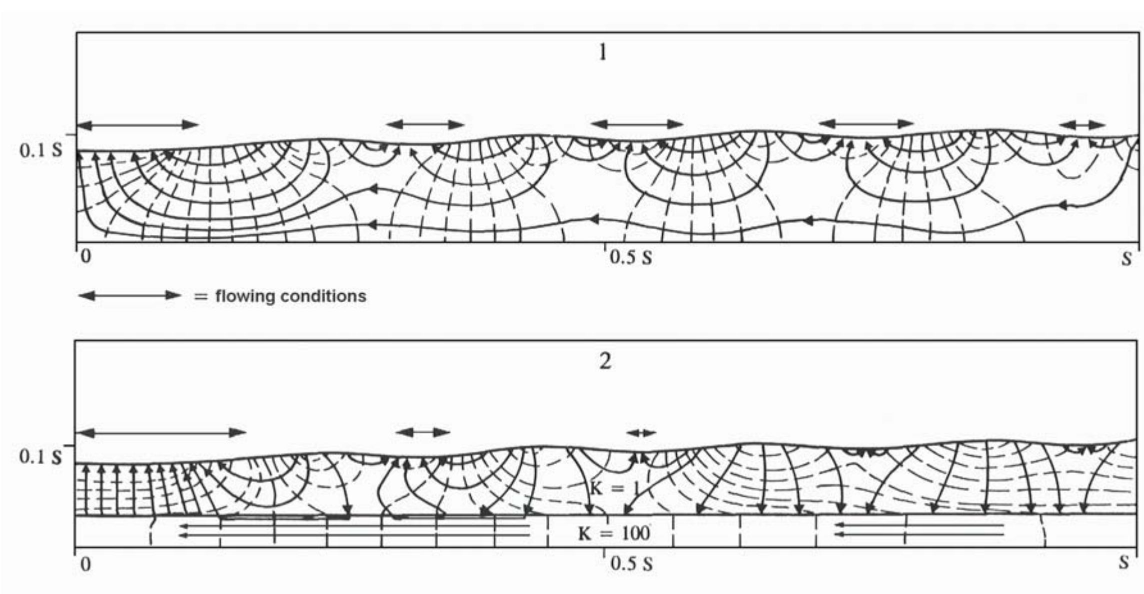

Figure 7. Effect of hummocky topography and layers with higher permeability on groundwater flow pattern and the location of recharge areas and artesian discharge areas (after Freeze and Witherspoon, 1967).

permeability are sufficient for the calculation of the flow pattern; however, actual values of dimensions and permeabilities should be used for the determination of mass flows and velocities.

Figure 7(1) clearly shows how strongly the topography determines the distribution of recharge and discharge areas, even in a homogeneous, isotropic medium. One can see a few flowlines extending underneath the local flow systems to the main discharge area. This flow pattern is strongly changed when a more permeable layer is introduced at the bottom of the section (Fig. 7(2)). This layer only has to be more permeable; it does not have to be an aquifer. Groundwater flow is now collected in the lower layer. Nevertheless one can see that more water still flows in the less permeable layer than in the more permeable layer (the "aquifer"). The quantity of water that is transferred from the less permeable layer to the "aquifer" determines the quantity flowing in the "aquifer". In the right-hand side of Figure 7(2) almost all water flows downward into the "aquifer"; in the left-hand side of the section we see water flowing parallel with the groundwater table in limited areas. Flow parallel with the groundwater table is thus not impossible. It only is a special case and it must therefore first be demonstrated that it occurs. As the topography and the geology are shaped differently almost everywhere, there is not the slightest justification to assume groundwater flow parallel with the watertable without having specific evidence.

In Figure 8 (1 and 2) the influence of permeability distribution (geology) on groundwater flow and on the distribution of recharge and discharge areas is shown for a simple topography. With the introduction of a permeability contrast of only 10:1 in a lens in the subsurface (Fig.8(1)), there is practically no groundwater flow parallel with the groundwater table anymore, even for this simple topography. In Figure 8(2), the permeability contrast of 1000:1 for a continuous layer in the subsurface causes the occurrence of only single large recharge and discharge areas. Notwithstanding the simple topography, groundwater flow parallel to the groundwater table occurs only in a very limited area. As a consequence of the continuity principle, and the extra flowlines in the confining layer, it is again clear: much more water flows through the overlying confining layer than through the "aquifer". In the aquifer, a one-time transport of the groundwater takes place, from the recharge area to the discharge area. In the overlying, low-permeability layer, the same quantity of groundwater is transported twice: (1) in the recharge area from the surface downwards into the aquifer, and (2) in the discharge area from the aquifer upwards to the groundwater table. Under natural conditions in such systems (Fig.8(2)), therefore, twice as much groundwater flows through the low-permeability layer as through the aquifer.

The principles demonstrated here are very important for several hydrogeological problems, e.g. for the protection of aquifer recharge areas; for the calculation of groundwater recharge; for the occurrence of groundwater flow through aquitards; for the migration of groundwater contaminants; and for the migration of mineral waters. The methods applied here differ fundamentally from more traditional current approaches and assumptions. They therefore also lead often to completely different results. Contrary to widespread assumption, in situations like those in Figures 7 and 8 the artesian aquifer does

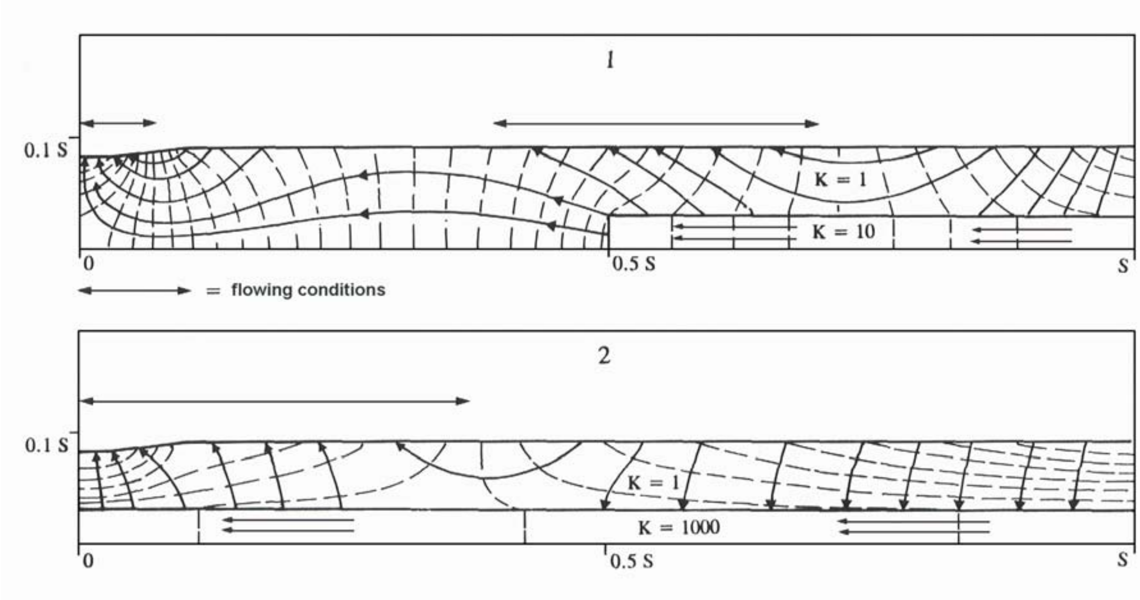

Figure 8. Effect of highly permeable aquifers on groundwater flow in confining layers and the occurrence of artesian conditions under gently sloping topography of the groundwater surface (after Freeze and Witherspoon, 1967).

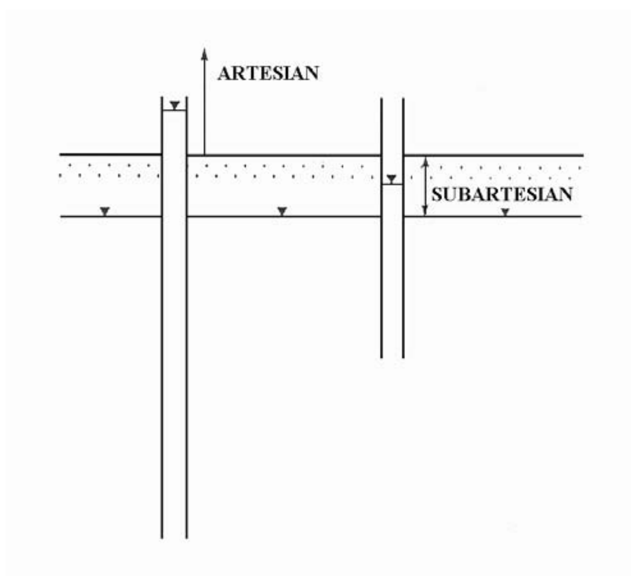

Figure 9. Definition of artesian and subartesian.

not have to crop out. Figure 10 shows the general case schematically. The water flows downwards to the aquifer under the high areas (recharge areas), and from the aquifer upwards under the valleys (discharge areas). Artesian conditions occur in the valleys. The head distribution in the aquifer reflects the topography. That has been confirmed in many studies around the world.

The conservative view of the occurrence of artesian water beneath "impermeable" layers (Fig.11) is poorly thought out and incorrect. Where should the water in the aquifer of Figure 11 then flow towards the right hand site of the diagram?

It is clear that the hydrodynamic principles and relationships outlined above contradict many of the generally adopted and often preached views about groundwater flow. Those principles and relationships shown here are, however, based on reality, and they have been applied and confirmed since the 1960s in numerous investigations worldwide.

1.5.2 Directional transport of pollutants dissolved in groundwater

In contemporary contaminant hydrogeology much emphasis is placed on chemical processes and calculations. These investigations are necessary and required by guidelines and criteria. What seems to be missing at many investigations is the additional and equally important emphasis on groundwater dynamics. Any polluted site is not static. There is constantly water seeping through the source of pollutants and dissolving contaminants. These contaminants are then carried away by groundwater flowing within three-dimensional systems called groundwater flow systems.

In fact these three-dimensional groundwater flow systems determine where the pollutants end up and where they re-enter the biosphere. This principle is demonstrated by the flowlines in Figures 4, 5, 6, 7, 8 and 9. These migration patterns can only be established in detail by adequate field installations. Estimations of their general pattern can be done by proper use of suited 2D-vertical numerical models. The results of this kind of inexpensive model calculation greatly help in the design of monitoring systems.

It appears that field investigations at industrial sites often seem to miss the groundwater dynamic rigor they need to return valid results for investments spent and to reduce expenditures on unnecessary monitoring equipment and chemical sampling.

Two high-profile European case histories of unnecessary expenditures on monitoring are presented by Weyer (2005a, 2005b). At these sites millions of dollars had been spent needlessly.

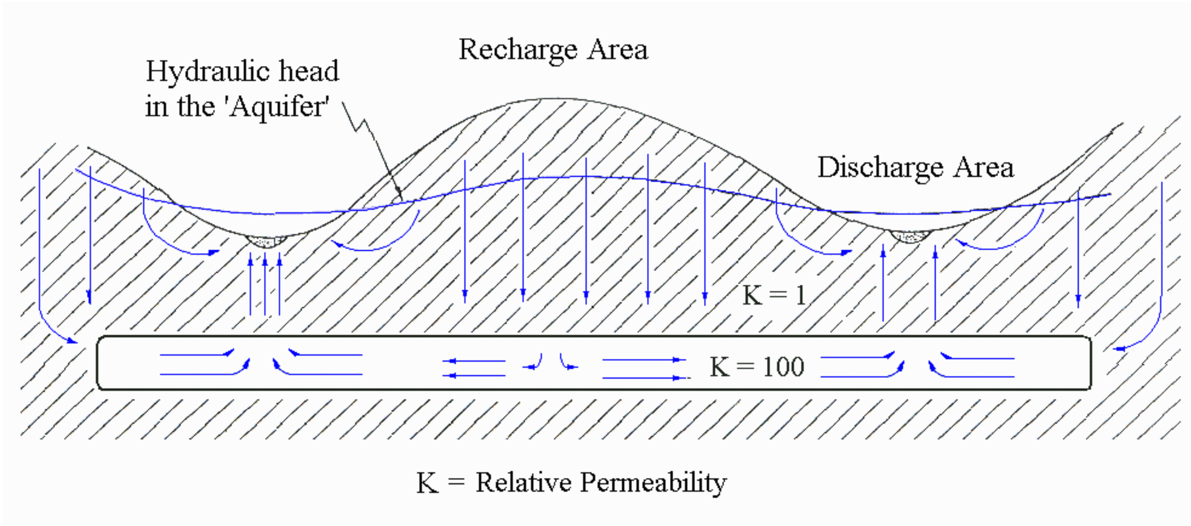

Figure 10. Schematics of groundwater flow in an artesian aquifer system. In this case the aquifer does not outcrop at the surface.

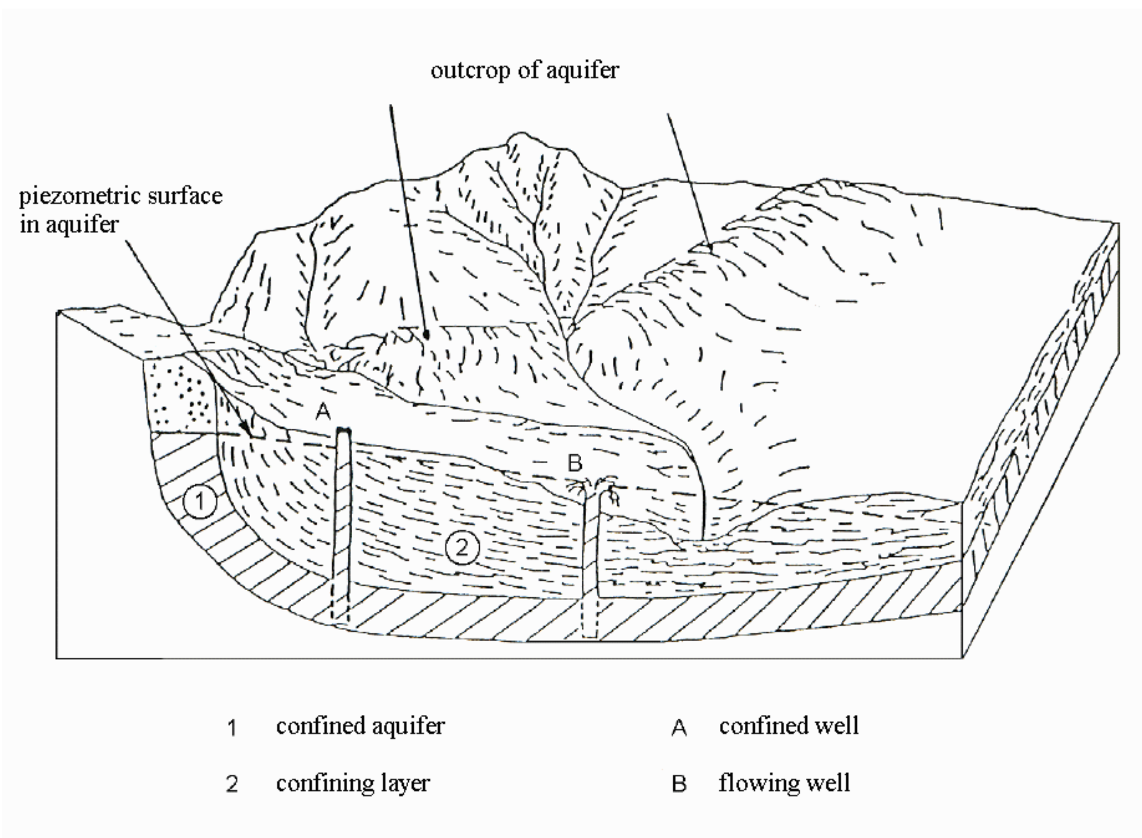

Figure 11. Traditional view of artesian (flowing) conditions (Walz, 1973, Fig. 6.1.3).

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